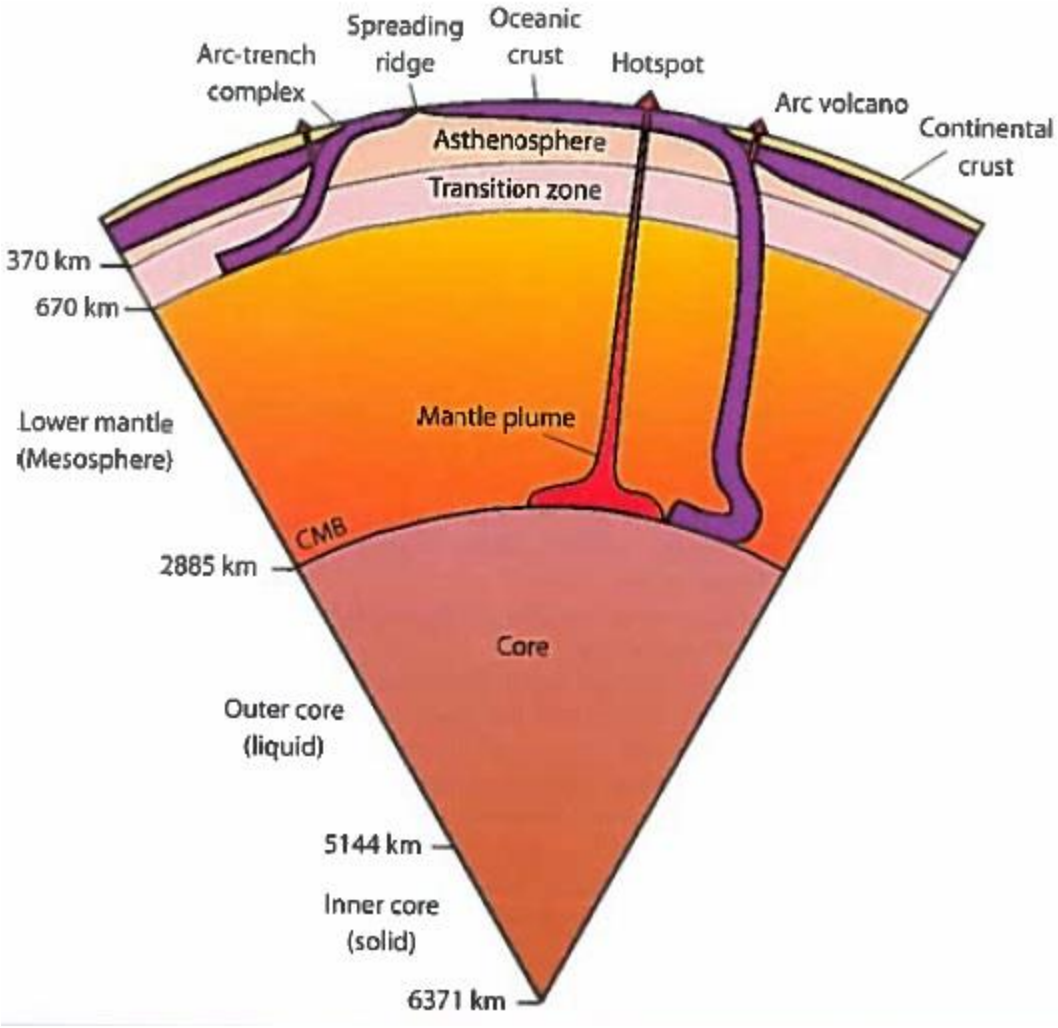
Geothermics

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

Convection

Convection: fluid flow driven by internal **buoyancy (B)**



$B = -gV\rho = -g\Delta m$

Δm = mass anomaly due to a volume V with a density difference ($\rho = \rho V - \rho$) induced by temperature and/or composition

Thermal regime of the deep interior

- Mantle is sufficiently compressible to originate adiabatic temperature gradients. Olivine—spinel and spinel—post-spinel transitions provided the benchmark temperatures of 1600°C at 400 km and 1700°C at 670 km.

$$(\mathrm{d}T/\mathrm{d}Z)_{\mathrm{s}} = T\alpha g/C_p \sim 0.5 - 0.6\,^{\circ}\mathrm{C/km}$$

- Solid melts when the thermal oscillation of atoms reaches a critical amplitude:

$$T_m = C\ m\ V^{2/3}\vartheta_D^2$$

Poirier, 1991

C=constant, *m* and *V* = mass and volume of the atoms ϑ_D =Debye temperature

Debye temperature ϑ_D is directly related to the maximum frequency of vibration of the solid v_m:

$$\vartheta_D = \frac{h\,v_m}{k_B}$$

$$\vartheta_D = 61.2\ k_l + 385$$

For silicate minerals (Horai and Simmons, 1970)

h = Planck constant (6.626×10^{-34} J s) and k_B = Boltzmann constant (1.381×10^{-23} J K^{-1})

Thermal regime of the deep interior

To calculate T_m at pressure p_m:

$$\frac{p_m}{p_o} = \left(\frac{T_m}{T_{m_o}}\right)^c - 1 \qquad c = (6\gamma + 1)/(6\gamma - 2)$$

Gilvarry, 1956

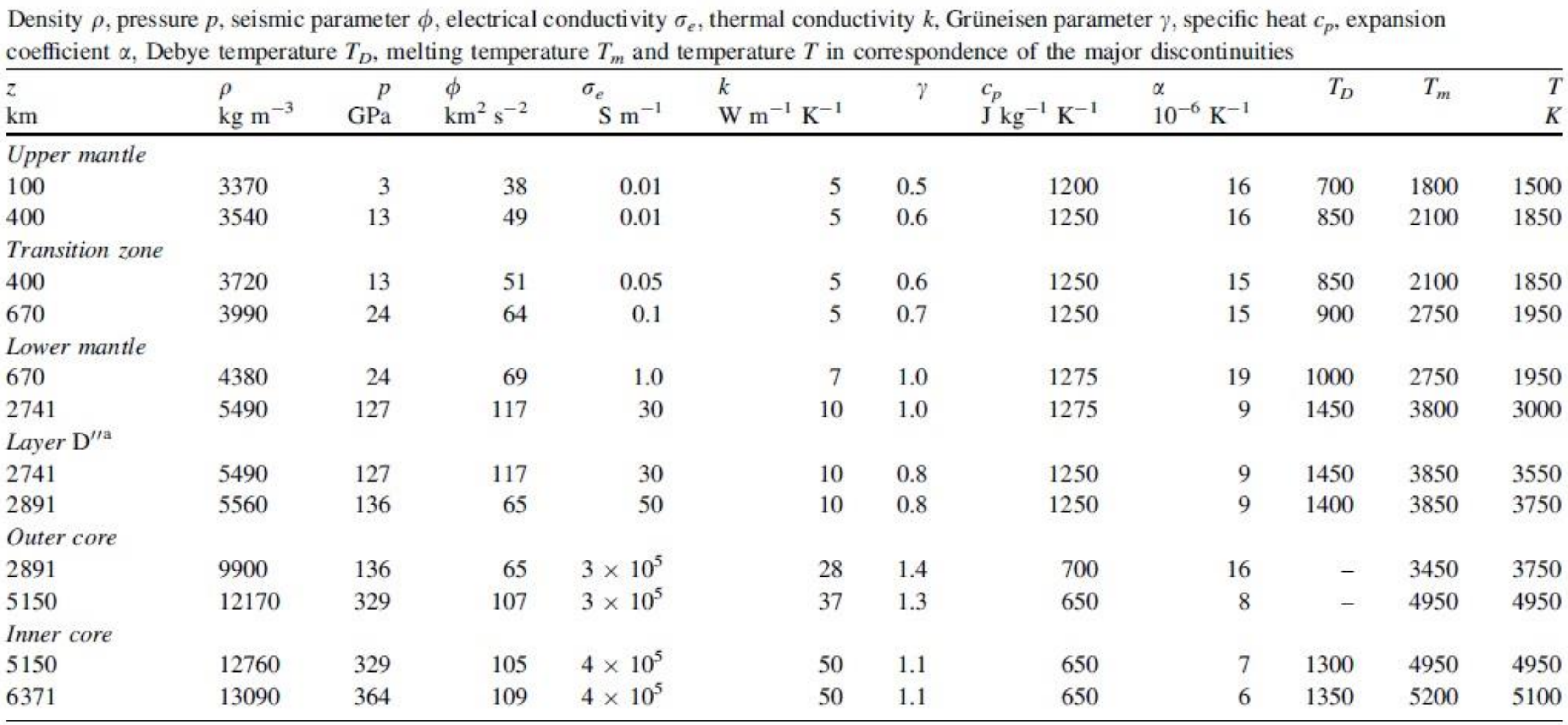
T_{m0}=melting *T* at zero pressure, p_0 = (negative) melting *P* at zero *T*, γ= Grüneisen parameter

$$\gamma = \frac{\alpha\ K_S}{\rho\ c_p} = \frac{\alpha\ K_T}{\rho\ c_V} = \frac{\alpha v_b^2}{C_P} \qquad K_S = \rho(\partial P/\partial \rho)_S$$

c_p and c_V are the specific heats at constant pressure *P* and volume *V*, K_S and K_T are bulk moduli at constant entropy *S* and temperature *T*, and α is the thermal expansion coefficient, v_b=bulk sound speed.

- **The Grüneisen parameter (γ) expresses the ratio between thermal pressure and thermal energy per unit volume (it measures the rate at which *P* increases as heat is input or the pressure required to prevent thermal expansion).**
- **γ is weakly dependent on pressure and temperature (for most minerals it varies between ~1.0 and 1.5).**

Thermal regime of the deep interior



Density ρ, pressure p, seismic parameter ϕ, electrical conductivity σ_e, thermal conductivity k, Grüneisen parameter γ, specific heat c_p, expansion coefficient α, Debye temperature T_D, melting temperature T_m and temperature T in correspondence of the major discontinuities

z km	ρ kg m^{-3}	p GPa	ϕ km^2 s^{-2}	σ_e S m^{-1}	k W m^{-1} K^{-1}	γ	c_p J kg^{-1} K^{-1}	α 10^{-6} K^{-1}	T_D	T_m	T K
Upper mantle											
100	3370	3	38	0.01	5	0.5	1200	16	700	1800	1500
400	3540	13	49	0.01	5	0.6	1250	16	850	2100	1850
Transition zone											
400	3720	13	51	0.05	5	0.6	1250	15	850	2100	1850
670	3990	24	64	0.1	5	0.7	1250	15	900	2750	1950
Lower mantle											
670	4380	24	69	1.0	7	1.0	1275	19	1000	2750	1950
2741	5490	127	117	30	10	1.0	1275	9	1450	3800	3000
Layer D$''^{a}$											
2741	5490	127	117	30	10	0.8	1250	9	1450	3850	3550
2891	5560	136	65	50	10	0.8	1250	9	1400	3850	3750
Outer core											
2891	9900	136	65	3×10^5	28	1.4	700	16	–	3450	3750
5150	12170	329	107	3×10^5	37	1.3	650	8	–	4950	4950
Inner core											
5150	12760	329	105	4×10^5	50	1.1	650	7	1300	4950	4950
6371	13090	364	109	4×10^5	50	1.1	650	6	1350	5200	5100

Electrical conductivity is linked to the motion of ions:

(in the upper crust) $\sigma_e = \sigma_{ef}\,\xi^n$

$$\sigma_e = \sigma_{e\infty}\exp\left(-\frac{E^*}{k_B\,T}\right)$$

(in the lower crust and at >> depths)

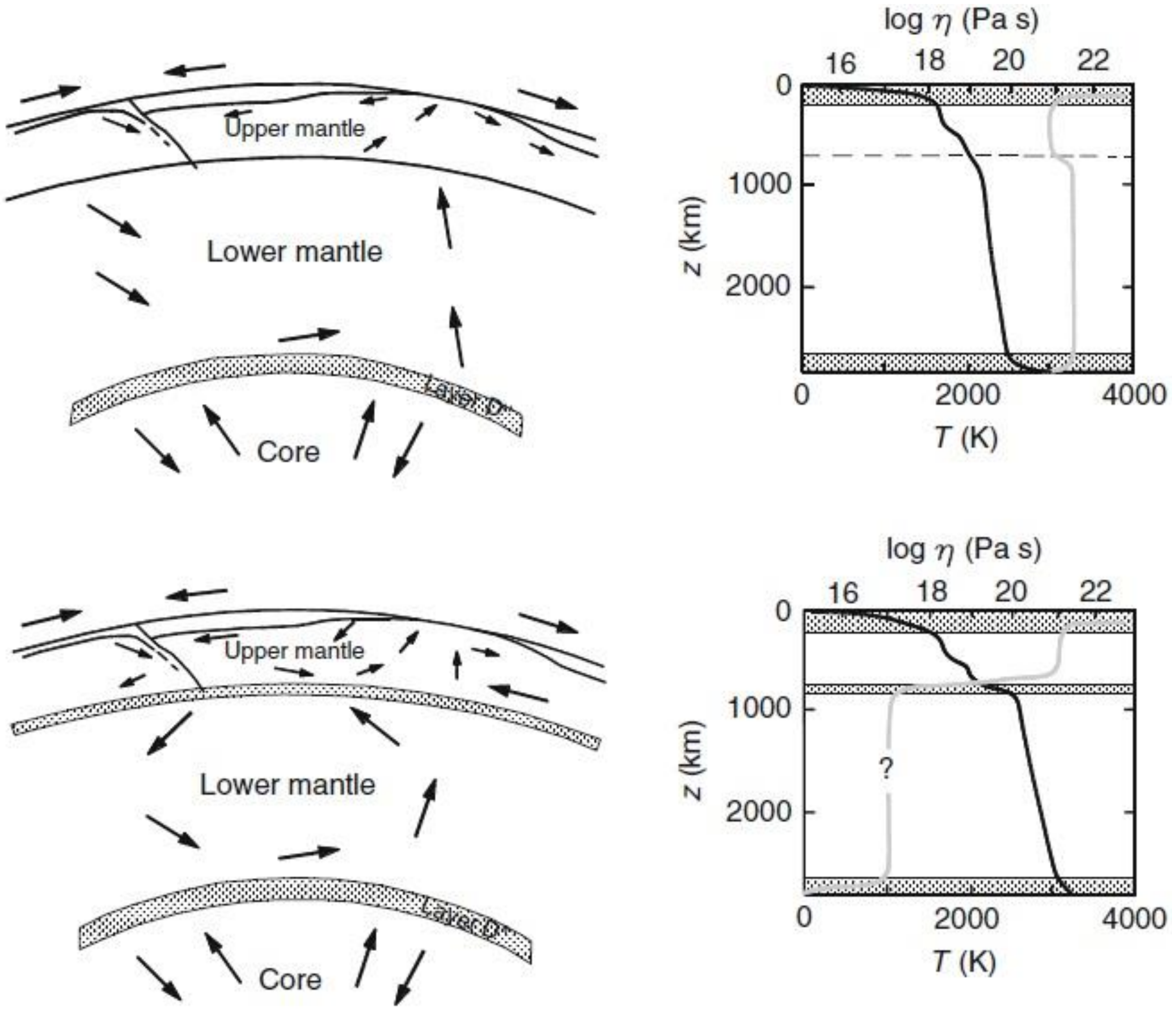
σ_{ef}= conductivity of the fluid, ξ=fraction of pores filled with the fluid, and *n* (1.3-2.5) a parameter that increases with compaction, cementation, and consolidation.

E*=Activation energy, k_b=Botzmann constant, $\sigma_{e\infty}$ extrapolated conductivity for T $\longrightarrow \infty$

In the core: $$\frac{k}{\sigma_e} = L\ T$$

L = 2.5 10^{-8} W $S^{-1}K^{-2}$ = Lorentz number (k =50 W $m^{-1}K^{-1}$ and *T* = 5000 K)

Thermal regime of the deep interior

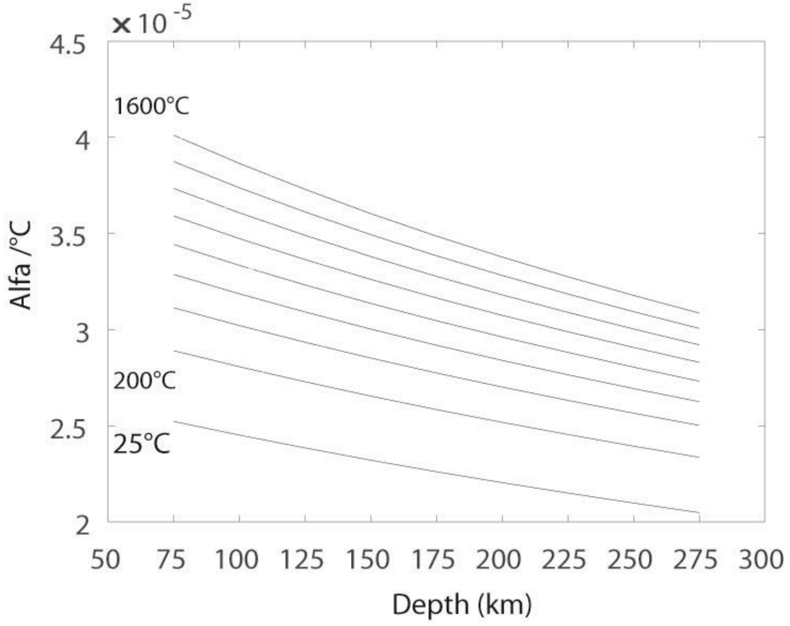


- Factors that could inhibit a mantle-wide convection are (1) a viscosity increase with depth, (2) a phase change with an inclination of the Clapeyron curve sufficiently negative and (3) the lack of mixing between chemically different layers.
- Phase changes are not an obstacle to convection, since Claperyon curve is positive for the transition olivine-spinel (dp/dt= 4MPaK^{-1}).

Buoyancy

Thermal buoyancy: $\rho = \rho_0[1 - \alpha(T - T_0)]$

With α about 3.0 x 10^{-5}/°C, a temperature contrast of 1000 °C gives rise to a density contrast of about 3%.



Thermal expansion coefficient

Examples:

1. A plume head 1000 km in diameter with ΔT 300 °C would have a B ~ 2 x 10^{20}N.
2. A Lithospheric slab subducted extended to the bottom of the mantle (3000 km), its B would be about -200 TN/m.
3. Subducting oceanic crust (7 km) gives a contribution to slab buoyancy of only ~20 TN/m.

Compositional buoyancy: induced by small variations in the upper/lower mantle composition

Larger density variations in the mantle are due to pressure-induced phase transformations of the mineral assemblage (not necessarily a source of buoyancy).

Convection model

Balance between buoyancy forces (B) and viscous resistance (R):

Motion of the plate (subducting slab) is resisted by the viscous stresses (proportional to velocity) accompanying mantle flow

Buoyancy B = $-gDd\rho\alpha\Delta T$

with ΔT the average difference in temperature between the descending lithosphere and fluid interior (T) ~$T/2$ and d, thickness of the subducting lithosphere (of the layer that diffused), depending on time (t) spent at the surface $t = D/v$:
$d = (kt)^{1/2} = (kD/v)^{1/2}$, with k diffusion coefficient.

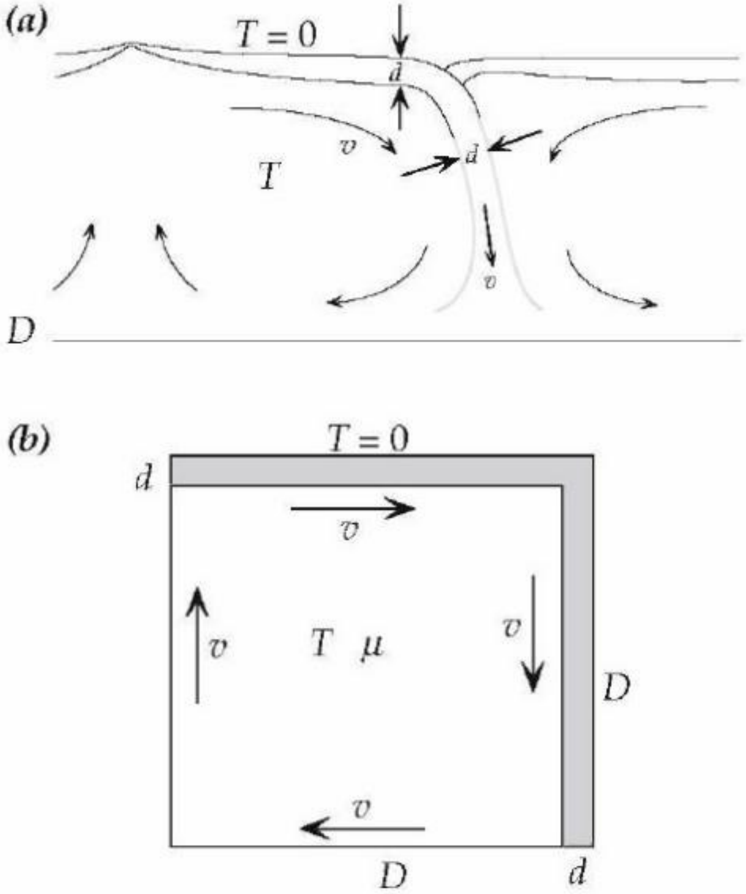
The resisting viscous stress σ acting on the side of the descending slab: $\sigma=\mu 2v/D$

Viscous resistance R (per unit length)= $D2\mu v/D= 2\mu v$

Balance between two forces: B+R=0 if $v =-gDdp\alpha T/4\mu$

$v=D(g\rho\alpha T(\kappa)^{1/2}/4\mu)^{2/3}$

For D= 3000 km, ρ= 4000 kg/m^3, $\alpha = 2 \times 10^{-5}$/°C, T = 1400 °C, k=10^{-6}m^2/s and $\mu=10^{22}$Pa s
$v = 2.8 \times 10^{-9}$m/s = 90 mm/yr (close to velocity of plate motion)



Convection model

Rayleigh Number (T): $R_a \sim (D/d)^3$ $R_a = g\rho\alpha T D^3/\kappa\mu$ $d/D = R_a^{-1/3}$ For the mantle $R_a = 3 \times 10^6$

$q = K\Delta T_q/d$ (total heat flux) $q_k = K\Delta T/D$ (the heat that would be conducted in the steady state in the absence of convection)

$\Delta T_q = qD/K$

Rayleigh Number (q): $R_q = g\rho\alpha\Delta T_q D^3/\kappa\mu$ $R_q = g\rho\alpha q D^4/K\kappa\mu$ $R_q/R_a = \Delta T_q/\Delta T \sim Ra^{1/3}$ $R_q = Ra^{4/3}$

Nusselt number $Nu \sim q/q_K = \Delta T_q/\Delta T \sim Ra^{1/3}$ For the mantle $Nu = 100$

Peclet number $Pe = v(D/\kappa)$ or $Pe \sim v/V$ $V = \kappa/D$ = velocity scale $Pe \sim Ra^{2/3}$ For the mantle $Pe = 9000$

Reynolds number $Re = uD/\nu$ with $\nu = \mu/\rho$ (kinematic viscosity ~ $10^{17} m^2 s^{-1}$), u=velocity of the flow ($30 mmyr^{-1}$) For the mantle $Re = 10^{-18}$

Prandt number: For the mantle $Pr = 10^{25}$

$$\mathrm{Pr} = \frac{\nu}{\alpha} = \frac{\text{viscous diffusion rate}}{\text{thermal diffusion rate}} = \frac{\mu/\rho}{k/(c_p\rho)} = \frac{c_p\mu}{k}$$

R_a= Onset and vigor of convection.

Pe = ratio between convective and diffusive heat transport rates.

Nu= ratio between the total heat flux (in the presence of convection) and heat flux that would be conducted with the same temperature difference across the layer. Thus, it express the efficiency of convection as a heat transport mechanism relative to conduction.

Re = turbulence of the flow : the flow is laminar if $Re < 10^3$.

Pr= ratio of momentum diffusivity (kinematic viscosity) to thermal diffusivity. For Pr << 1 the thermal diffusivity dominates, while for Pr >> 1, the momentum diffusivity dominates the behavior. For magma and Earth mantle convection is very effective in transferring energy in comparison to pure conduction, so momentum diffusivity is dominant.

Convective time scale (time needed for a fluid to cross the depth of the fluid layer at the convective velocity v) decreases with increase of Ra.

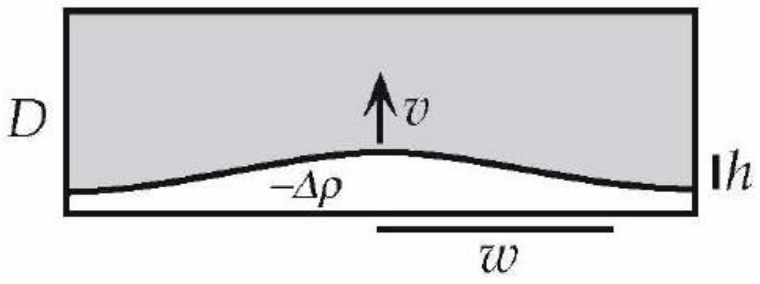
$\tau_v = D/v = \tau_k R_a^{-2/3}$ with $\tau_k = D/V = D^2/k$ τ_k=it estimates the time it would take the fluid layer to cool by conduction in the absence of convection.

Then if $R_a = 3 \times 10^6$ and u= 0.2m/yr, τ_k= 14 Myr, while for $R_a = 10^6$ and u= 0.02m/yr, τ_k= 130 Myr.

Convection model

- For a fluid layer heated uniformly on a lower horizontal boundary, there is a minimum amount of heating below which convection does not occur.
- The transition from conduction to convection, just at the point of instability, is called **marginal stability**, which occurs when Ra ~ 1000.

Marginal stability



From the balance of buoyancy (B) and viscous resistance (R_s) force:

$B = g\Delta\rho wh$ if $w \ll D$ $R_s = \mu(v/w)w = \mu v = \mu dh/dt$

we get: $dh/dt = g\Delta\rho wh/\mu$ and $h = h_o exp(t/\tau_S)$, where $\tau_S = \mu/g\Delta\rho w$ h_0=constant

The bulge grows exponentially with a time constant τ_S, because the interface is unstable (Rayleigh-Taylor instability). Since τ_s gets smaller as *w* gets bigger, broader bulge grows more quickly.

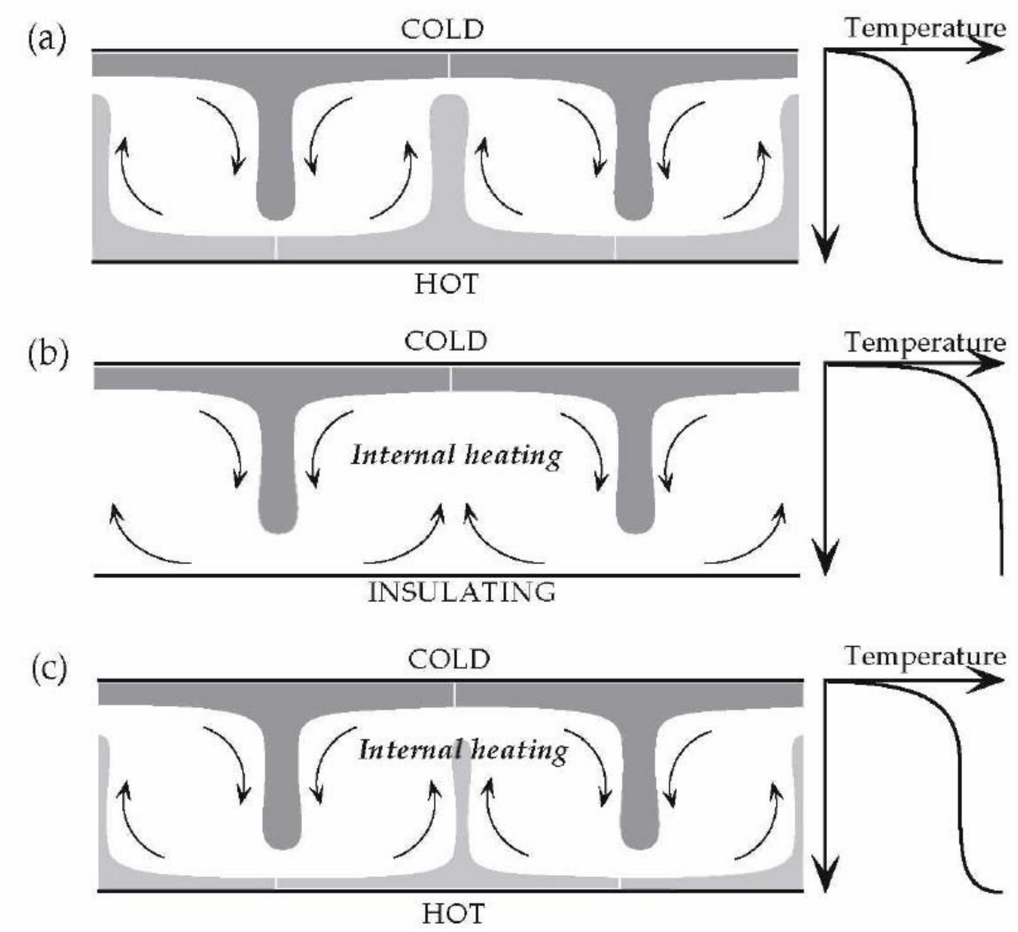
if $w \gg D$ $uD = vw$ $R_1 = \mu(u/D)w = \mu v w^2/D^2$ $\frac{\partial h}{\partial t} = \frac{g\Delta\rho D^2}{\mu w} h$ $\tau_1 = \frac{\mu w}{g\Delta\rho D^2}$ τ_l gets bigger for larger w

If w=D the time scale of the growth of the instability gets smaller (fastest growing bulge) $\tau_{RT} = \mu/g\Delta\rho D$ and considering that $\tau_k = D^2/\kappa$

$R_a = \tau_k/\tau_{RT} = g\Delta\rho D^3/\kappa\mu$ for the mantle, $R_a = 3 \times 10^6 \gg 1000$ (critical value)

In the mantle heat does not diffuse very far in the time it takes the fluid to become unstable and overturn (thermal diffusion time scale is very long)

Heating modes



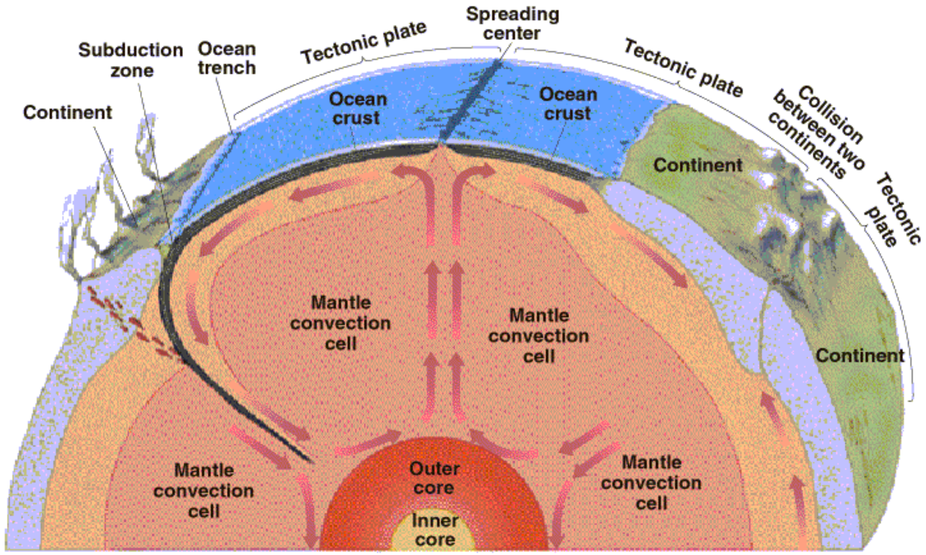
Active upwelling: heat entering from below and there is no heat generated within.

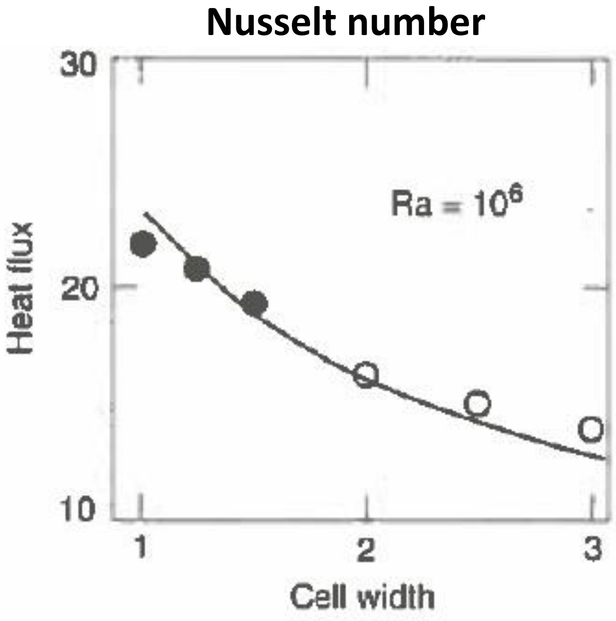
Passive upwelling: a fluid layer is heated from within by radioactivity and the cool fluid sinking form the top boundary layer drives circulation. In this condition, upwellings would be a passive response rather than involving positively buoyant material.

Active and passive upwelling: the heat input to the fluid layer might be a combination of heat entering from below and heat generated within. The upper thermal boundary layer conducts outward both the basal and internal heat and thus develops a greater temperature drop than the basal boundary (no symmetry between upwellings and downwellings).

If the layer covects more vigorously, the thickness of the layer having homogeneous temperature will increase, the thermal boundary layers will be thinner and the temperature gradients through them will be higher, driving larger heat fluxes.

Convection model



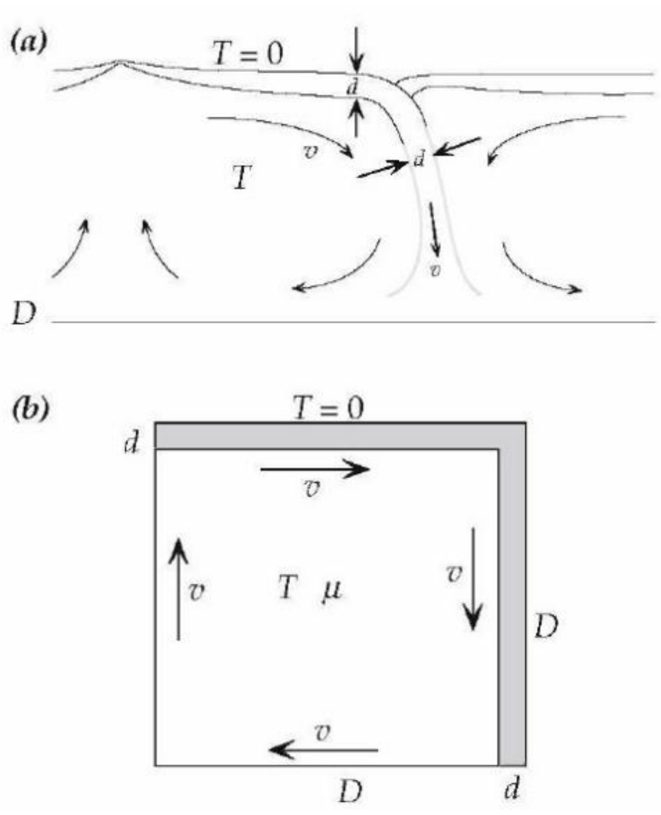


- For Rayleigh-Benard convection in a fluid layer, convection regime takes the form of cells that are about as wide as the layer thickness h, as long as $Ra<10^5$.
- Mantle convection cells change size based on continental/oceanic areas distribution: the wider is the continental plate the wider is the cell.
- The Earth today has cells of aspect ratio (L/h) ~3 (horizontal heat transport can be neglected), so that its rate of heat loss is ~ half that for cells of L/h=1 (cooling of the Earth is less efficient in presence of continents).
- The main effect for an aspect ratio > 1 (wide cells) is that the Nusselt number and average heat flux across the cell decreases as *L* increases.
- Continents perturb the convective flow pattern as their lithospheric roots divert mantle flow.

Convection model

In a thin lid with large aspect ratio vertical heat transport dominates:

$$q = +\lambda_L \frac{T - T_o}{d} \qquad \left(\lambda \frac{\partial T}{\partial z}\right)_{z=0} = \lambda_L \frac{T - T_o}{d}$$

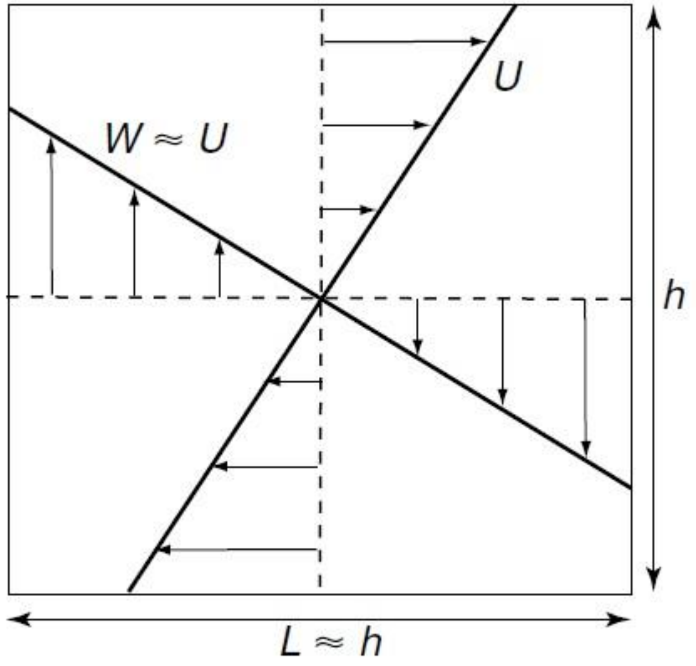


If we neglect the lid thickness in comparison with the fluid layer depth, the heat flux at the top of the fluid, i.e. at *z* = 0

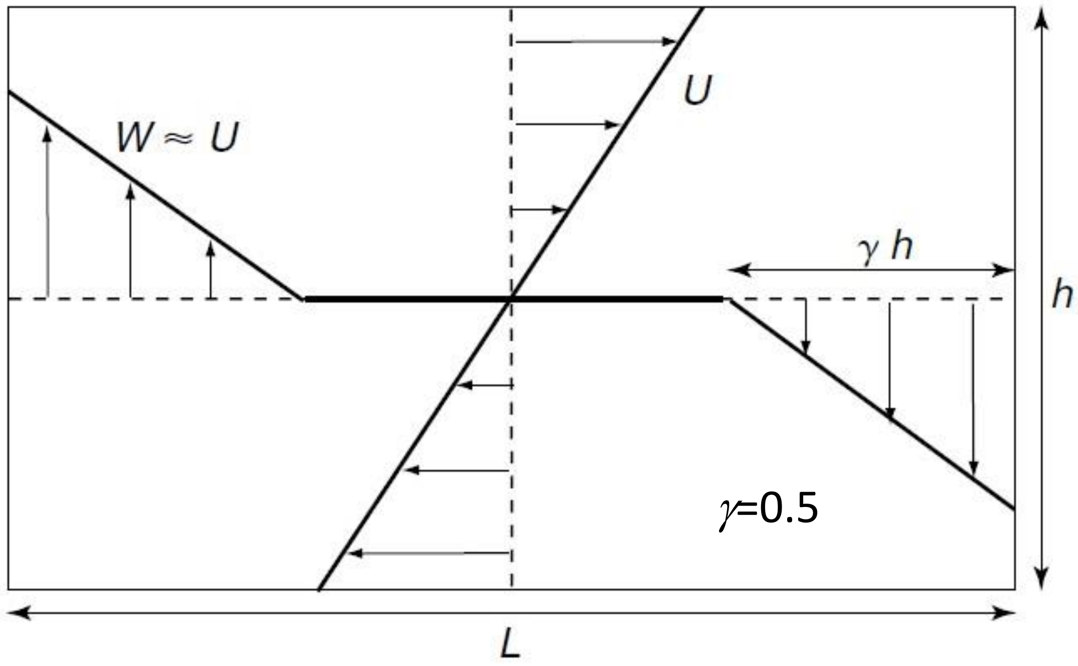
$$\left(\frac{\partial T}{\partial z}\right)_{z=0} - \mathrm{B}T = 0 \qquad \mathrm{B} = \frac{\lambda_L}{\lambda}\frac{h}{d}$$

h=*D* (thickness of the fluid layer) ~ 3000 km
d=thickness of the lid ~ 300 km
λ_L=thermal conductivity of the lid, λ=thermal conductivity of the fluid, λ_L/λ ~ 1
B=Biot Number (B=10, since h/d=10) is small, since it leads to heat fluxes that are small fraction of those of free convective cells (e.g., q~ 12mWm^{2} at the base of the cratonic lithosphere, q~ 100mWm^{2} close to the ridges).
For B <<1 (d>> or λ_L <<) heat flux approaches zero.

Convection model



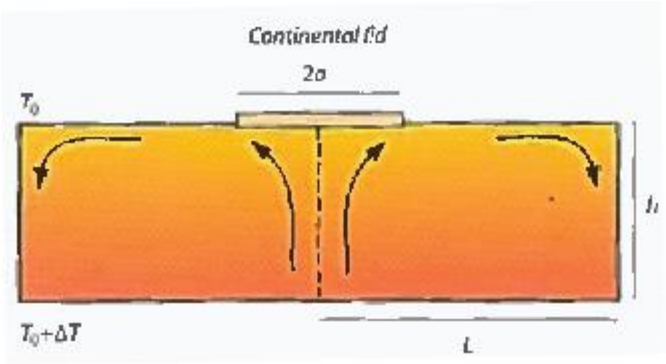
Convection cell of aspect ratio ≈ 1 (Rayleigh–Benard convection)

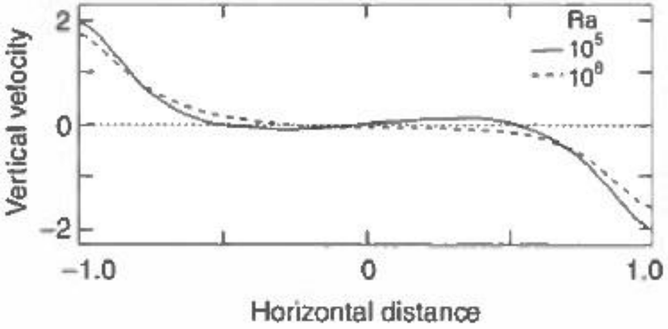


Convection cell of aspect ratio >1

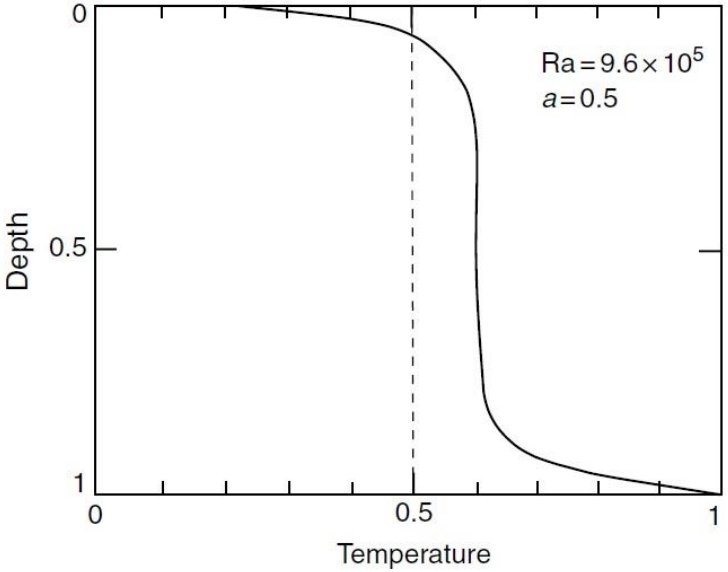
- Kinetic dissipation may be split into two different components: one is generated by the horizontal shear flow with a vertical gradient of horizontal velocity, while the other component is associated with the vertical flow, which is such that velocity drops from the maximum value W over horizontal distance $\delta = \gamma h$.
- If the cell of aspect ratio >1 the horizontal velocity field stretches over the whole layer thickness, whereas the vertical velocity field extends over kinetic boundary layers of width γh.

Convection model





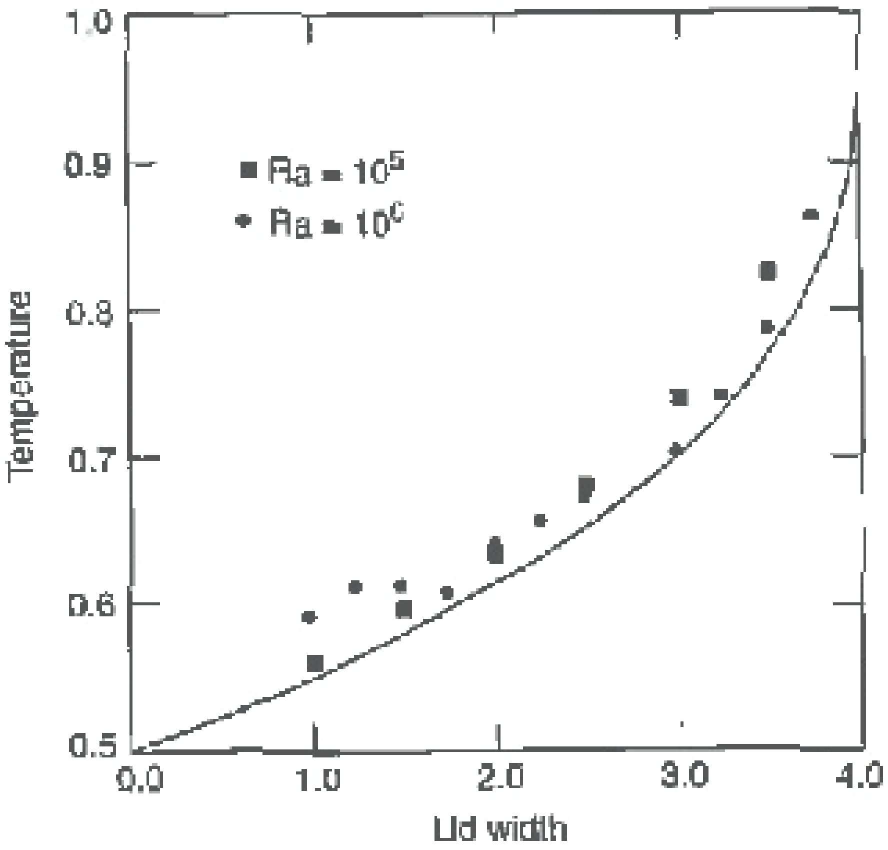
- In presence of poorly conductive lid, the base of the lid is at higher *T* than the top of the fluid away from the lid at T_0. Then, the fluid beneath the lid is heated and becomes involved in an upwelling centered on the continent that feeds horizontal flow extending to large distances. The upper and lower horizontal surfaces of the cell move in the horizontal direction with velocity U and –U at the top and bottom, respectively, while the velocity is zero in the middle of the cell.
- Aspect ratio of the cells (*L*/*h*) depends on the width of the lid and Ra: for Ra> 10^6 the flow involves small-scale instabilities superimposed on a larger-scale circulation.
- Because of the difference in the efficiency of heat transport at the upper and lower boundaries, the average temperature of the fluid layer is larger than *To*+Δ*T*/2.

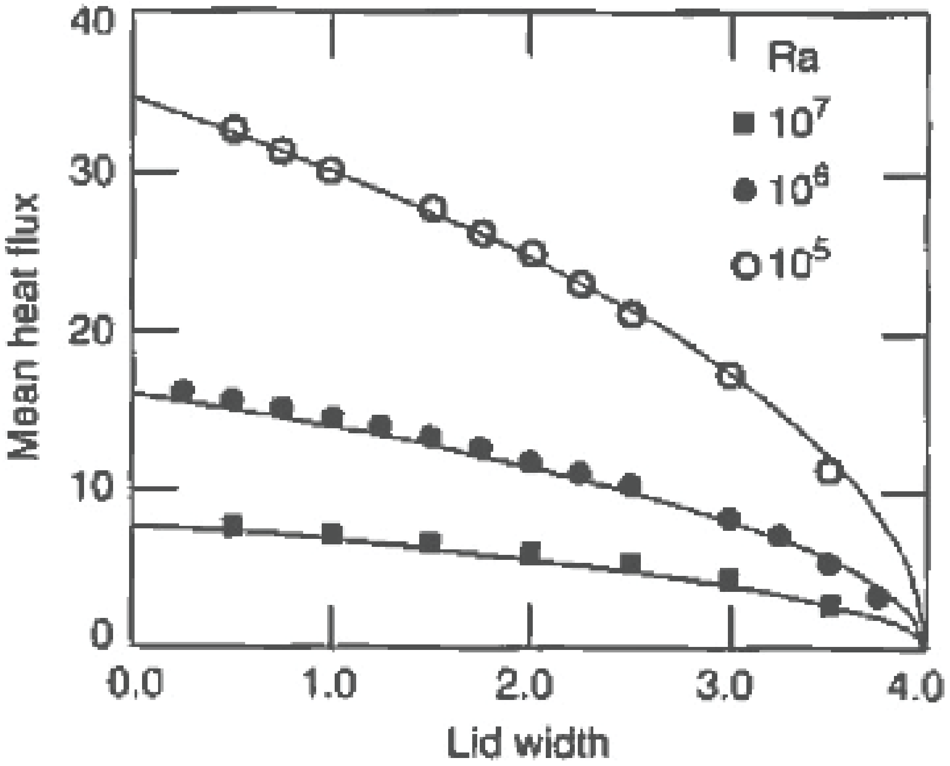


a=dimensionless lid width

Convection model

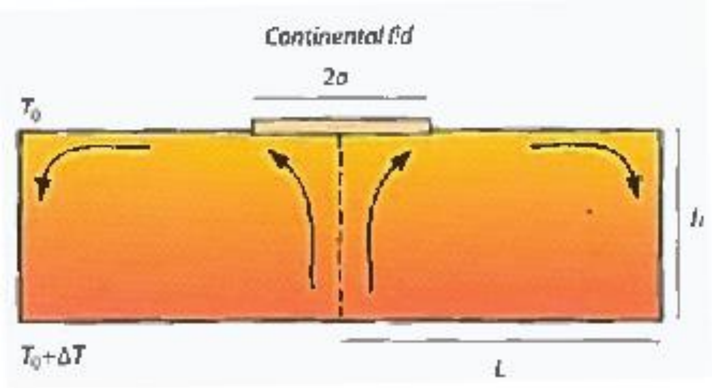
Continental insulation acts to increase the average mantle temperature and decrease its heat flux





Dynamic topography

(due to supercontinental assembly and dispersal)

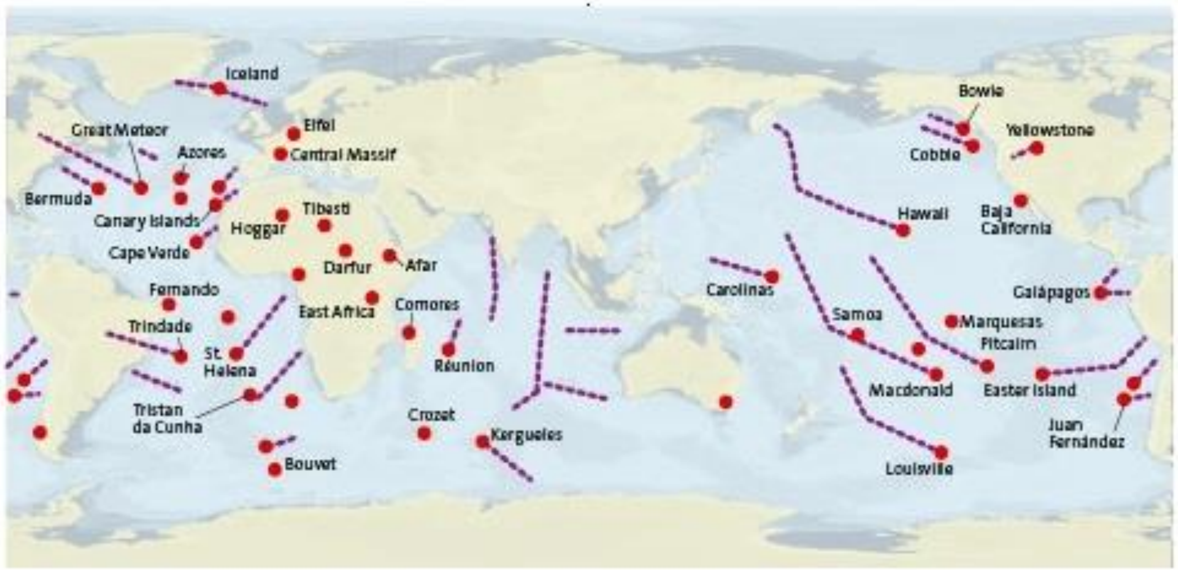


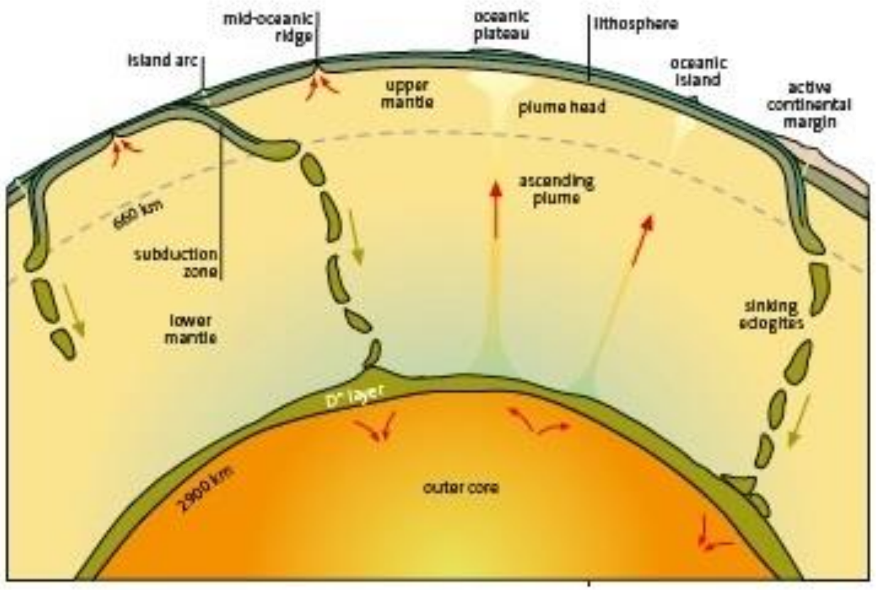
A wide insulating lid (during supercontinent assembly) warms the mantle beneath it and the entire interior of the Earth:

- it may cause plumes localization and would explain the flood basalts with supercontinental brak-up.
- Presence of the lid prevents the heat from escaping to the surface (as under the oceans).
- It makes the cooling effect of subduction absent.
- It influences the formation of elongated large-scale cells with a size dependent on the width of the insulating lid (it increases the mean temperatures of the convective fluid layer).

Very large continent can cause drip-type instabilities associated to small-scale convention cells (λ < 1000 km) and thus form intracratonic basins depocenters.

Hot Spot, mantle plumes, and Large Igneous Provinces (LIPs)

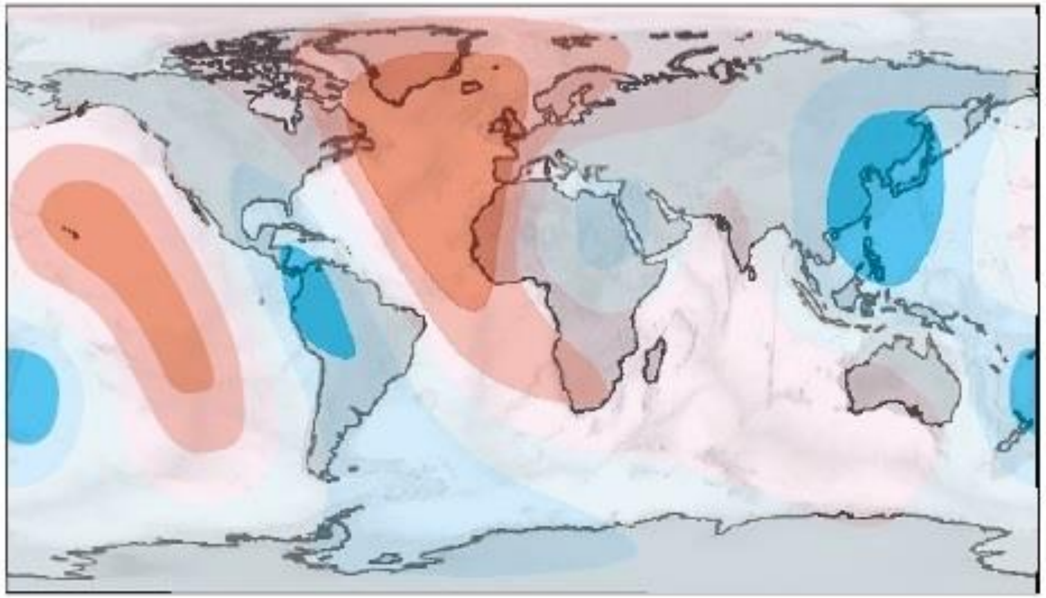




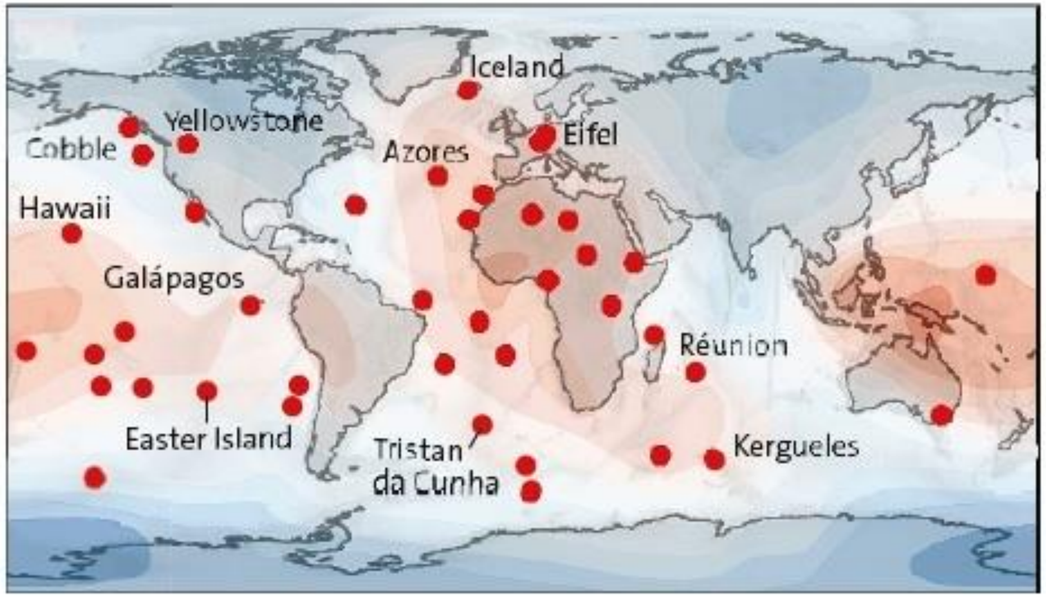
Plumes

- Hot mantle rocks (300 °C higher than surrounding mantle), not molten, because of the high pressure, arising from the lower mantle (or shallower depths).
- Depth formation of mantle plumes is imaged by the seismic tomography and reflected by the geochemistry.
- Hot spot lifetime is about 100 Myr.
- The influence on the Earth energy budget of the heat flux carried by mantle plumes is of ≈ 2–4 TW < 10% of the global heat flow.

Hot Spot and geoid anomalies

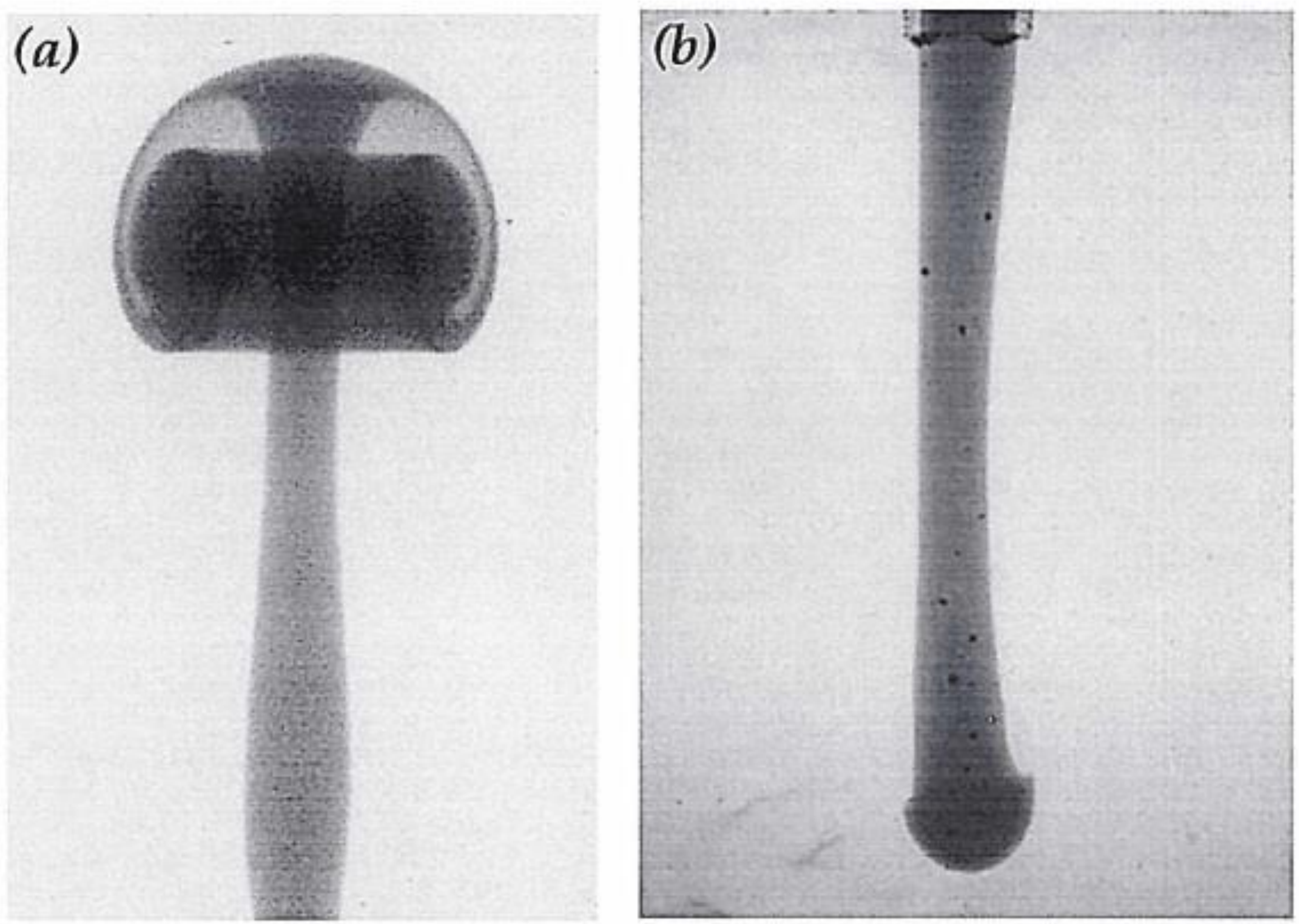


- Hot spot are concentrated in correspondence of geoid bulges



- Oceanic hot spot are rich of incompatible elements and high isotopic ratio of ^{87}Sr/^{86}Sr, ^{187}Os/^{188}Os, and ^{3}He/^{4}He indicating the influence of a lower mantle source.

Plume generation



(a) The buoyant fluid is hot, and the plume viscosity is about 1/300 times that of the surrounding fluid (the column has a large, nearly spherical head at the top with a very thin conduit or tail connecting it to source).

(b) The injected fluid is cooler and thus denser and more viscous than the ambient fluid (the diameter of the buoyant columns is fairly uniform over its height).

Plume generation

- **Plume's diameter depends on (1) the volume flux and temperature excess of the source (2) thermal and viscosity properties of the lower mantle into which plume starts to ascend.**
- **Plume's head grows by entrainment as it ascends through the mantle (then as a function of distance travelled).**

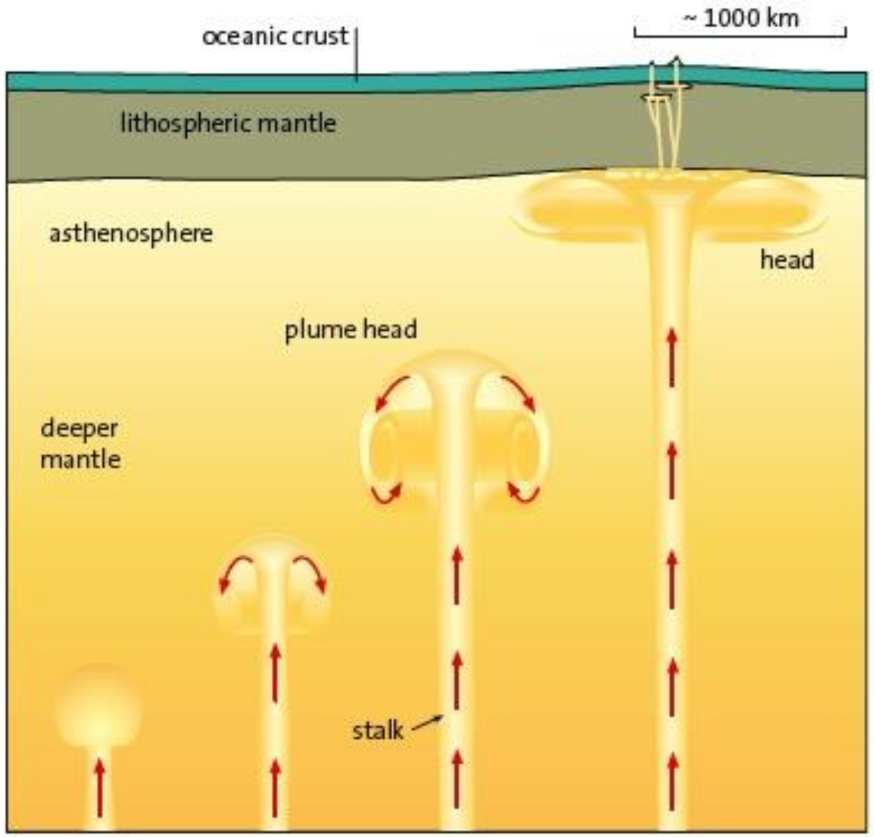
Plume heads slow down by the viscosity of the surrounding mantle:

- **Expansion of the head (slower upward motion compensated by the faster tail)**

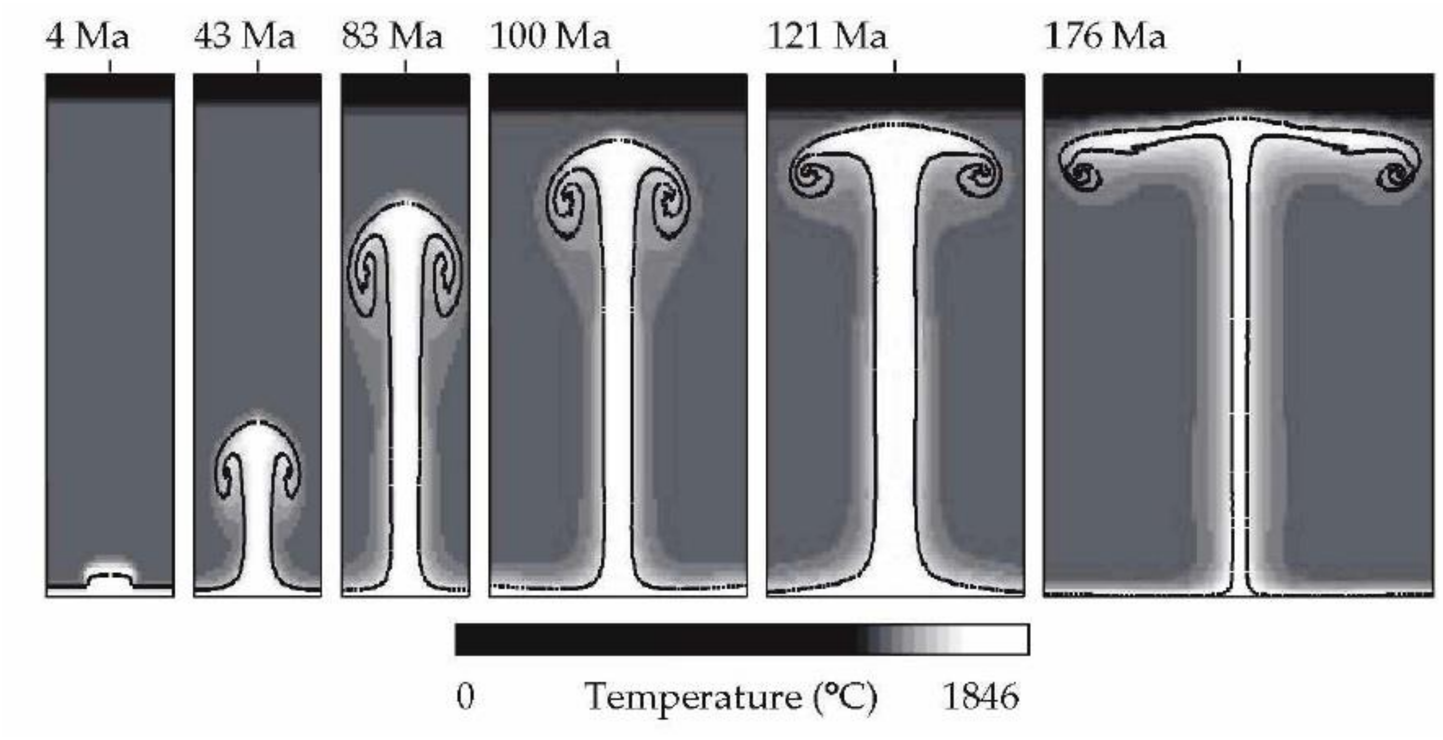
$$\mu = a\exp\left(\frac{E+pV}{RT}\right)$$

μ=viscosity
a=constant
P=pressure
v=activation volume

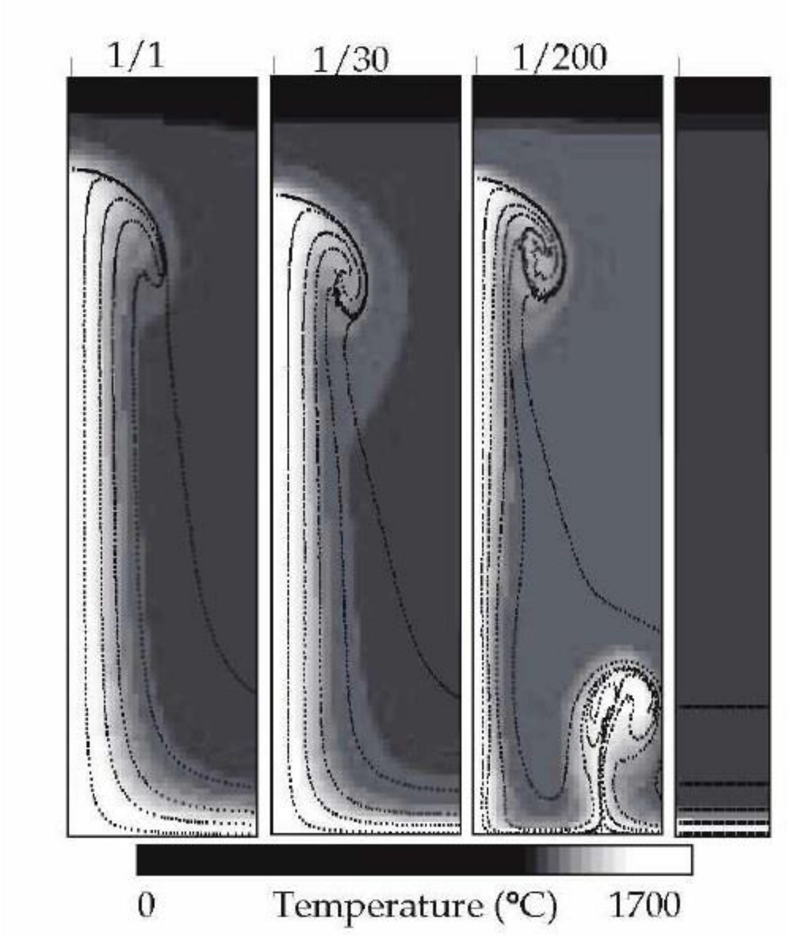
Activation enthalpy: H=E+pv



Plume generation



Plume tails become thinner by decreasing plume’s viscosity



Plume heads have a larger buoyancy which is more capable of penetrating resistance than a narrow column.

Laminar and Turbolent Plumes

Hot plume material loses heat to the surrounding fluid by diffusion (plume grows wider as it rises).

Laminar Regime: $W \propto \frac{R^2 \Delta\rho g}{\mu} \propto \frac{R^2 \rho_0 \alpha \Theta g}{\mu}$

W = plume velocity, μ = viscosity, R = radius, Θ = temperature anomaly

Plume Energy Flux

$$P \propto \rho_0 C_p W \Theta R^2 \qquad W \propto \sqrt{\frac{\alpha g P}{\mu C_p}} \qquad R \propto \sqrt{\kappa \frac{z}{W}}$$

Since radius increases with height due to diffusion: $R \approx \sqrt{\kappa t}$ $\quad t \propto z/W$

Since the plume dimensions change with height, Reynold number varies with height (increases approching the surface):

$$\mathrm{Re}(z) = \frac{\rho W R}{\mu} \propto z^{1/2}$$

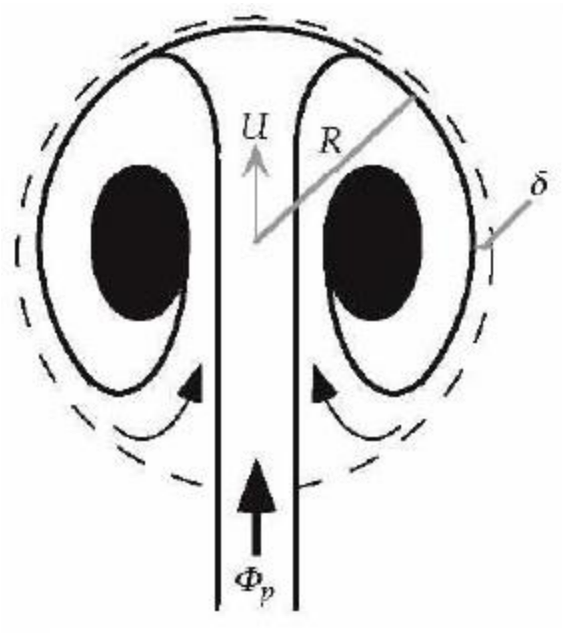
Reynold Number for Laminar

$$\mathrm{Re} = \frac{\rho W R}{\mu} \propto z^{2/3}$$

Reynold Number for Turbolent

Close to the source Reynold number is small and the flow cannot be turbolent

Plume generation



ϕ_P=πr^2U

(ϕ_P=volumetric flow rate at which the swell volume is created)

- The heat goes partly outwards, to form the thermal boundary layer around the head, and partly inwards, to further heat the entrained material.
- The thickness, δ, of the thermal boundary layer adjacent to the hot plume head depends on the time the adjacent fluid is in contact with the passing plume head.

$$\delta = \sqrt{\kappa t} = \sqrt{\frac{2\kappa R}{U}}$$

t=2R/U

$$U = \frac{g\rho\alpha\Delta T R^2}{3\mu}$$

U=(velocity for a low-viscosity sphere)

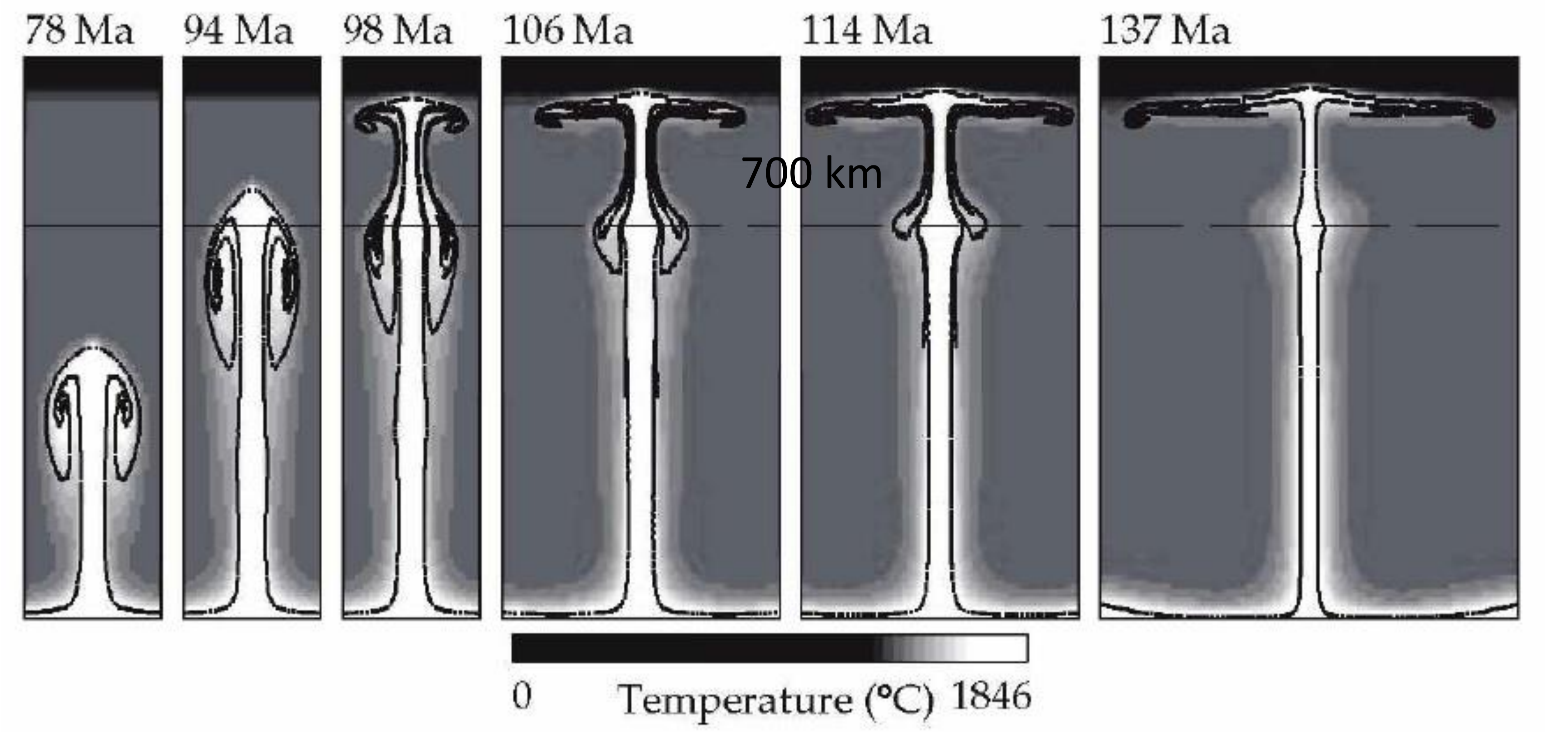
$$\Phi_e = 2\pi R\delta U$$

ϕ_e=The rate at which boundary layer fluid flows through the horizontal cross-sectional area of the boundary layer near the head's equator

The rate of increase of the head radius due to entrainment is: $$\frac{\partial R}{\partial t} = \frac{\Phi_e}{4\pi R^2}$$

Plume generation

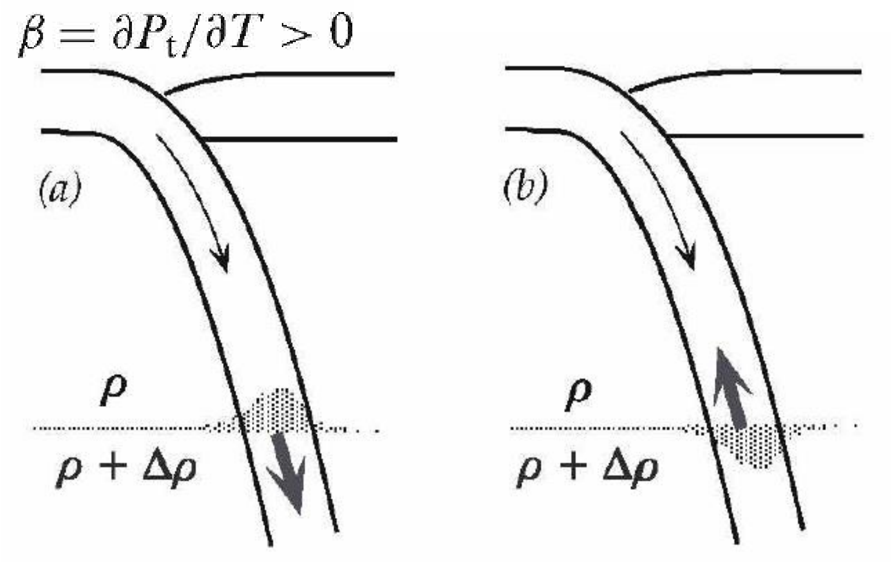
Effects of depth dependence of viscosity and a phase transformation on the plume



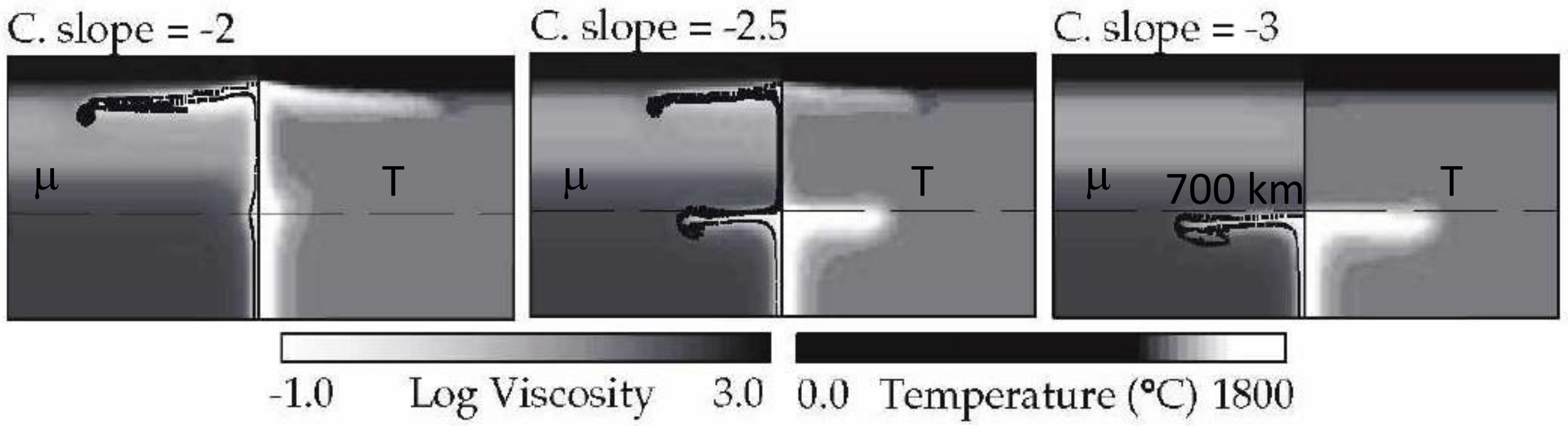
- The viscosity increases by a factor of 20 at 700 km and exponentially by a factor of 10 below.
- The effect of a phase transformation at 700 km depth is simulated with a moderately negative Clapeyron slope of -2MPa/K.

Plume generation

Effects of depth dependence of viscosity and a phase transformation on the plume

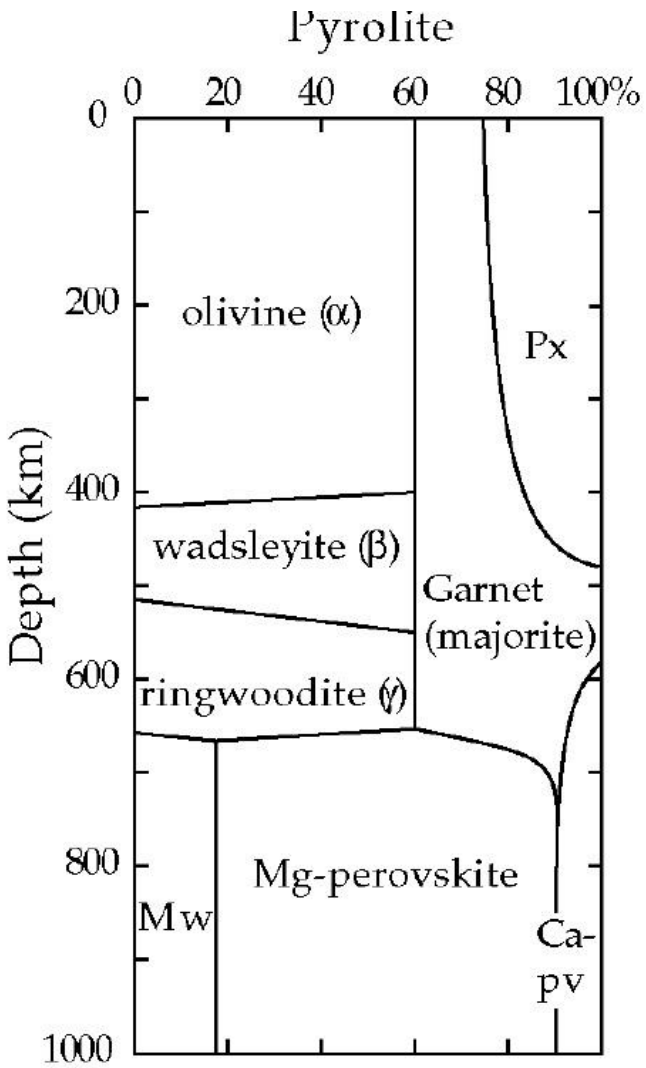


- A poitive Clapeyron slope causes a negative buoyancy force (broad arrow) that would add to the negative thermal buoyancy of the cold slab, aiding its descent.
- A negative Clapeyron slope delays the transformation to greater pressure and depth within a descending slab, producing a positive buoyancy that would oppose the slab's descent.
- In cold subducted lithosphere *T* may be too low for the thermally activated reactions to occur. Thus, the phase would persist metastably, producing a positive buoyancy that would always oppose the descent of the slab.
- The mechanical strength of subducted lithosphere may be sufficient for stresses from 410 km to be transmitted to 660 km, so the opposing buoyancies from the different depths will also tend to cancel.



- In a hot rising column of mantle a positive Clapeyron slope would cause the transformation at a greater depth, yielding a positive buoyancy that would enhance the column's rise, while a negative Clapeyron slope would inhibit its rise.

Phase Transformations and Clapeyron Slope



Phases transformation at 410 km:

- Olivine -> wadsleyite reaction has a positive Clapeyron slope (~3 MPa/K)
- Pyroxene -> majorite transformation may have a strongly negative slope (net effect unclear).

Phases transformation at 660 km:

- Ringwoodite -> perovskite + magnesiowiistite transformation has a negative Clapeyron slope of ~ -2 MPa/K.
- Garnet -> perovskite transformation may have a strongly positive Clapeyron slope (about 4 MPa/K) and a substantial density increase, yielding a negative buoyancy in opposition to the transformation of the ringwoodite component (net effect unclear).

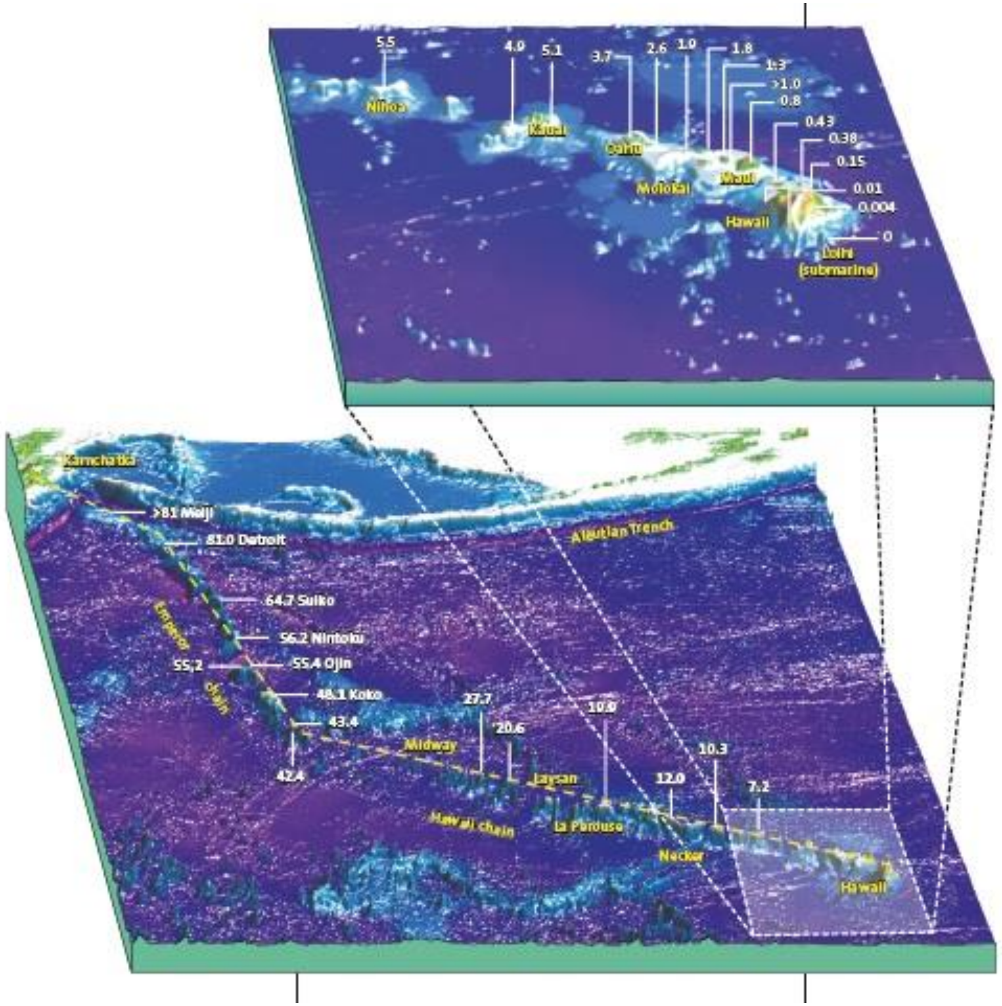
The effect of a phase transformation may be significant in 2D, constant viscosity model, but less significant in 3D and with *T* dependent viscosity

Hot spot tracks

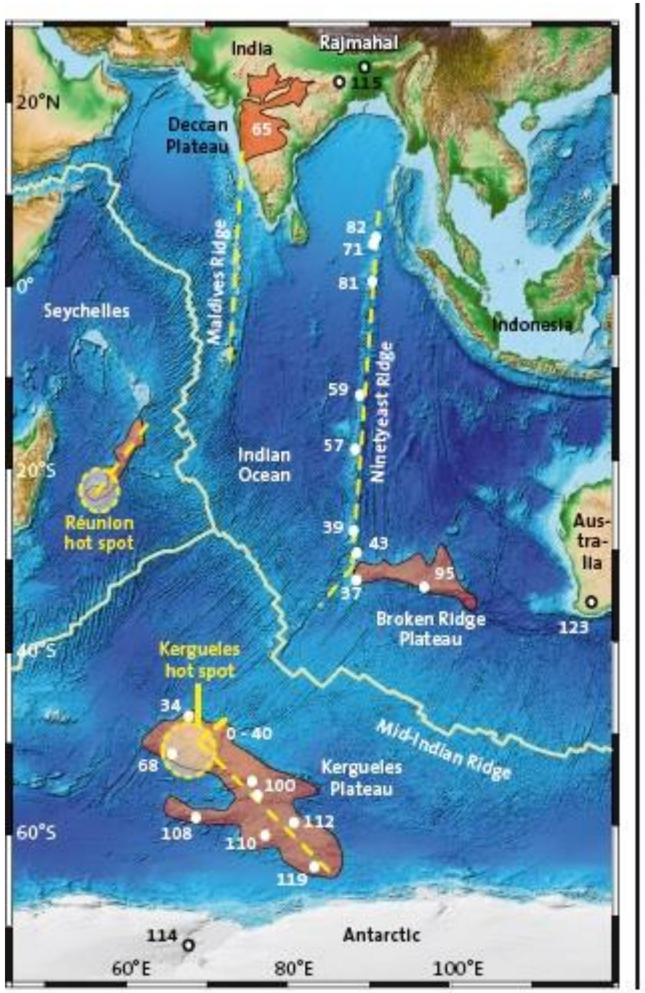
Hot spots form tracks on the ocean floor in response to the motion of the ocean plate

Pacific Plate drifts over the plume at a rate of ca. 100 km in 1Myr

Hawaai (6000 km-long track)

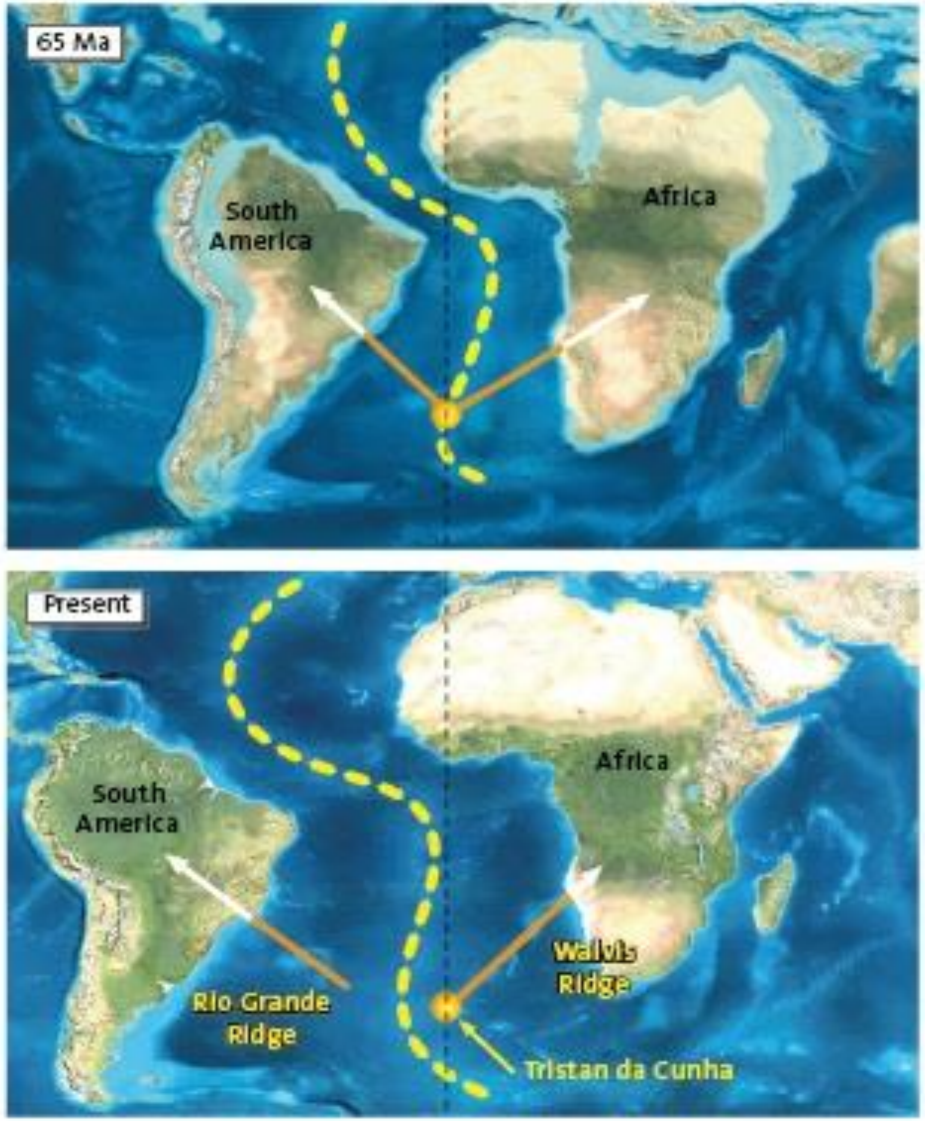


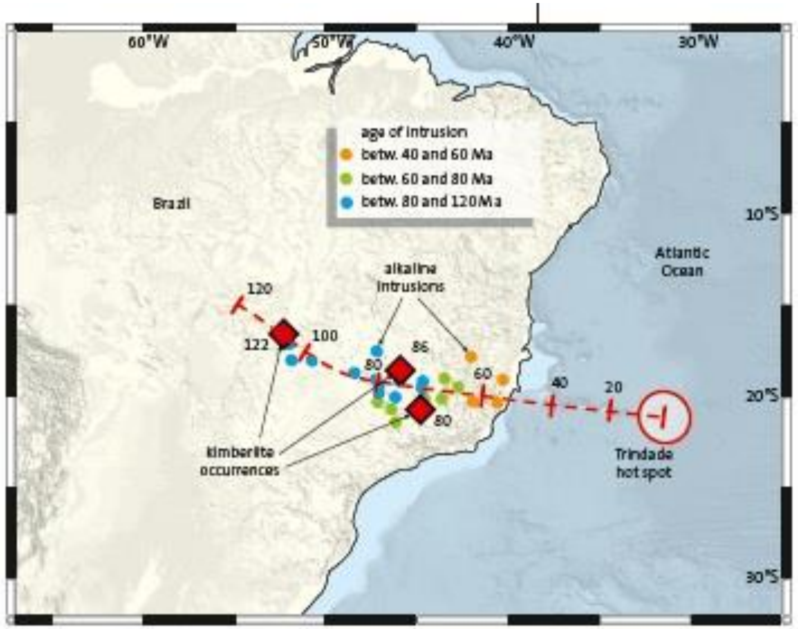
India

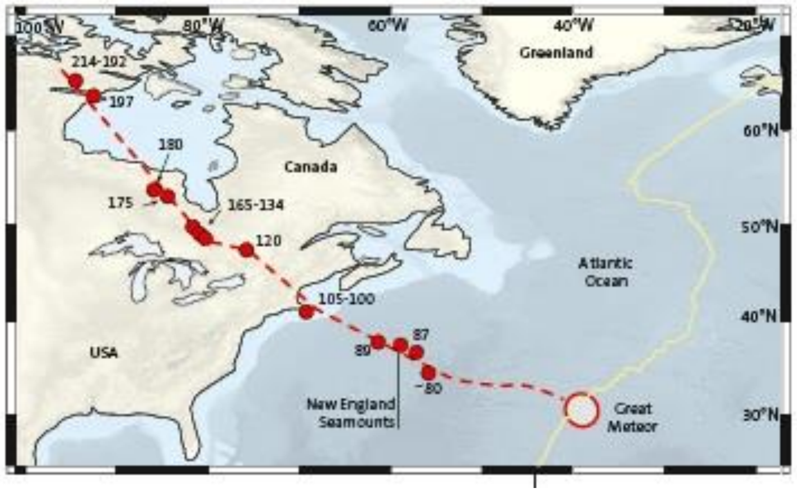


Maldives Ridge and the Ninetyeast Ridge were generated by the hot spots of Réunion and the Kergueles.

Hot spot tracks



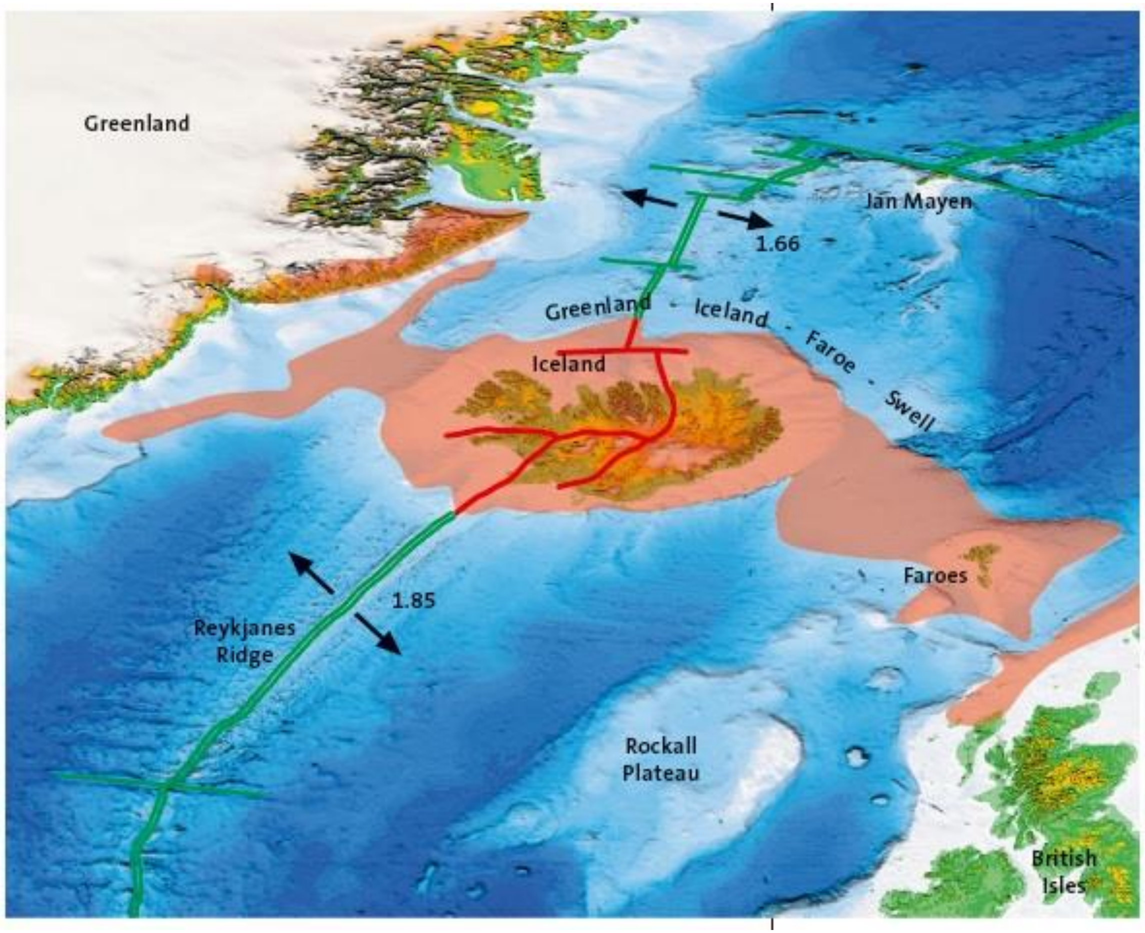




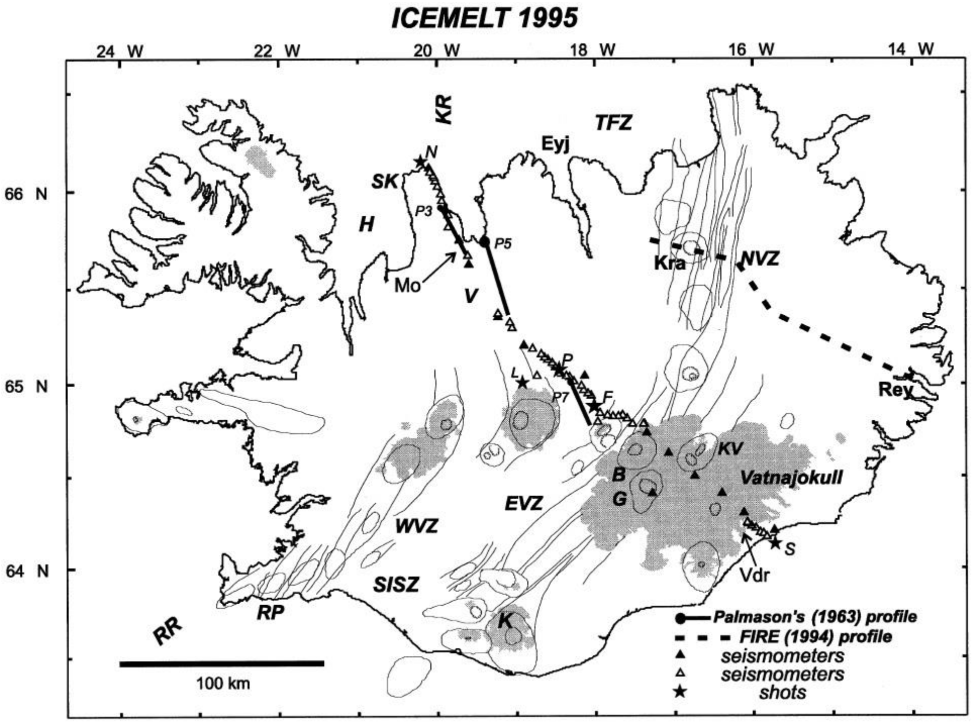
The plate boundary shifted 30 Ma, and the hot spot Tristan da Cunha tracked across only the African Plate.

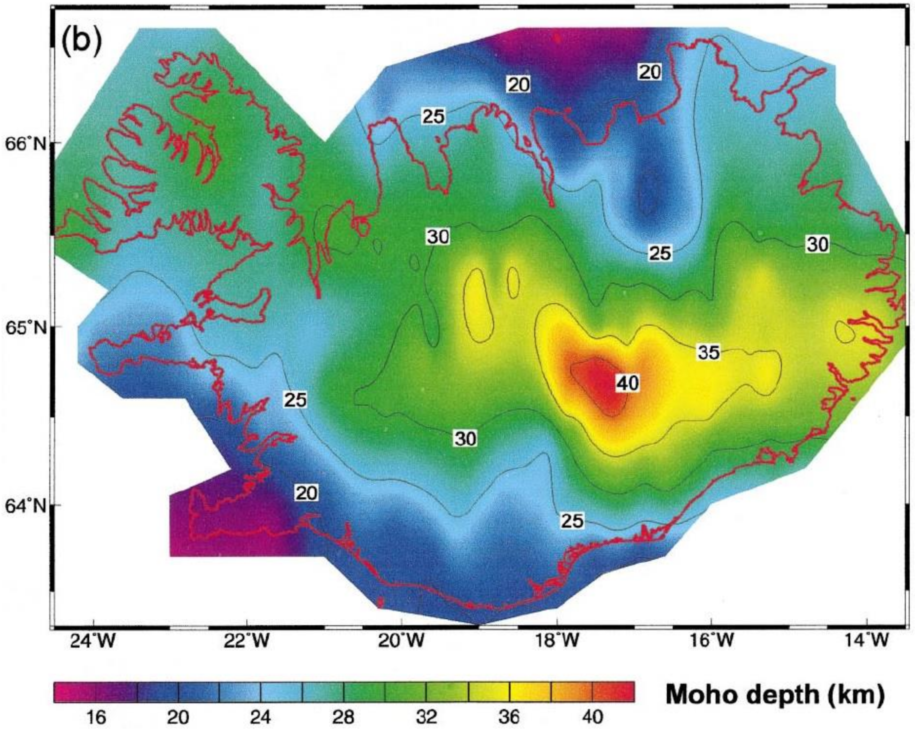
Iceland hot spot

- Iceland is a product of the activity of a mantle plume, which ca. 70 Ma (Late Cretaceous) was located below Western Greenland and the Canadian Arctic.
- The young mid-ocean ridge came under the influence of the mantle plume in the Eocene ca. 40 Ma.

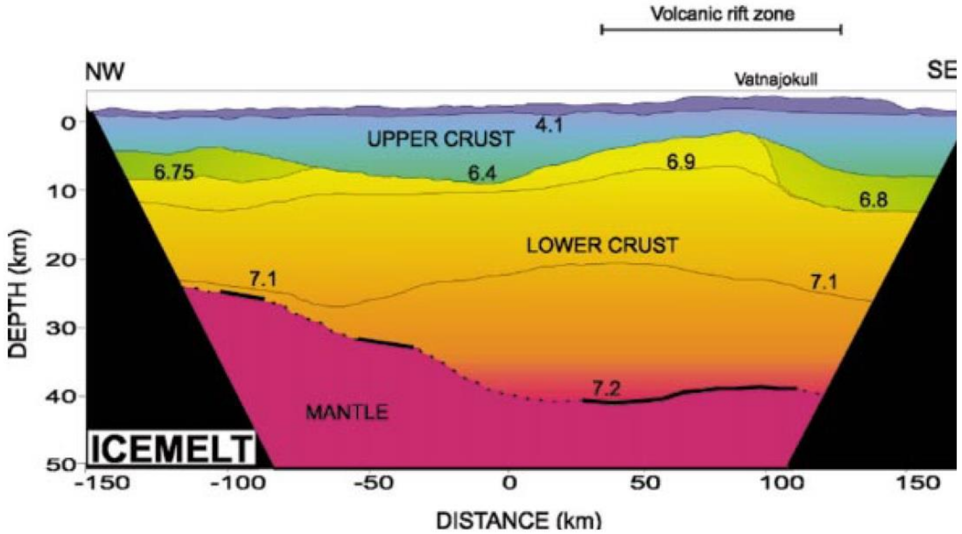


Iceland Crust



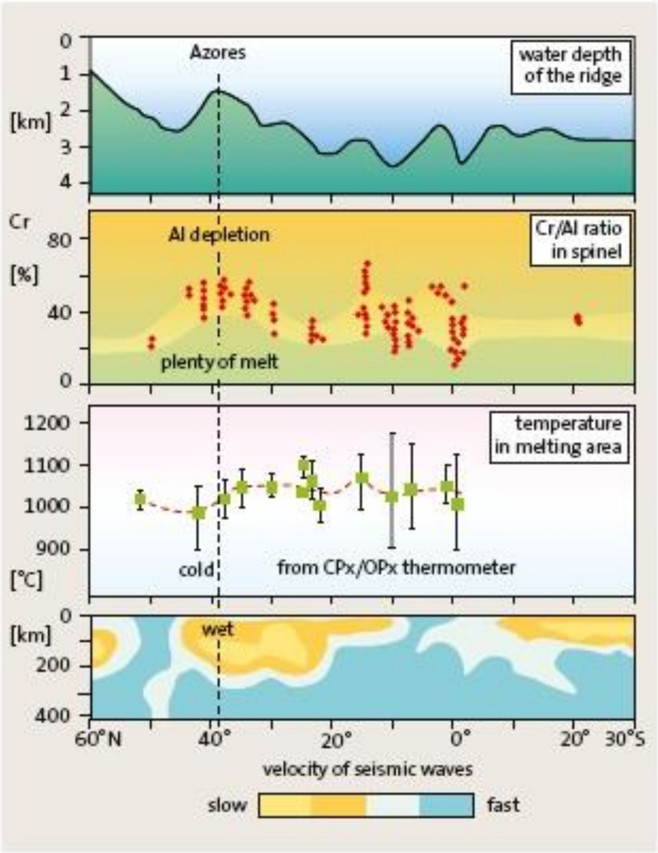


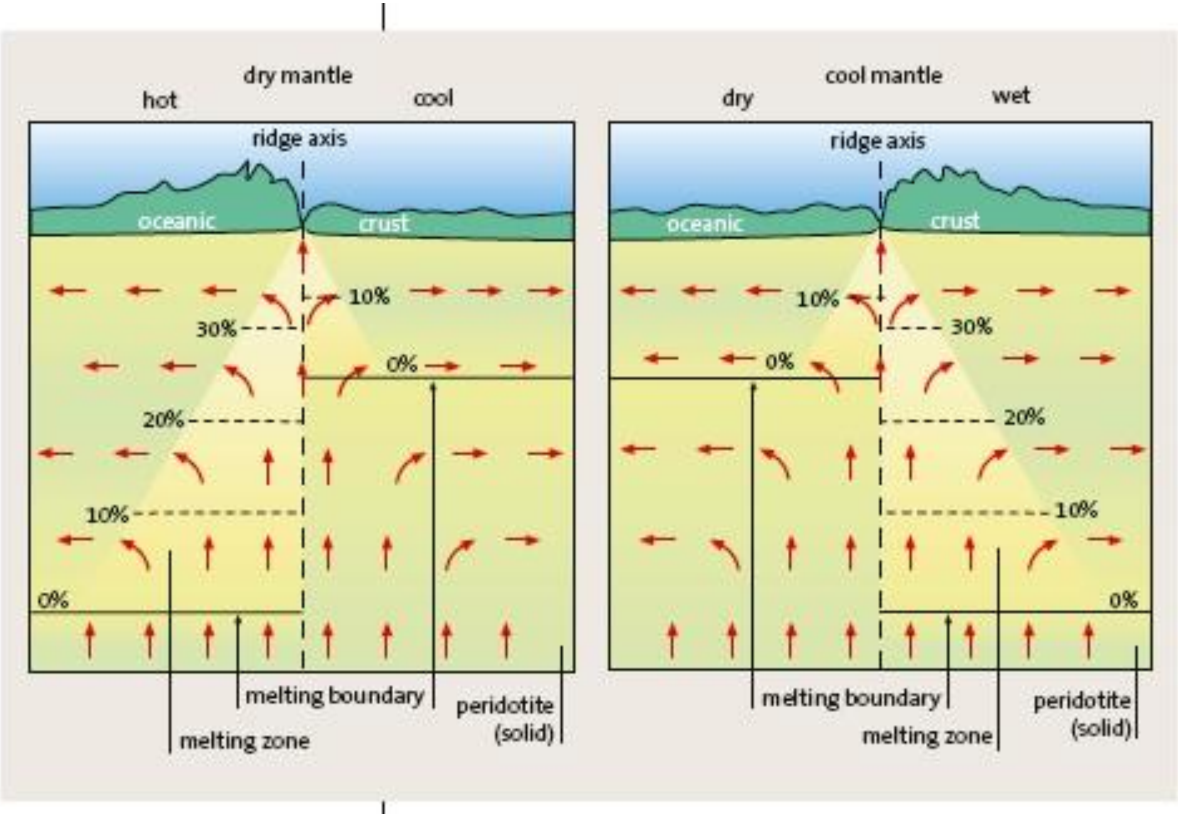
Darbyshire, 2000, EPSL, 181



Darbyshire, 1998, GJI, 135

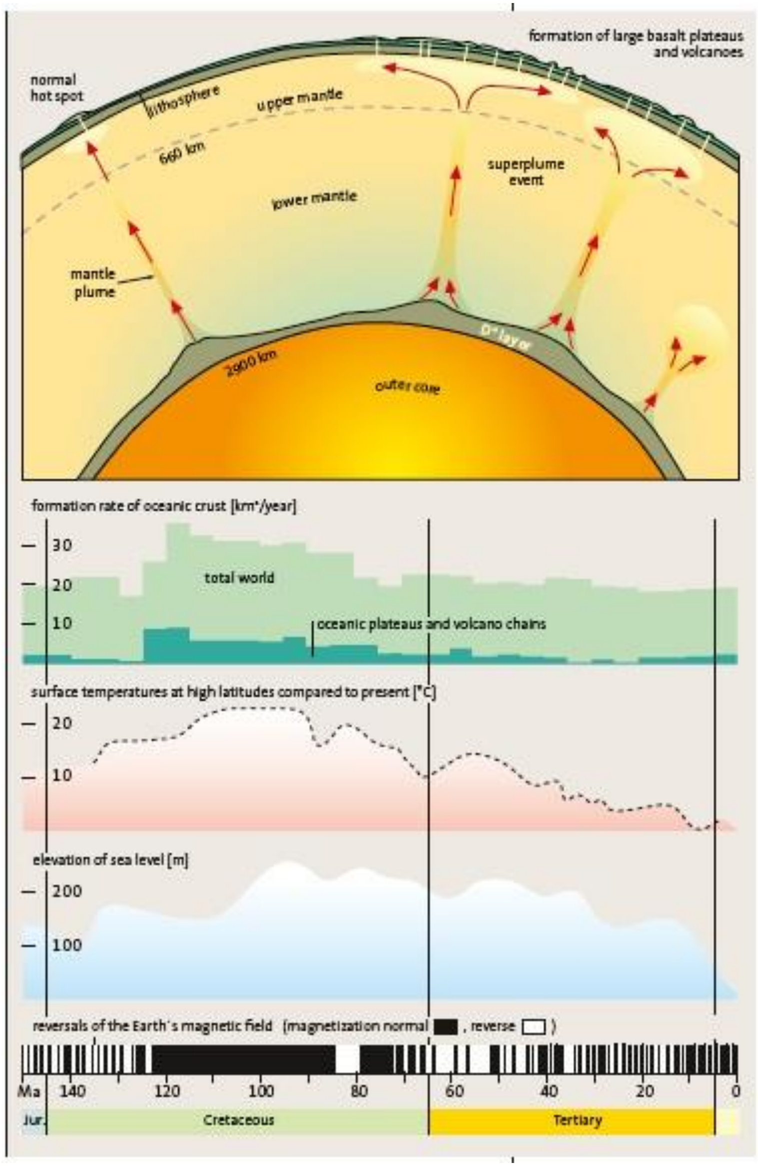
Azores: hot or cold spot?



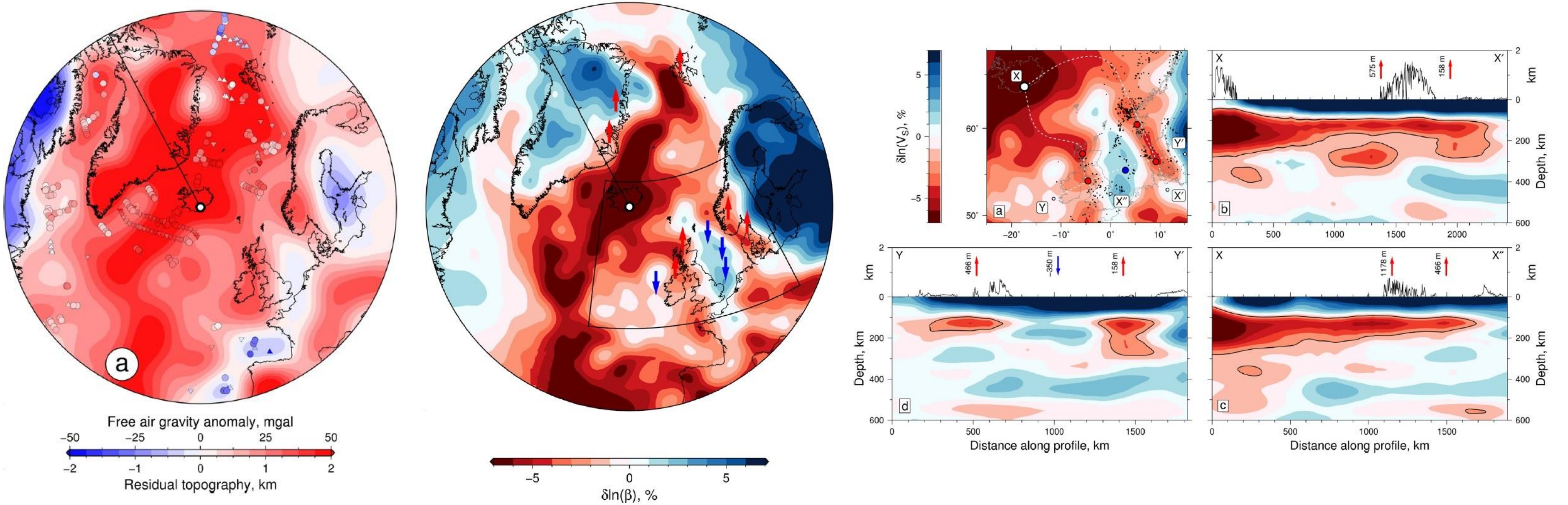


Cretaceous Superplume event

- The Cretaceous from ca. 125 Ma to 85 Ma was a time of extremes: extreme conditions were caused by the exceptionally high activity of mantle plumes that resulted in many large and unusually productive hot spots on the surface.
- The total production of oceanic crust increased, within a time interval of only a few million years, from around 20 km^3 per year to ~35 km^3 per year.



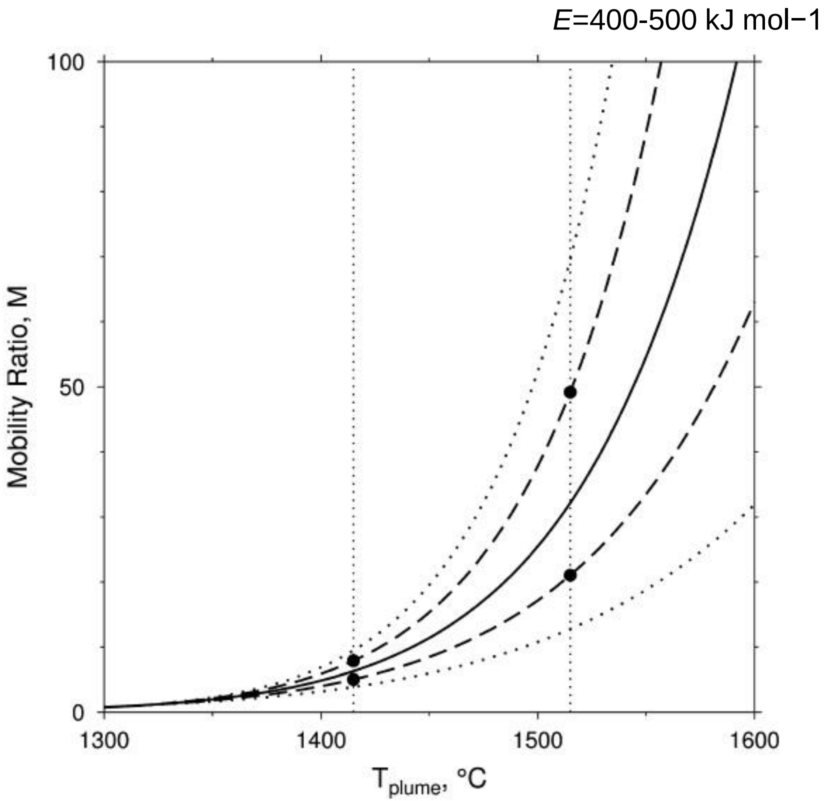
Radial viscous fingering generated by plumes



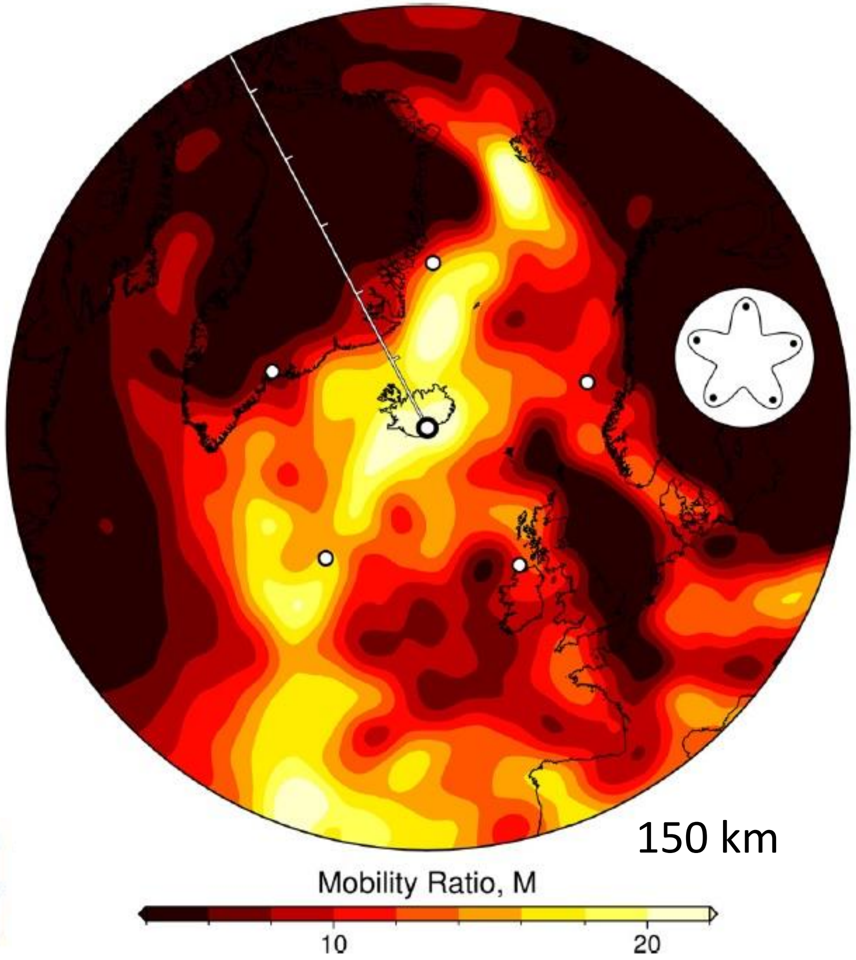
Schoonman et al., 2017, EPSL 468

There are narrow, slow velocity fingers (low velocity anomaly > 2%) that protrude beneath the fringing continental margins (British Isles and western Norway)

Radial viscous fingering generated by plumes



Schoonman et al., 2017, EPSL 468



$$M = \frac{\eta_r}{\eta} \qquad \eta = \eta_r \exp\left\{ \frac{E}{R} \left(\frac{1}{(T + 273)} - \frac{1}{(T_r + 273)} \right) \right\}$$

η_r= mantle viscosity

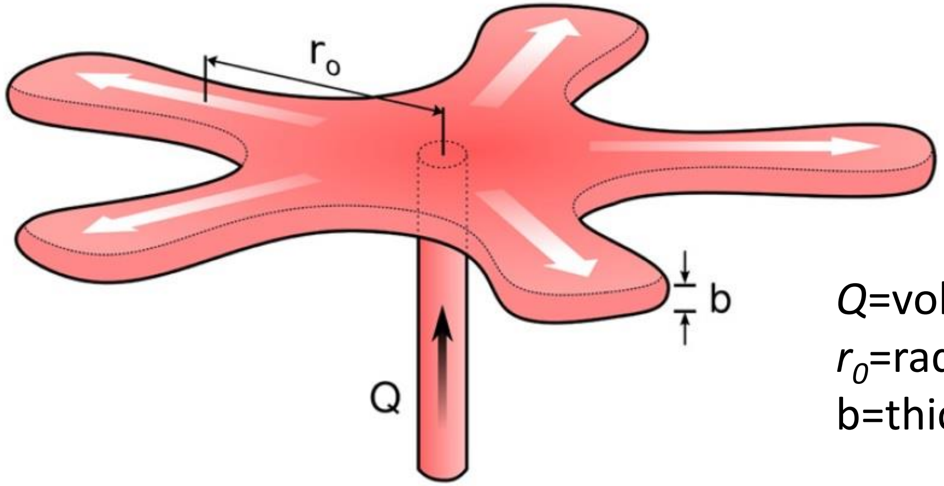
η= plume viscosity

Tr=mantle temperature

T= plume temperature

Saffman-Taylor instability: When a less viscous fluid displaces a more viscous fluid, the boundary between the two fluids can become unstable and promote viscous fingering.

Radial viscous fingering generated by plumes



Q=volume flux
r_0=radius of central portion of plume
b=thickness of the layer

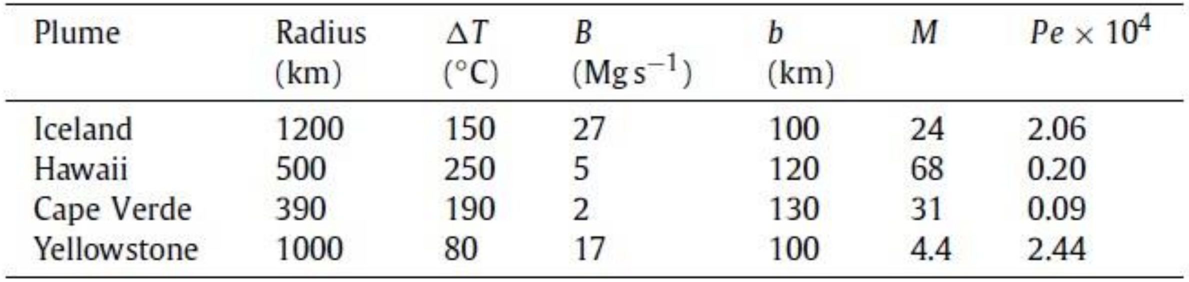
Schoonman et al., 2017, EPSL 468

Radial fingers are generated by a phenomenon known as the Saffman–Taylor instability

- Wavelength and number of fingers are controlled by the mobility ratio (i.e. the ratio of viscosities η_m/η_p), by the Péclet number (i.e. the ratio of advective and diffusive transport rates), and by the thickness of the horizontal layer into which fluid is injected.
- Iceland plume has an irregular planform due to small-scale convective circulation (radial fingers) that can generate and maintain surface deformation on short length scales.

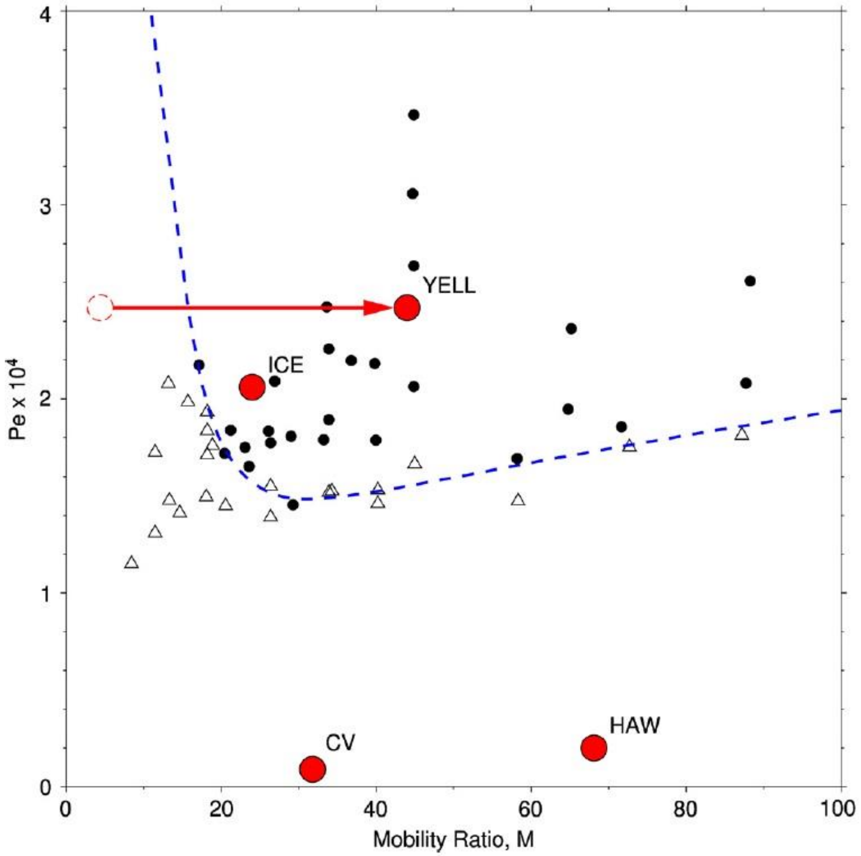
Radial viscous fingering generated by plumes

(experimental analysis)



Plume	Radius (km)	ΔT (°C)	B ($\mathrm{Mg\,s^{-1}}$)	b (km)	M	$Pe \times 10^4$
Iceland	1200	150	27	100	24	2.06
Hawaii	500	250	5	120	68	0.20
Cape Verde	390	190	2	130	31	0.09
Yellowstone	1000	80	17	100	4.4	2.44

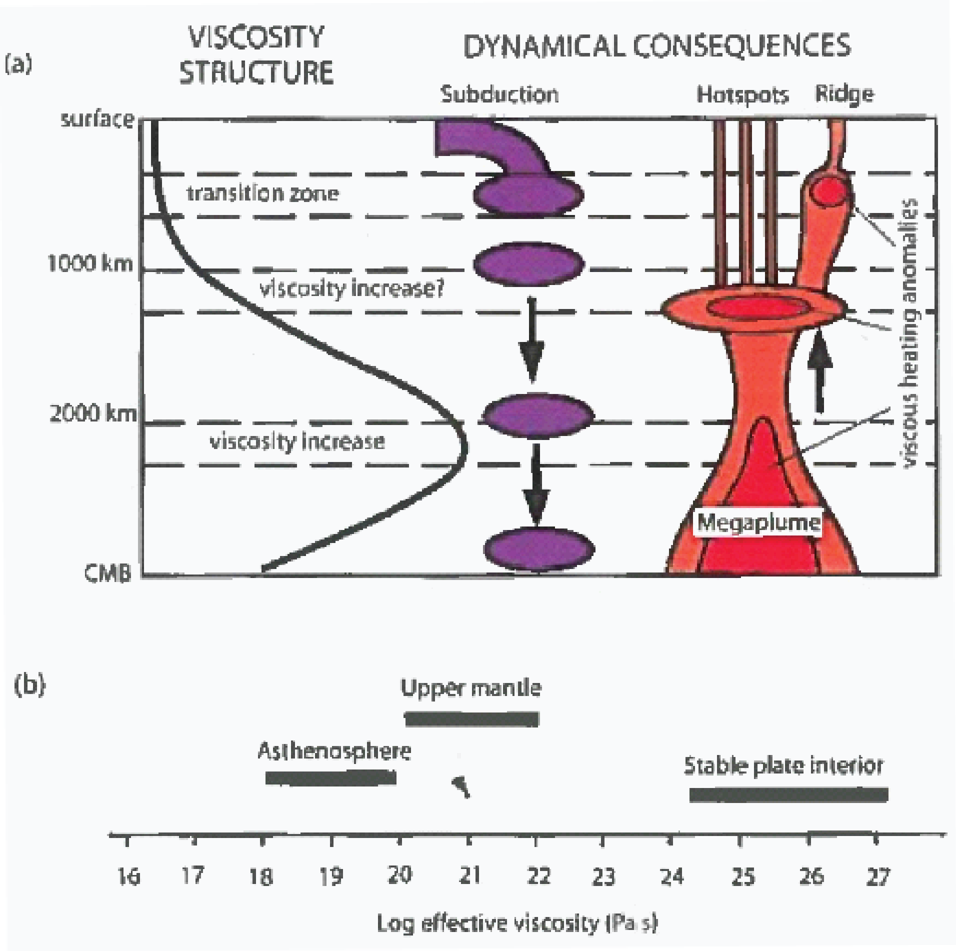
Schoonman et al., 2017, EPSL 468



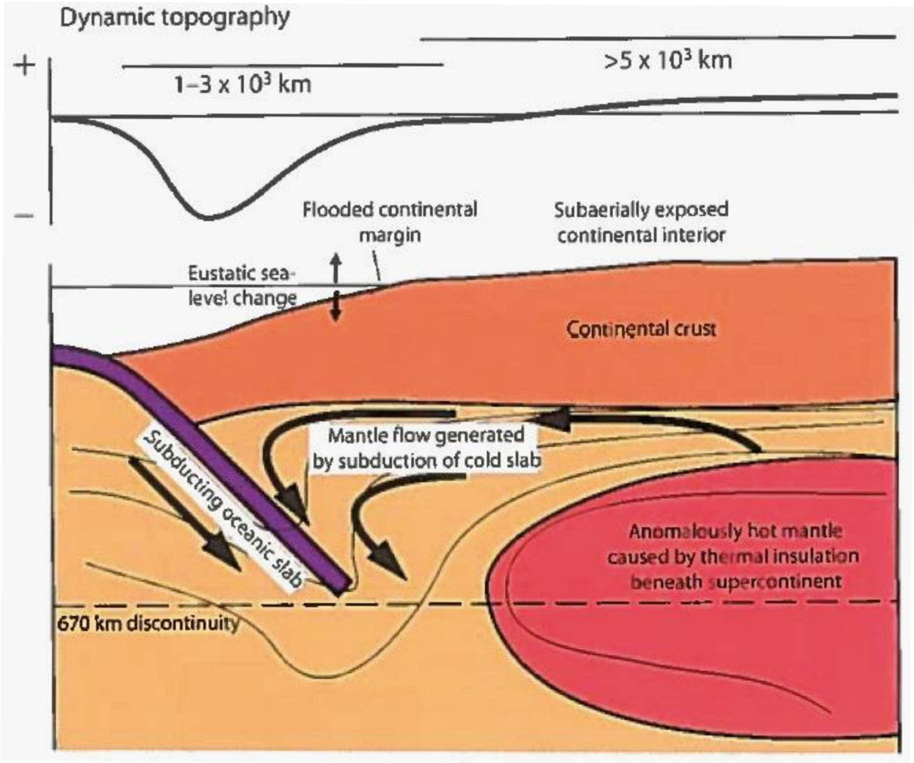
- Absence of fingering is principally a consequence of smaller buoyancy fluxes (Hawaiian and Cape Verde plumes).
- Yellowstone plume has likely a high mobility because of the presence of minor fractions of hydrous melt (this plume has an excess asthenospheric *T* of not more than 55–80◦C).

Dynamic topography

Vertical displacement of the Earth's surface generated in response to flow within the mantle

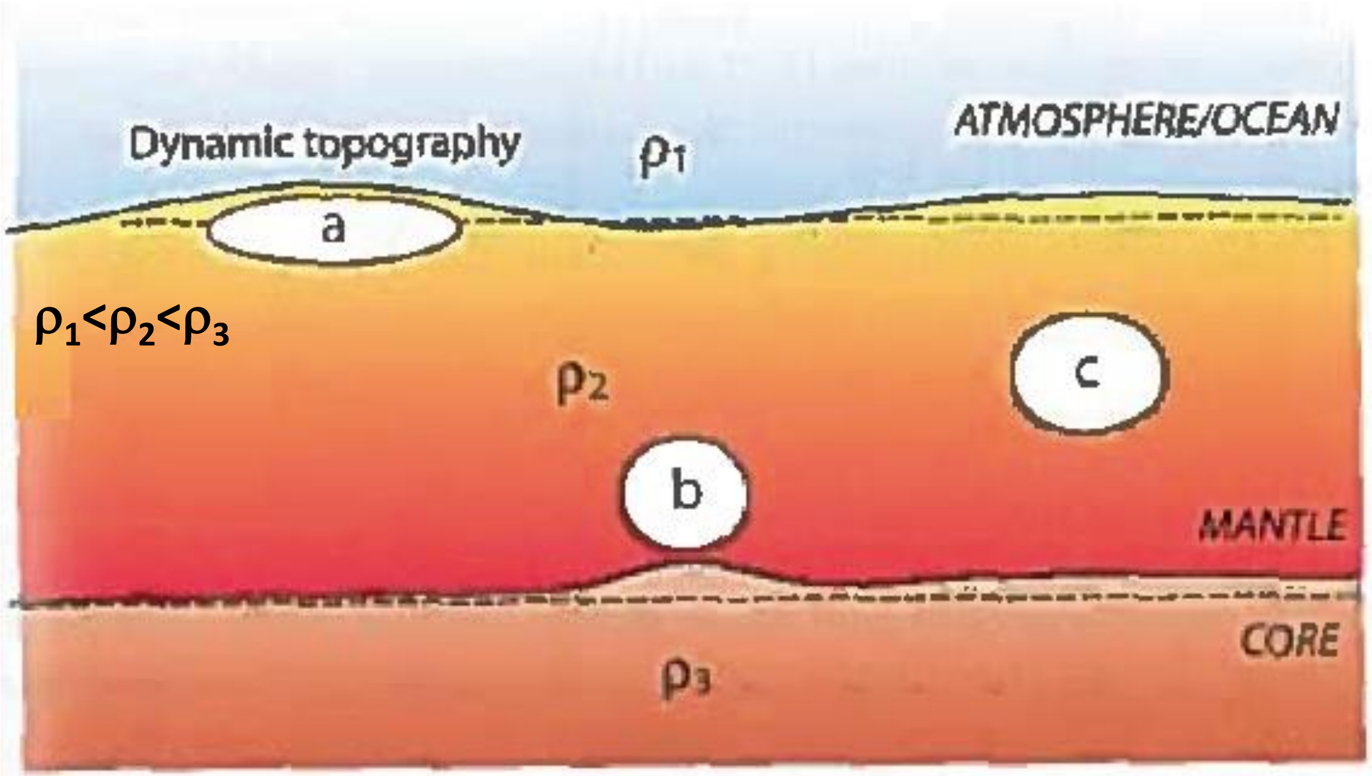


- Dynamic topography is transient and is usually of the order of few hundred meters (300-500 m).
- Response time of the mantle to a disturbance depends on its viscosity and wavelengths of the anomalous body (e.g., λ 1-3 x 10^3 km, t= 10^4 yr).
- It is estimated by removing from the topography the isostatic effects of lithospheric thickness changes and thermal subsidence.



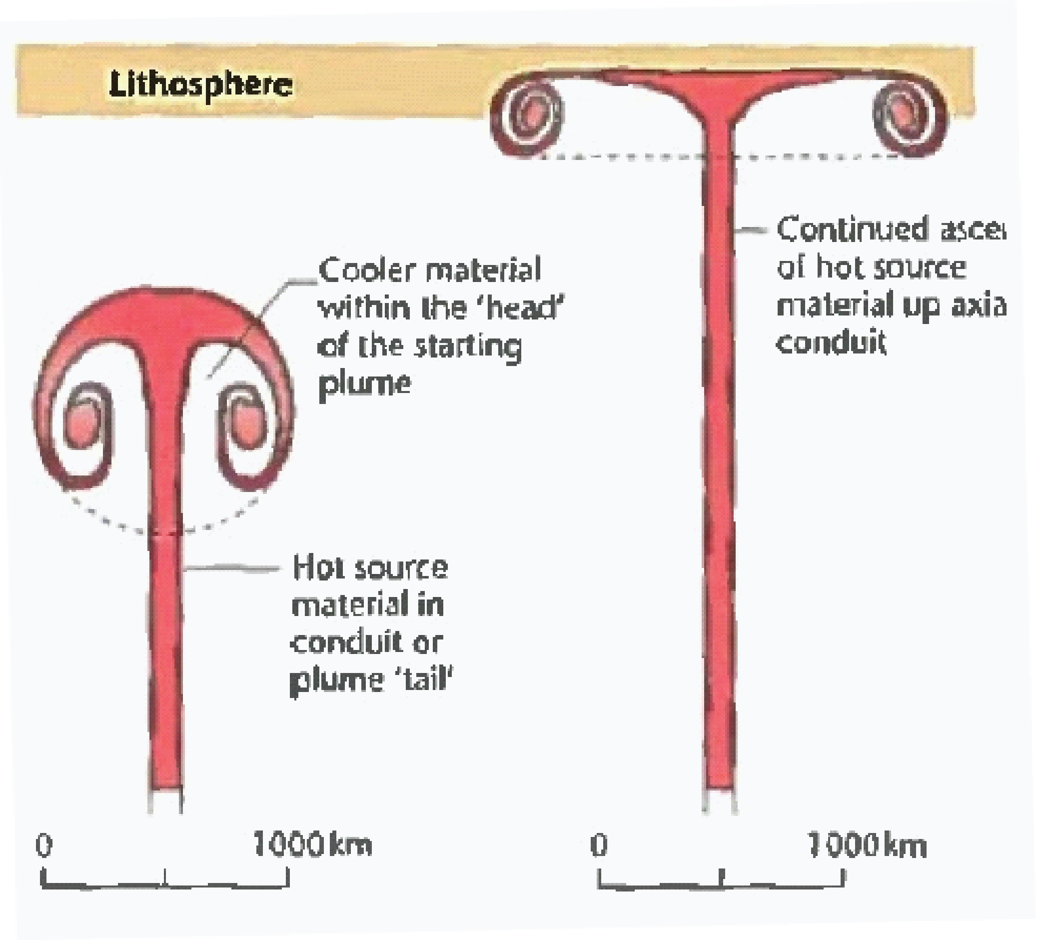
Dynamic topography

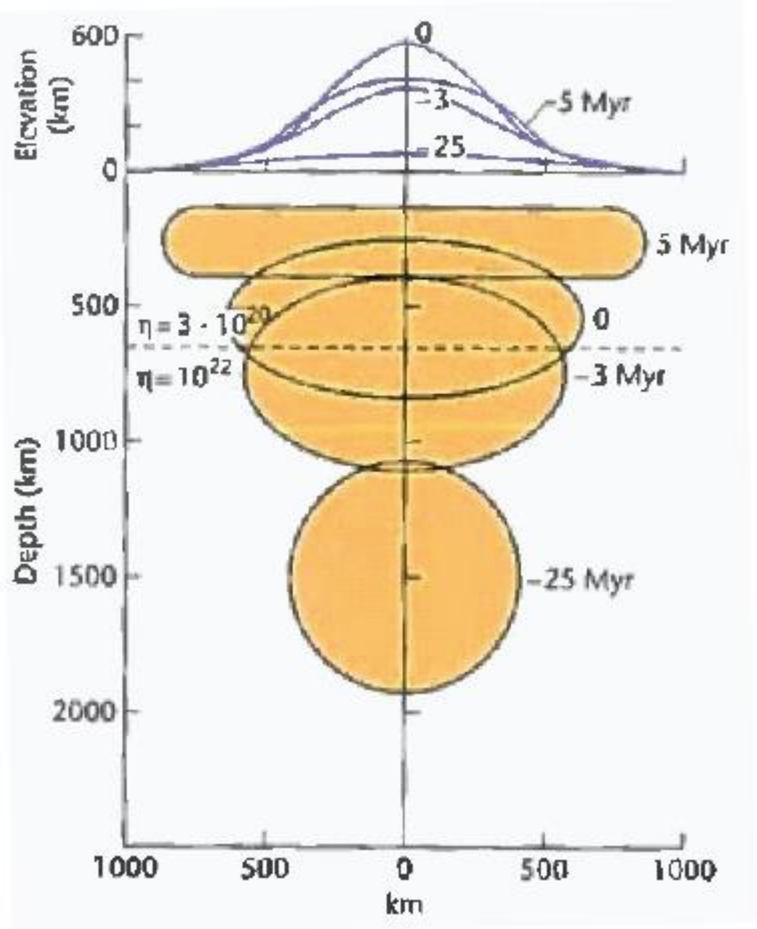
- Buoyancy in a fluid layer deflects both the top and the bottom surfaces of the fluid and the combined weight of the topographies balances the internal buoyancy.
- The amount of deflection of each surface depends on the magnitude of the viscous stresses transmitted to each surface, which depends on the distance from the buoyancy to the surface and viscosity.



Dynamic topography

(due to plumes)



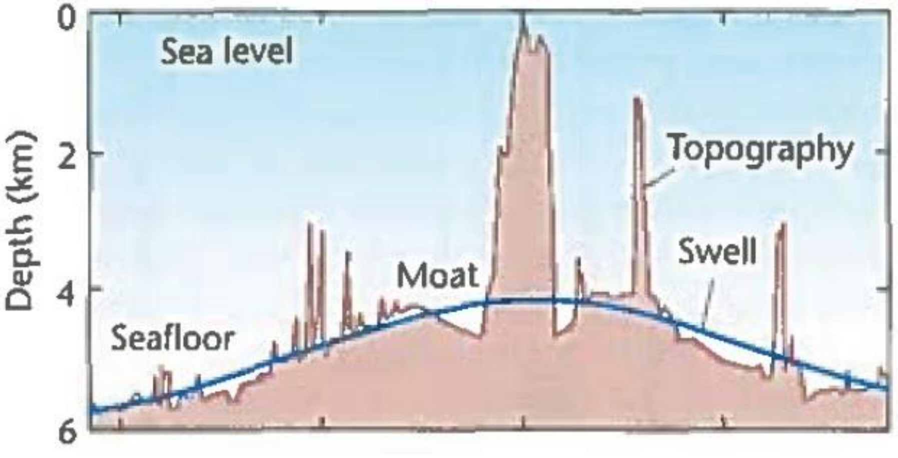


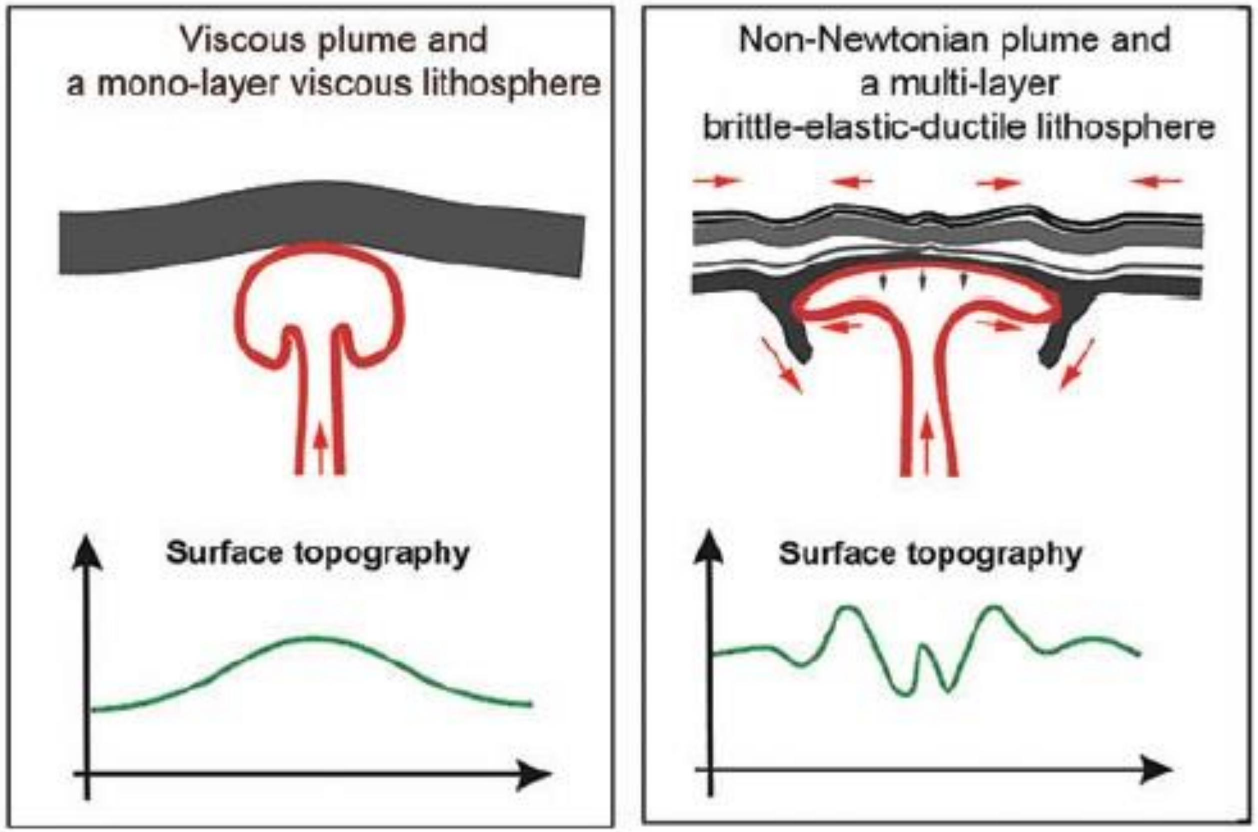
Buoyancy flux: $3\text{x}10^4$N s^{-1} ΔT=300 K

- Surface is weekly uplifted if the plume head is within the lower mantle (-25 Myr).
- Surface uplift takes place rapidly when the head's plume reaches the asthenosphere.
- Maximum uplift depends on (1) penetration of the hot plume into the cold lithosphere (2) a volume increase caused by melting (when it reaches shallow depths).

Plume and topography

Hawaiian Ridge

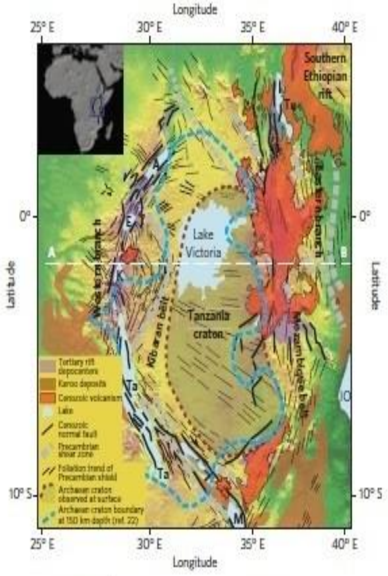


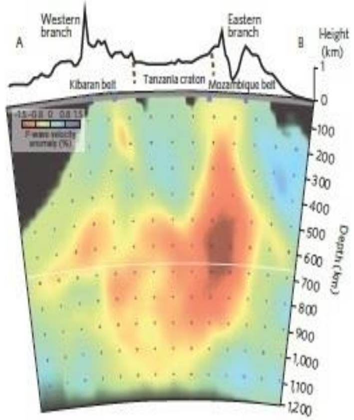


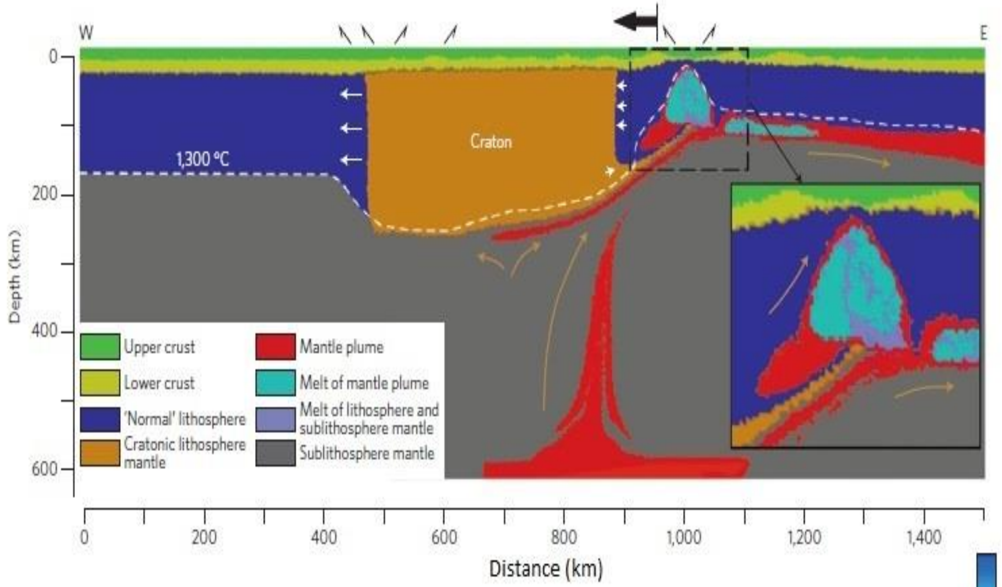
Burov and Cloetingh, 2009, Geophys. J. Int., 178

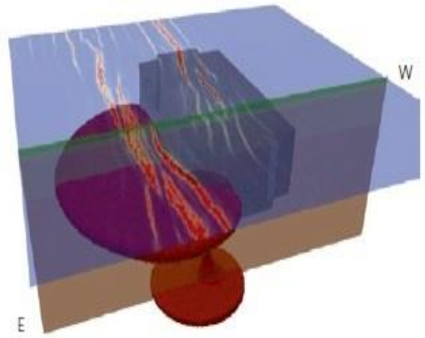
- The conventional models predict only long-wavelength (controlled by the plume head size) isostatic topography due to plume impact.
- Accounting for plate rheology and multilayer lithosphere structure yields a more complex response, with several short-wavelengths generated by intraplate deformation, tectonic-style deformation at surface and strong lithospheric mantle erosion at depth.

Plume beneath a craton
(Tanzania Craton)

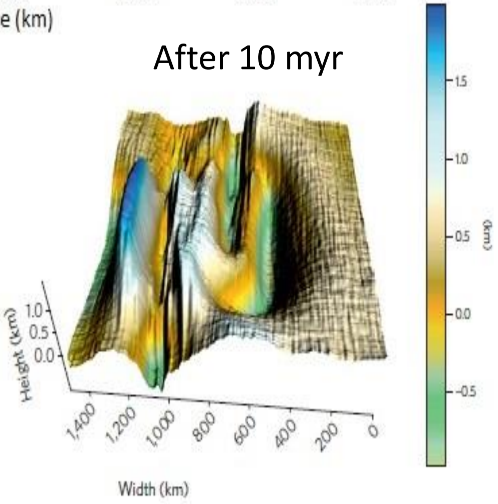








After 10 myr



- **The Tanzanian craton is surrounded on both sides by active rift branches**: **(1)** the magma-poor western rift exhibits low-volume volcanic activity, large (*M*>6.5) magnitude earthquakes, and hypocentre depths reaching 30–40 km, and **(2)** the magma-rich eastern rift is characterized by a broad zone of shallow (5–15 km) and lower magnitude seismicity, but voluminous Cenozoic volcanism.
- Surface topography first reacts by domal uplift, soon after (<1Myr) replaced by subsidence and coeval initiation of long and narrow rifted basins on either side of the craton.
- These basins form above a thinning lithosphere, creating channels for the subsequent migration of mantle plume material.

Koptev et al., 2014, Nature Geoscience

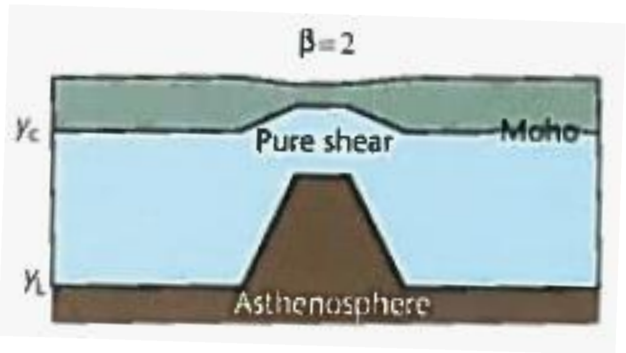
The plume is deflected by the cratonic keel and preferentially channelled along one of its sides, leading to the coeval development of magma-rich and magma-poor rifts along opposite craton sides, fed by melt from a single mantle source.

Melt generation during continental extension

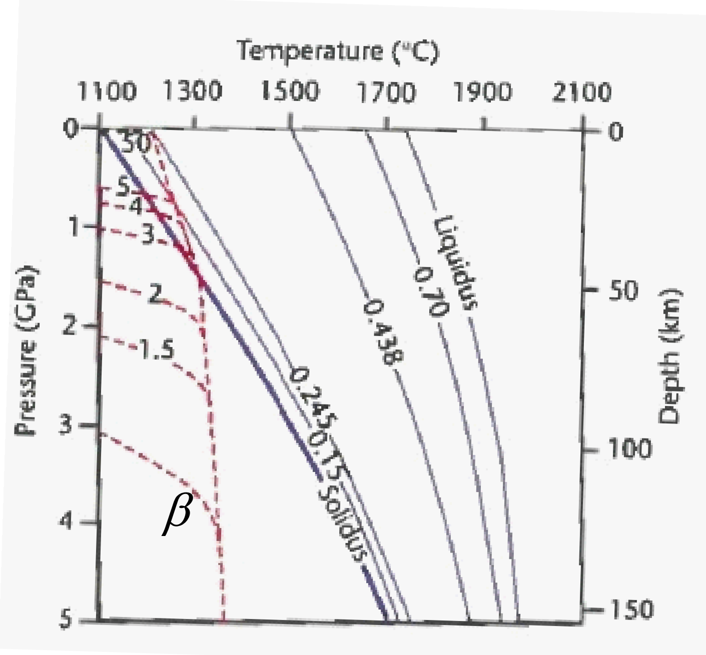
Amount of melt generated during the lithospheric stretching depends on the T_P of the asthenosphere (T that the asthenosphere would have if brought to the surface adiabatically without melting) and amount of stretching.

For β=2, T_p= 1400°C (due to plume activity), and L= 100 km, $\text{Thick}_{\text{MELT}}$= 2 km

Large T_p increases the amount of MgO and decreases NA_2O (from alkali basalts to tholeites)

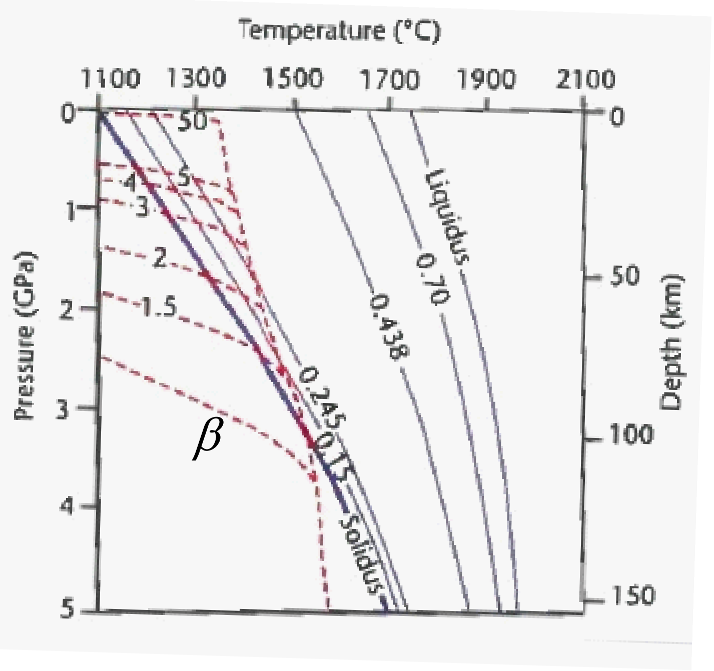


In absence of plume



T_p= 1280°C L=100 Km ν=4x10^{15} m^{2}s^{-1}

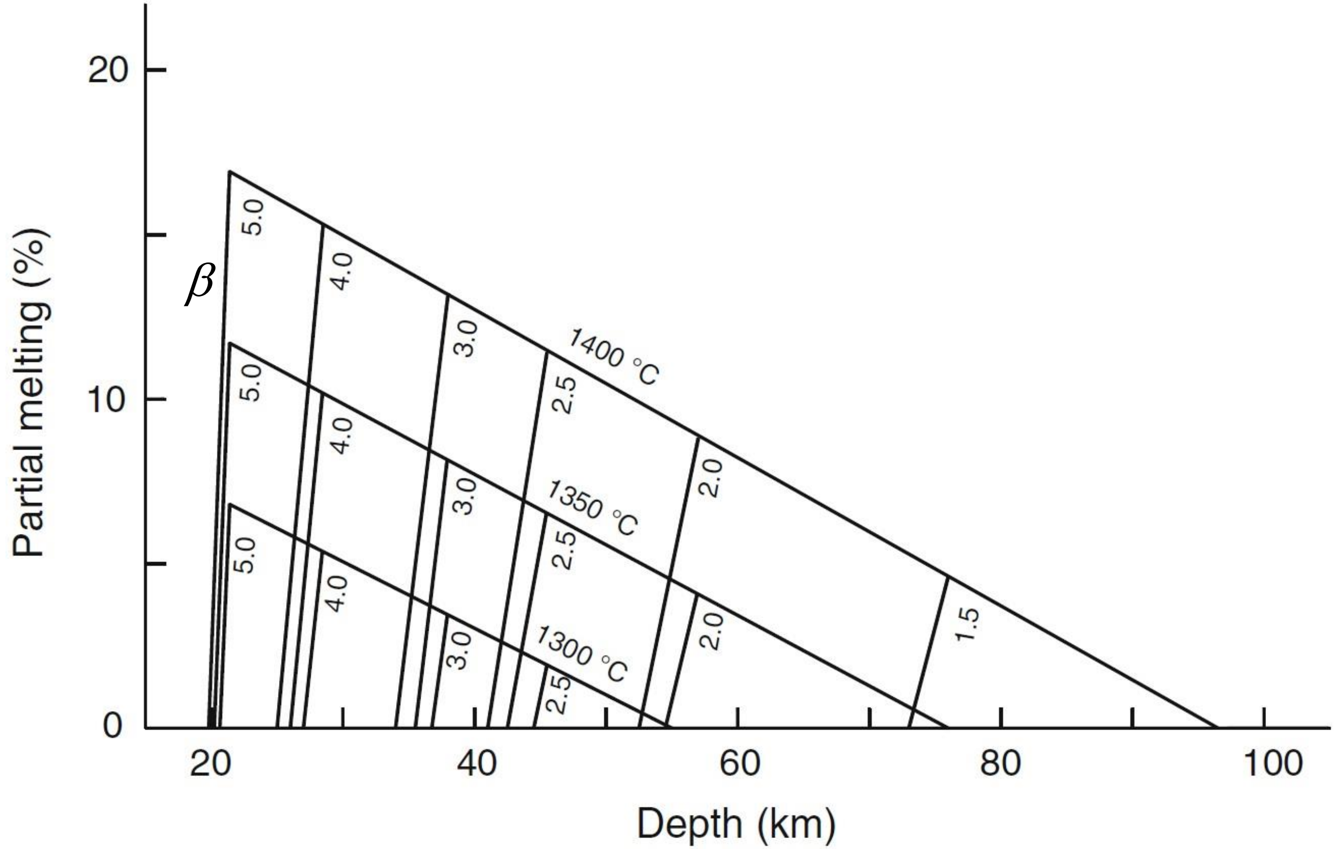
In presence of plume



T_p= 1480°C L=100 Km ν=4x10^{15} m^{2}s^{-1}

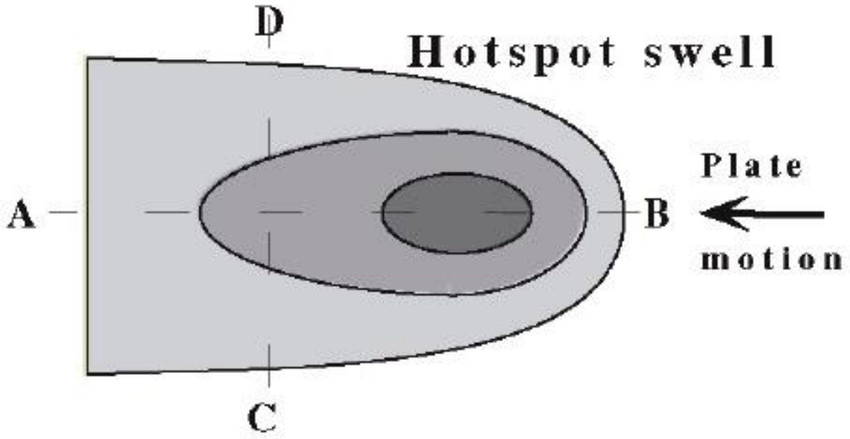
Melt generation during continental extension

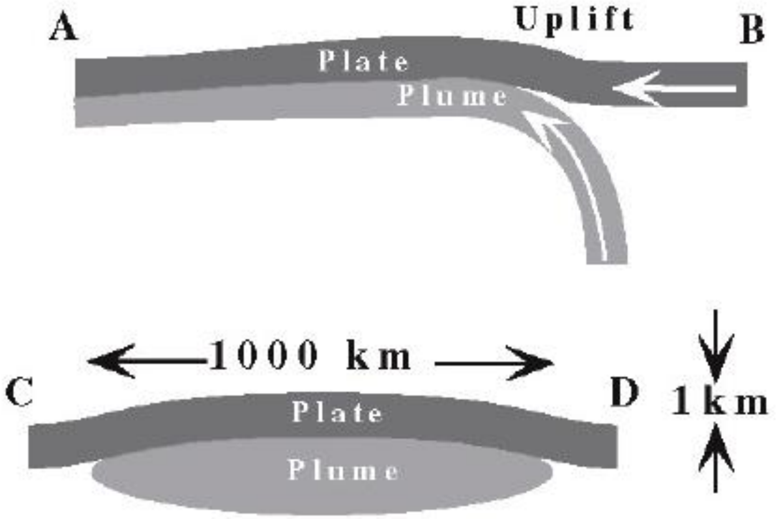
- For a thermally normal asthenosphere (T_p = 1300°C), no melting occurs unless stretching exceeds values of 2.0.
- The contribution of heat flow from melt cooling lasts several years after the rift episode and even for larger periods than the characteristic thermal time or the solidification time of the melt layers



Partial melting as a function of depth at various amounts of extension β (1.5, 2.0, 2.5, 3.0, 4.0 and 5.0) and for asthenosphere T_p (1300, 1350, and 1400 °C)

Buoyancy flow rate and heat flow rate





There is a close isostatic balance between the weight of the excess topography created by this uplift and the buoyancy of the plume material under the plate.

Buoyancy flow rate: $b = g\Delta\rho \cdot \pi r^2 u$ $\Delta\rho = (\rho_\mathrm{p} - \rho_\mathrm{m})$ *r*=radius of vertical cilinder *u*=average velocity

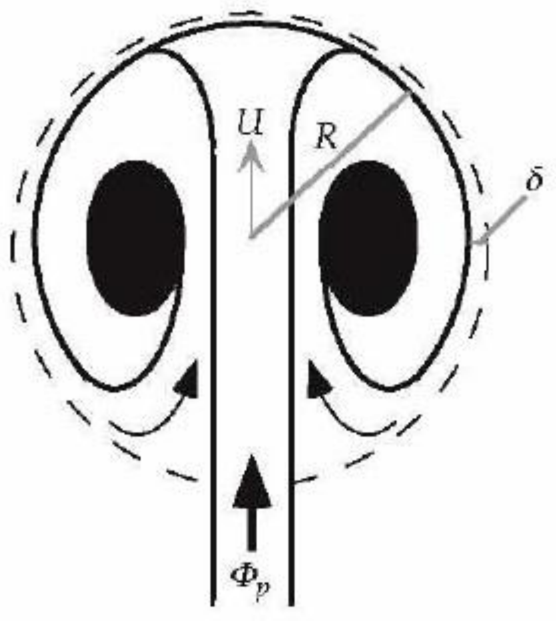
$\Delta T = T_\mathrm{p} - T_\mathrm{m}$ $\rho_\mathrm{p} - \rho_\mathrm{m} = \rho_\mathrm{m}\alpha\Delta T$ $b=g\rho_m\alpha\Delta T\pi r^2 u$

The addition to swell topography each year is equivalent to elevating by a height h = 1 km a strip of sea floor with a width w = 1000 km (the width of the swell) and a length $v\delta t$ = 100 mm.
The rate of addition to the weight (negative buoyancy) of the new swell is: $W = g(\rho_\mathrm{m} - \rho_\mathrm{w})wvh = b$

(the effective difference in density is between the mantle, ρ_m, and see water, ρ_w)

Heat flow rate: $Q = \pi r^2 u \rho_\mathrm{m} C_P \Delta T$ $Q = C_P b / g\alpha$

Volumetric flow rate and heat flow rate



The volumetric flow rate is related to the buoyancy flow:

$$\Phi_{\mathrm{p}} = b/g\rho_{\mathrm{m}}\alpha\Delta T$$

$$b=g\rho_m\alpha\Delta T\pi r^2u$$

b is also related to the rate at which the swell volume is created, ϕ_s, through the weight of topography, *W:* $W = g(\rho_{\mathrm{m}} - \rho_{\mathrm{w}})wvh = b$

$$\Phi_{\mathrm{s}} = wvh = W/g(\rho_{\mathrm{m}} - \rho_{\mathrm{w}}) = b/g(\rho_{\mathrm{m}} - \rho_{\mathrm{w}})$$

The plume volumetric flow rate is related to the swell volumetric rate of creation through:

$\phi_P=\pi r^2 u$

(ϕ_P=volumetric flow rate at which the swell volume is created)

u=average velocity in the conduit

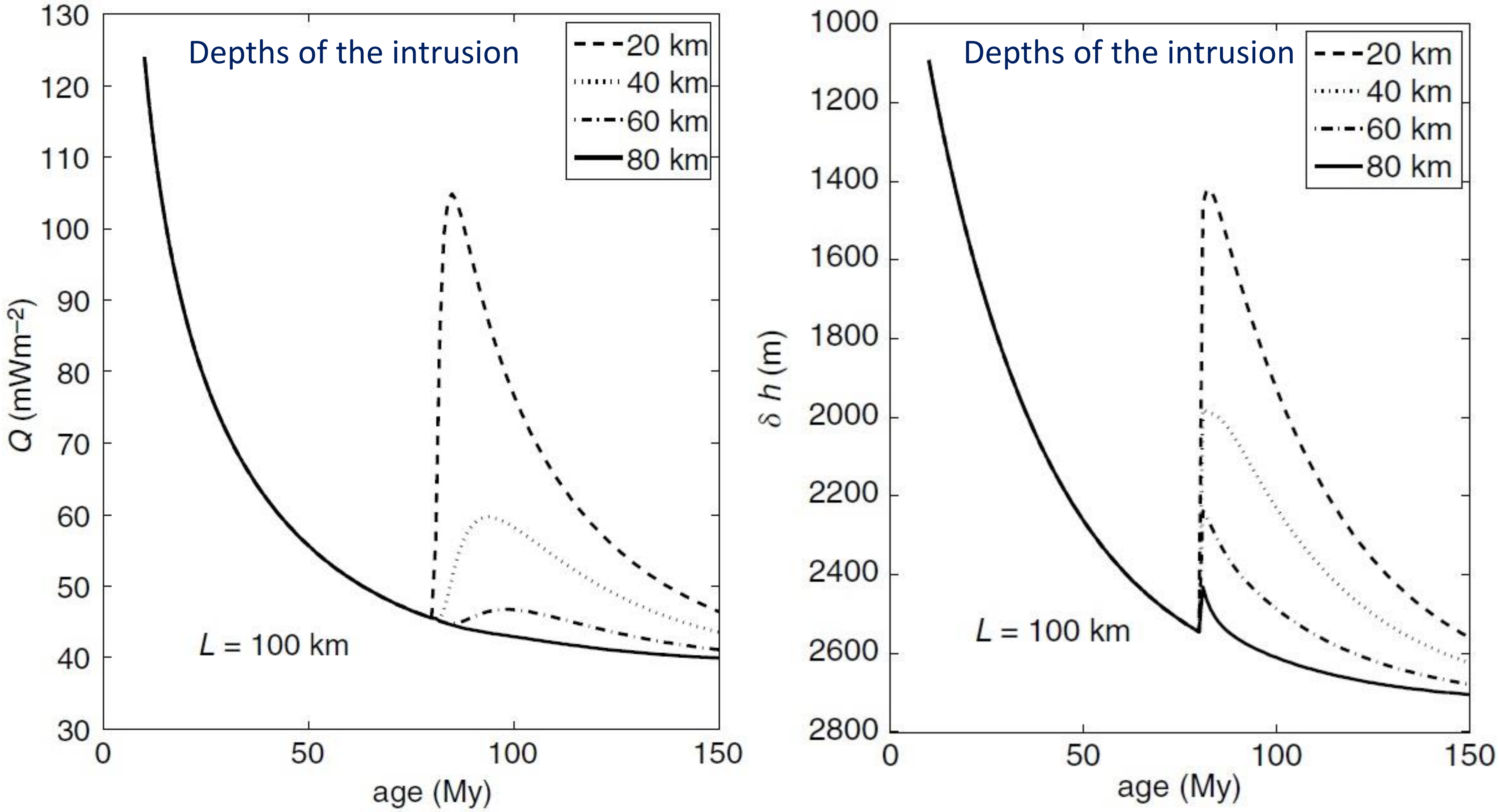
$$\Phi_{\mathrm{p}} = \Phi_{\mathrm{s}}(\rho_{\mathrm{m}} - \rho_{\mathrm{w}})/\rho_{\mathrm{m}}\alpha\Delta T$$

e.g.: ϕ_s = 0.1 km^3/yr, ρ_m = 3300 kg/m^3, ρ_w = 1000 kg/m^3, α = 3x 10^{-5} /°C, and ΔT = 300 °C
$\phi_p = \phi_s(\rho_m - \rho_w)/ \rho_m \alpha \Delta T$ = 7.5km^3/yr (75 times the rate of uplift of the swell)

The magmas usually show evidence of being derived from 5-10% partial melting of the source, which means that about 80-90% of the plume material does not melt at all.

Plume thermal perturbation

The emplacement of a plume below an oceanic plate perturbs the temperature, heat flux, and topography.



Variation of the heat flux (a) and bathymetry (b), following replacement of the lowermost lithosphere by hot material at 80My.

A swell height of δh relative to sea floor of the same age must be supported by the thermal expansion of the lithospheric column $\delta h = \alpha h\langle T\rangle$, where $\langle T\rangle$ is averaged over the entire lithospheric column.

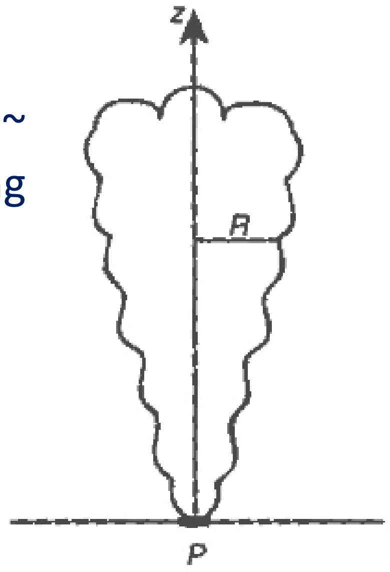
This corresponds to an input of heat $\rho_m C_p h\langle T\rangle$:

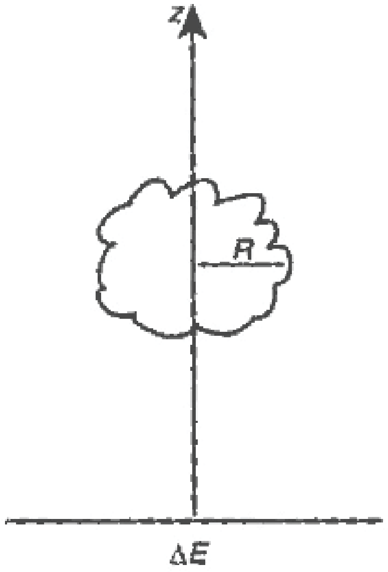
$$\Delta Q = \frac{\rho_m C_p \delta h}{\alpha}$$

The maximum of heat flux anomaly (usually < 10mWm^{-2}) is retarded by several Myr, while the bathymery changes instantanealy.

Thermals

Plume: a constant energy *E* is released for ~ 100 Myr and thus a narrow upwelling structure extends vertically.





Thermal: A fixed amount of energy *E* is released and a volume of heated fluid detaches from the source.

$$\Delta E = \rho_o C_p \Delta T V$$

ΔE is the energy needed to heat a volume V of fluid to temperature $To + \Delta T$

The thermal grows due to diffusion or turbulent entrainment, but the total amount of energy transported remains costant:

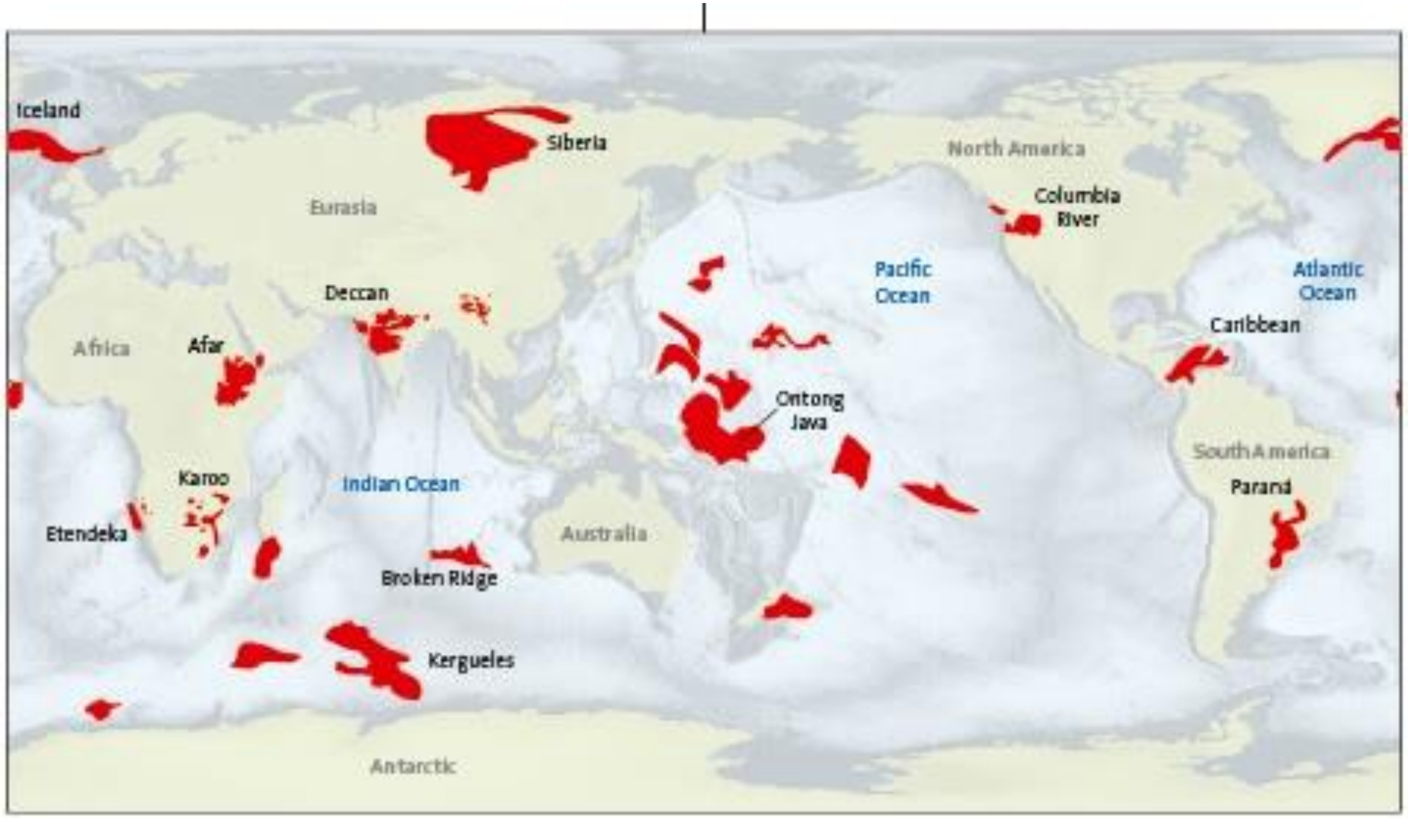
$$B = g(\rho_o - \rho)V = \rho_o g \alpha \Delta T V = \frac{g\alpha}{C_p}\Delta E$$

$$\mathrm{Ra}_E = \frac{\rho_o g \alpha \Delta T V}{\kappa \mu} = \frac{g \alpha \Delta E}{C_p \kappa \mu}$$

With Ra_E: strength of the flow, analougus to Rayleigh number

LIPs (Large igneous provinces): Large Magma Volume in few Myr

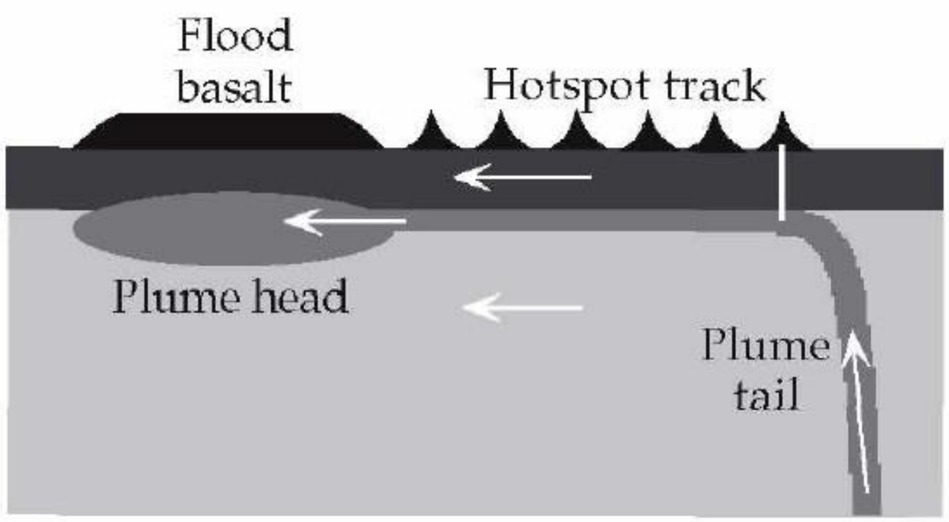
Possible origin: higher basaltic composition in the plume head from subducted oceanic slab

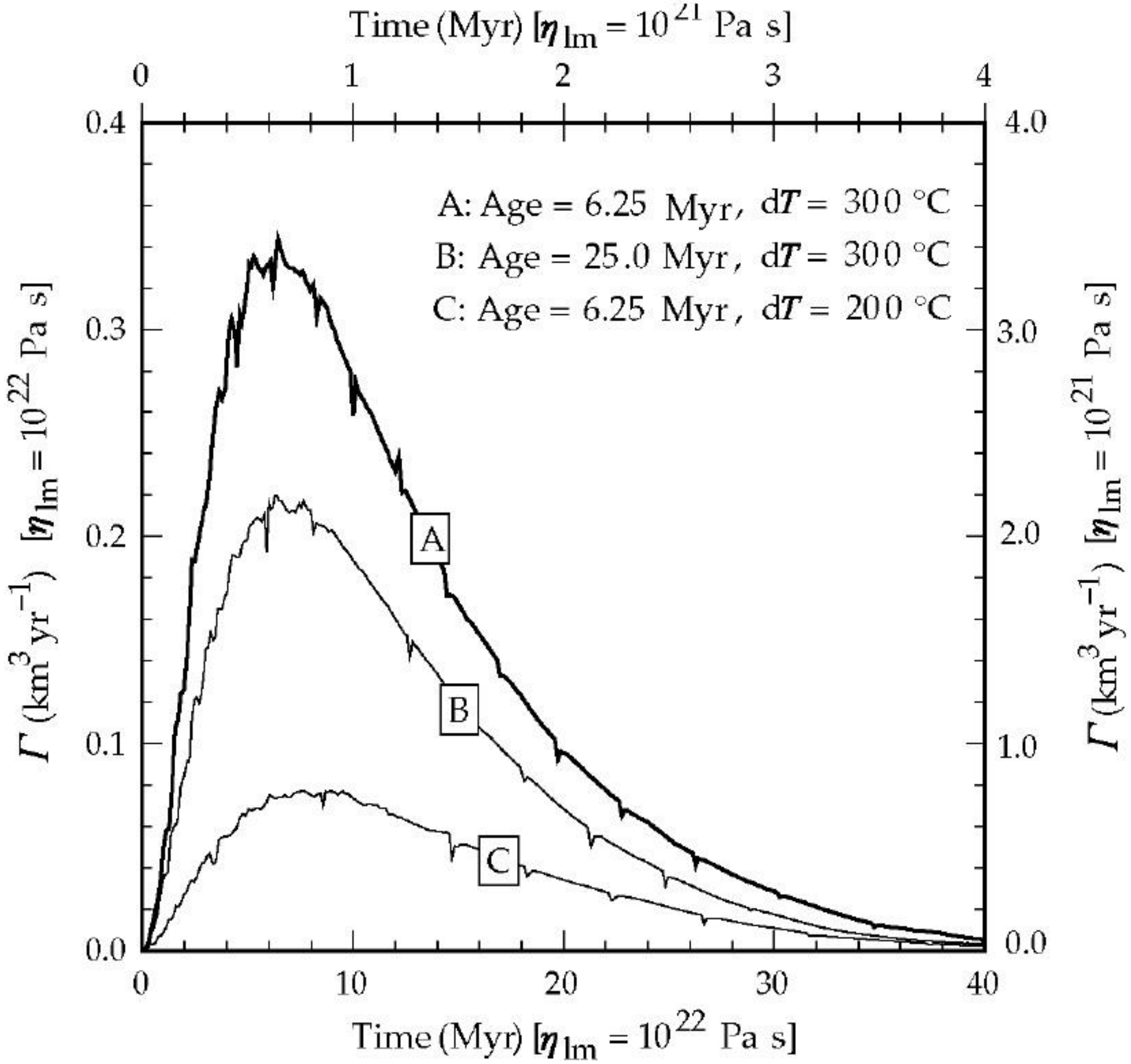


They extend up to 2000km across, several km thick, 10 million km^3 of volcanic products

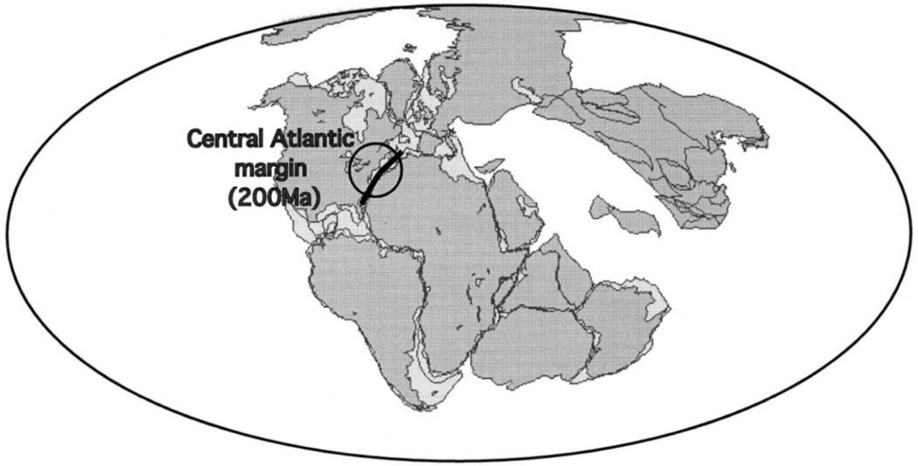
- Siberian Flood Basalts > 4 mln. km^3
- Deccan Traps ~2 mln. km^3
- North Atlantic Province >2-4 mln.km^3
- Columbia River Province ~ 0.3 mln. km^3
- Onthong-Java Plateau > 40mln. km^3

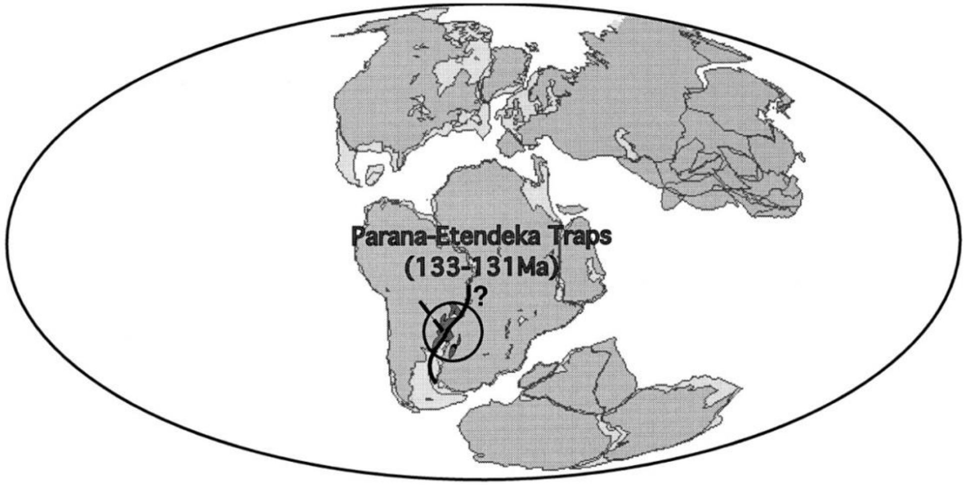
LIPs (Large igneous provinces)

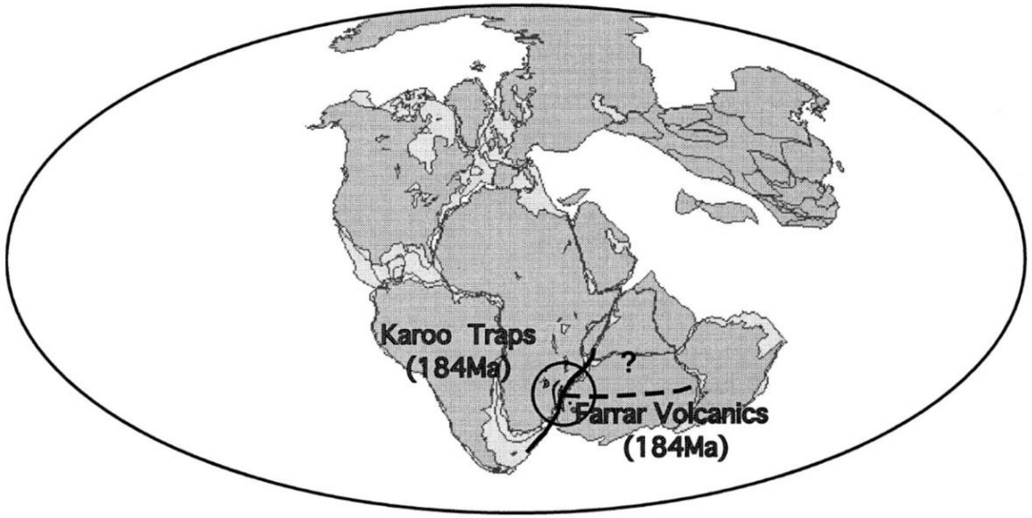




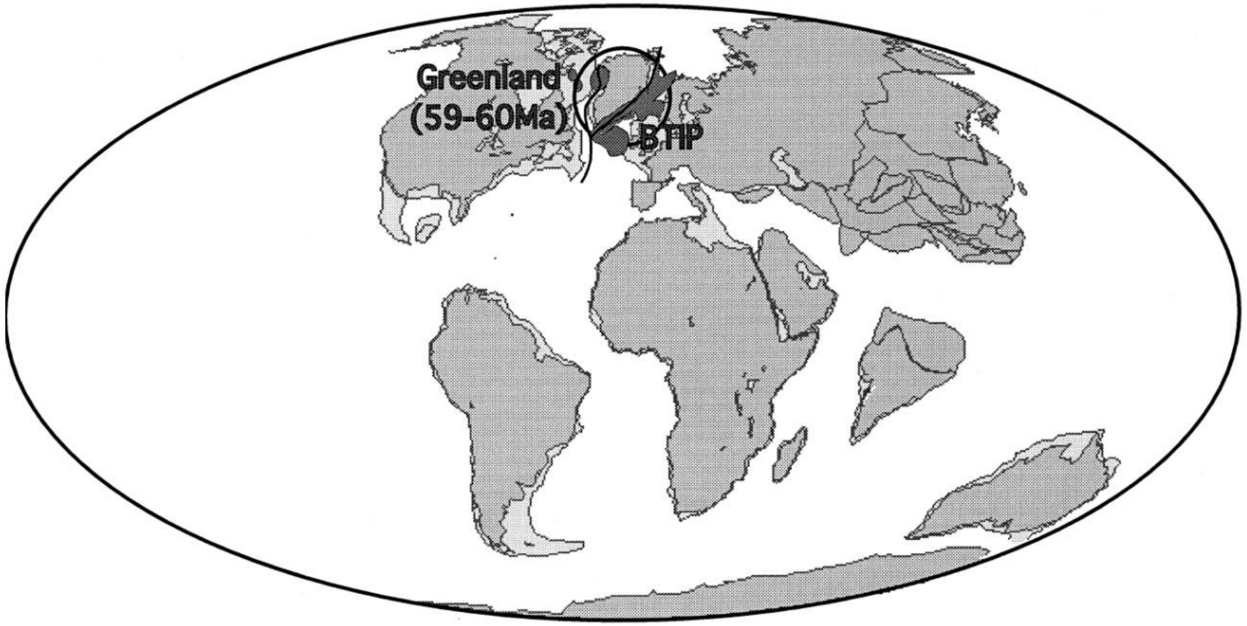
LIPs often predate continental break-up

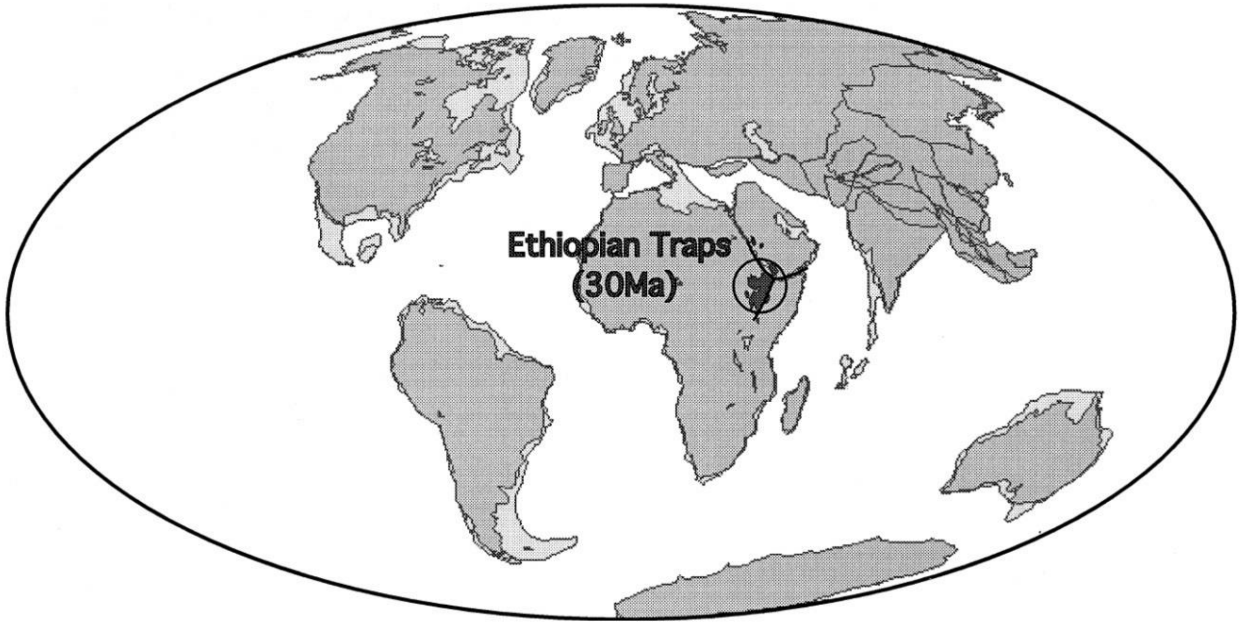






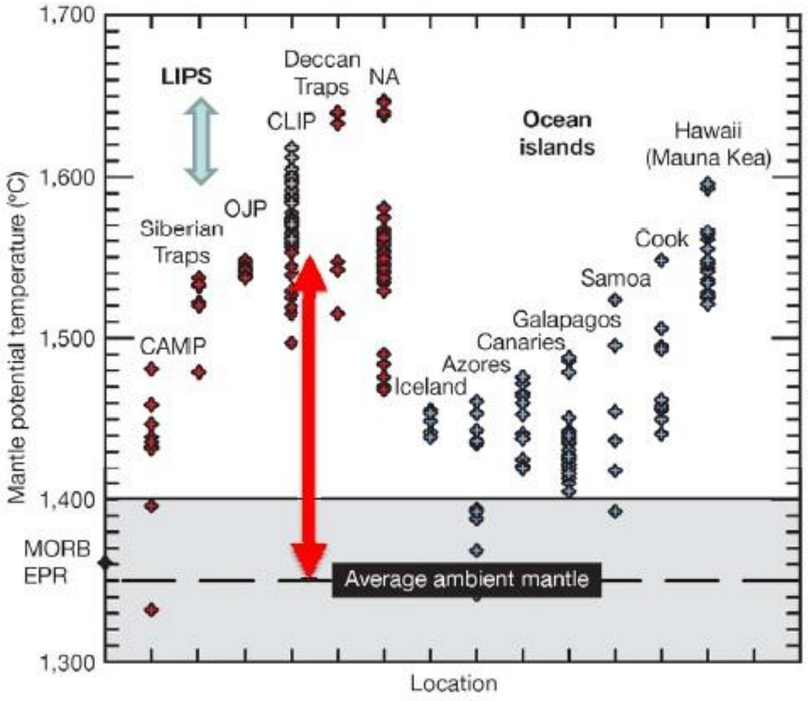
LIPs often predate continental break-up





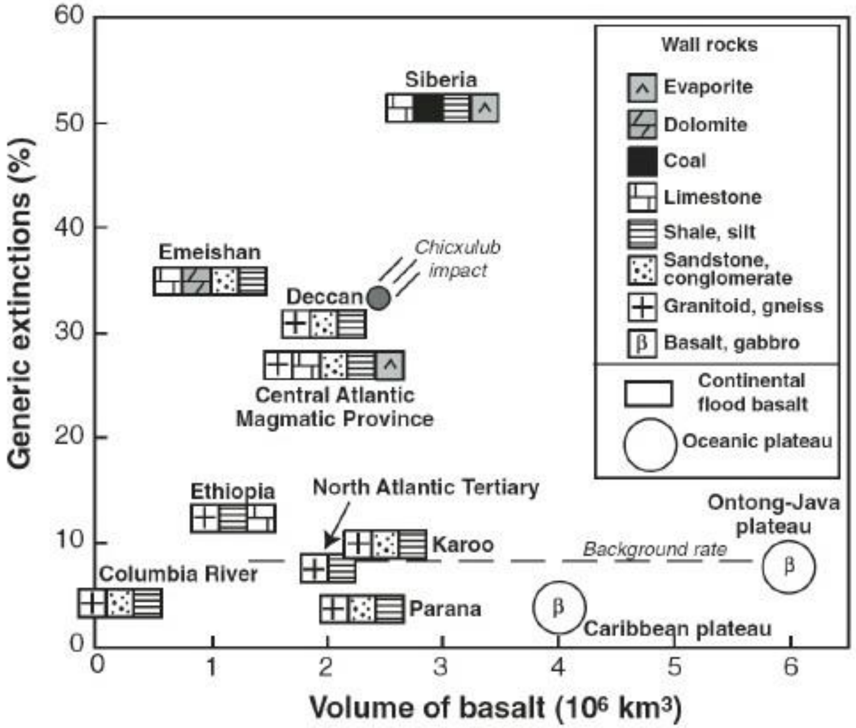
LIPs Main Characteristics

LIPs source has high temperature

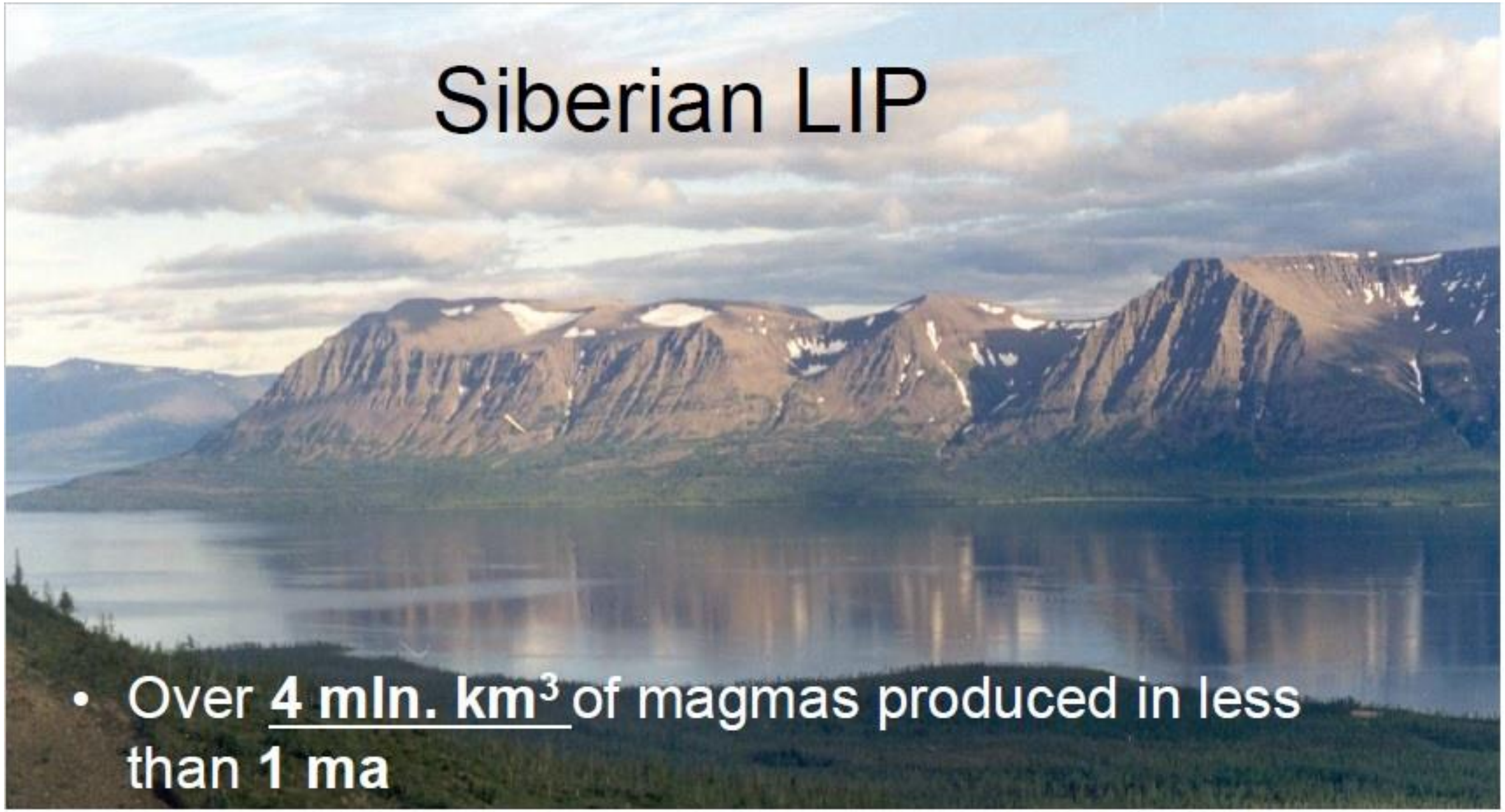


Herzberg and Gazel, 2009

LIPs correlate with mass extinction events



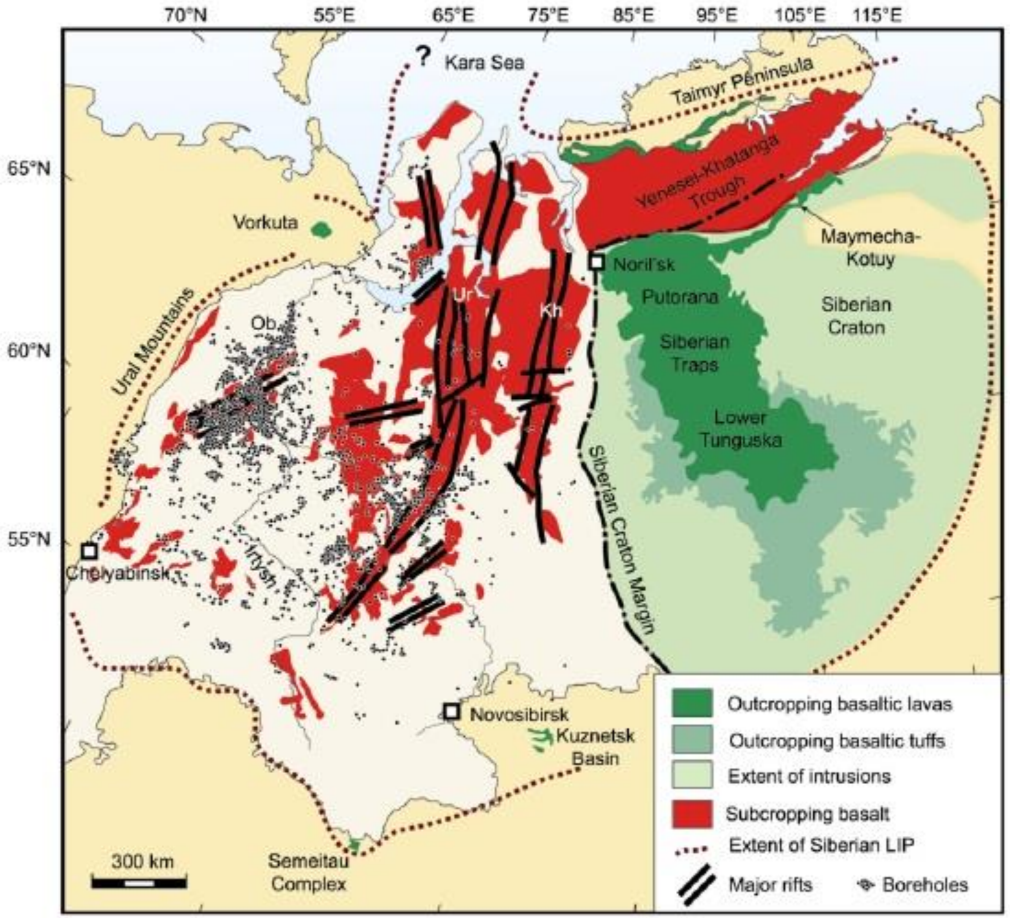
No Correlation with LIPs volume



Siberian LIP

- Over **4 mln. km^3** of magmas produced in less than **1 ma**

Siberian Traps



White and Saunders (2005)

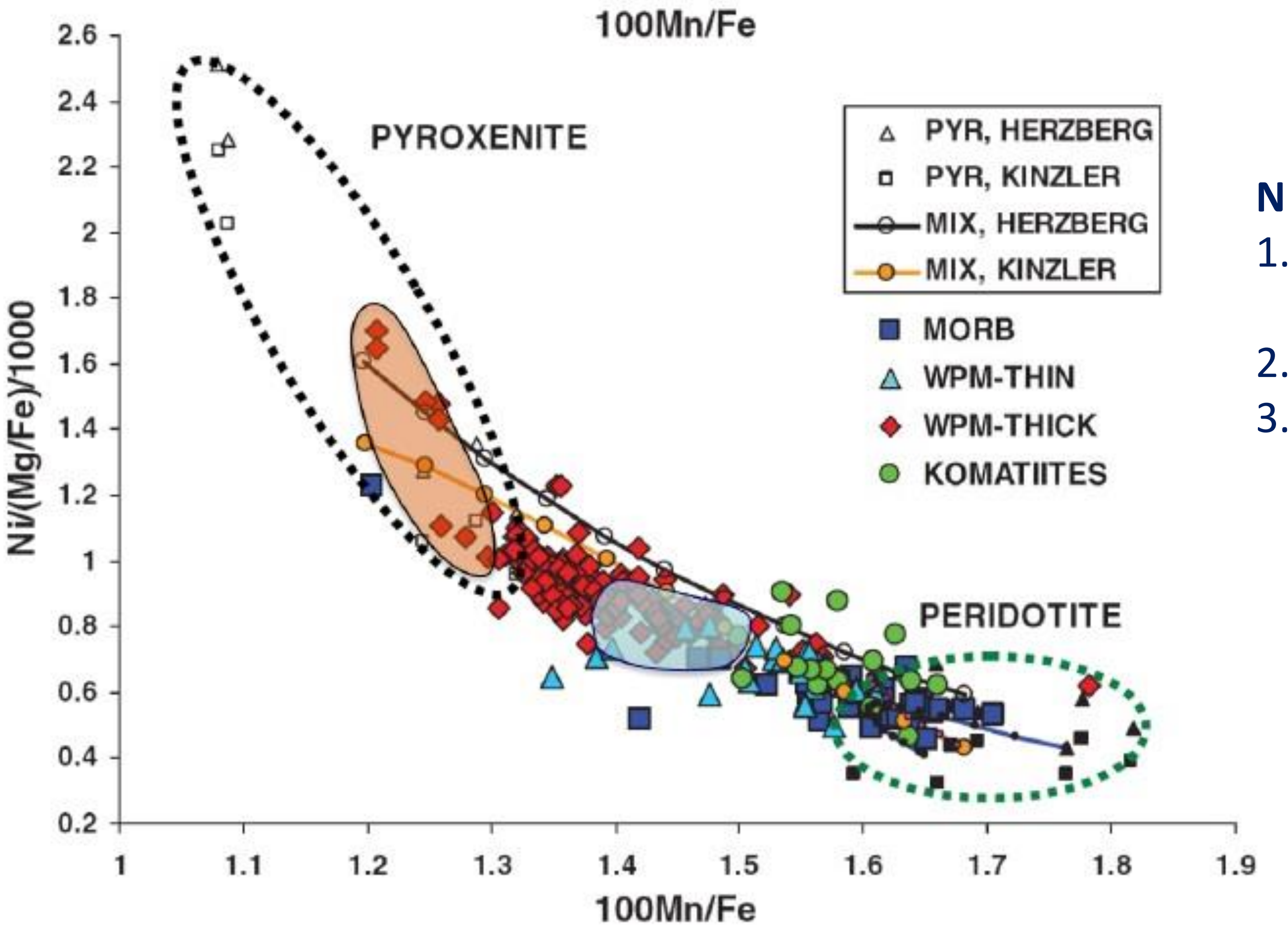
- Why no pre-magmatic uplift?

- Why large volume of magmas erupted at thick cratonic lithosphere without extreme extension?

- How lithosphere was thinned by >50 km during only few 100 thousand years?

- What was the source of large volumes of CO_2 and other gases that triggered P-T mass extinction?

Piroxenite formation



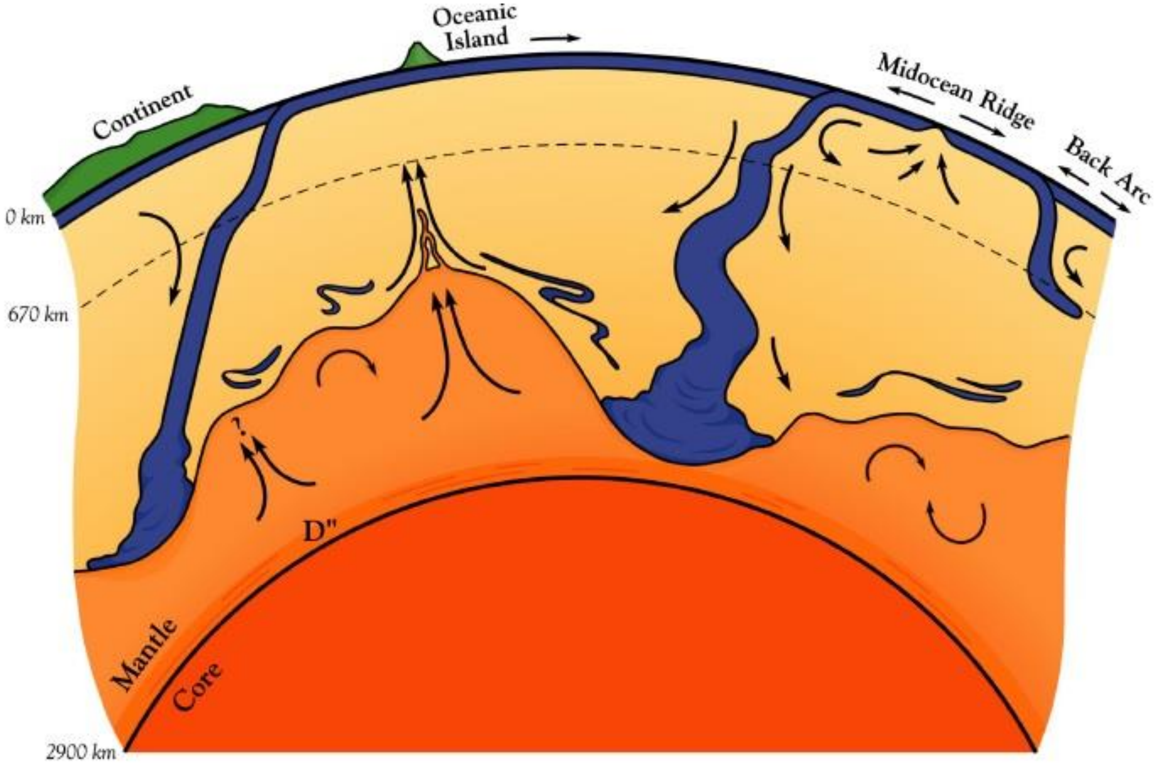
Ni excess and Mn depletion is interpreted as:

1. the result of the contribution of olivine free pyroxenite lithology in their source.
2. Effect of clinopyroxene crystallization.
3. Contribution of core material to the mantle source.

Sobolev et al, *Science*, 2007

Piroxenite formation

Crustal recycling

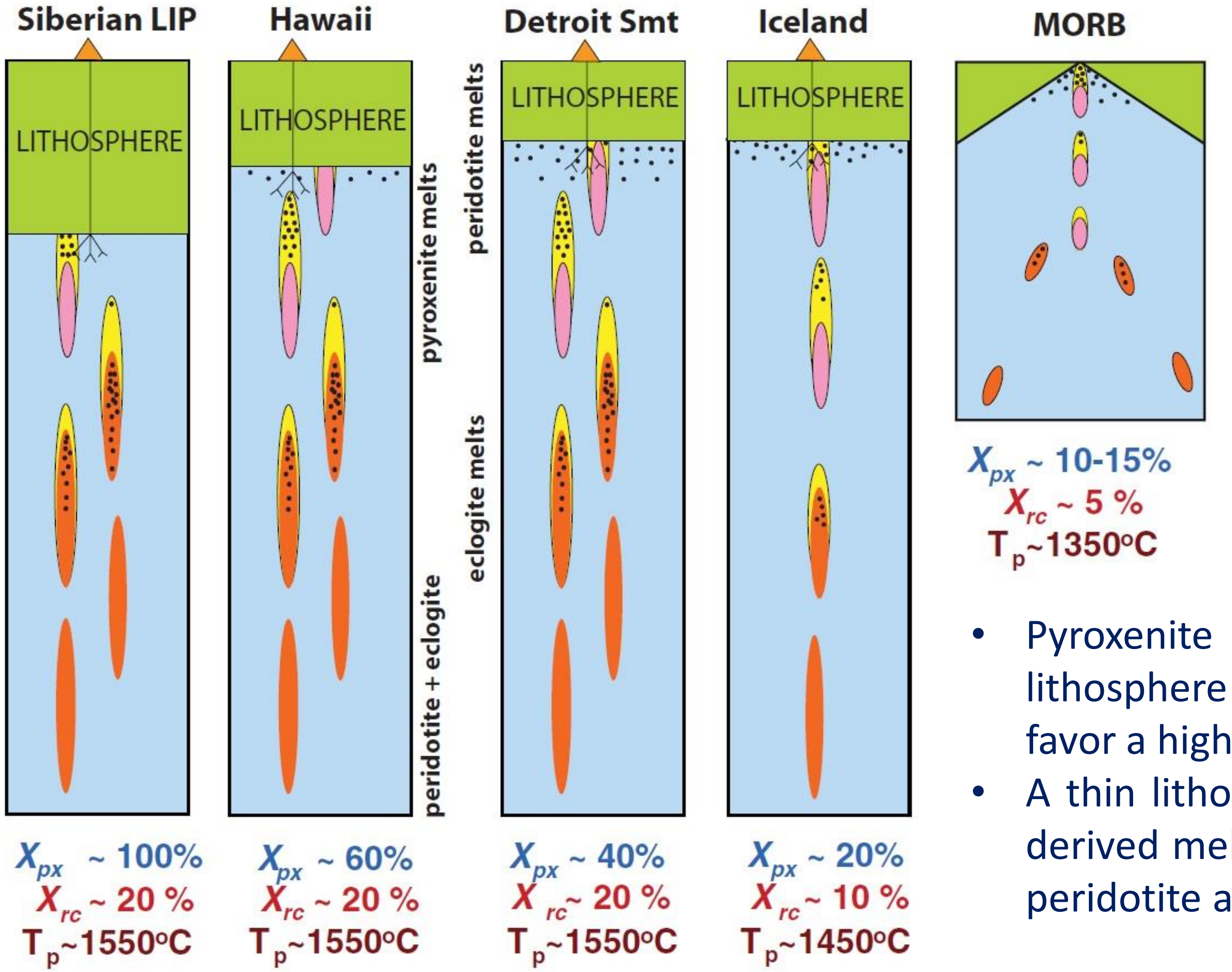


Kellogg et al., 1999

- In subduction at P > 2.5 GPa, the basaltic and gabbroic portions of the oceanic crust are transformed completely to eclogite (clinopyroxene and garnet) with a free SiO_2 phase.
- In the ascending, the silica-oversaturated eclogite starts melting at higher pressures than the peridotite and produces high silica melt, which reacts with olivine from peridotite, producing pyroxenes and garnet.

Eclogite

Piroxenite formation



Blue = upwelling peridotitic mantle
Red=recycled oceanic crust (eclogite with free SiO_2 phase)
Black dots = melting
Yellow= reaction zones forming hybrid pyroxenite
Pink= refractory restite after eclogite melting

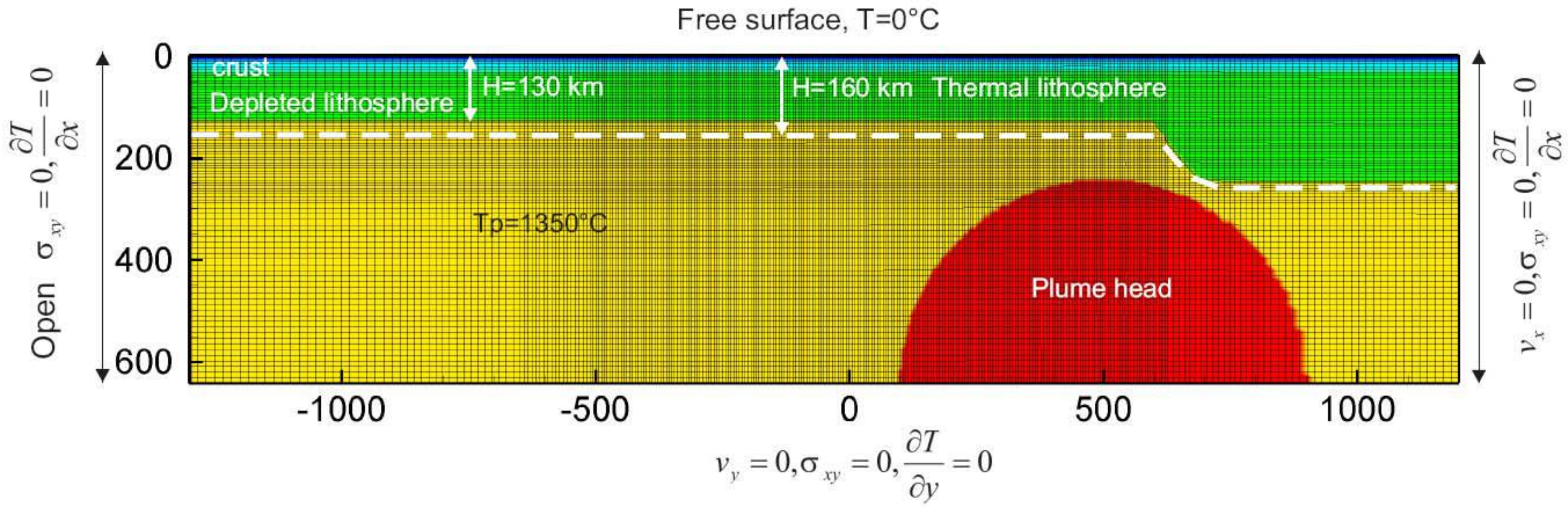
- Pyroxenite melts at higher *P* than peridotite, a thick lithosphere will suppress low-depth peridotite melting and favor a high proportion of pyroxenite derived melts.
- A thin lithosphere favors a higher proportion of peridotite-derived melt because of the increasing degree of melting of peridotite at shallower depths.

Sobolev et al, *Science*, 2007

Plume numerical models

Model Setup

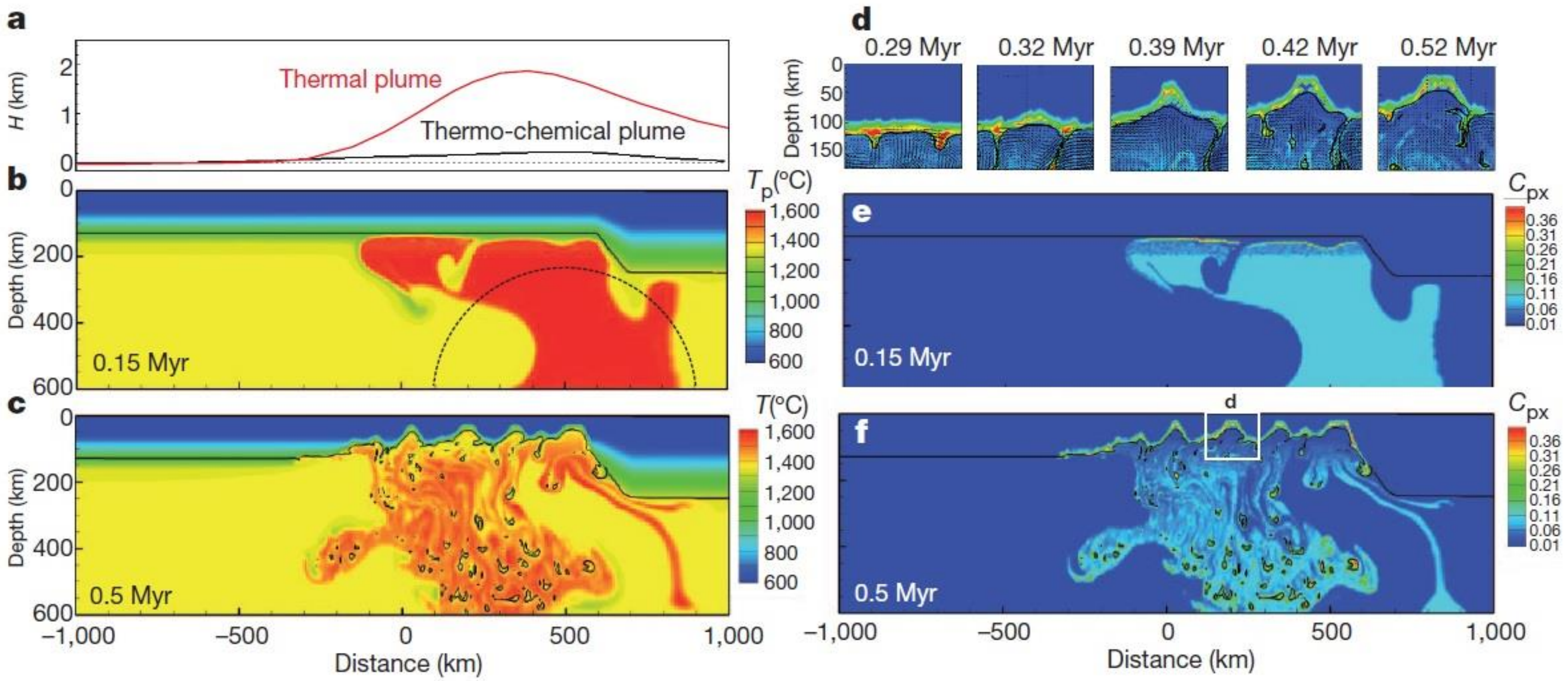
- Plume potential temperature Tp=1600°C
- Eclogite content in plume 10-20wt% (15wt%)
- Initial lithospheric thickness = 130 km



Finite element size is 5 X 5 km in the best resolved part of the model

Sobolev et al., 2011, Nature

Plume numerical models



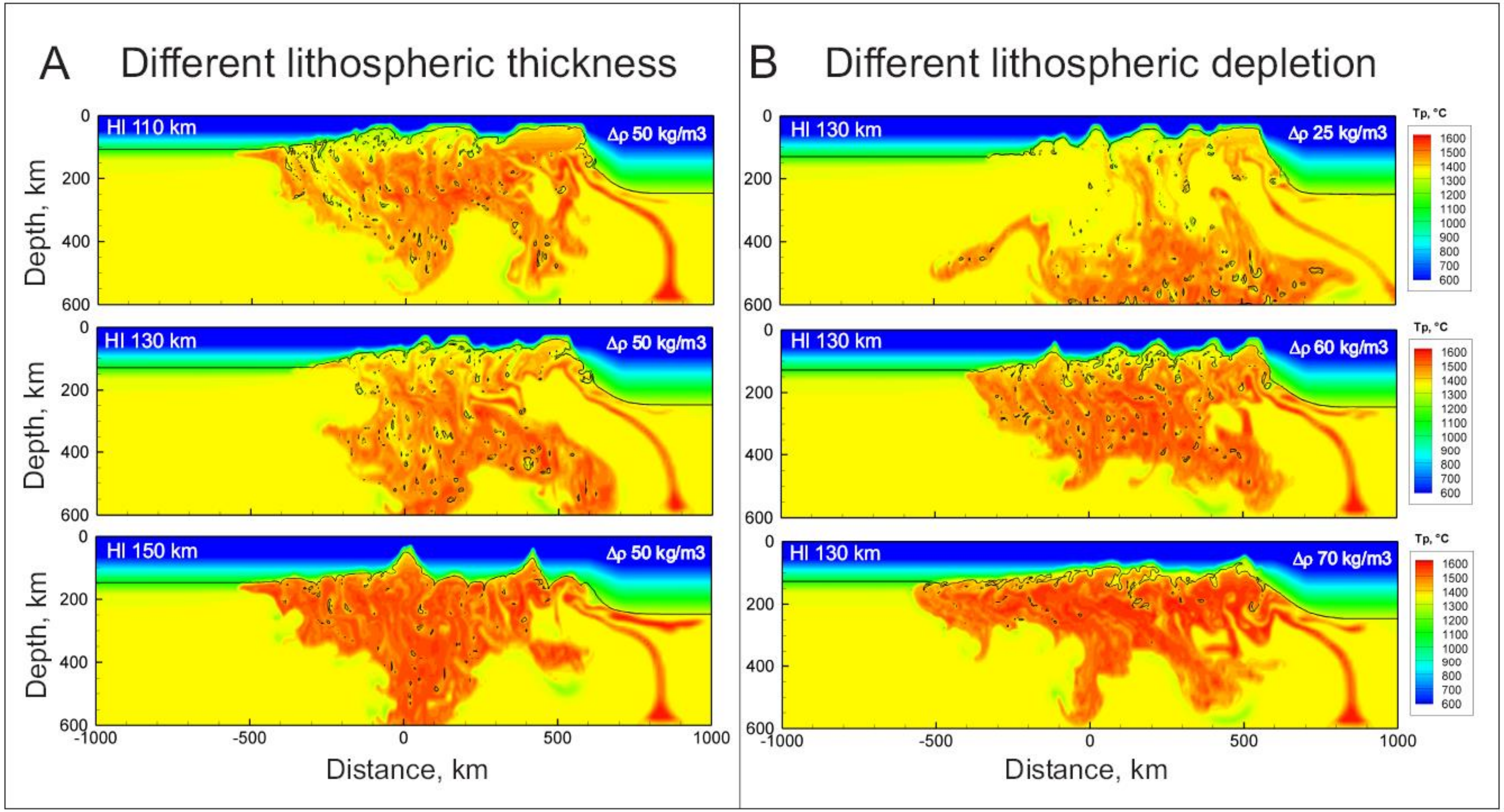
Sobolev et al., 2011, Nature

Magma intruded into the crust = 6–8 x10^6km^3

Large fraction (15 wt%) of dense recycled material is present within the plume, its buoyancy is strongly decreased, resulting in little regional uplift (250 m).

1. The plume head erodes the lowest part of the thermal lithosphere and rapidly spreads below the more refractory depleted lithosphere.
2. Plume ascent leads to progressive melting of recycled eclogitic material in the plume and to the formation of reaction pyroxenite, which melts at depths of 130–180 km (well before the peridotite).
3. The melt intrude into the lower lithosphere, cools and crystallizes to dense eclogite. It also strongly heats, weakens and mechanically erodes the lithosphere, promoting Raleigh–Taylor instabilities.
4. Enriched in eclogite, the lithospheric material in the boundary layer above the plume escapes to the sides of the plume and then downwards, allowing the plume to ascend.
5. The plume reaches its minimum depth of about 50 km crystallizing to a garnet-free assemblage, having a density lower than that of the ambient mantle (no formation of Raleigh–Taylor instabilities).

Effect of plume on the intensity of the lithospheric destruction

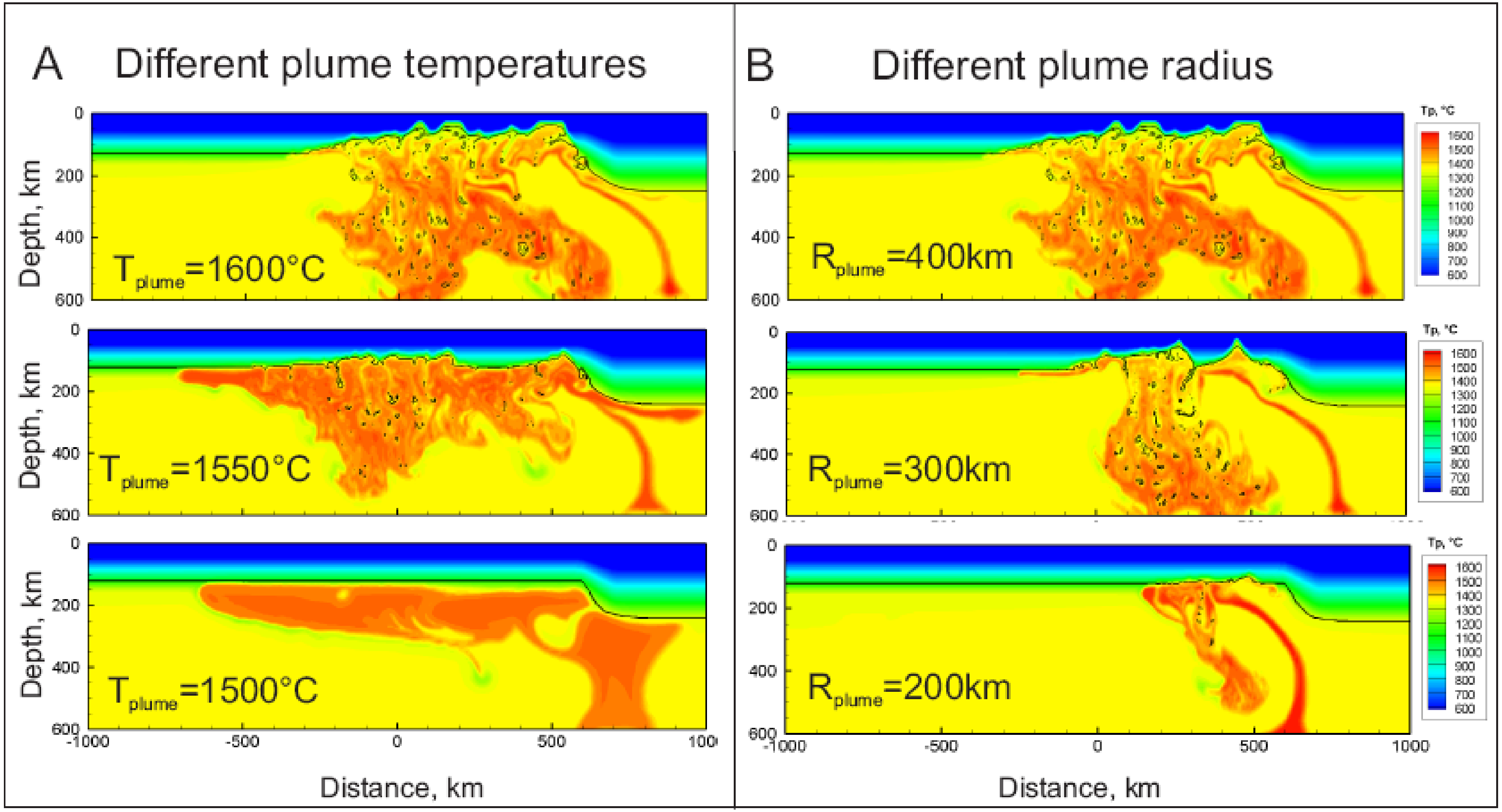


Hl= thickness of the depleted lithosphere

model time 1.0 Myr

Sobolev et al., 2011, Nature

Effect of plume on the intensity of the lithospheric destruction

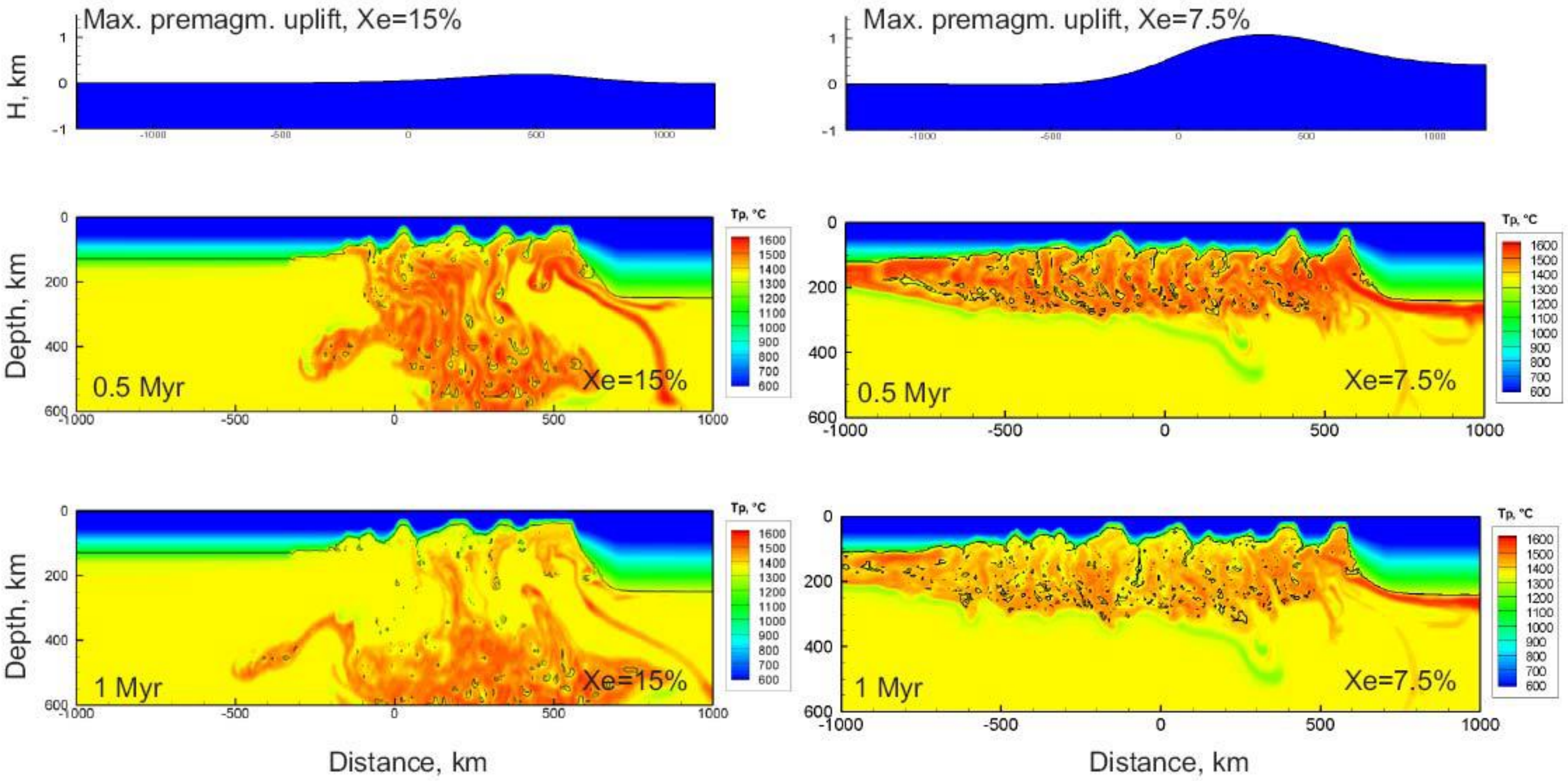


model time 1.0 Myr

Sobolev et al., 2011, Nature

Effect of plume on the surface topography and intensity of the lithospheric destruction

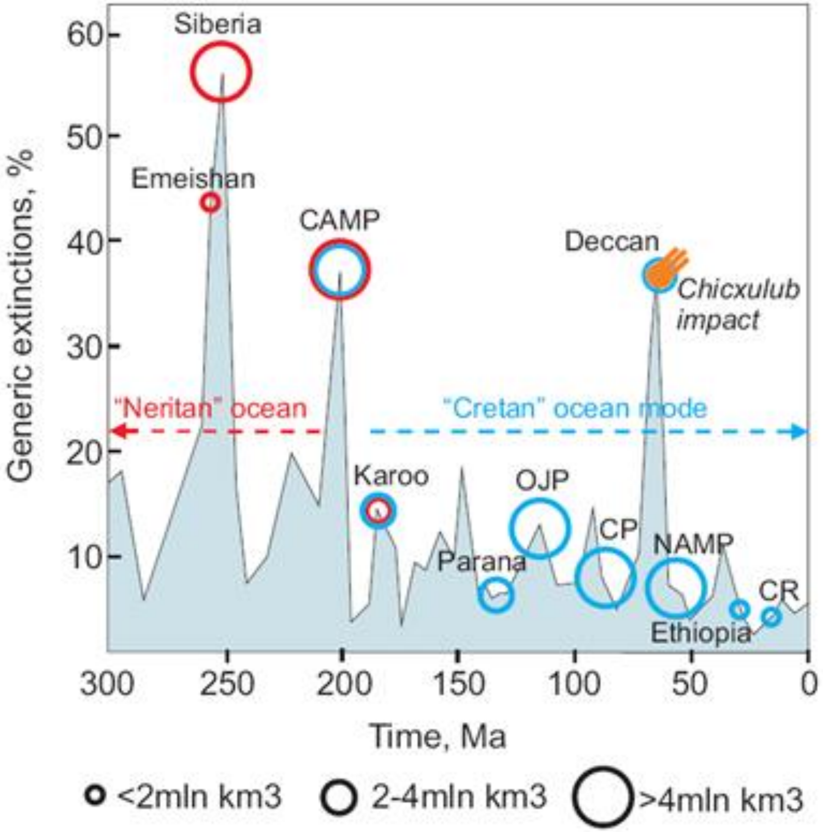
Different plume composition



Sobolev et al., 2011, Nature

Results of plumes numerical models

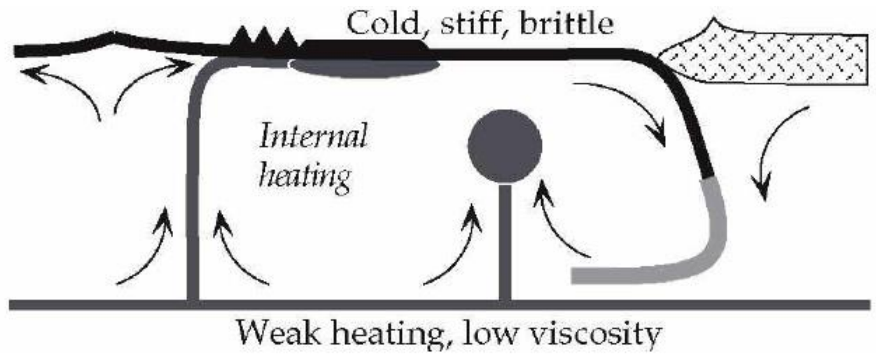
- Thermochemical plume rich in recycled crust does not generate significant pre-magmatic uplift of the lithosphere.
- Such a plume is able to thin dramatically cratonic lithosphere without extension and to generate several mln km^3 of melt in few 100 thousand years.
- Massive CO_2 and HCl degassing from the plume could alone trigger the Permian-Triassic mass extinction and before the main volcanic phase.



CAMP, Central Atlantic Magmatic Province; NAMP, Northern AtlanticMagmatic Provinces, OJP, Ontong Java; CP, Caribbean Plateaux; CR, Columbian River basalts.

Conclusions

Convection mode depends mostly on:



- The presence of cold high viscous lithospheric plates (plate-scale flow)
- The viscosity variations in the mantle
- The presence of transformation phases in the mantle

  - The plates are an integral part of a convection system: stresses are transmitted through the viscous mantle as well as through the elastic plates.
  - Evidence from seismic tomography and the gravity field supports the possibility that there is a large mass flow through the mantle transition zone, and that mantle convection occurs as a single layer rather than two.

- ☐ The hotspot swells constrain the buoyancy flux of plumes and indicate that plumes transport less than about 10% of the mantle heat budget (secondary mode of mantle convection).
- ☐ Plumes are produced in the lowermost mantle adjacent to the core within the so-called D'' layer, due to a difference in T (active upwelling), while mantle rising under midocean occurs without T difference (passive upwelling).
- ☐ Plumes probably do not influence plate motions very much, though they may trigger some changes in favourable circumstances, such as major rifting.