Course of Geodynamics

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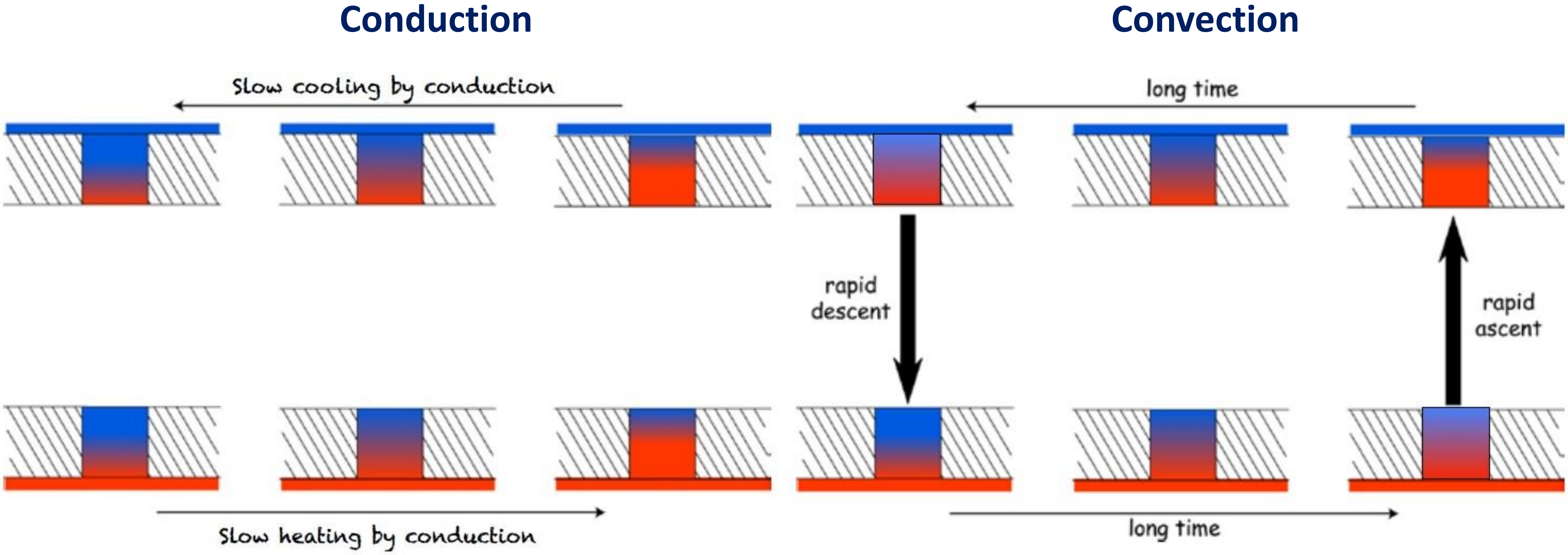
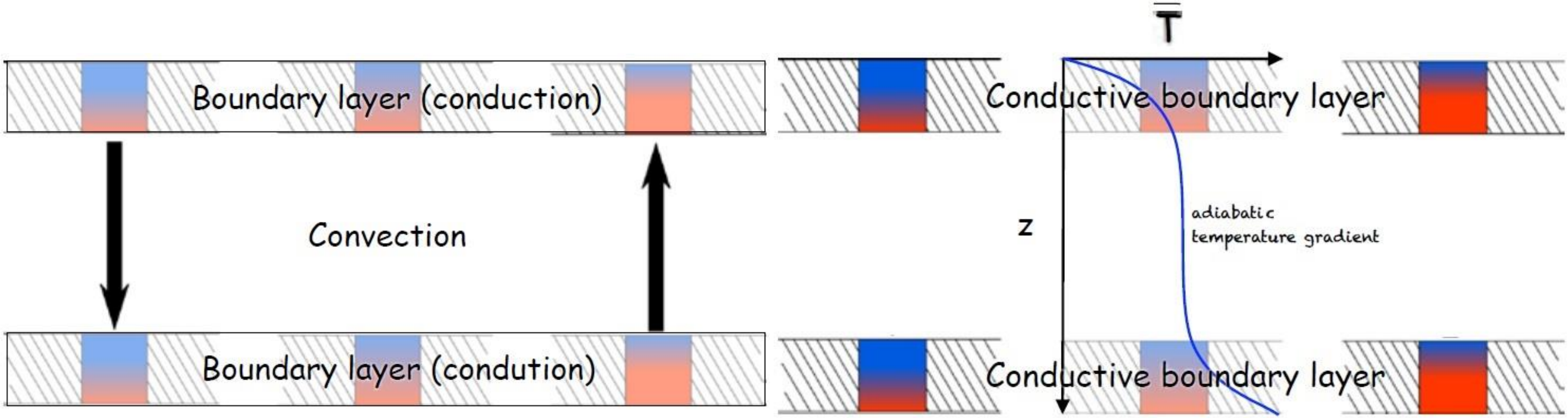
Course Outline:

1. Thermo-physical structure of the continental and oceanic crust
2. Thermo-physical structure of the continental lithosphere
3. Thermo-physical structure of the oceanic lithosphere and oceanic ridges
4. Rheology and mechanics of the lithosphere
5. Plate tectonics and boundary forces
6. **Hot spots, plumes, and convection**
7. Subduction zones systems
8. Orogens formation and evolution
9. Sedimentary basins formation and evolution

Convection and Conduction



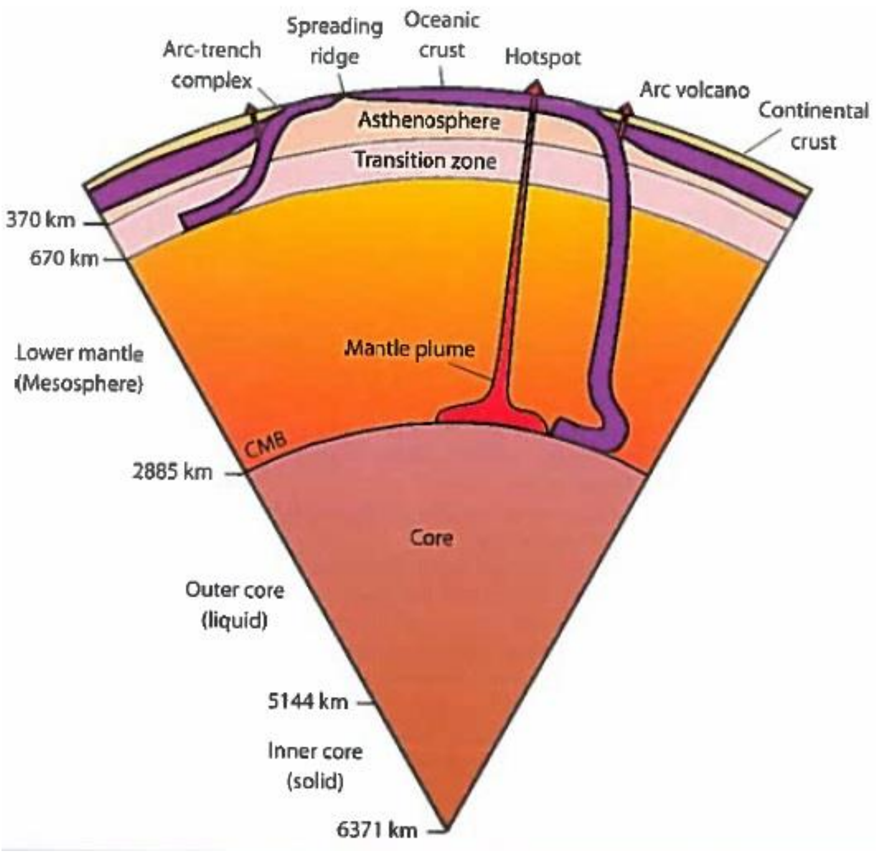
Convection and Conduction



Speed of plate tectonics (~10 cm/j) is determined by the time it takes to cool and heat the boundary layers

Convection

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



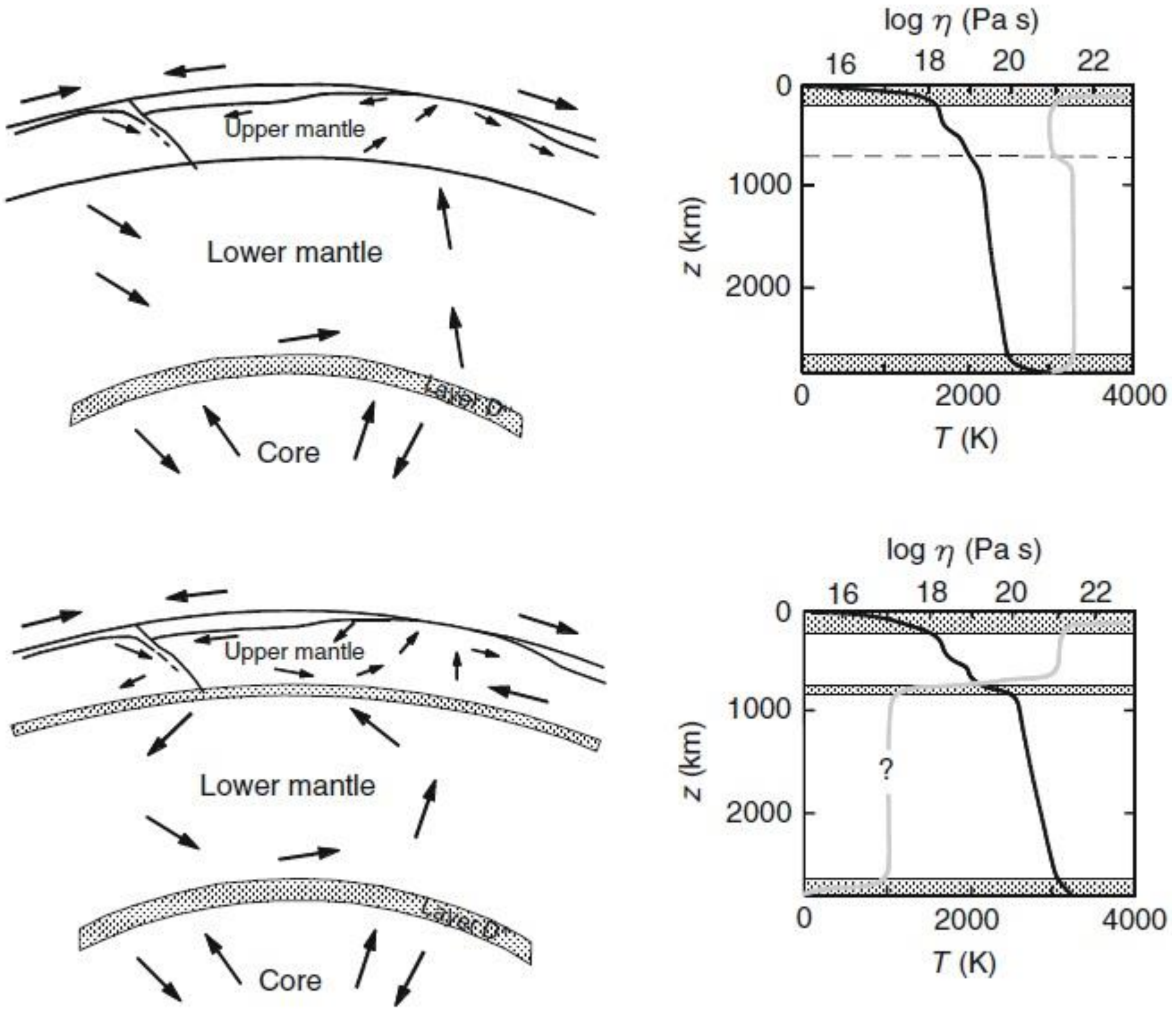
$B = -gV\rho = -g\Delta m$

Δm = mass anomaly due to a volume V with a density difference induced by temperature, $\rho = \rho_0[1-\alpha(T-T_0)]$, and/or composition

- 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.

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

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= 3MPaK^{-1}) and slightly negative for the transition spinel-perovskite (dP/dT= -2MPaK^{-1}) .

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) ~$T/2$ ($B = -gDd\rho\alpha T/2$)
and d, thickness of the subducting lithosphere (of the layer that diffused), depending on time (t) spent at the surface $t = D/v$:
$d = (\kappa t)^{1/2} = (\kappa D/v)^{1/2}$, with κ thermal diffusivity.

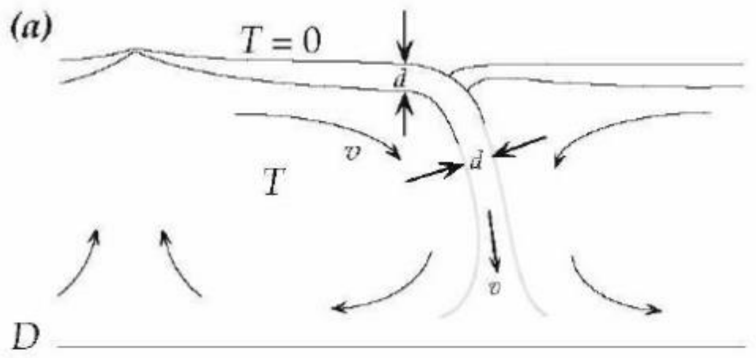
The resisting viscous stress σ acting on the side of the descending slab is estimated from a characteristic velocity gradient ($2v/D$): $\sigma=\mu 2v/D$

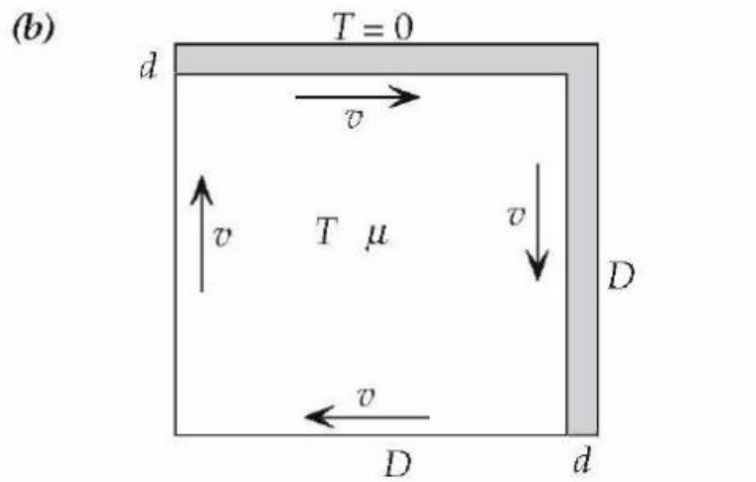
Viscous resistance R (per unit length) $= D2\mu v/D = 2\mu v$

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

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

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)

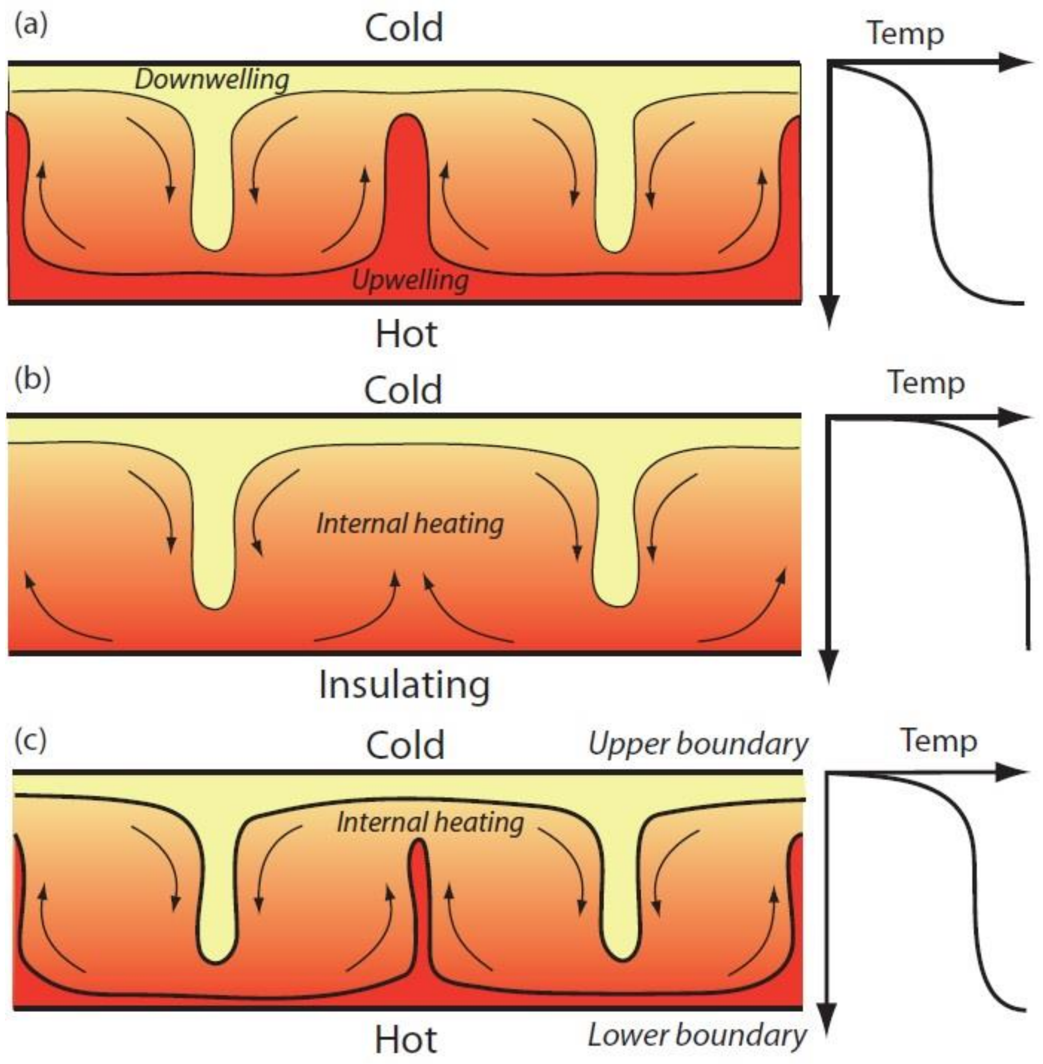




The onset of convection is given by the Rayleigh Number:

Rayleigh Number: $R_a = g\rho\alpha TD^3/\kappa\mu$ For the mantle $R_a = 3 \times 10^6$

Heating modes



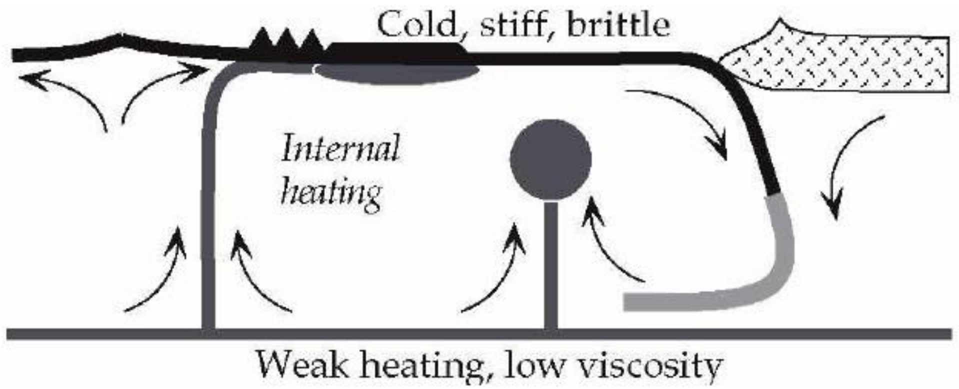
Active upwelling: heat enters from below and there is no heat generated within.

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 T will increase, the thermal boundary layers will be thinner and the temperature gradients through them will be higher, driving larger heat fluxes.

Convection and plate tectonics



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

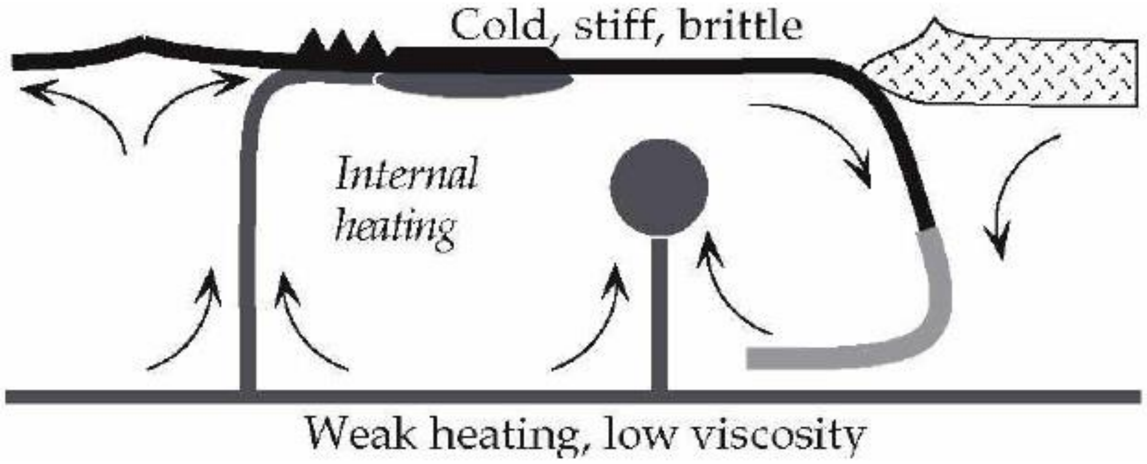
Tensile stresses in the interior of supercontinents depend on the size of the plate, the Rayleigh number of mantle convection, the viscosity profile of the mantle, and the amount of radioactive heat present.

- 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.

Several mechanisms produce convection patterns that promote the growth and dispersal of supercontinent:

- The insulating properties of large masses of continental lithosphere create mantle upwelling beneath their interiors.
- Large plates also prevent the mantle beneath them from being cooled by subduction, which further promotes upwelling.
- Sites of downwelling may be controlled by the intrinsic buoyancy of continental lithosphere, which tends to concentrate subduction zones along continental margins.

Convection and plate tectonics

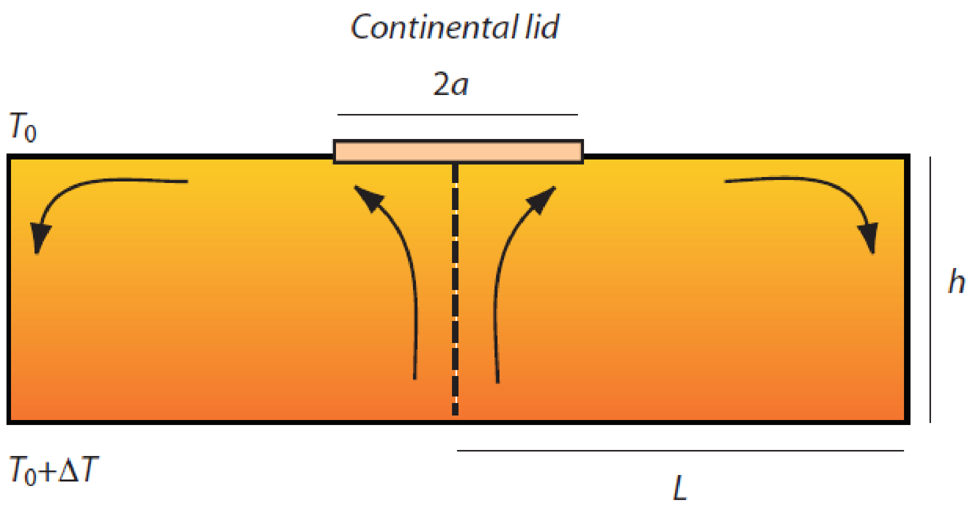


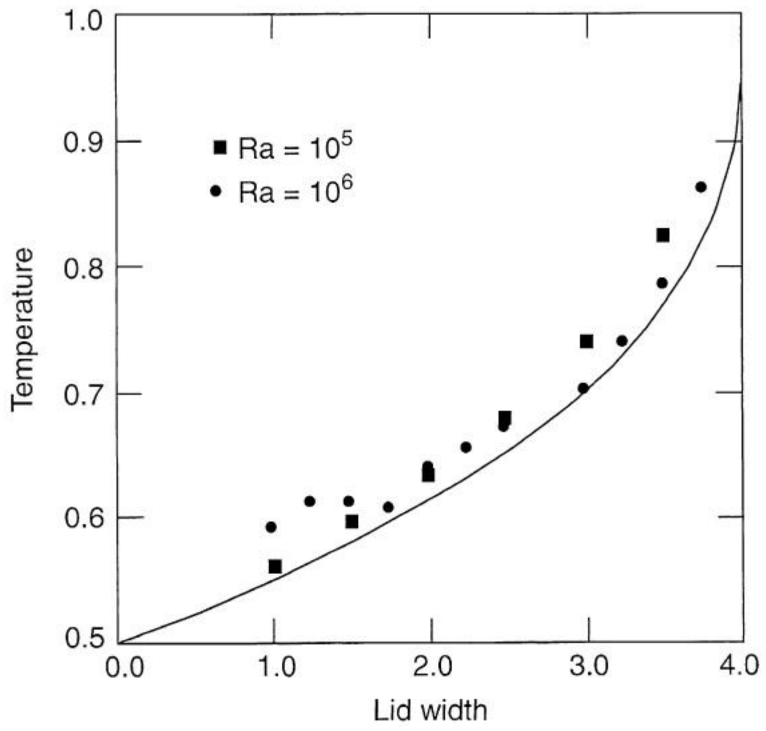
Role of internal heating:

Due to internal heating, plate motion is characterized by episodic reversals in direction as mantle circulation patterns change from clockwise to counterclockwise and vice versa (plates suddenly change direction on timescales of some 300 Myr):

- The downwelling of cold material at one edge of a plate can entrain hot material that is trapped below the plate and drag it into the lower mantle. The hot, buoyant material then begins to ascend as the drag of the cold downwelling wanes.
- The ascent of hot material pushes the plate laterally and induces new cold downwelling on the other side of the plate, beginning a new cycle of upwelling and plate motion in the opposite direction.

Convection and plate tectonics



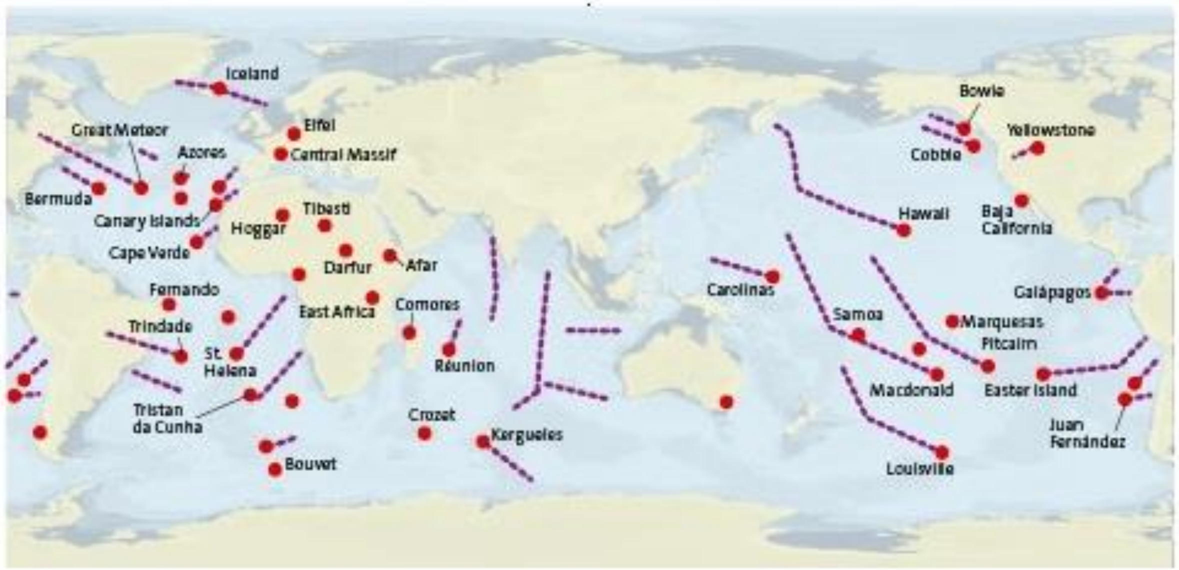


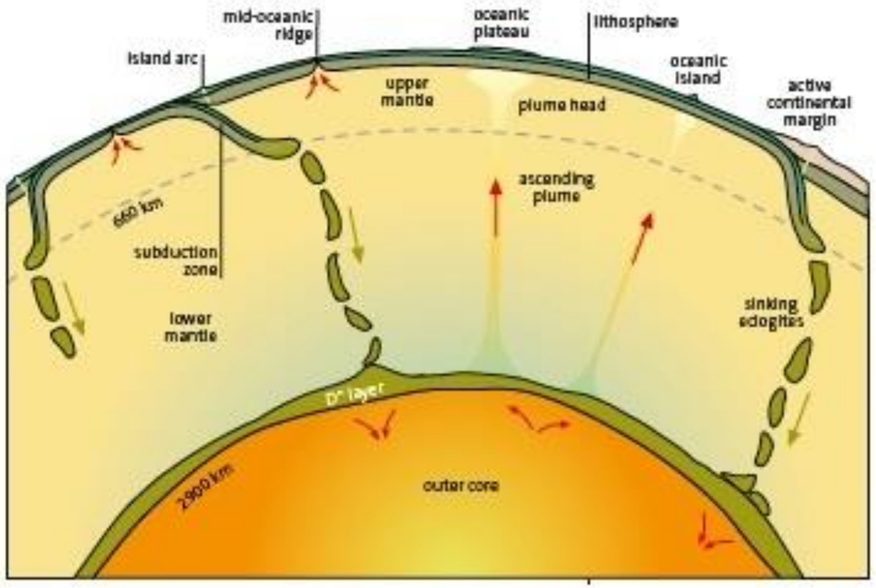
A wide insulating lid (during supercontinent assembly) warms the mantle beneath it and the entire interior of the Earth:

- It may cause plumes localization and would explain the flood basalts with supercontinental break-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)



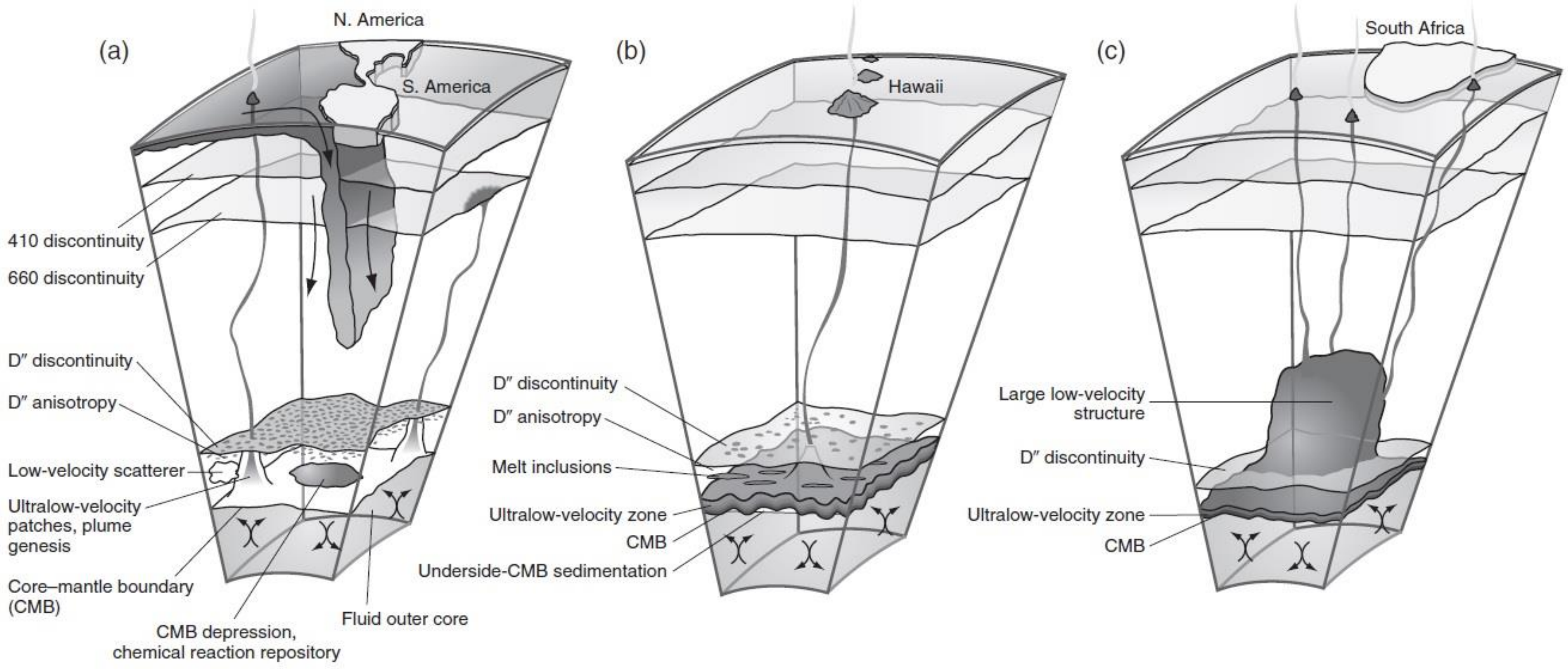


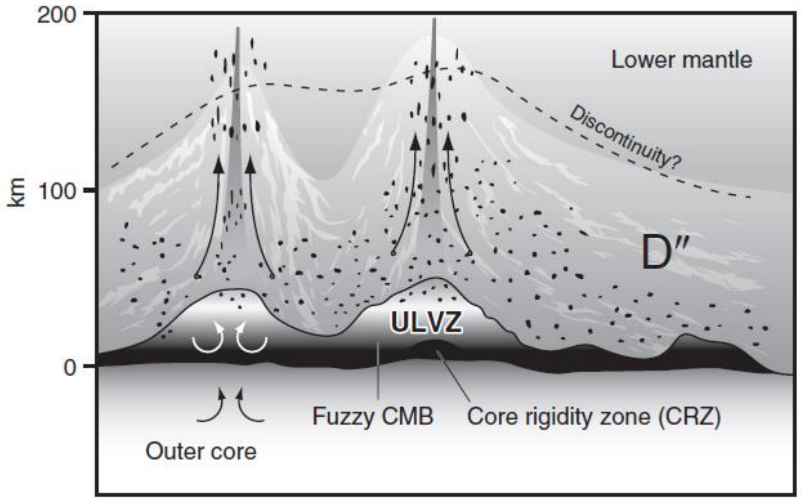
Plumes

- Hot mantle rocks (200-300°C higher than surrounding mantle), not molten, because of the high pressure, arising from the base of the lower mantle (D’’ Layer) 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.

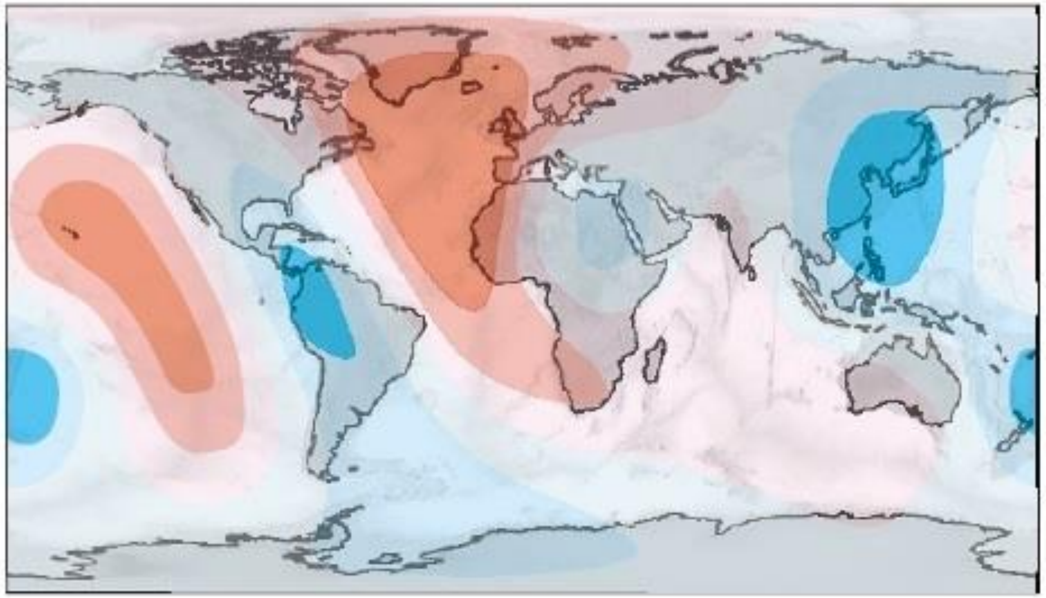
Nature of layer D’’

- The upper boundary of the layer D’’ is characterized by a velocity discontinuity, below which there may be an increase (e.g., beneath regions where there are subducting slabs) or decrease in the seismic velocities.
- Mineralogical explanations of the D” discontinuity focuses on the transformation from perowskite to the postperowskite phase.
- In a 5 to 50-km-thick layer immediately above the core–mantle boundary there is often a zone of ultra-low seismic velocities, with decreases in the shear wave velocity of 10–50% (most extensively developed beneath major hotspots). This implies partial melting with more than 15% melt.
- Lateral and vertical variations within layer D” may be caused by variations in chemical composition, mineralogic phase changes (due to the mixing of molten iron from the core with the perovskite of the mantle to form new high-pressure minerals) and/or varying degrees of partial melting and *T* differences. The result would be a chemically distinct, high-density layer but with a low viscosity.

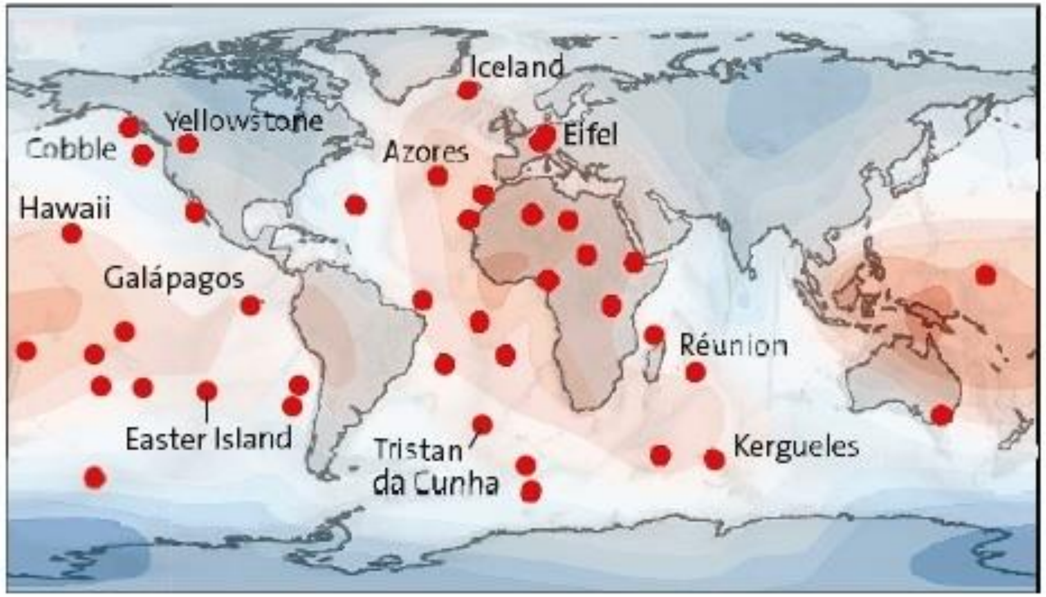




Hot Spot and geoid anomalies



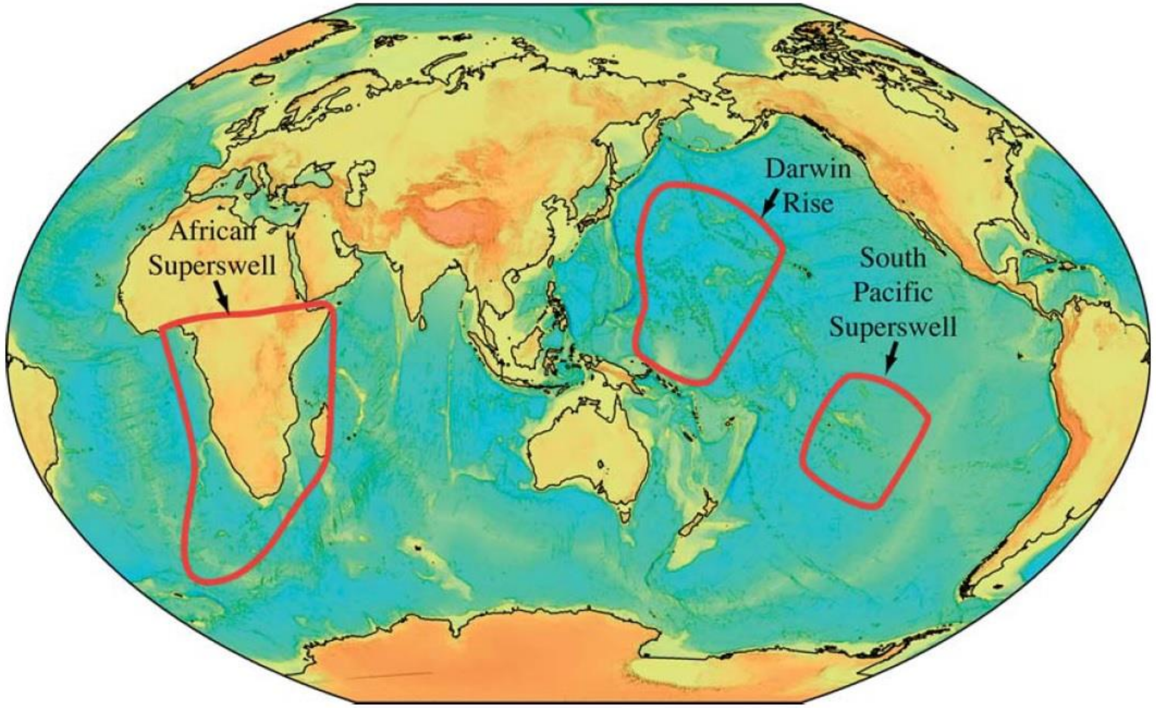
- Hot spot are concentrated in correspondence of geoid bulges



- 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.

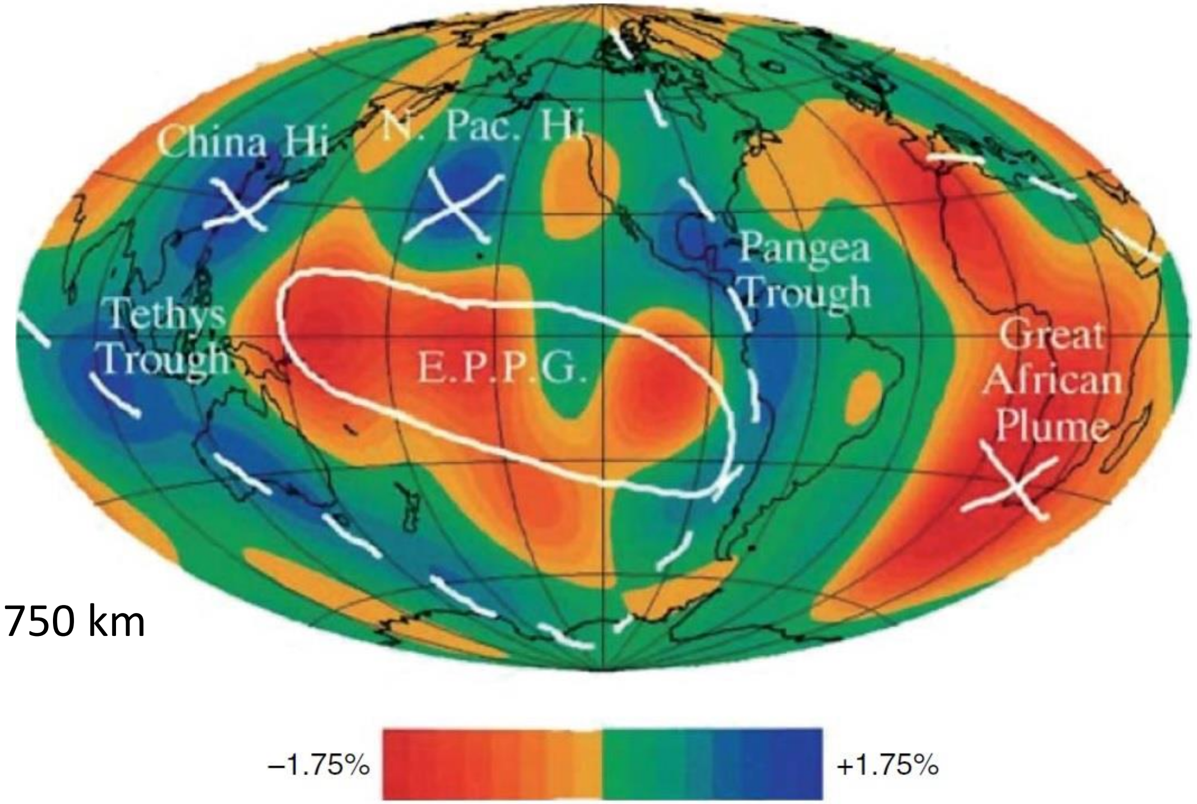
Superswells

- Superswells represent clusters of hot spots in a restricted geographical region of anomalously high elevation, extending several hundred km beyond the area of excess volcanism, which cannot be explained by crustal thickening, examples are the Africa and the South Pacific, characterized by slow seismic velocities both in the upper and lower mantle, related to the anmalous high temperatures.
- The hot spot in the south Pacific have anomalously low seismic wave speeds in the mantle, suggesting an origin involving anomalously low densities but without substantially thinned lithosphere. The African Superswell, on the other hand, does not show a seismic anomaly in the upper mantle but rather a broad columnar zone of slow velocities in the lower mantle, both of compositional and thermal origin.
- In general, if a swell forms at a hot spot, it decreases in height with time and can no longer be detected when reaches an age of 80 Myr, (or even <50 Myr). On the other hand, the young Madeira and Canary hot spots do not show any swell.



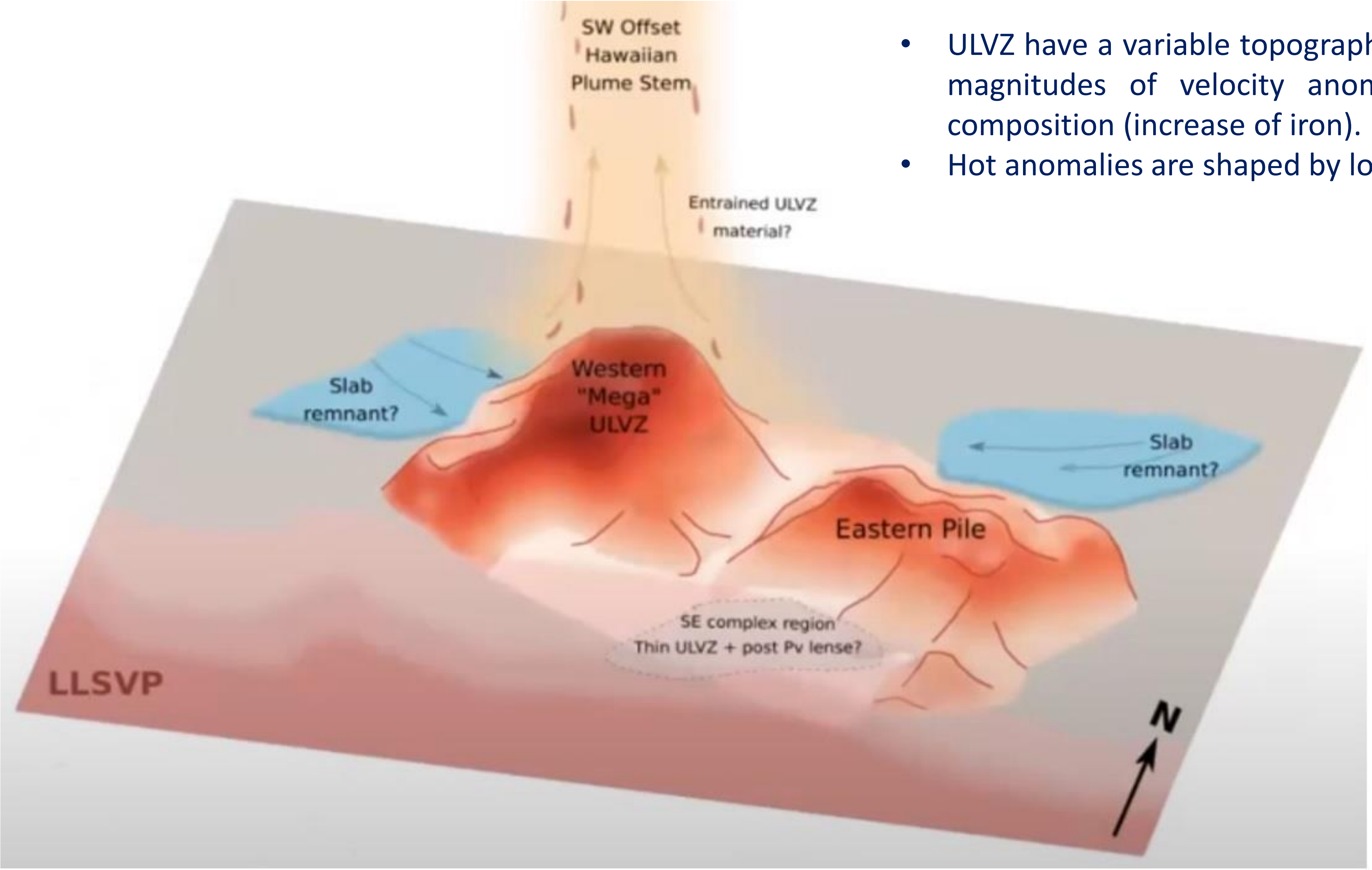
McNutt and Caress, 2007, Treatise ofGeophysics, vol. 1

Global seismic tomographic image of shear-wave velocities



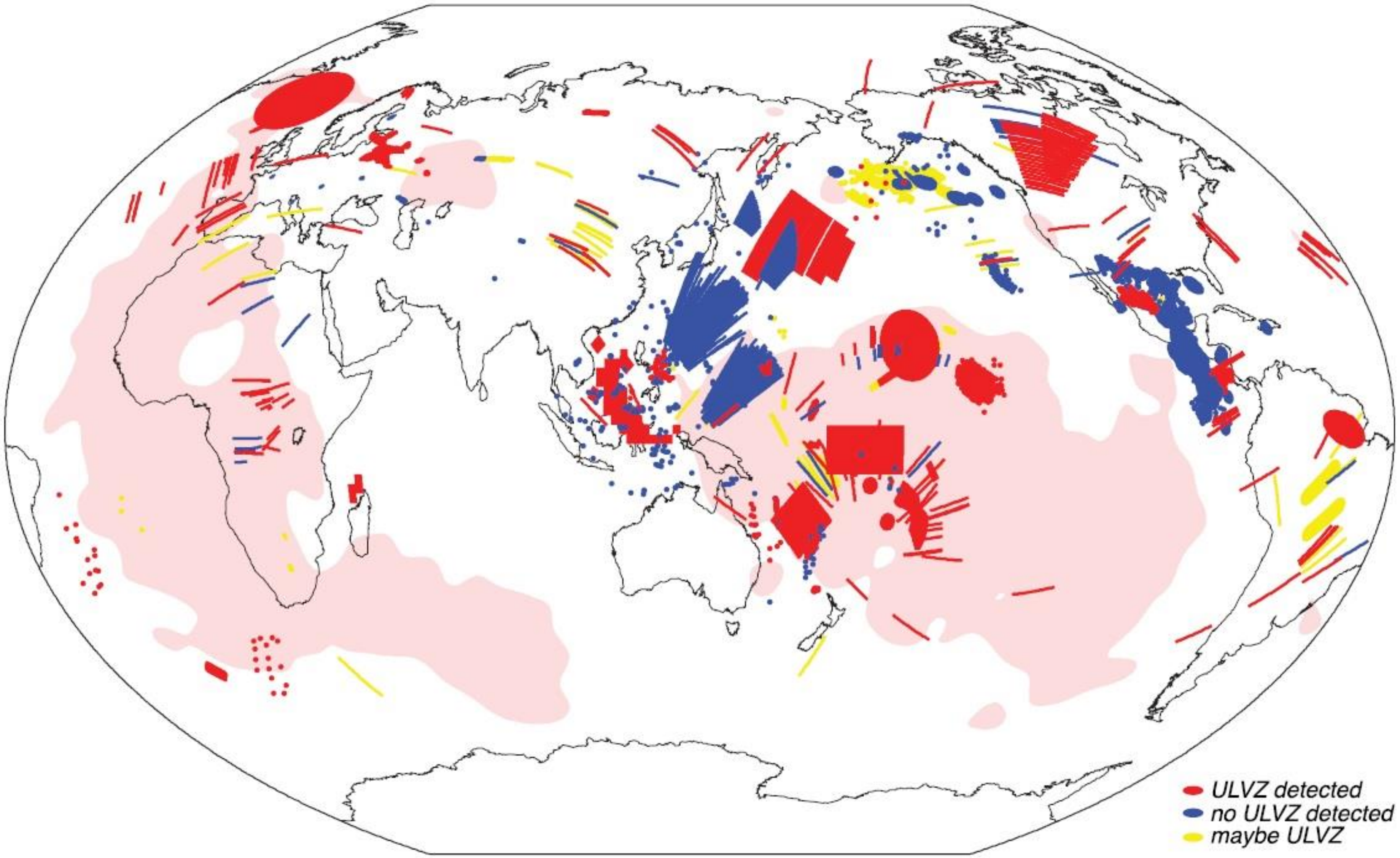
2750 km

Ultra Low Velocity Zones (ULVZ)



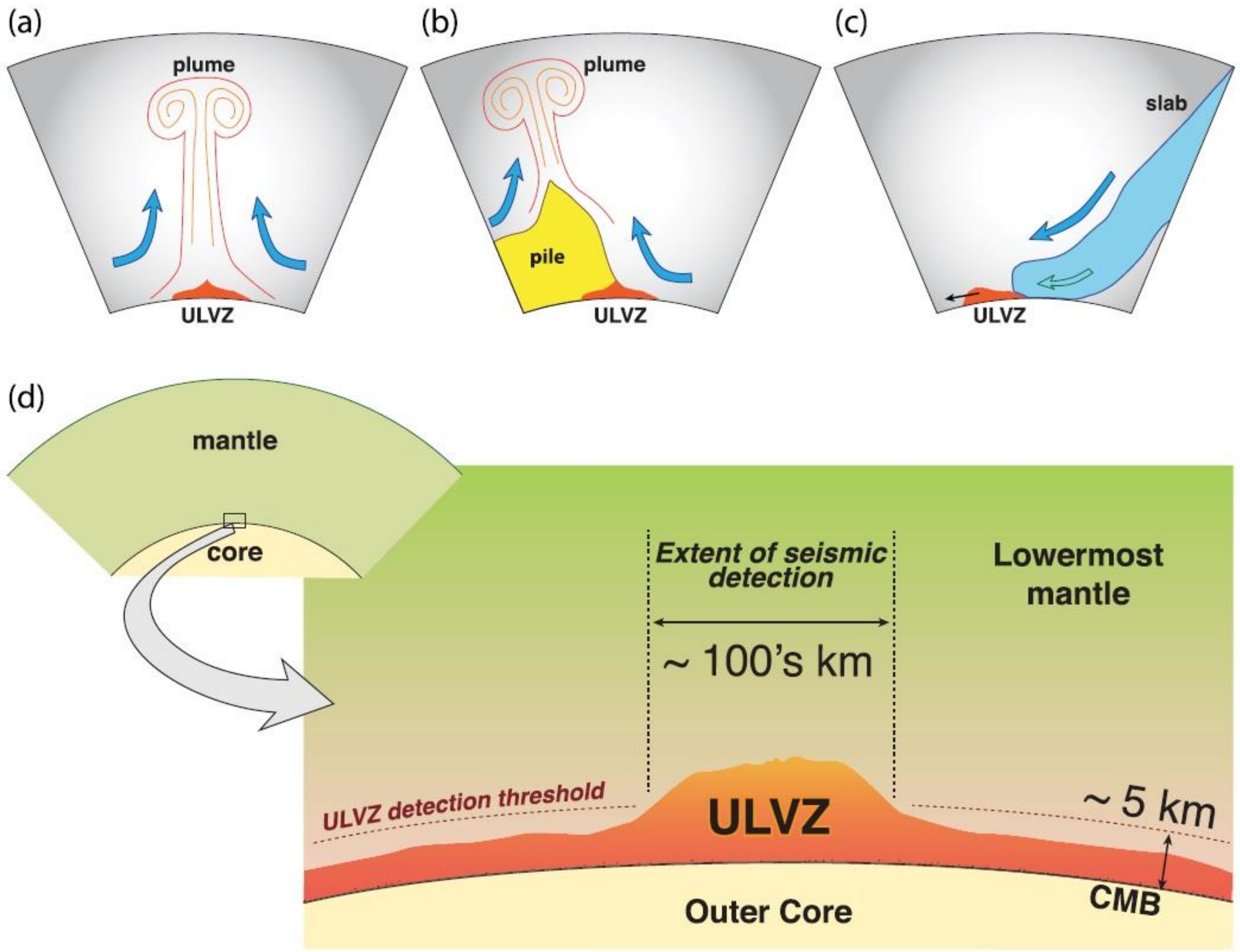
- ULVZ have a variable topography: 5-25 km, which result in different magnitudes of velocity anomalies, related to temperature or composition (increase of iron).
- Hot anomalies are shaped by lower mantle processes.

ULVZ Location



Yu and Garnero, 2018, G3, 19.

ULVZ Origins



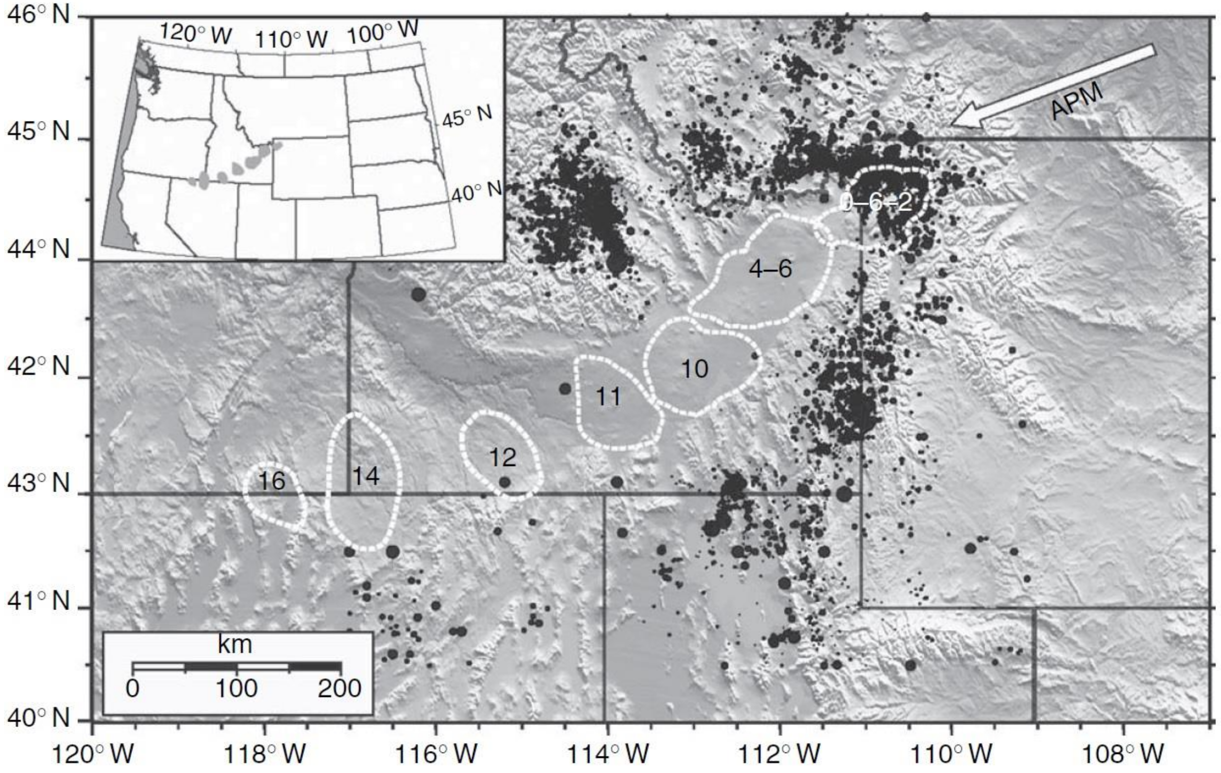
Yu and Garnero, 2018, G3, 19.

a) ULVZs have been reported to exist beneath surface hot spots, associated with mantle plumes. (b) Compositionally distinct ULVZs advect to the margins of thermochemical piles (c) ULVZs in relatively cold regions might be due to deeply subducted oceanic crust, or possible accumulated products of chemical reactions between the core and mantle.(d) The possibility of widespread thin ULVZs.

Geophysical characteristics of hot spots

Orientation and Age

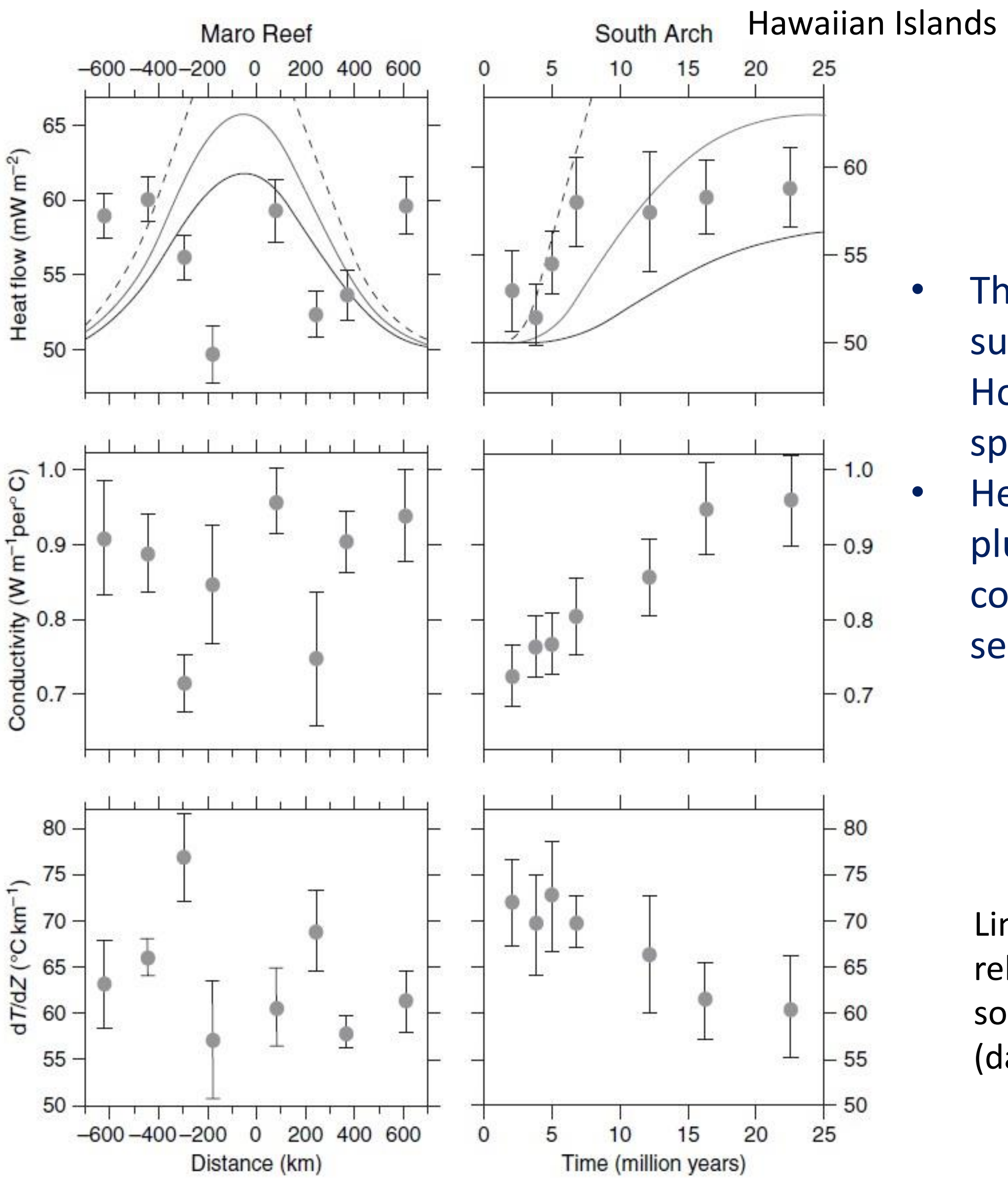
- Hot spots erupting on the Pacific Plate tend to form linear chains of volcanoes, often having a monotonic age progression. In some cases the most common age pattern is nearly synchronous volcanism at several volcanic centers along the chain, with short age progressions. However, the sequence of hot-spot eruptions is not completely random in space or time.
- Long-lived (>50My) age-progressive volcanism occurs in 13 hot spots, defining a kinematic reference frame that is deforming at rates lower than average plate velocities. Over geologic time there has been significant motion between the Indo-Atlantic hot spots, the Pacific hot spots, and Iceland. Short-lived (<22My) age progressions occur in at least 8 volcano chains.
- The Yellowstone hot spot is the only continental hot spot showing a clear age progression. The effective speed of the hot-spot track is 4.5 cm yr^{-1}, which is interpreted to include a component of the present-day plate motion (2.5 $cmyr^{-1}$) and a component caused by the Basin and Range extension.



Ito and van Keken, 2007, Treatise of Geophysics, vol. 7

Geophysical characteristics of hot spots

Heat Flow



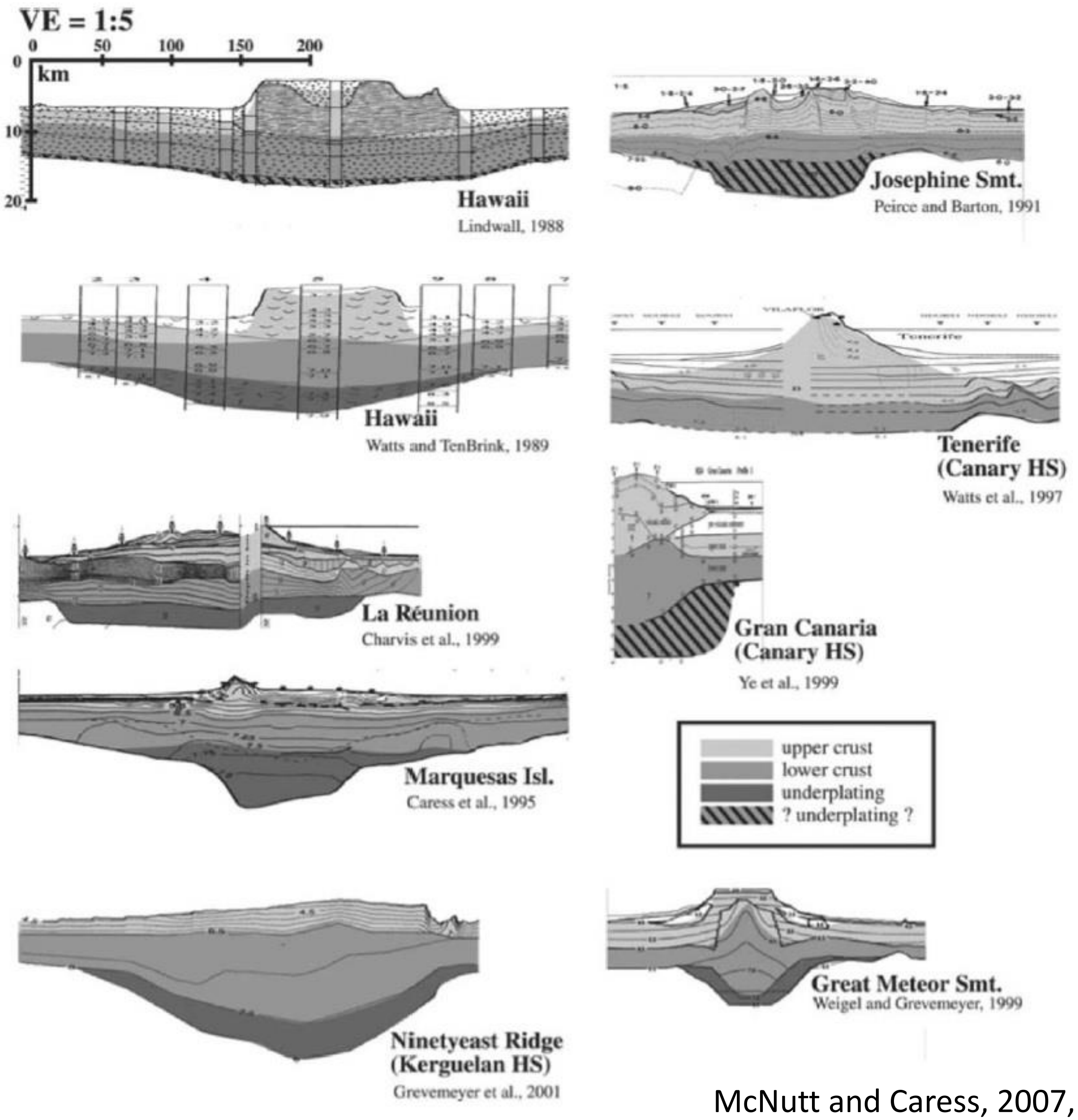
McNutt and Caress, 2007, Treatise ofGeophysics, vol. 1

- There are modest (10-25%) heat flow anomalies associated with hot spots, suggesting that the low-density material is at least partially thermal in origin. However, the effects of near-surface fluid circulation effectively mask much of the spatial and temporal pattern of any thermal disturbance.
- Heat flow variations over Hawaiian swell controlled by near-surface processes, not plume properties: areas of high conductivity act as radiators while areas of low conductivity act as insulators, producing nonuniform conduction of heat to the seafloor that is controlled by variations in thermal conductivity.

Lines on the upper panels show the theoretical variations in heat flow expected if the hot spot reheated the lower lithosphere to asthenospheric values (1350°C) as it passed over the thermal source. The amount of thinning assumed is 60km (solid line), 50km (dotted line), and 40km (dashed line).

Geophysical characteristics of hot spots

Magmatic Underplating



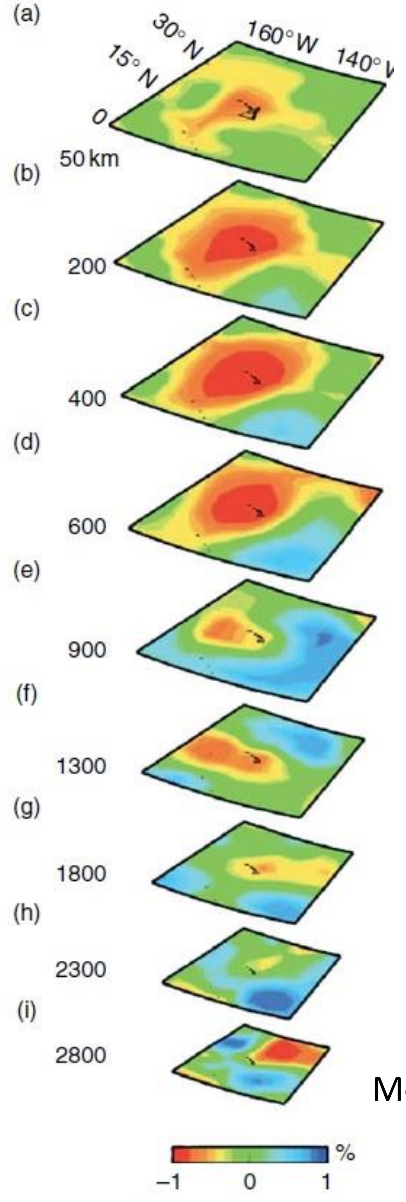
- Intrusive volcanism and underplating are characteristic for most of the volcanic provinces, which were located above a hot spot at the time of their formation.
- The seismic velocity of the underplated material suggests that the hot spot produced melt buoyant enough to rise through the upper mantle to the base of the crust, but not sufficiently buoyant to rise above the crust–mantle boundary.
- Underplating may be associated with later stages of hot-spot volcanism (e.g., Hawaai and Ninetyeast Ridge), but the existence of underplating beneath the currently volcanically active La Reunion demonstrates that subcrustal intrusions can occur during the primary edifice building stage of hot-spot volcanism.

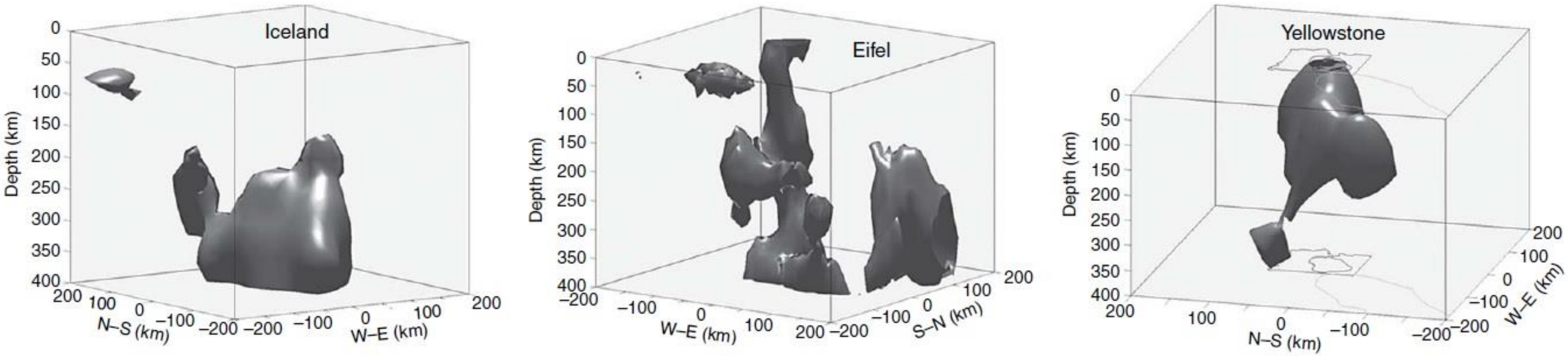
McNutt and Caress, 2007, Treatise ofGeophysics, vol. 1

Geophysical characteristics of hot spots

Seismic Velocities

- A hot spot having a plume origin must show continuous low shear velocity in the underlying upper and lower mantle.
- There are only a few hot spots that match this criterion (Afar, Bowie, Easter, Hawaii, Iceland, Louisville, McDonald, and Samoa), out of the 37 considered.
- In several cases, the anomalous seismic structure extends well below 410 km producing a deepening of the discontinuity defining the top of the transizion zone, but not the updoming of the discontinuity of defining its bottom.
- Using broader criteria based on seismological, other geophysical, and helium isotope data, Afar, Easter, Hawaii, Reunion, Samoa, Louisville, Iceland, and Tristan are likely to be of plume origin.





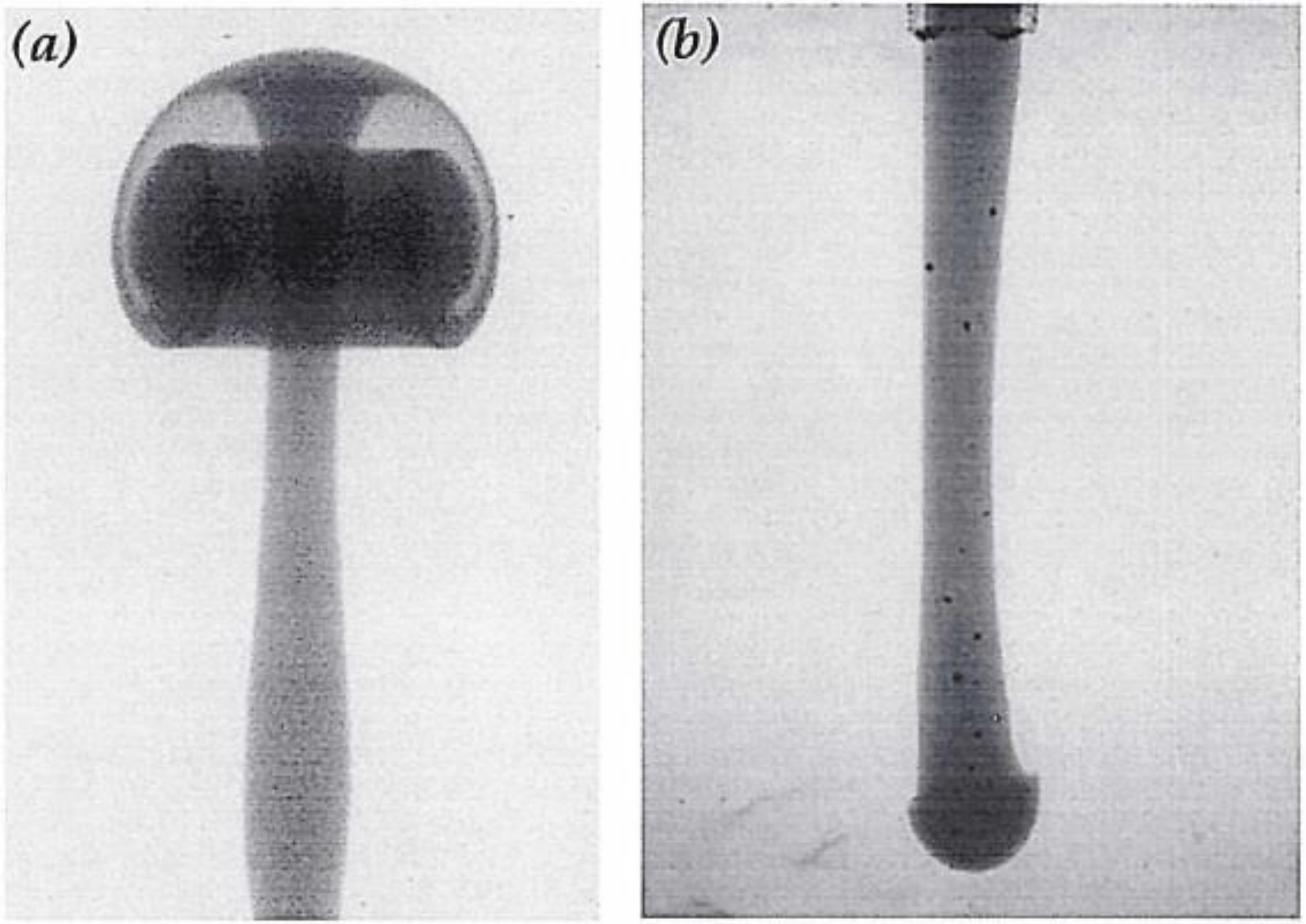
Contours of -1% P-wave velocity anomaly

McNutt and Caress, 2007, Treatise ofGeophysics, vol. 1

Ito and van Keken, 2007, Treatise ofGeophysics, vol. 7

P-wave tomographic images in map view around the Hawaiian hot spot

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).**

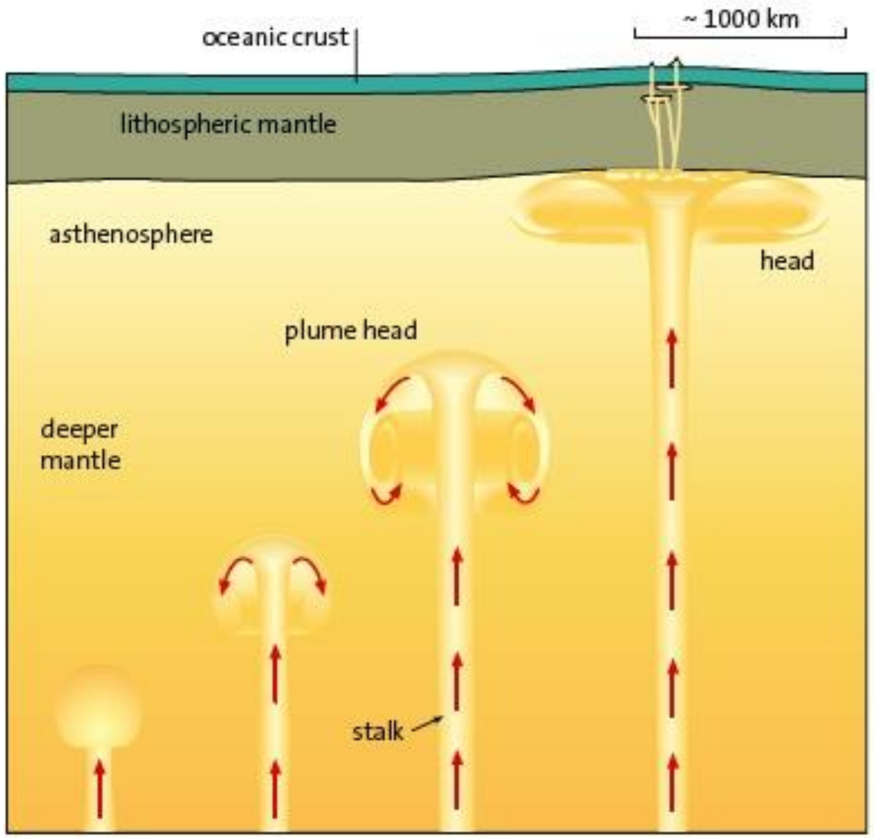
Plume heads slow down by the viscosity of the surrounding mantle:

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

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

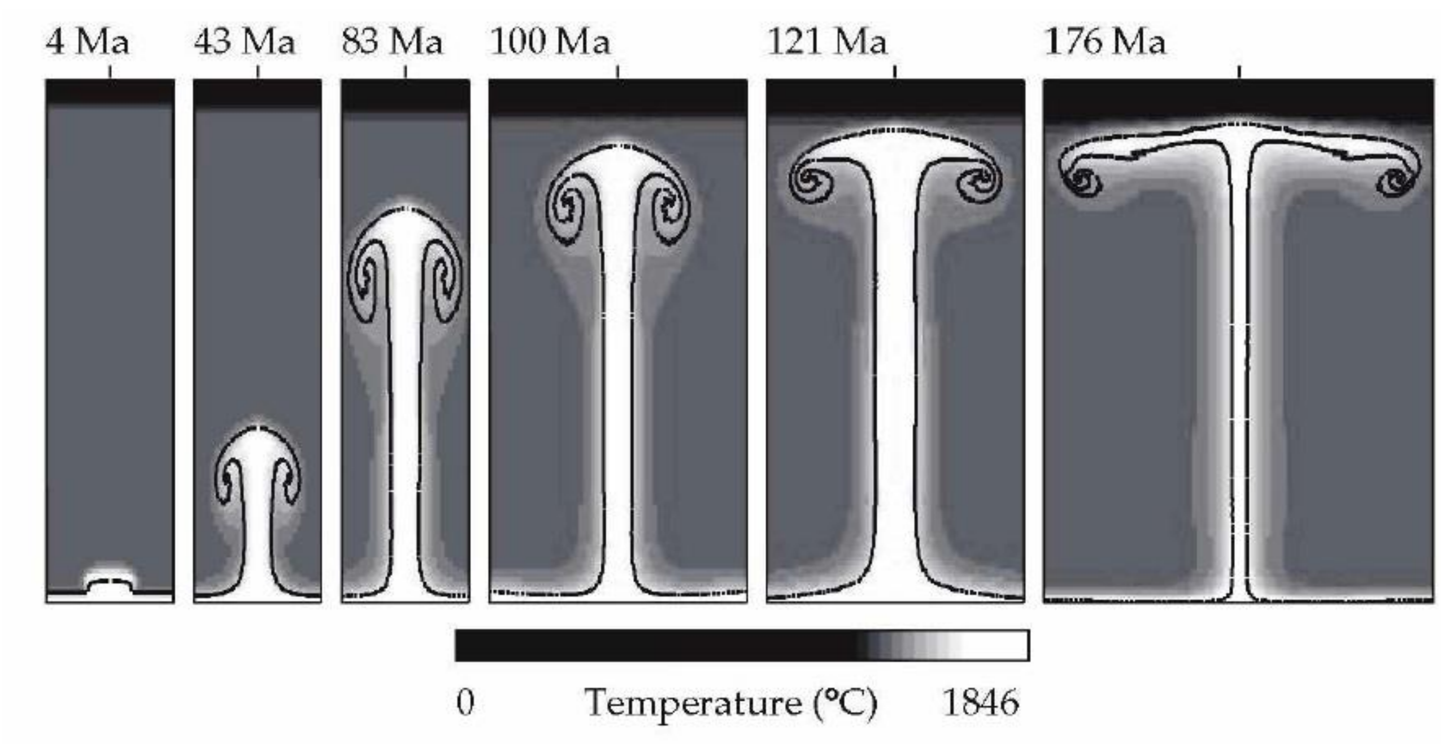
μ=viscosity
a=constant
P=pressure
V=activation volume
E=activation energy

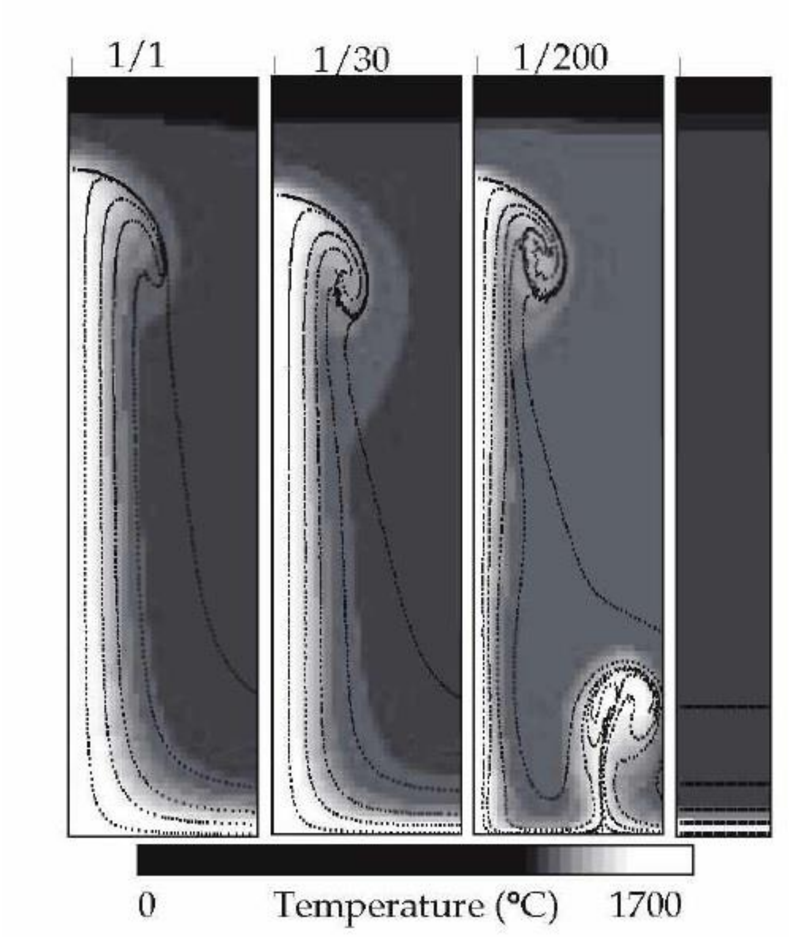
Activation enthalpy: H=E+PV



Plume generation

Plume tails become thinner by decreasing plume's viscosity

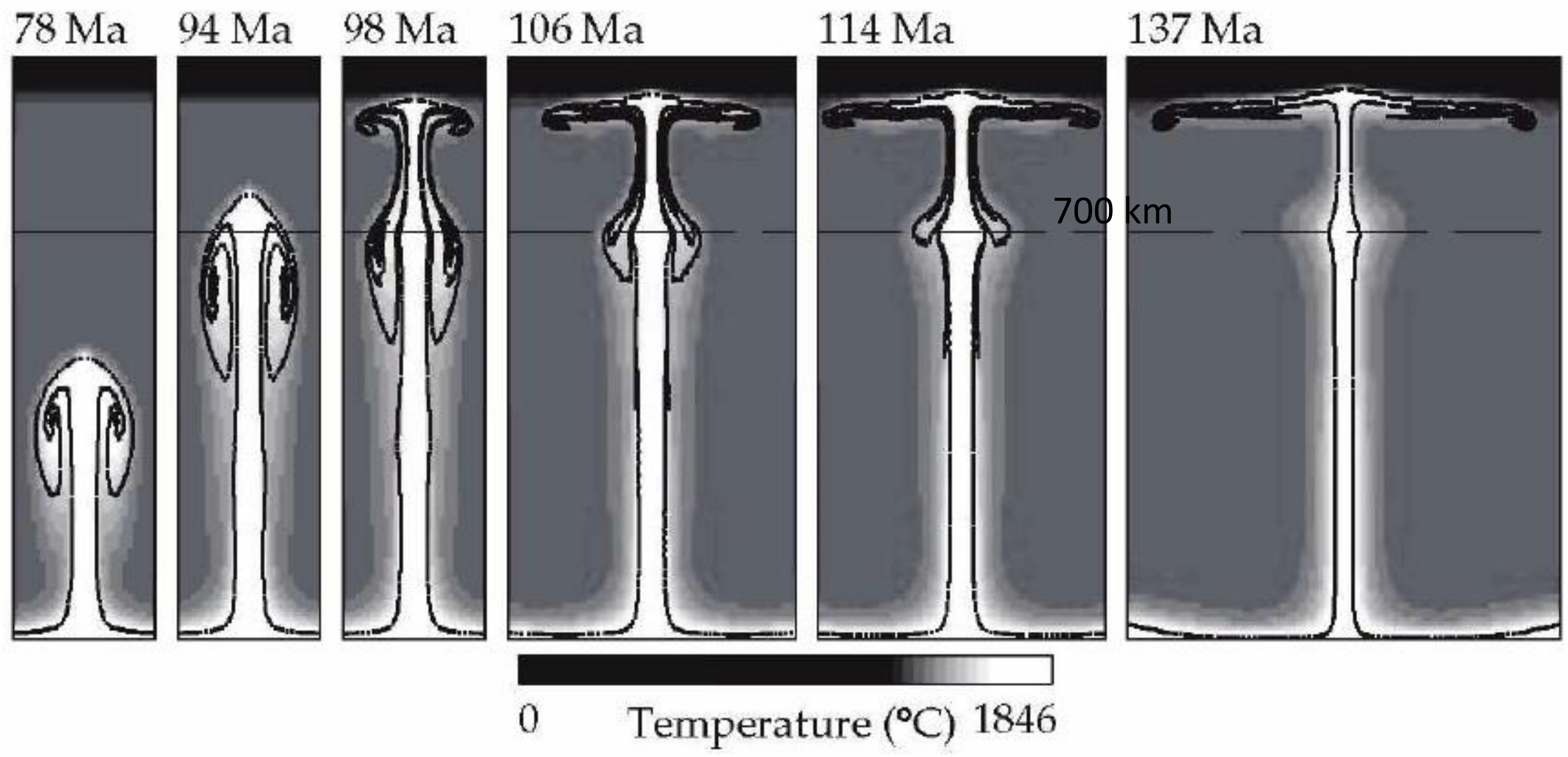




Plume heads have a larger buoyancy which is more capable of penetrating resistance than a narrow column.

Plume generation

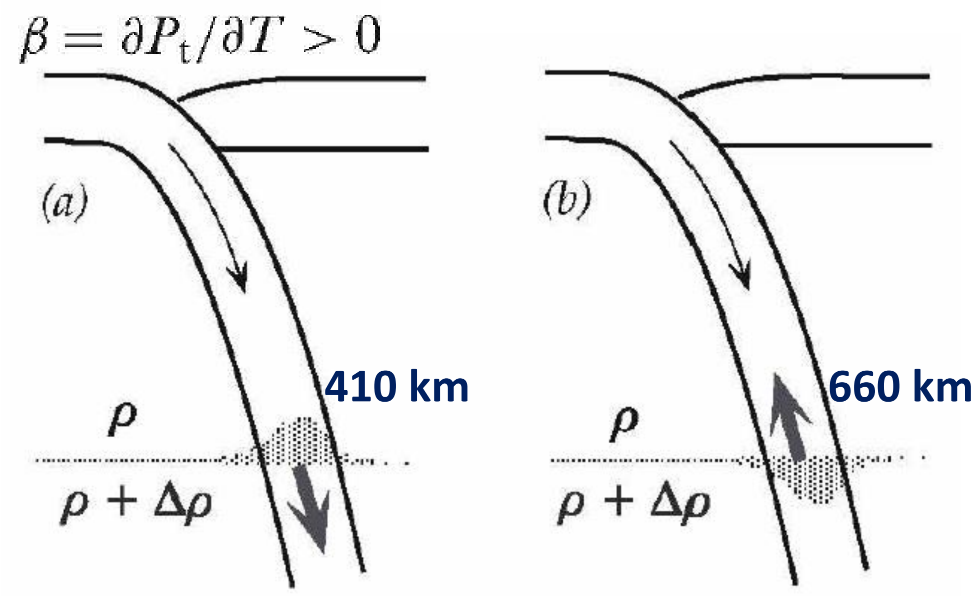
Effects of depth dependence of viscosity and a phase transformation on the plume



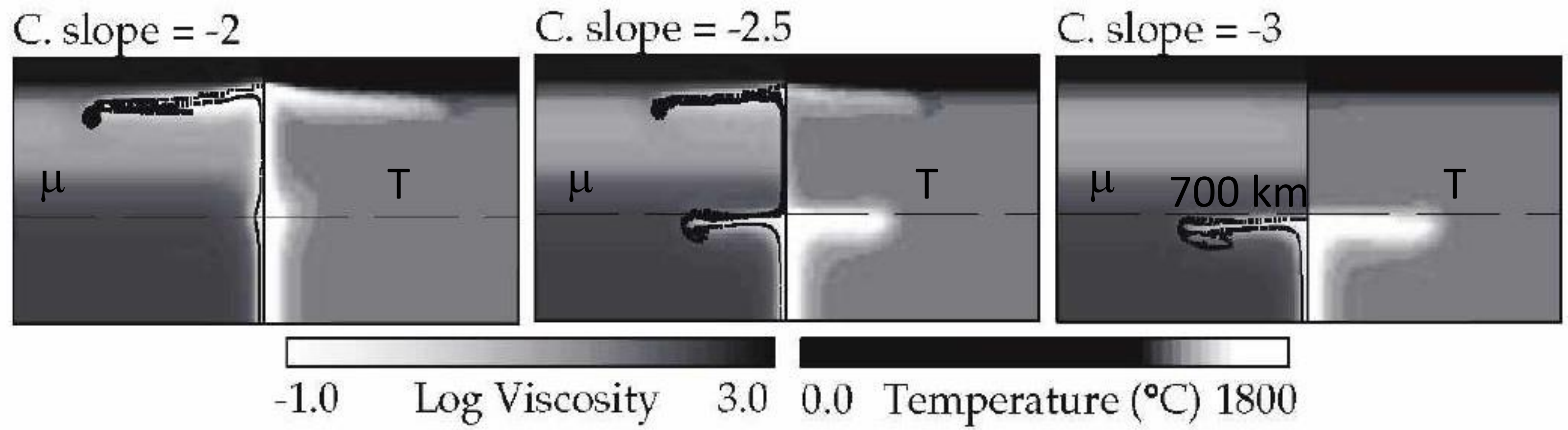
- The viscosity increases by a factor of 20 at 700 km and exponentially by a factor of 10 below.
- The effect of a phase transformation at 700 km depth is simulated with a moderately negative Clapeyron slope of -2MPa/K.

Plume generation

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

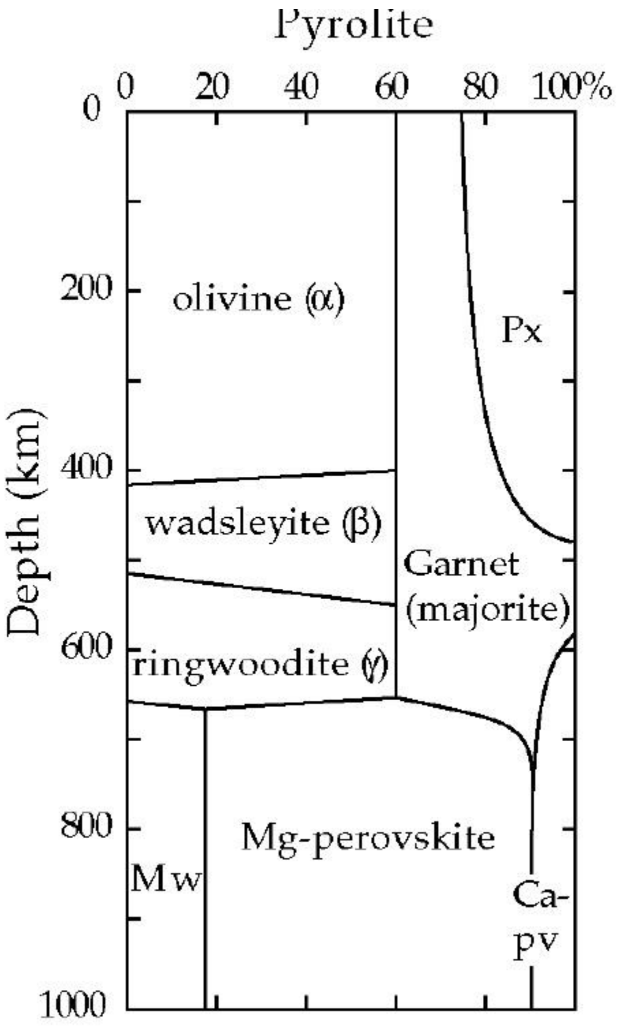


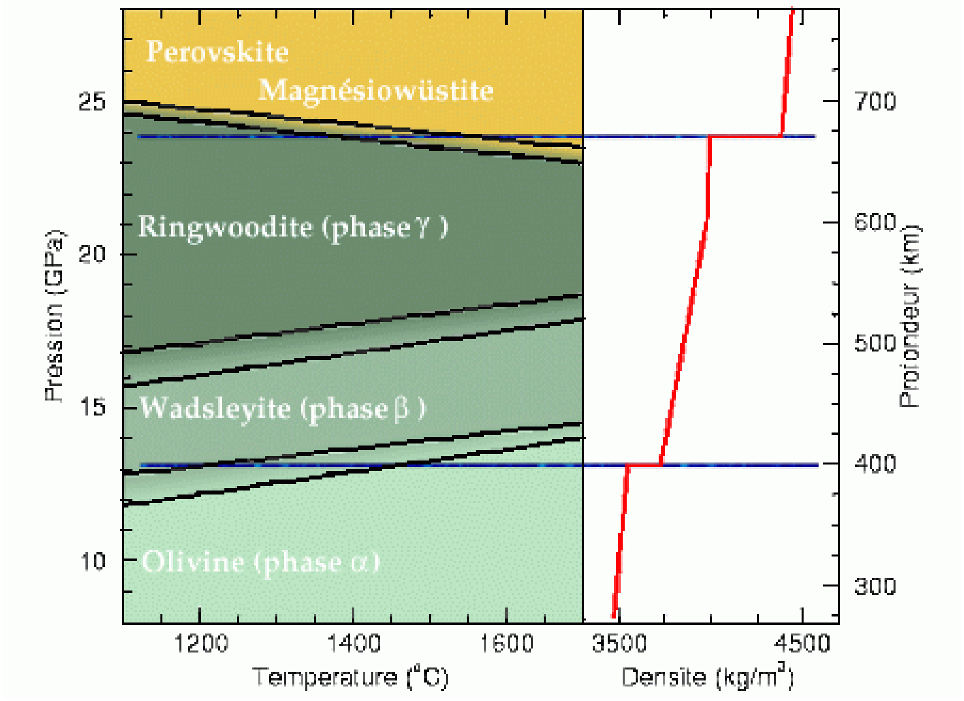
- A poitive Clapeyron slope causes a negative buoyancy force (broad arrow) that would add to the negative thermal buoyancy of the cold slab, aiding its descent.
- A negative Clapeyron slope delays the transformation to greater pressure and depth within a descending slab, producing a positive buoyancy 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 of ~3 MPa/K (esothermic reaction)
- 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 (endothermic reaction).
- 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).

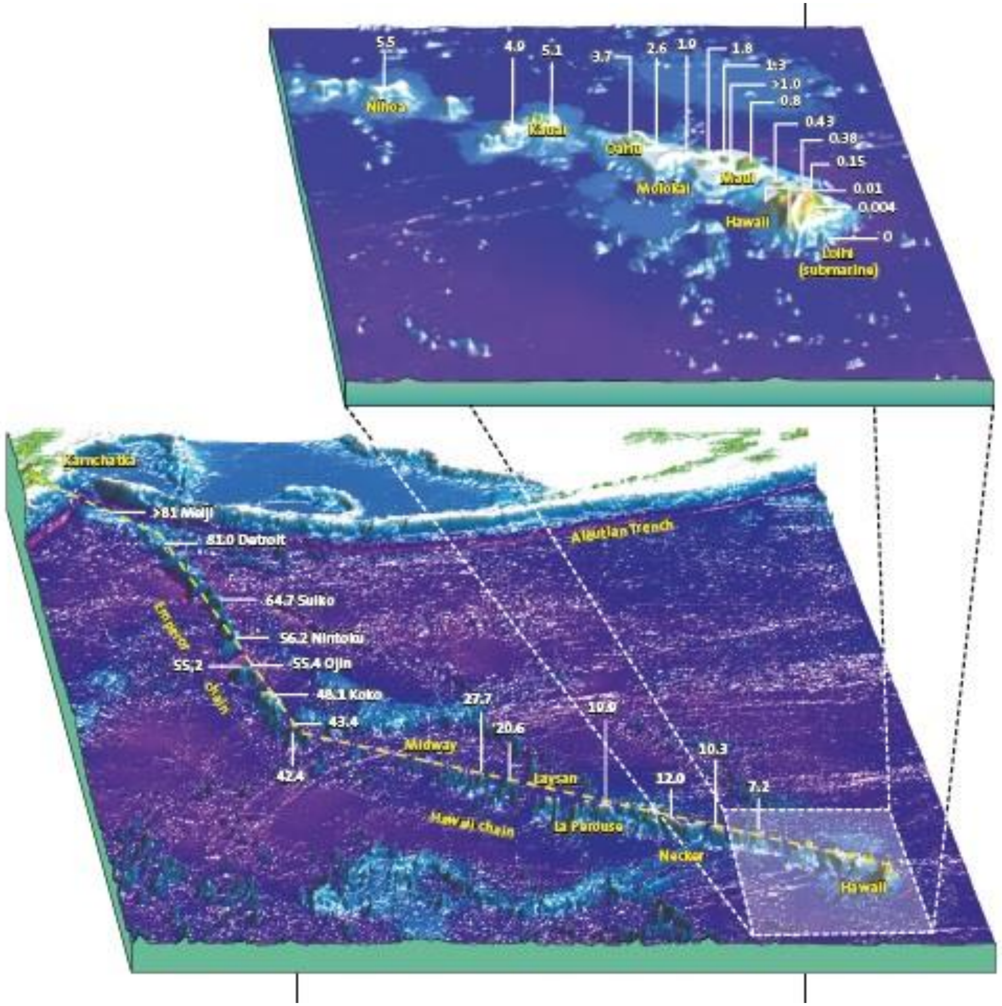
The effect of a phase transformation may be significant in 2D, constant viscosity model, but less significant in 3D and with *T* dependent viscosity.

Hot spot tracks

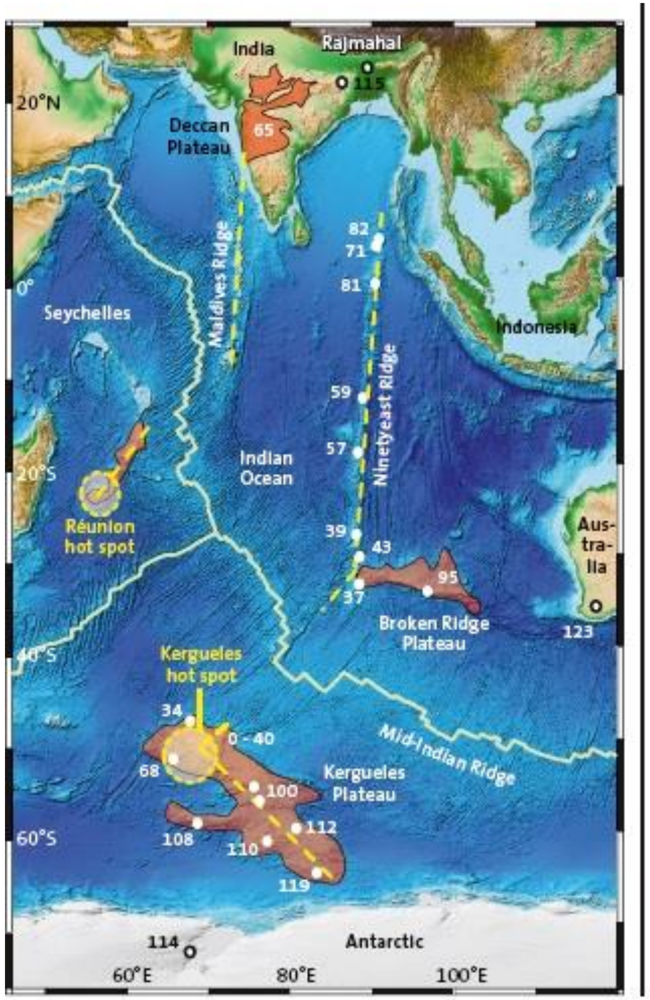
Hot spots form tracks on the ocean floor in response to the motion of the ocean plate

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

Hawaai (6000 km-long track)

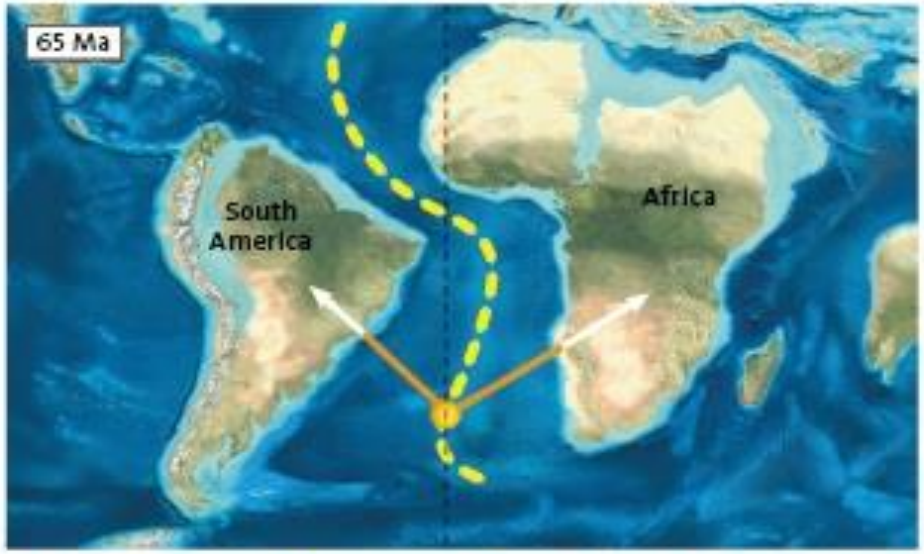


India

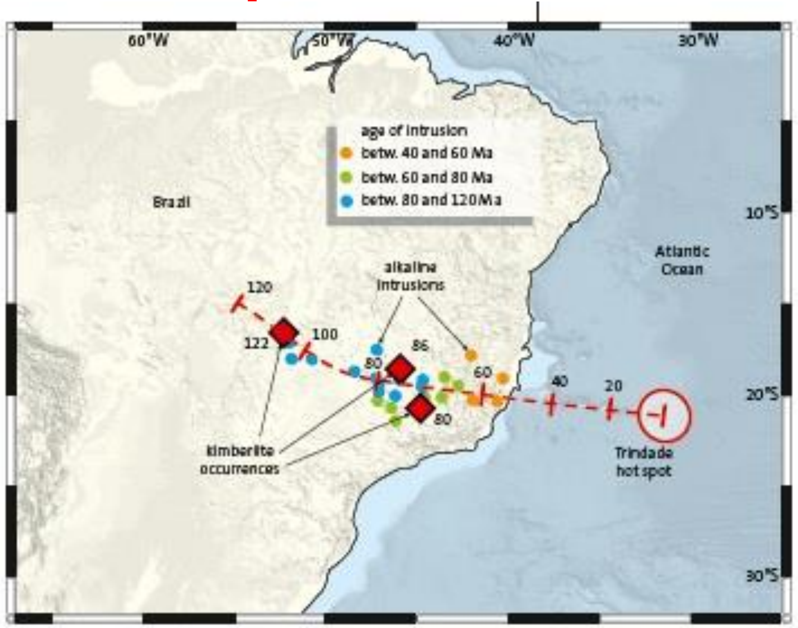


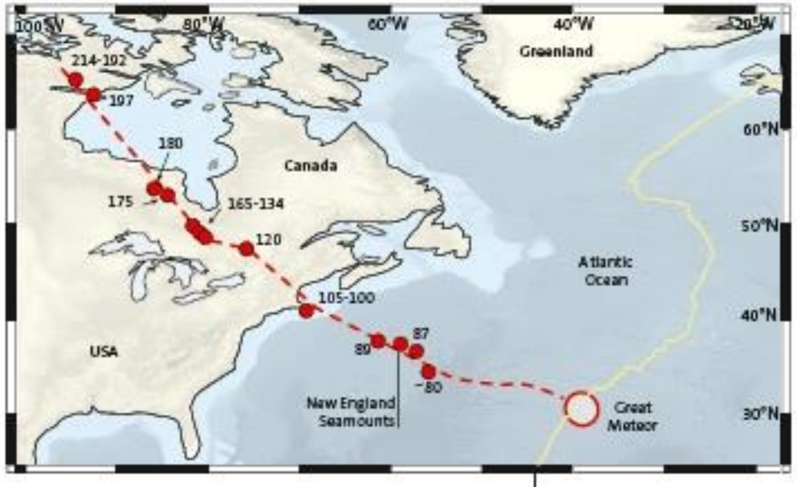
Maldives Ridge and the Ninetyeast Ridge were generated by the hot spots of Réunion and the Kergueles.

Hot spot tracks



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



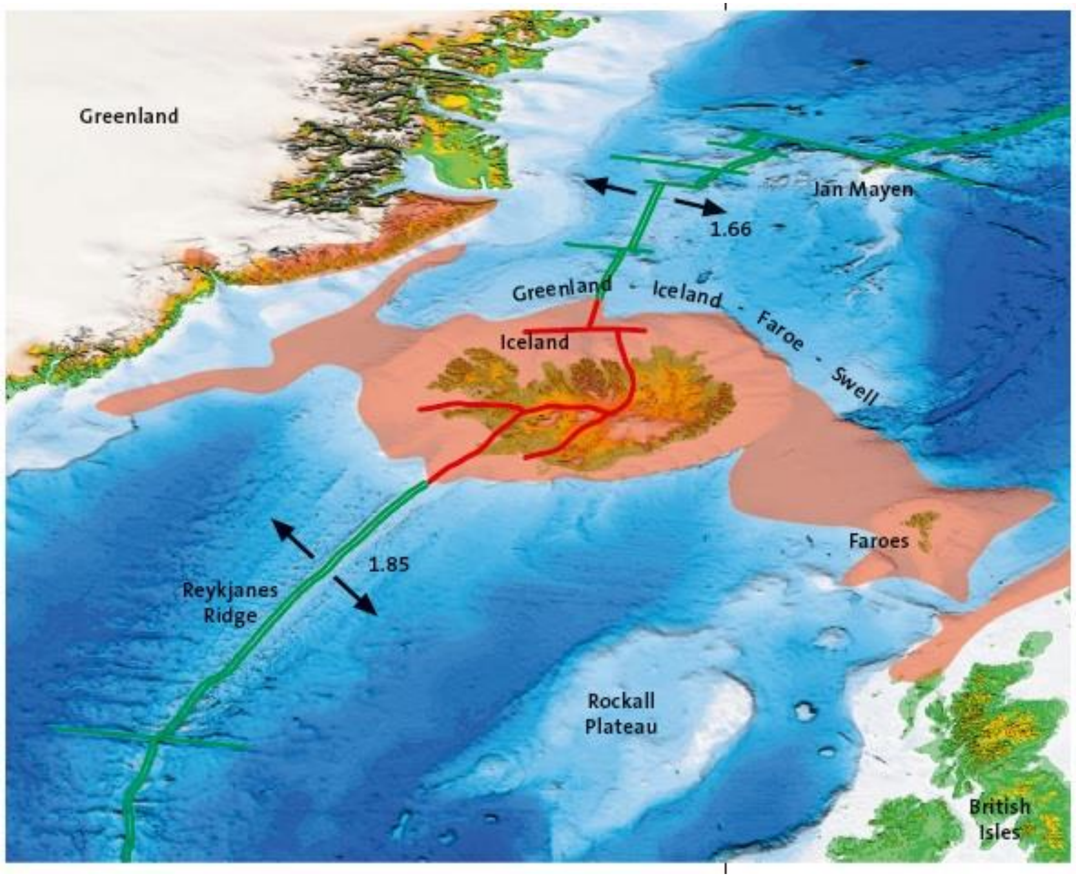


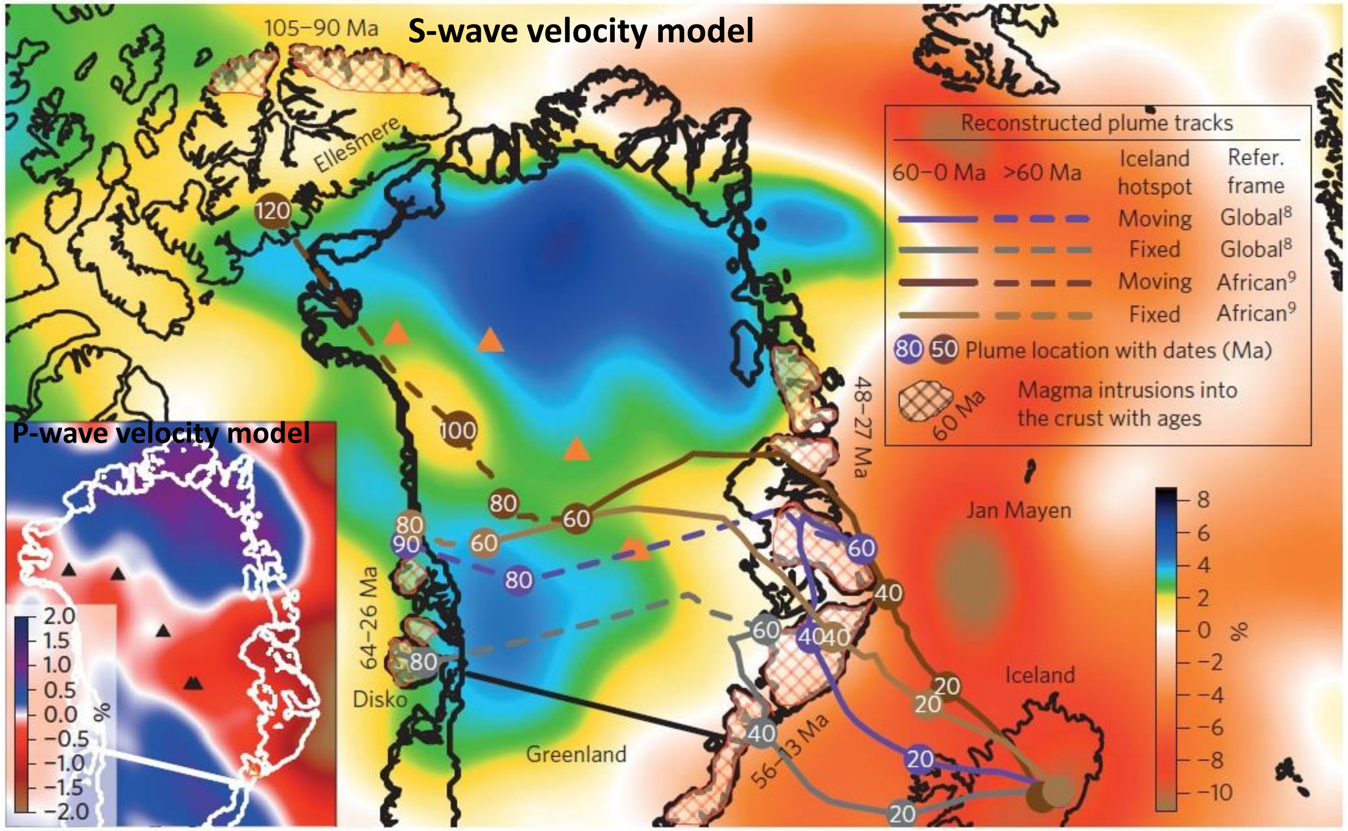
- The hot spot of Trindade in the South Atlantic left a track during the Cretaceous when the Brazilian Shield drifted over the hot spot.

- As North America moved across the hot spot for 100 Myr during the Jurassic and Early Cretaceous, a 4000 km-long track was formed that extends from Hudson Bay to New England.

Iceland hot spot

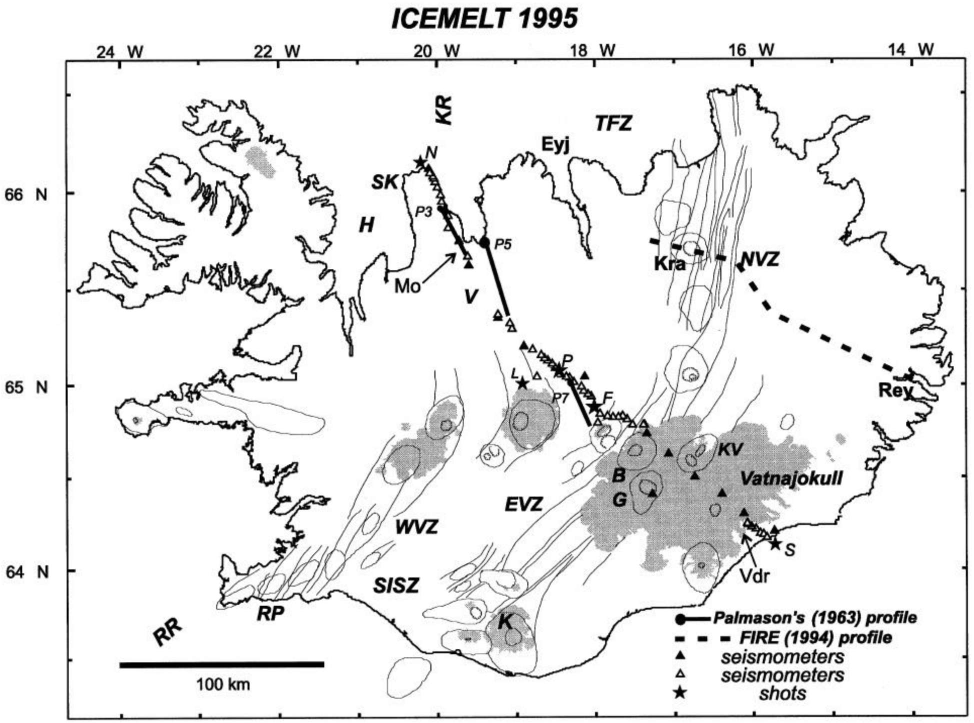
- Iceland is a product of the activity of a mantle plume, which ca. 80 Myr (Late Cretaceous) was located below Western Greenland and the Canadian Arctic.
- The young mid-ocean ridge came under the influence of the mantle plume in the Eocene ca. 30-40 Myr.

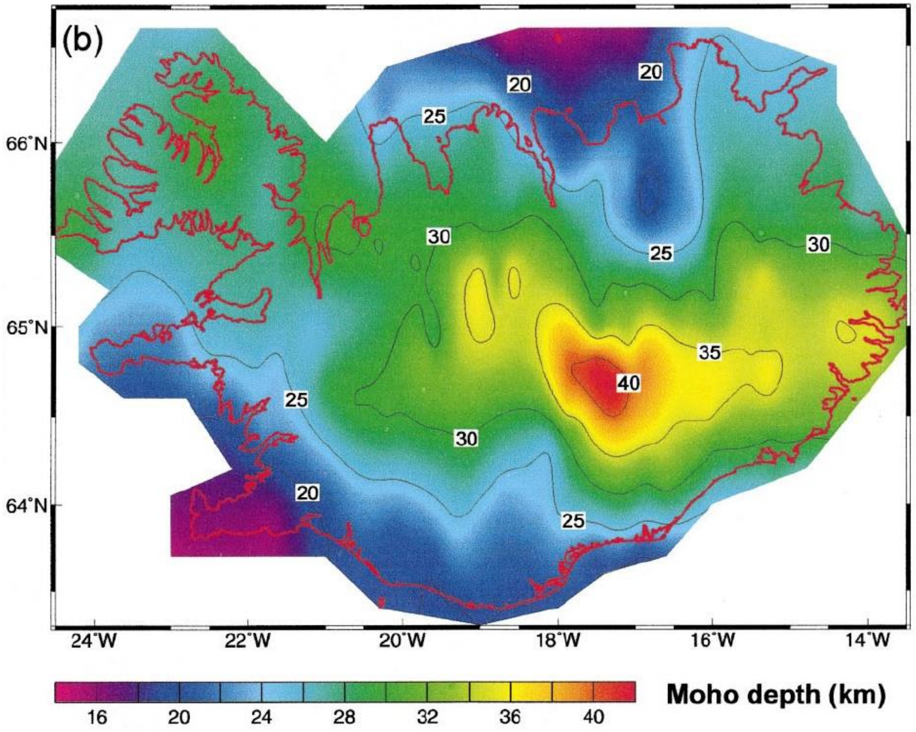


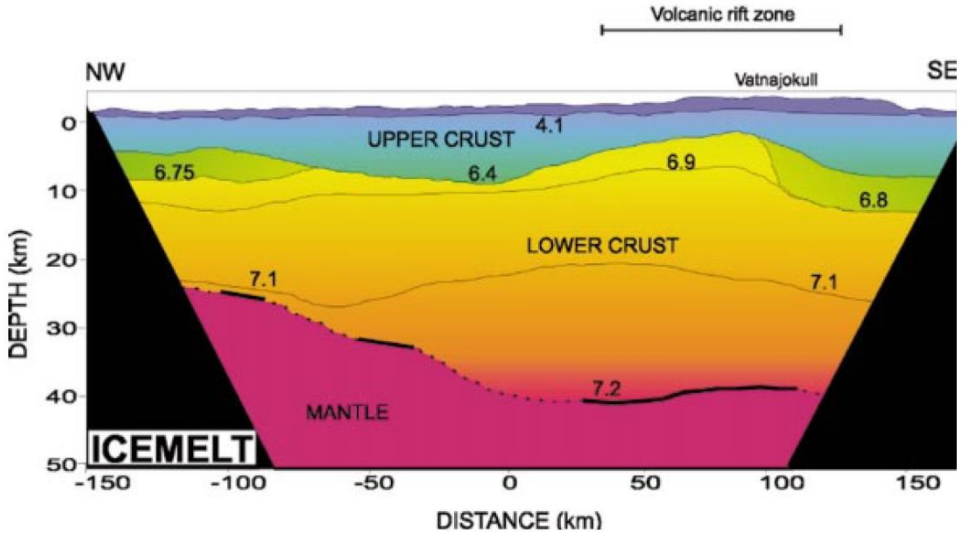


Rogozhina et al., 2016

Iceland Crust



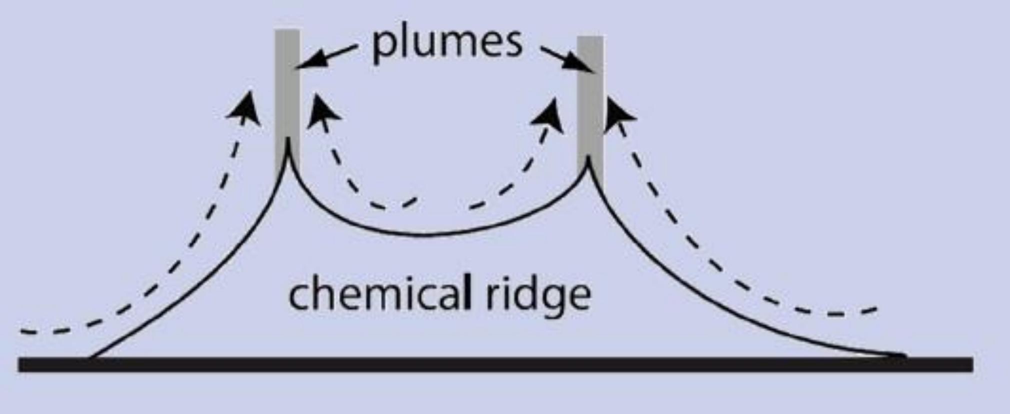


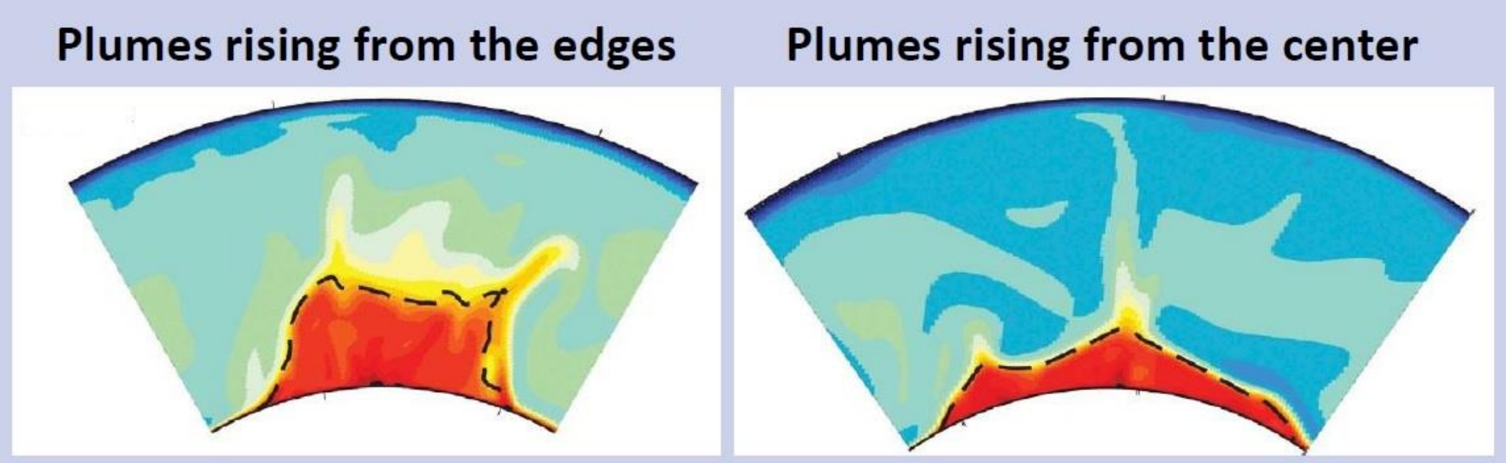


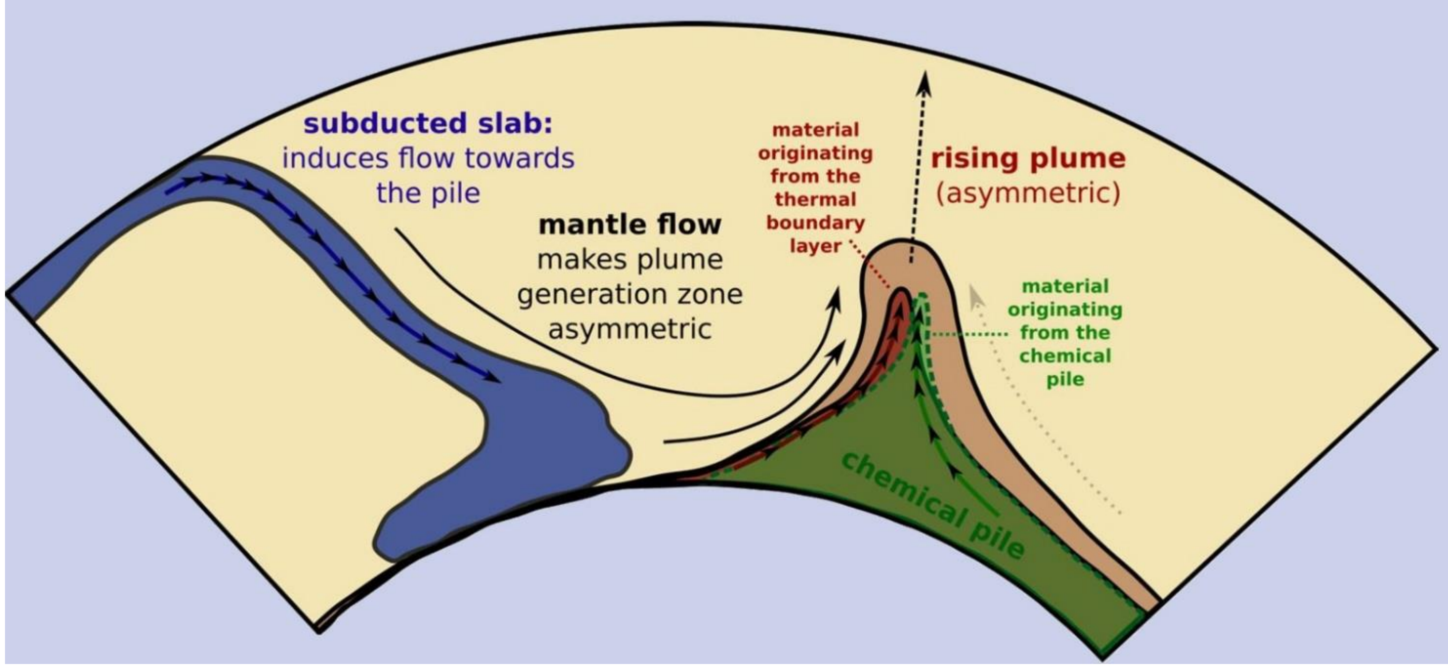
Darbyshire, 2000, EPSL, 181

Darbyshire, 1998, GJI, 135

Symmetric and Asymmetric Plumes

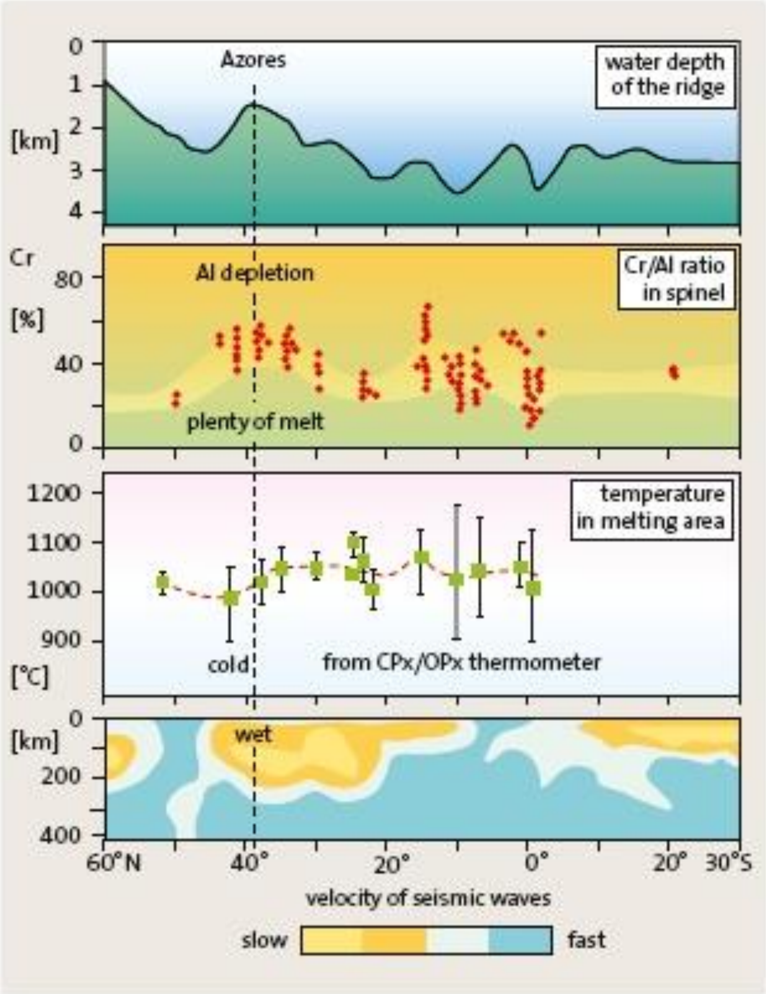


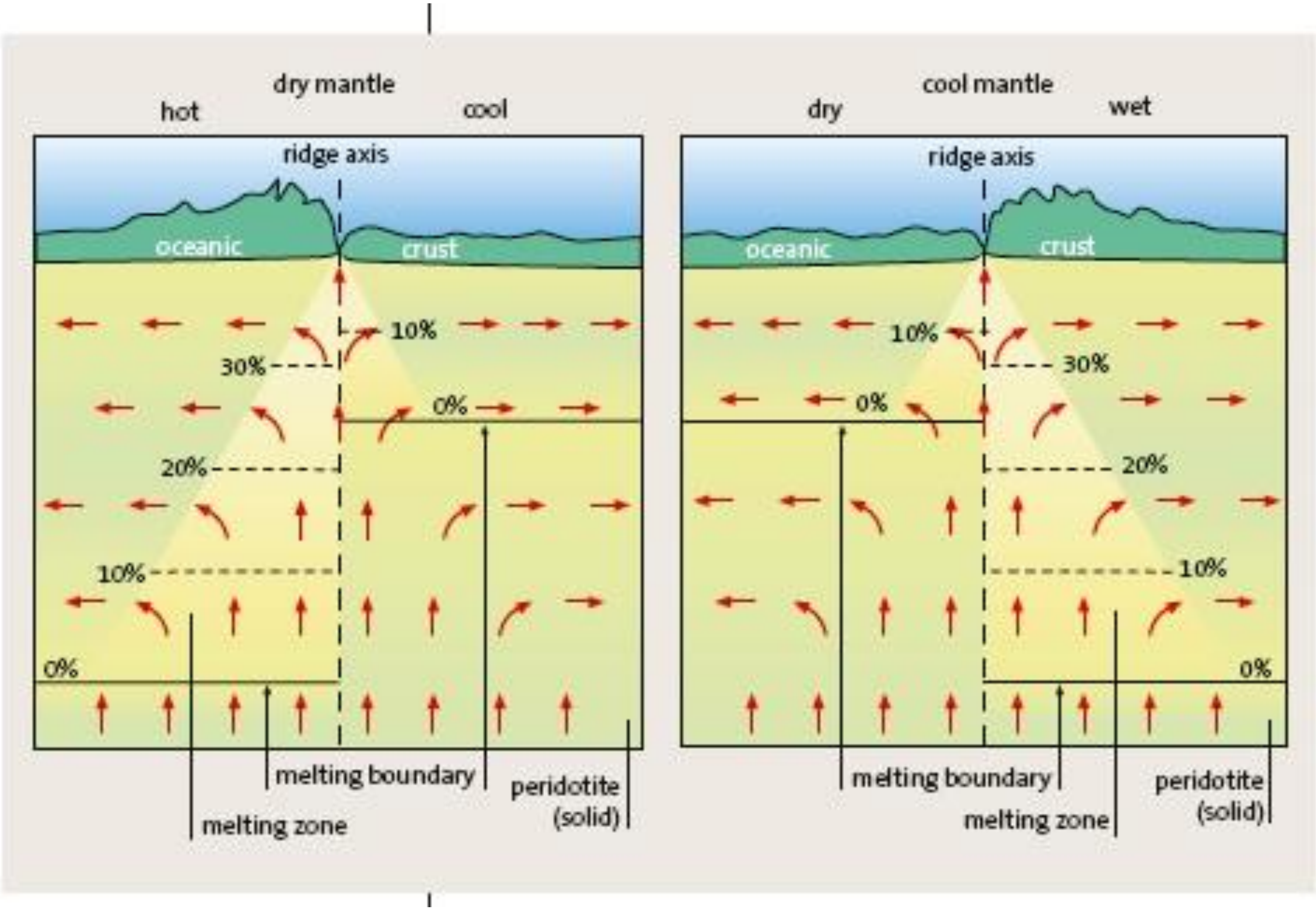




- **If the density anomaly has steep edges, the plumes will rise preferentially near the edges.**
- **The mantle flow and the subducting slabs influence the symmetry of the pumes.**

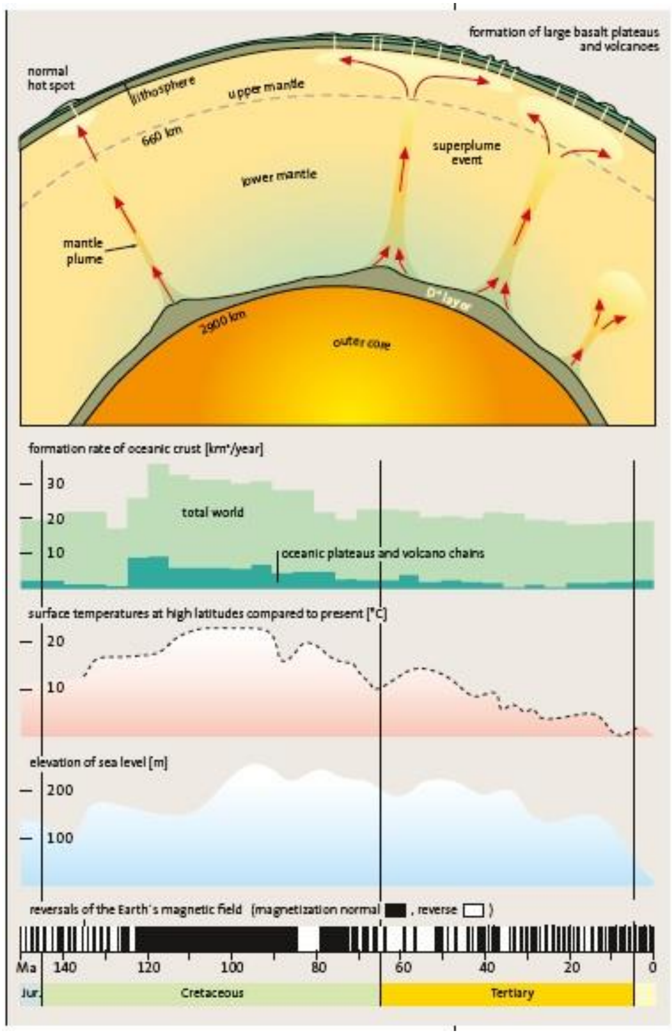
Azores: hot or cold spot?

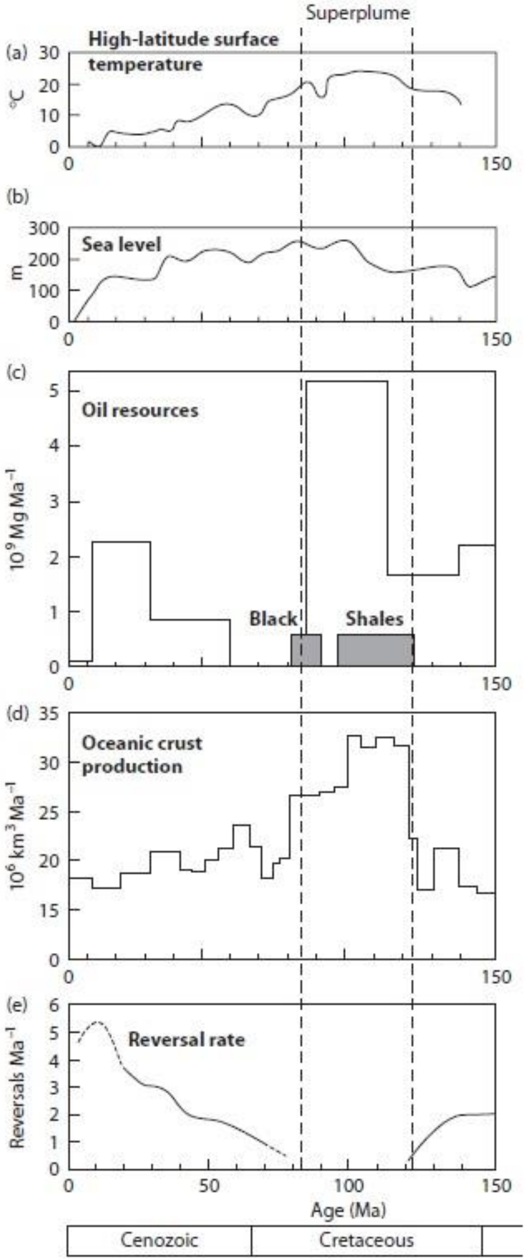




- Basalts of the Azores contain three to four times the amount of water as normal basalts of mid-ocean ridges indicating a “wet” melting region.

Cretaceous Superplume event





- The Cretaceous from ca. 125 Myr to 85 Myr 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/yr to ~35 km^3/yr.
- At the high latitudes the surface *T* of the Earth increased by about 10°C (as shown by oxygen isotope measurements). This effect was probably caused by the release of large amounts of CO_2 during the volcanic eruptions, which created an enhanced “*greenhouse*”.
- The rates of carbon and carbonate sequestration in organisms increased due to the larger area of shallow seas and the increased *T*.
- Alternatively these processes can be ascribed to a general reorganization of plates on a global scale associated with the break-up of Pangea and reorganization of the Pacific plate, which determined a passive upwelling of the asthenosphere.

Other possible mechanisms for intraplate volcanism

Fixed hot-spot model fails to explain:

(1) Observed departures from linearity of individual volcanic chains and inconsistent orientations among multiple chains which lie on the same plate;
(2) Short-lived chains and ones which fluctuate in size;
(3) Violations of predicted along-chain age versus distance behavior;
(4) Some linear volcanic ridges in the Pacific form much more rapidly than would be predicted by fixed hot-spot model and display intermittent volcanic activity with longevities shorter than 40 Myr.

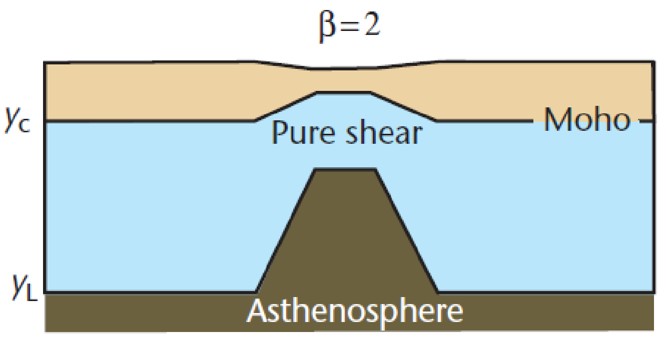
Other Hypotheses:

- Diffuse plate extension has been thought as a possible mechanism of intraplate volcanism or as alternative (1) cracking of the lithosphere under the action of thermal contraction (inconsistent with low seismic velocities beneath the volcanoes), (2) decompression melting in small-scale convective upwellings.
- In oceanic lithosphere, decompression melting may be possible only in regions where a large-scale mantle upwelling can counteract conductive cooling, keeping the mantle at its solidus temperature over some depth range.
- In the absence of a large-scale upwelling, buoyant melting may not occur spontaneously but may be triggered by some initial upwelling due to relief on the bottom of the lithosphere.
- Melting beneath spreading centers should produce a compositional lithospheric layer that is both more viscous (because of de-hydratation) and compositionally buoyant of a thickness comparable to the maximum depth of melting beneath a spreading center.

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

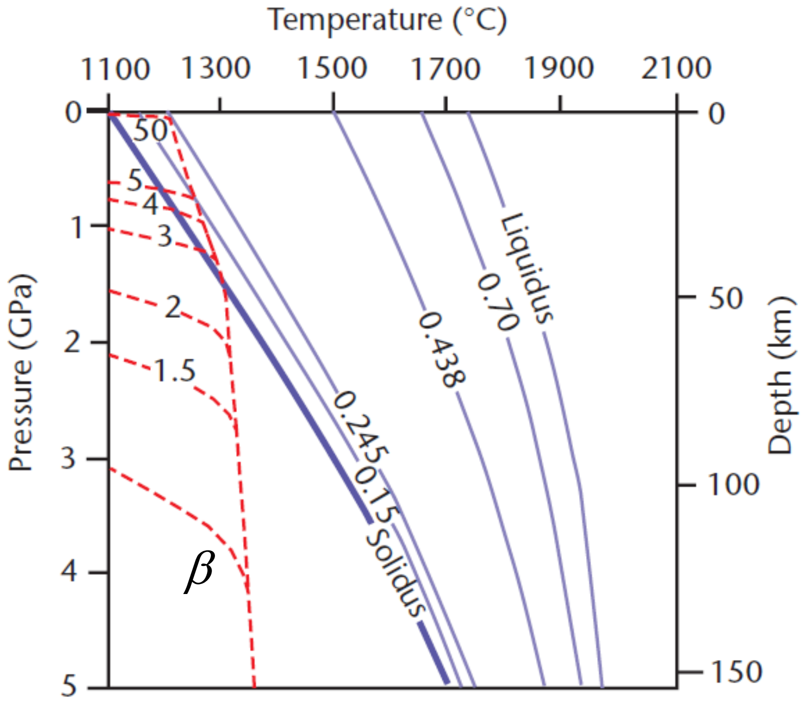
For β=2, T_p= 1400°C (due to plume activity), and L= 100 km, Thick_{MELT}= 2 km



$\nu=\eta/\rho$

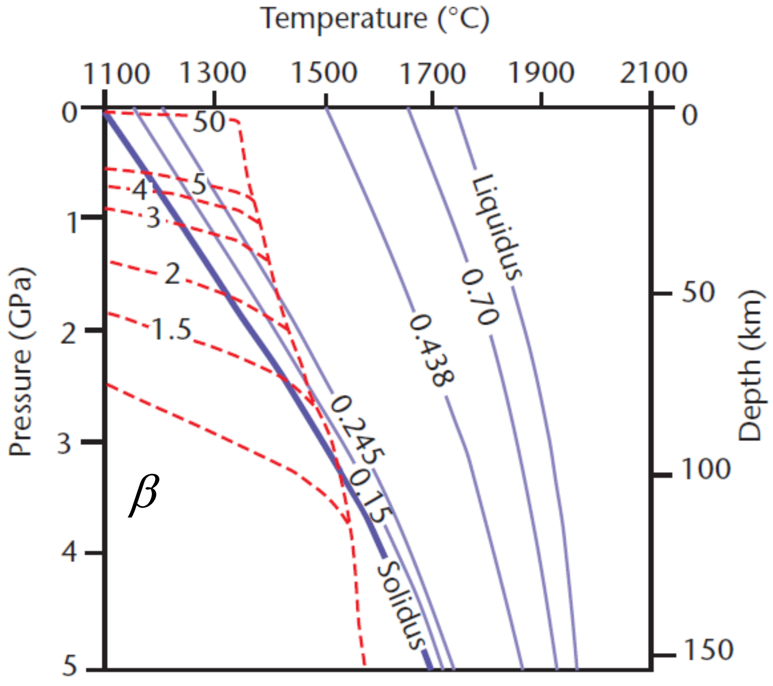
ν=kinematic viscosity

In absence of plume



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

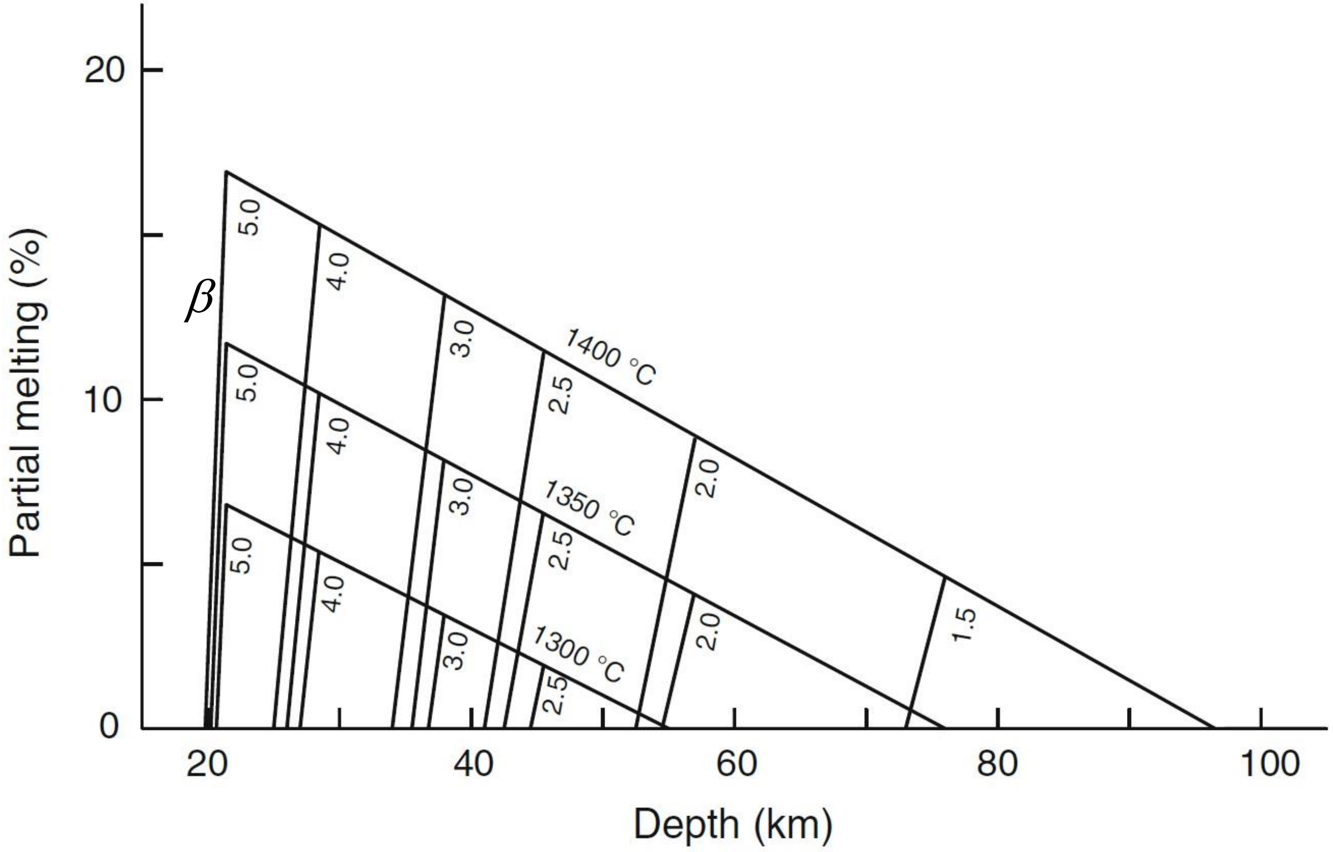
In presence of plume



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

Melt generation during continental extension

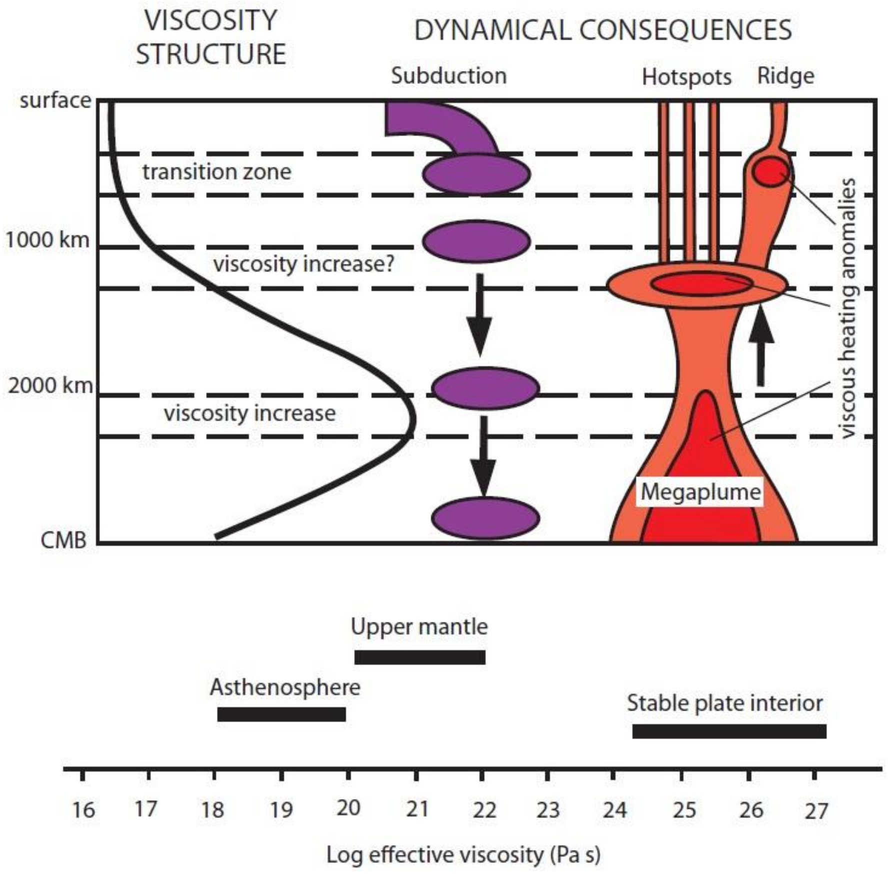
- For a thermally normal asthenosphere (T_p = 1300°C), no melting occurs unless stretching exceeds values of 2.0.
- The contribution of heat flow from melt lasts several years after the rift episode and even for larger periods than the characteristic thermal time or the solidification time of the melt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



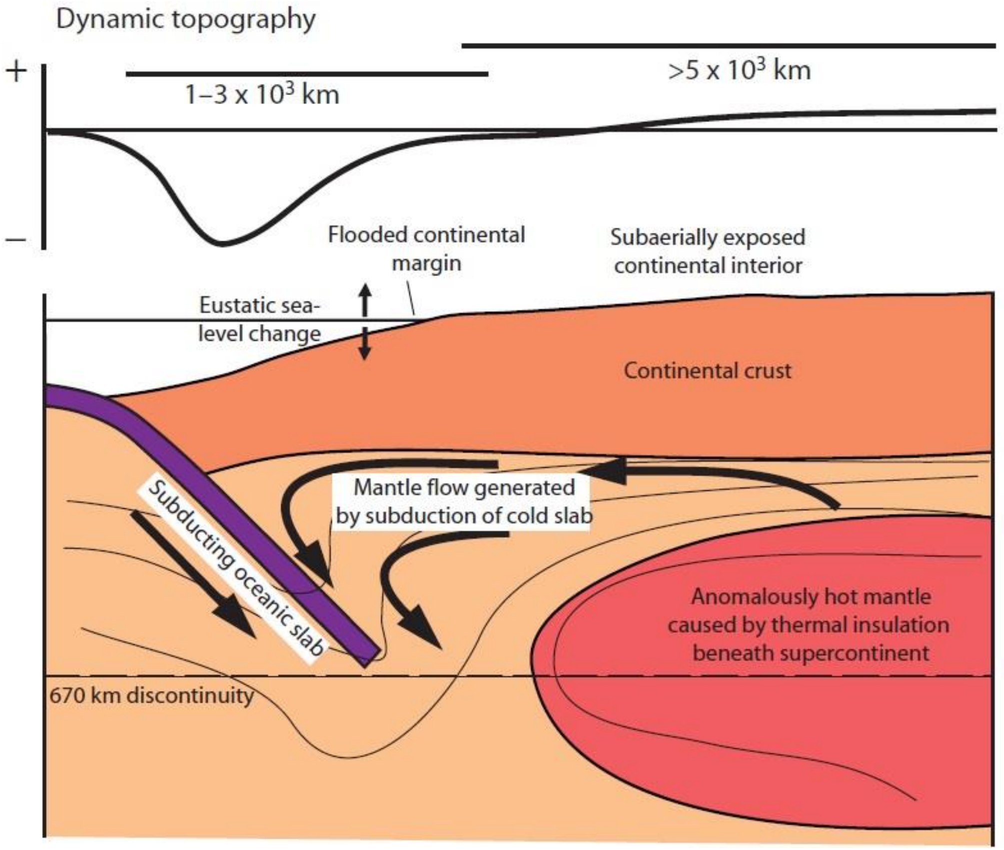
Partial melting as a function of depth at various amounts of extension β (1.5, 2.0, 2.5, 3.0, 4.0 and 5.0) and for asthenosphere T_p (1300, 1350, and 1400 °C).

Dynamic topography

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

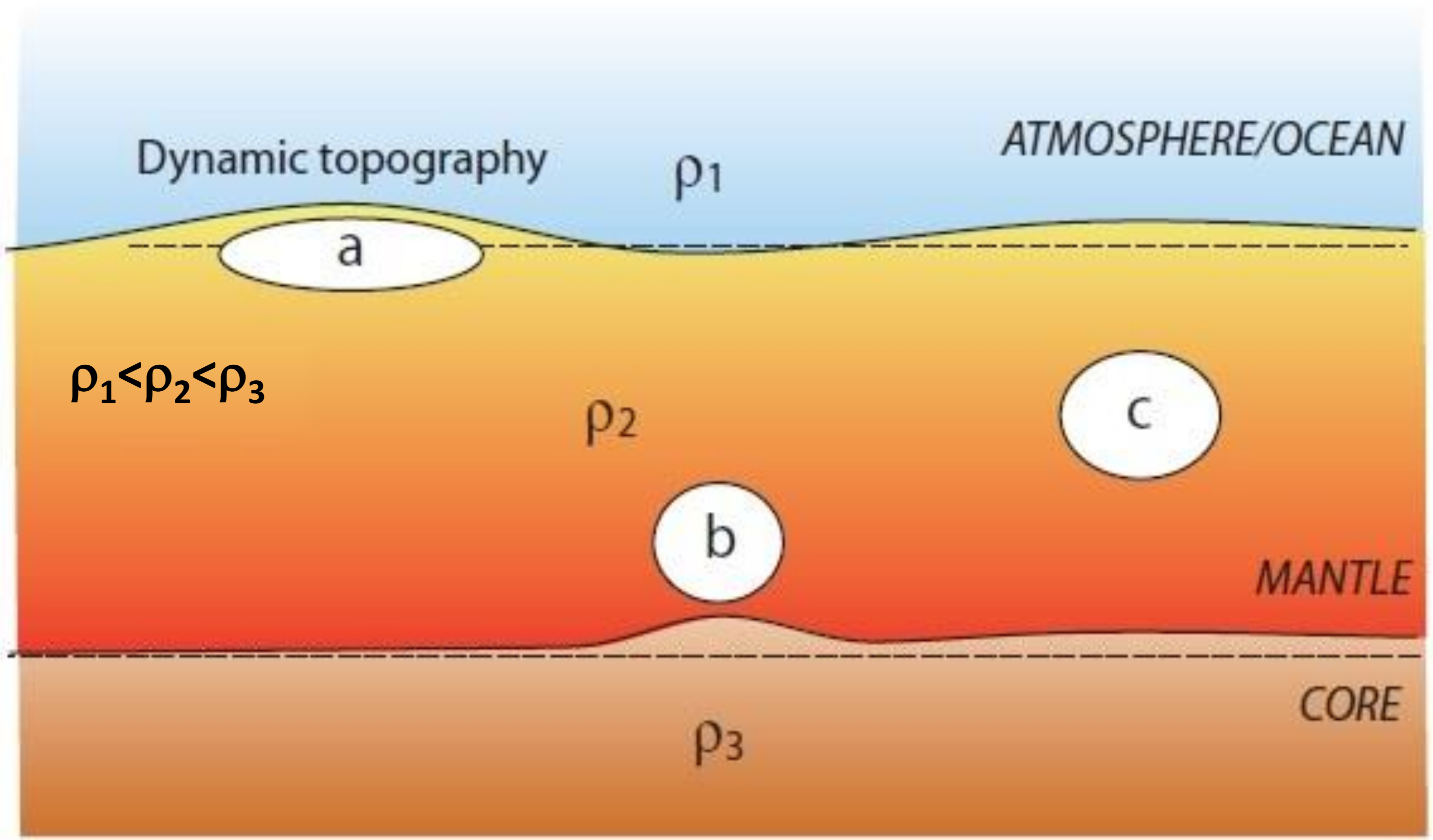


- Dynamic topography is transient and is usually of the order of few hundred meters (300-500 m).
- Response time of the mantle to a disturbance depends on its viscosity and wavelengths of the anomalous body (e.g., λ 1-3 x 10^3 km, t= 10^4 yr).
- It is estimated by removing from the topography the isostatic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

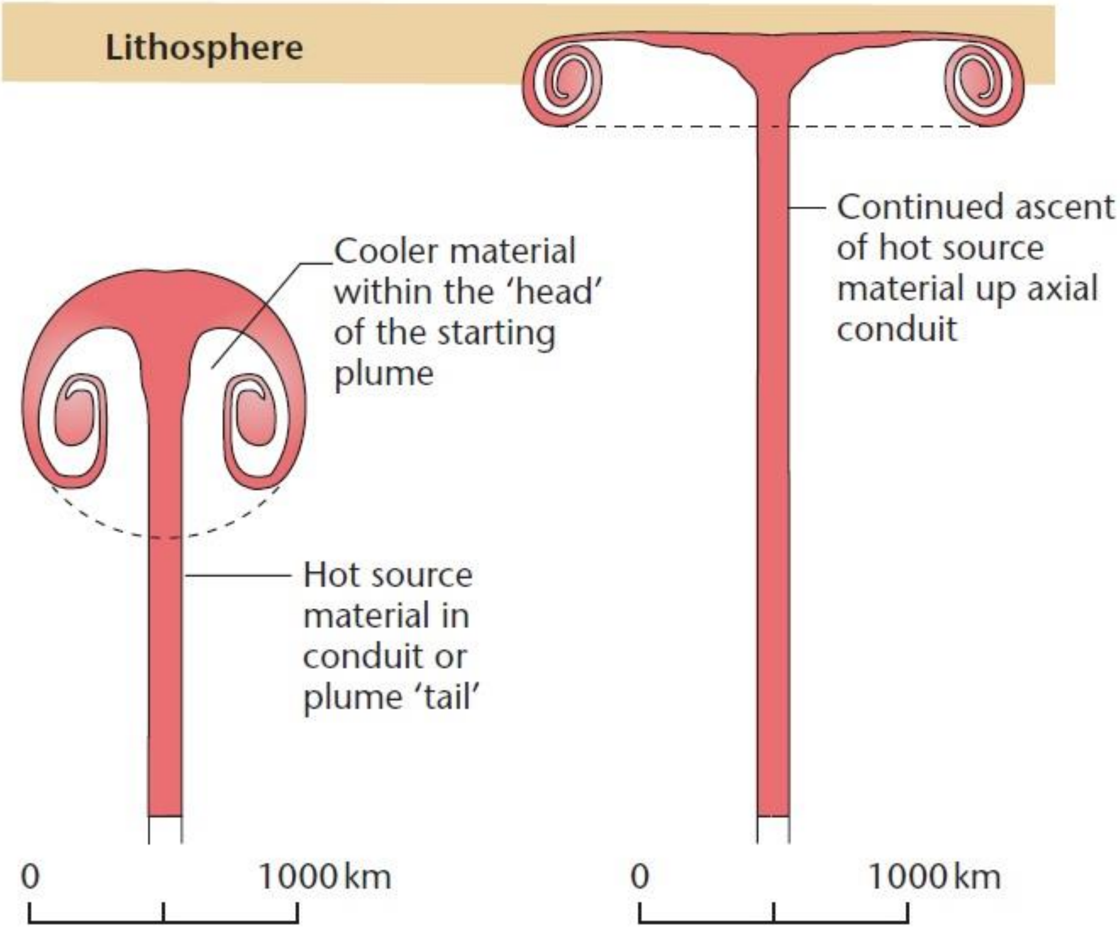


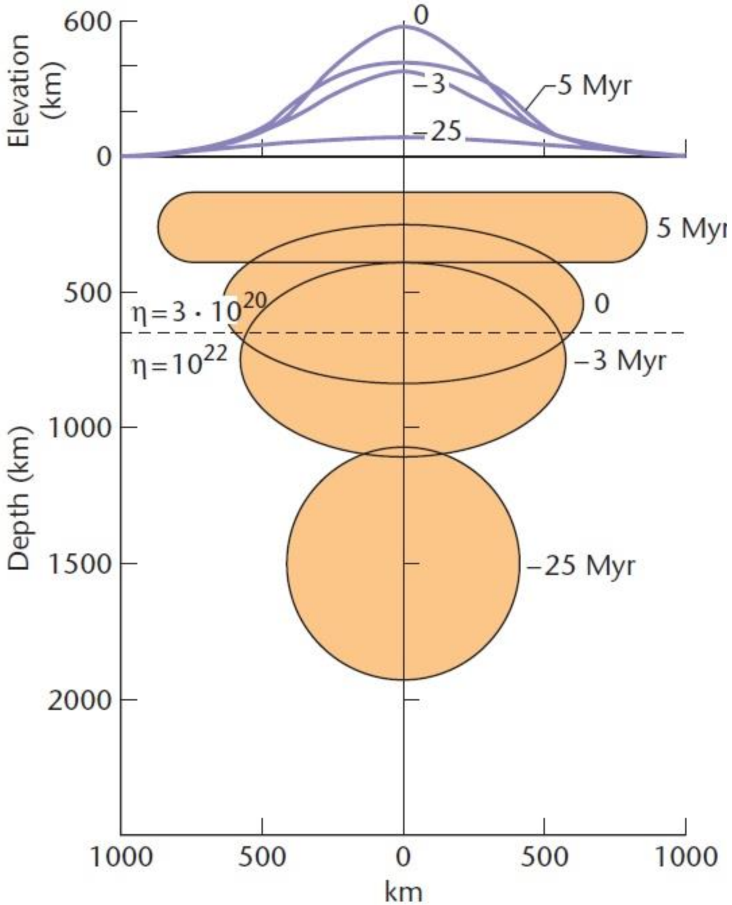
Dynamic topography

- 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



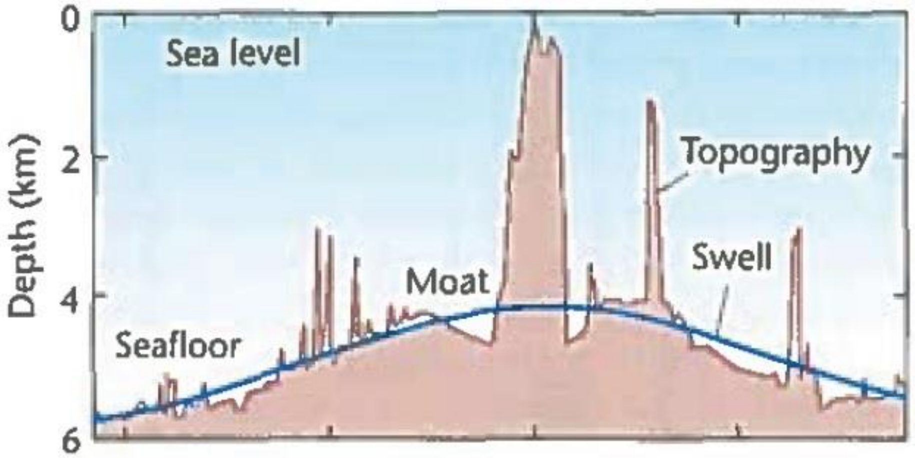


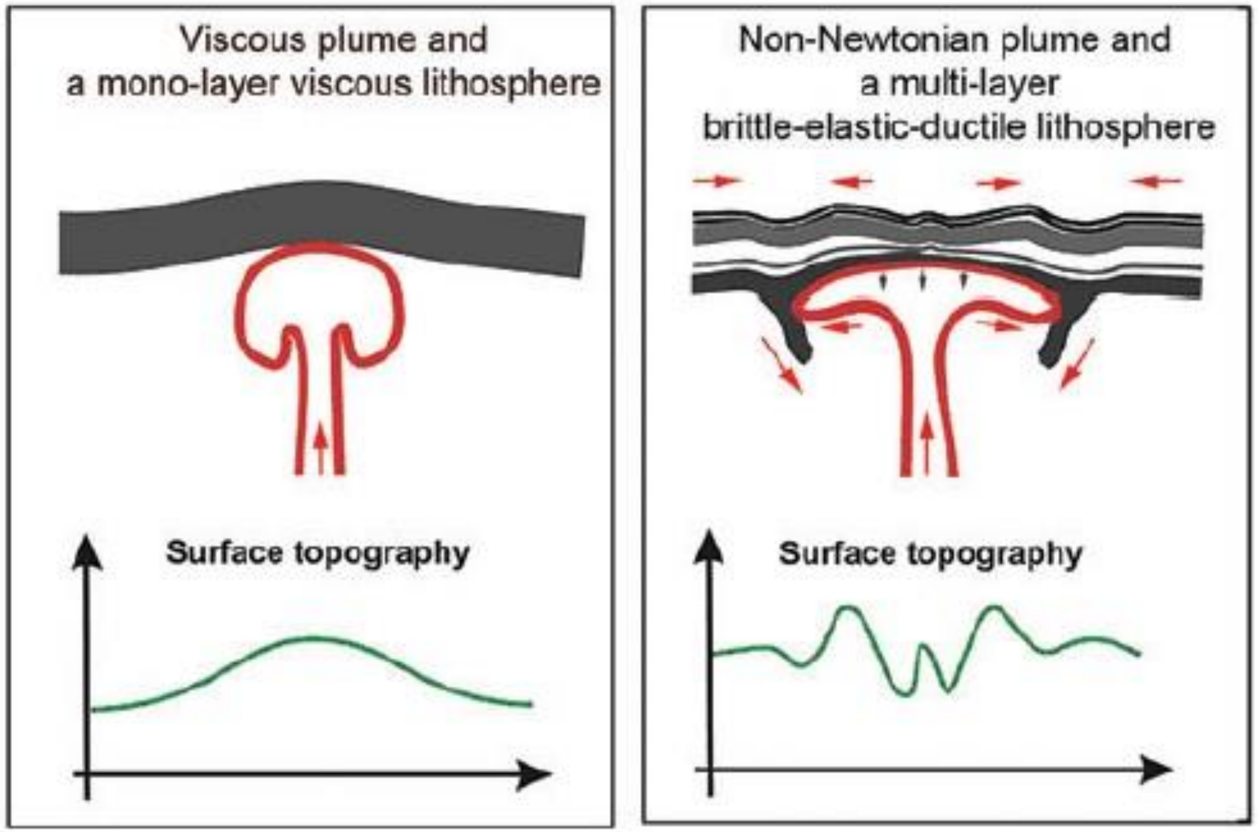
Buoyancy flux: 3×10^4 N s^{-1} ΔT=300 K

- 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

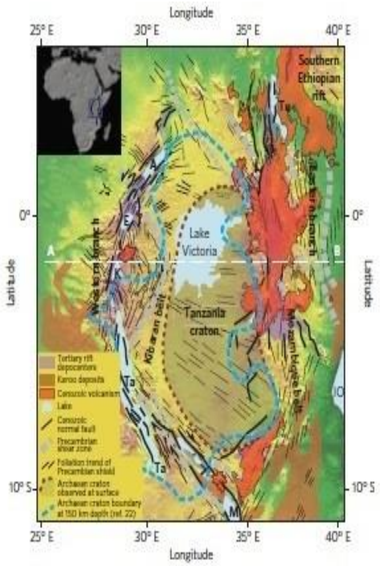


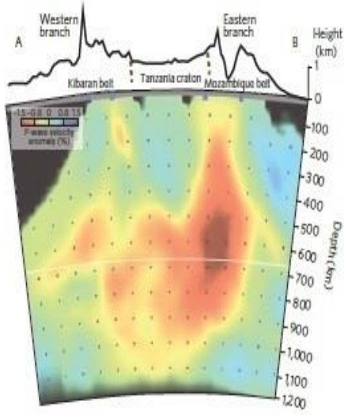


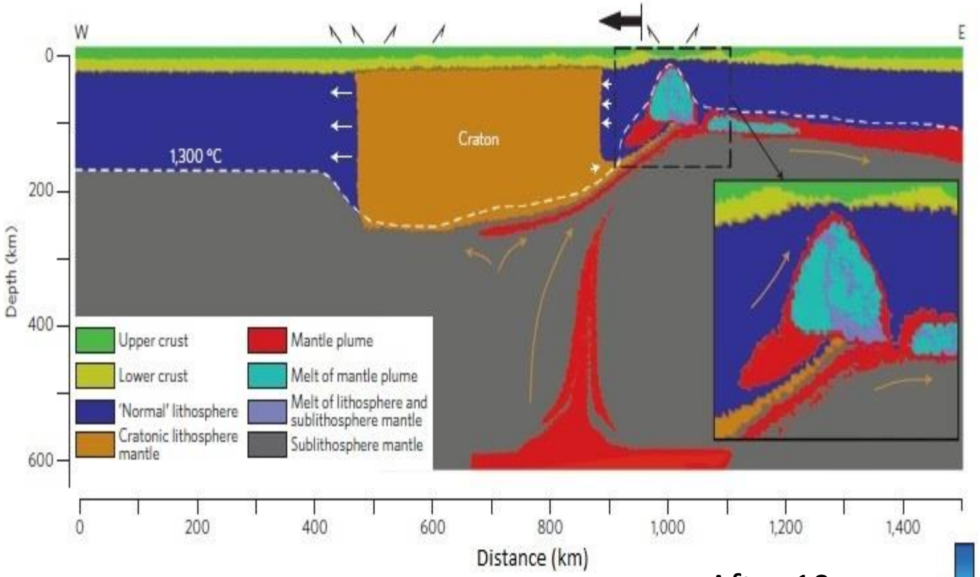
Burov and Cloetingh, 2009, Geophys. J. Int., 178

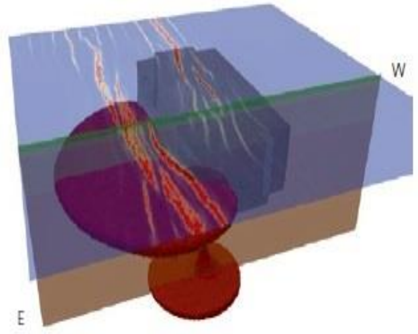
- The conventional models predict only long-wavelength (controlled by the plume head size) isostatic topography due to plume impact.
- Accounting for plate rheology and multilayer lithosphere structure yields a more complex response, with several short-wavelengths generated by intraplate deformation,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

Plume beneath a craton
(Tanzania Craton)

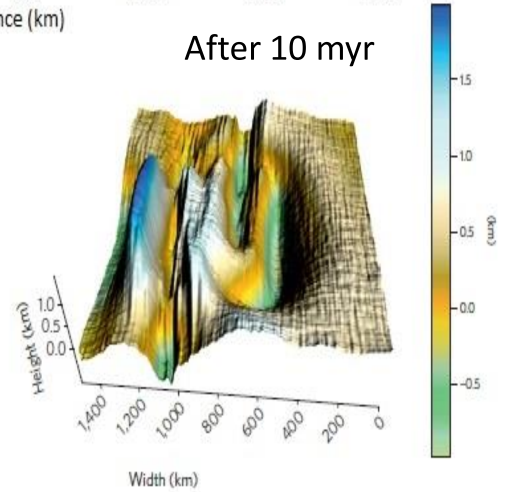








After 10 myr



- **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.

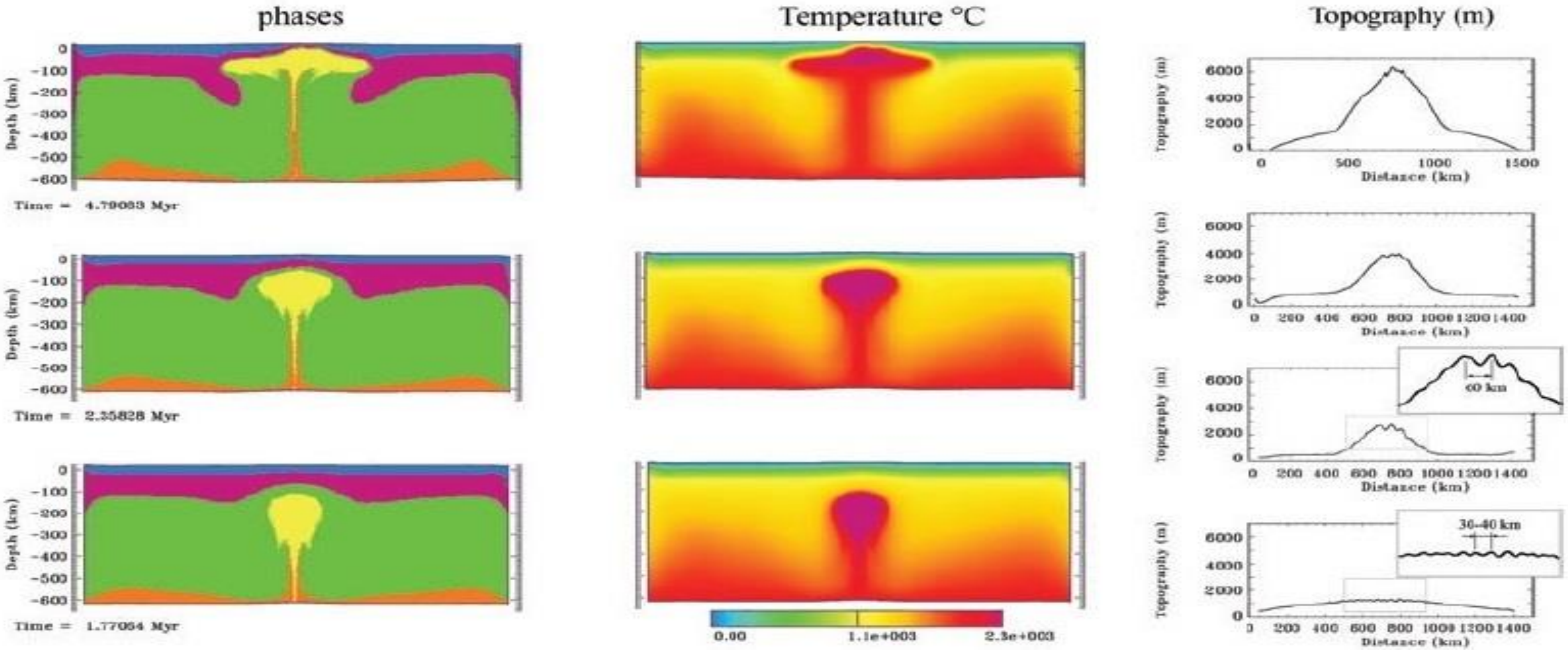
Koptev et al., 2014, Nature Geoscience

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.

Plume and topography

Burov and Cloetingh, 2009, GJI, 178

Young lithosphere (60 Myr old), applicable for early Archean plates



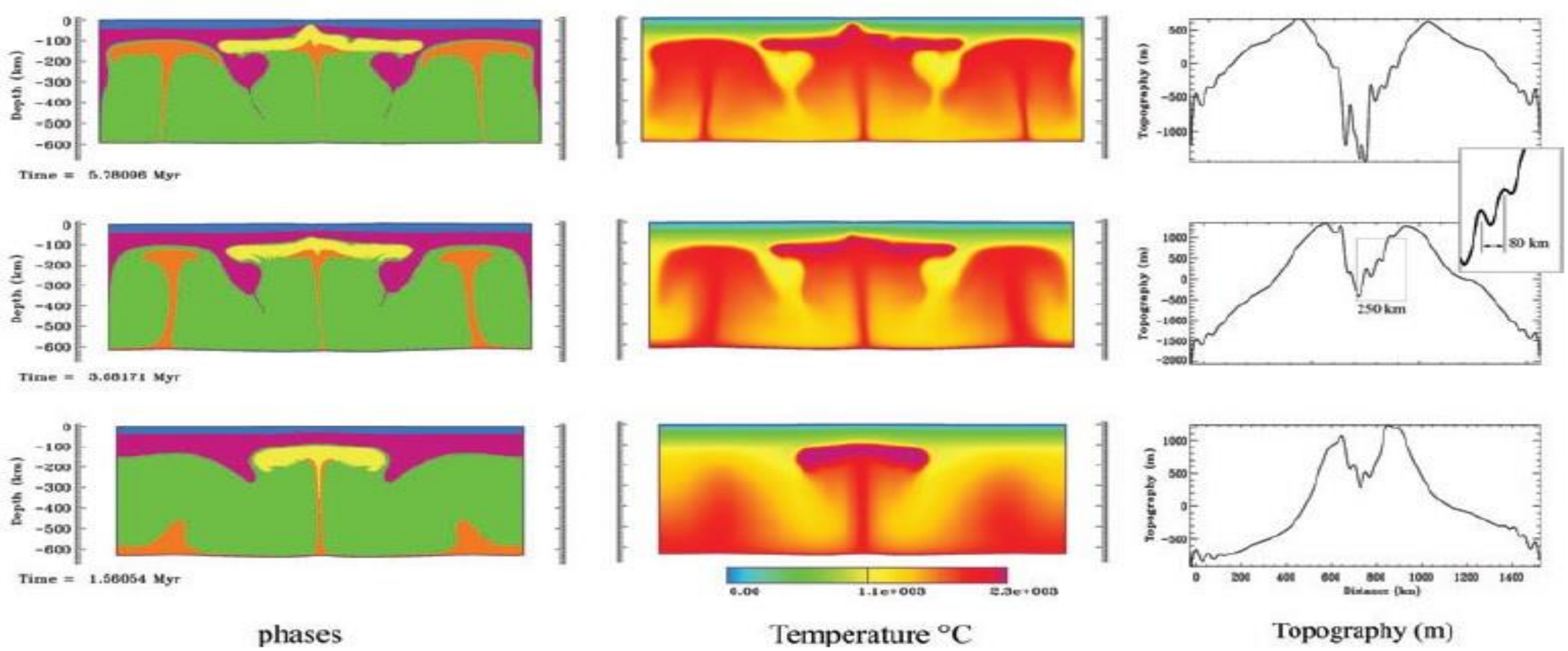
blue – upper and lower crust; purple – lithosphere mantle; green – sublithosphere mantle; yellow – plume; orange – marker layer at the bottom of the mantle.

- The ascending plume head first (<1 Myr) produces large scale (λ>1000 km) relatively small amplitude a (h <1 km) uplift superimposed by short-wavelength tensional and locally compressional crustal instabilities (λ<50 km, h <100 m). This deformation is soon followed (after 1 Myr) by a higher amplitude tectonic scale uplift (λ=250–300 km, h ~2 km), superimposed by amplified short-wavelength crustal deformation (λ <50 km, h ~ 300 m).
- To preserve high topography, the effective viscosity μ of the sub-brittle lithospheric layers should be high enough (e.g. >10^{23} Pa s), differently the topography would flatten by gravity driven flow in less than 1–2 Myr.

Plume and topography

Burov and Cloetingh, 2009, GJI, 178

Old lithosphere (150 Myr)



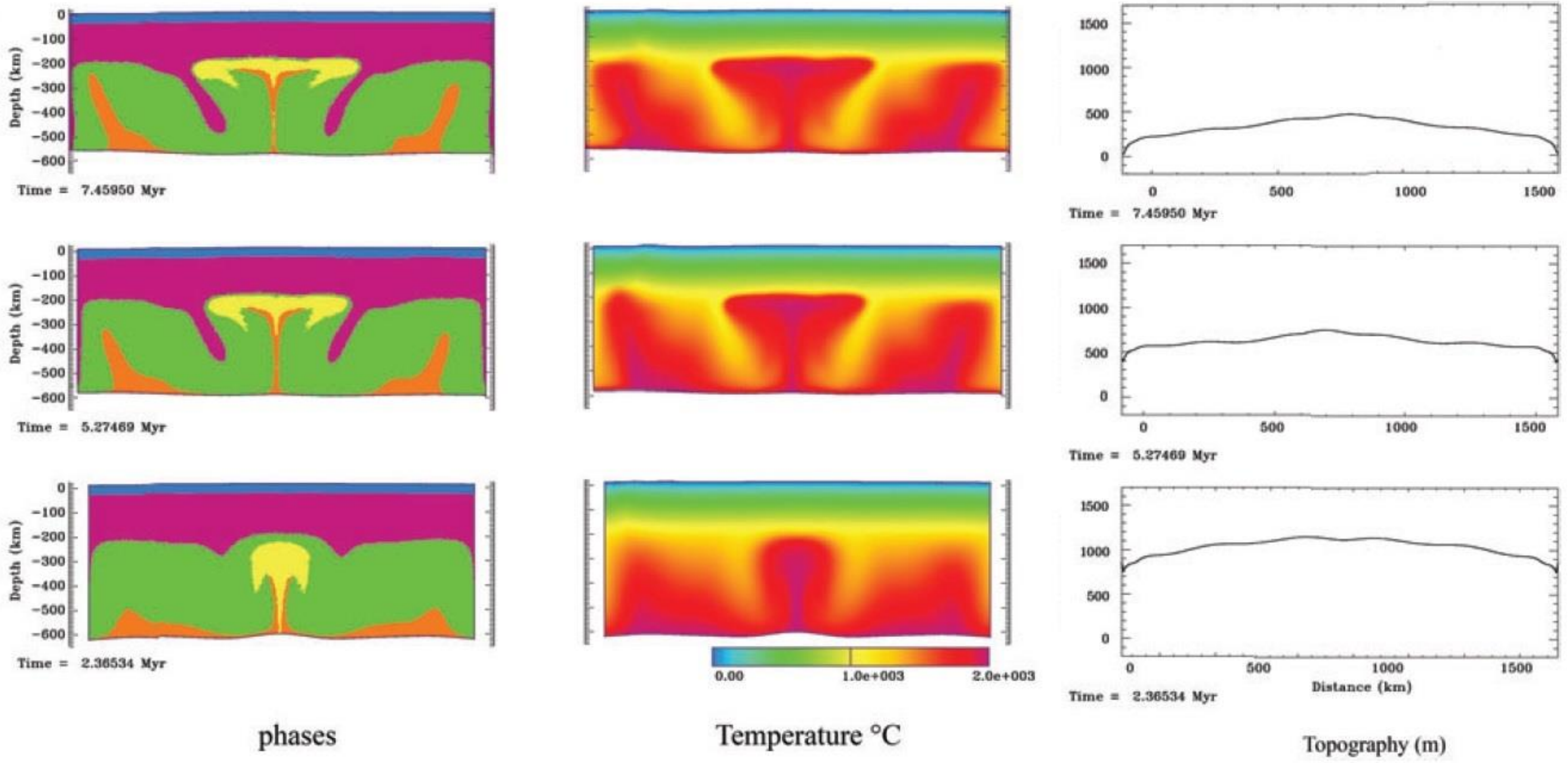
- The plume head intrudes at a large scale in the lithosphere and produces rifting with flexural-scale rift shoulders.
- The topography has a large-scale uplift of 600 km wide, with amplitude of up to 3–4 km, with inside a 2 km deep rift basin of 250–300 km width.
- The plume continued spreading at 5.6–6 Myr, moving further toward the surface and producing a subduction-like down thrusting of the mantle lithosphere to 400 km depth and a large wavelength uplift in an area of >1000 km wide, with a vertical amplitude of 3 km, overprinted by a 300 km wide rift-type basin with a depth of about 2 km.

Plume and topography

Final surface wavelengths are controlled by the mechanical properties of the lithosphere

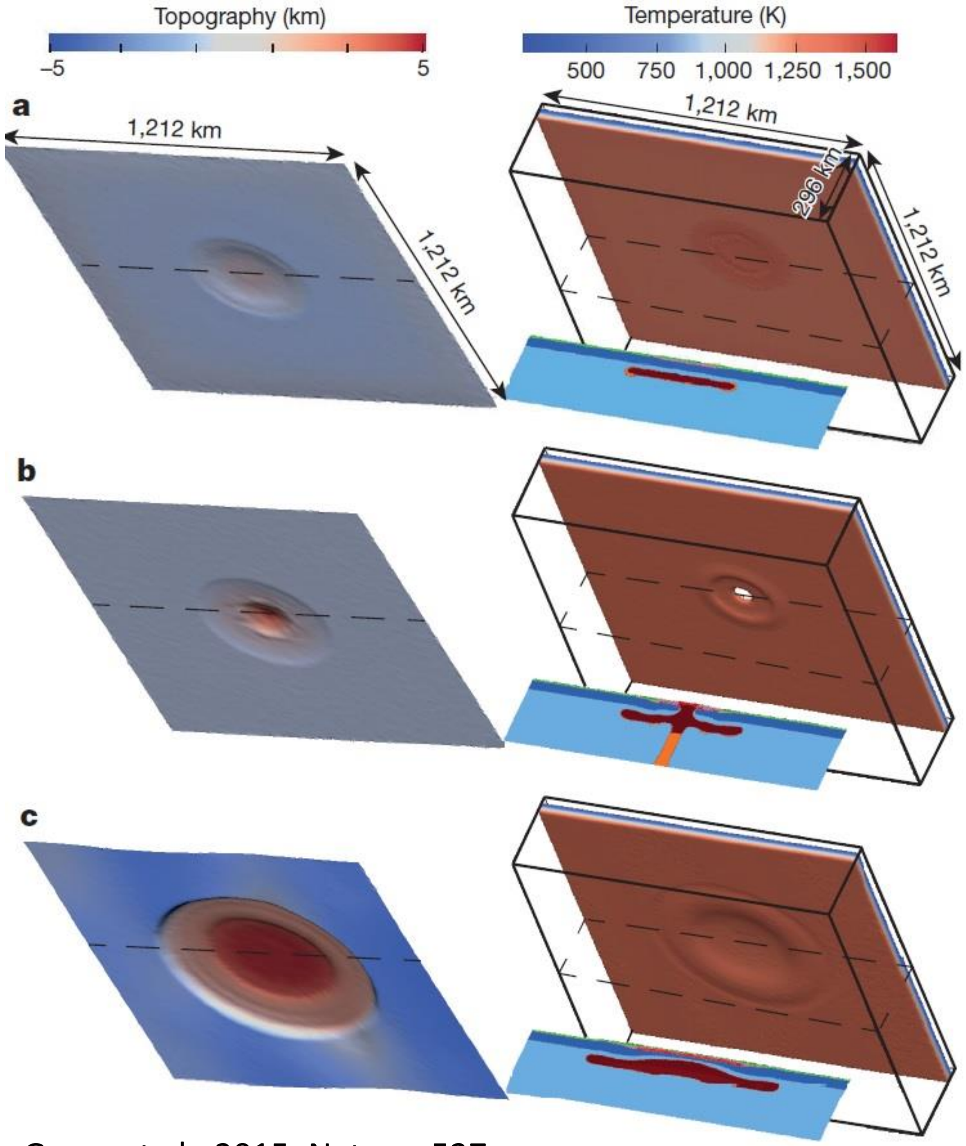
Burov and Cloetingh, 2009, GJI, 178

Very old lithosphere: 1000 Myr



- As the age of the lithosphere increases, the mantle-downthrusting produced by the plume head remains very significant but the amplitude of the surface expression is smaller (0.5 km) while its wavelength is larger (>600 km).
- Periodic surface undulations with a wavelength of 300 km (produced by very strong crust) are also observed.

Can a mantle plume induce subduction?



Gerya et al., 2015, Nature, 527

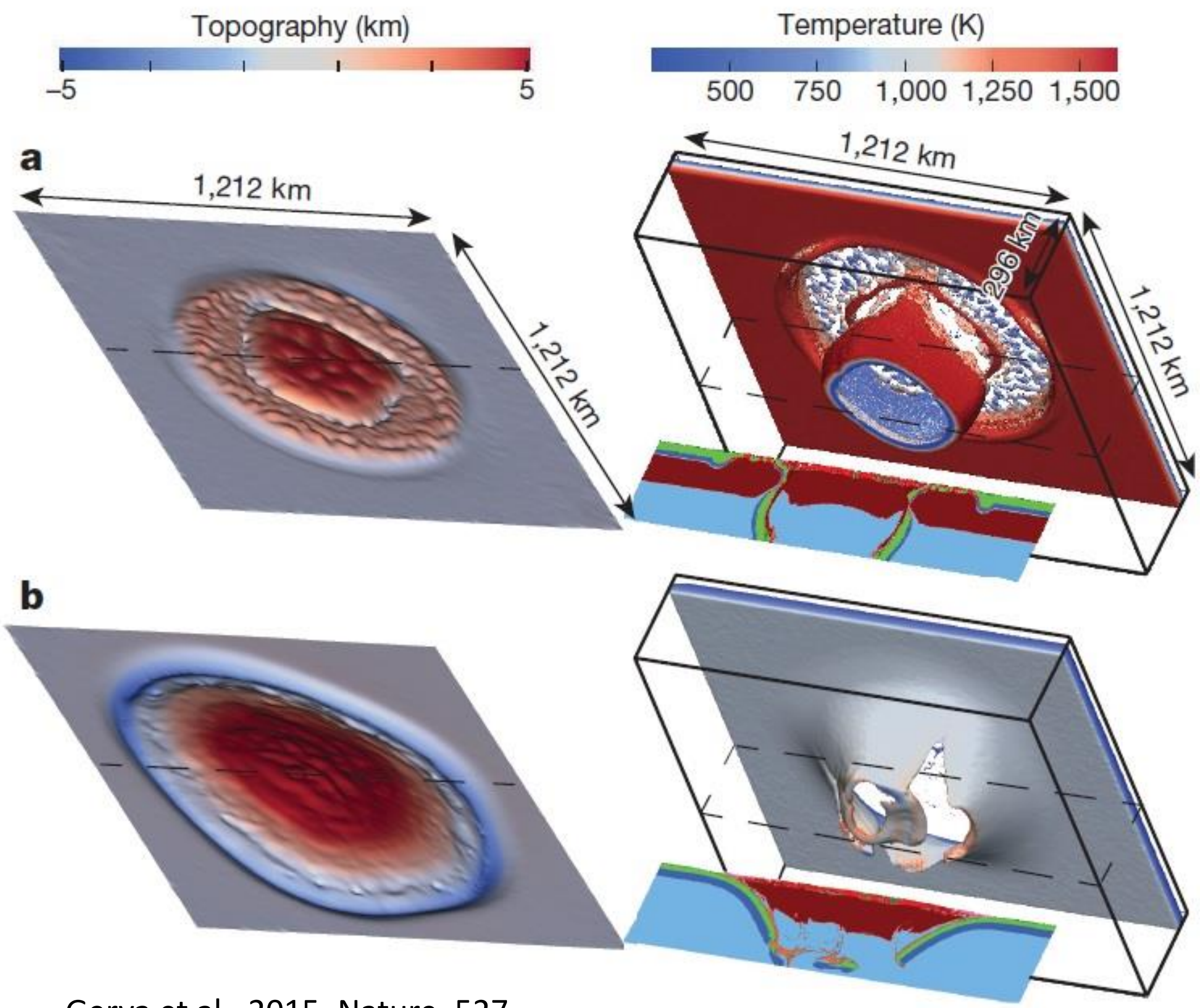
Subduction initiation induced by a mantle plume is favored by:

(1) Old age (>40 Myr) and hence large negative buoyancy of the precursor oceanic lithosphere.
(2) Longevity of plume–lithosphere interaction (duration of at least 56 Myr).
(3) Very high plume temperature (up to 1,620 °C).

- Plume interaction with young lithosphere results in oceanic plateau formation followed by nearly circular sheet-like lithospheric ‘drips’ that are driven by thickened lower-crust eclogitization and are terminated by shallow break-off of the descending oceanic lithosphere.
- Eclogite dripping is short-lived and does not create coherent retreating slabs. Repetitive drips quickly remove pre-existing mantle lithosphere, so that thickened, hot mafic crust comes to overlie convecting asthenosphere.

a Failed subduction initiation caused by insufficient magmatism induced weakening of the plate. **b** Influence of a long-lived plume with a conduit providing additional heat and mass supply through time. **c** Influence of increased plume size and temperature.

Can a mantle plume induce subduction?



Gerya et al., 2015, Nature, 527

- Plume interaction with old lithosphere results in self sustained subduction, which is further assisted by strong densification of thick oceanic crust, owing to its eclogitization.
- Compositional buoyancy of the depleted Archaean oceanic lithosphere with thick crust does not preclude subduction, but requires greater cooling ages (>60–70 Myr) for plate subductability compared to present-day mantle temperature conditions (> 30 Myr).

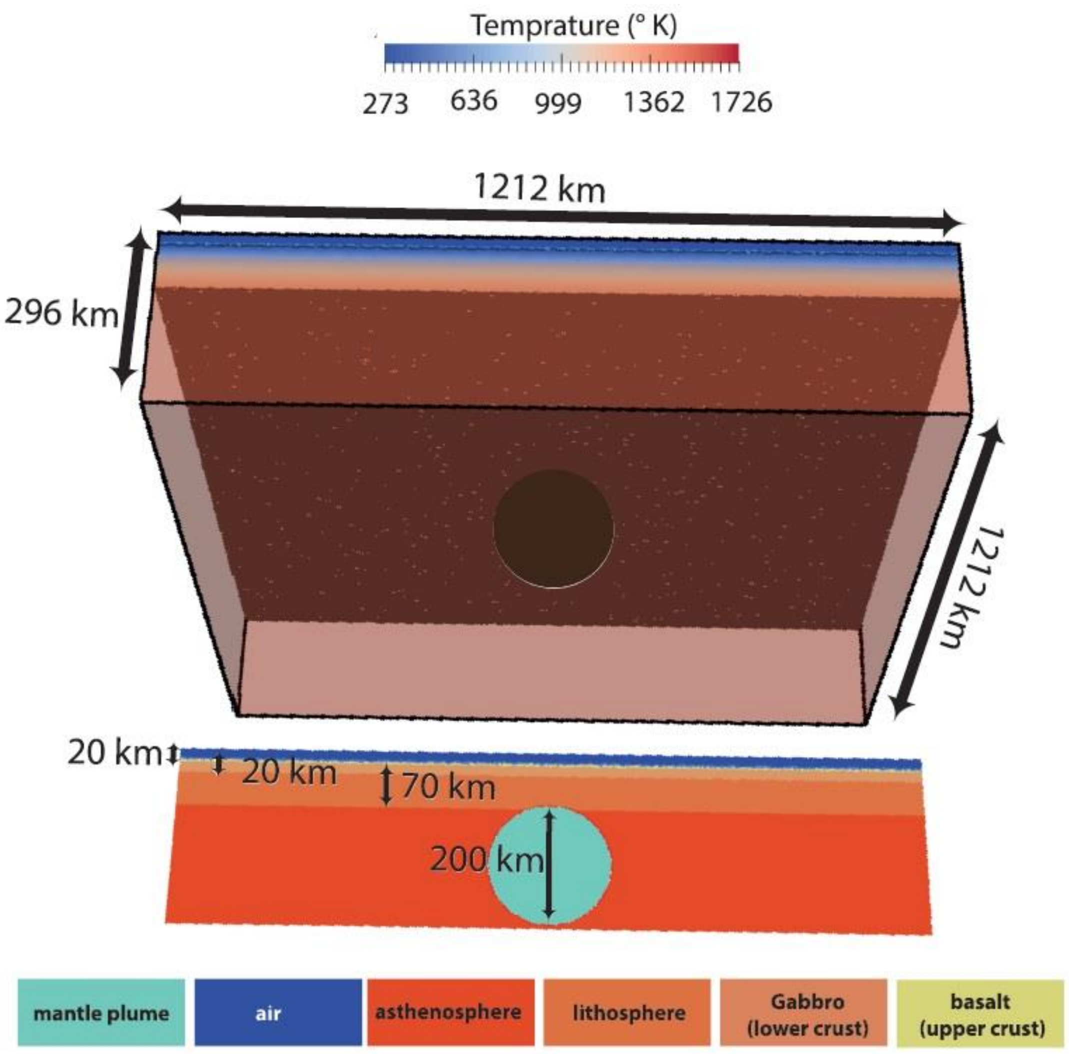
a Development of plume-induced lithospheric drips for 20 Myr old oceanic plate with 30 km thick crust. **b**, Development of plume-induced self-sustaining subduction for 80 Myr old oceanic plate with 20 km thick crust.

Response of the lithosphere to plume-lithosphere interaction

Parameters to investigate for different oceanic lithospheric age:

(1) effect of thickness of the crust, (2) mantle temperature, (3) extension rate, and plume tail.

Initial Model Setup



Baes et al., 2020, G3, 21

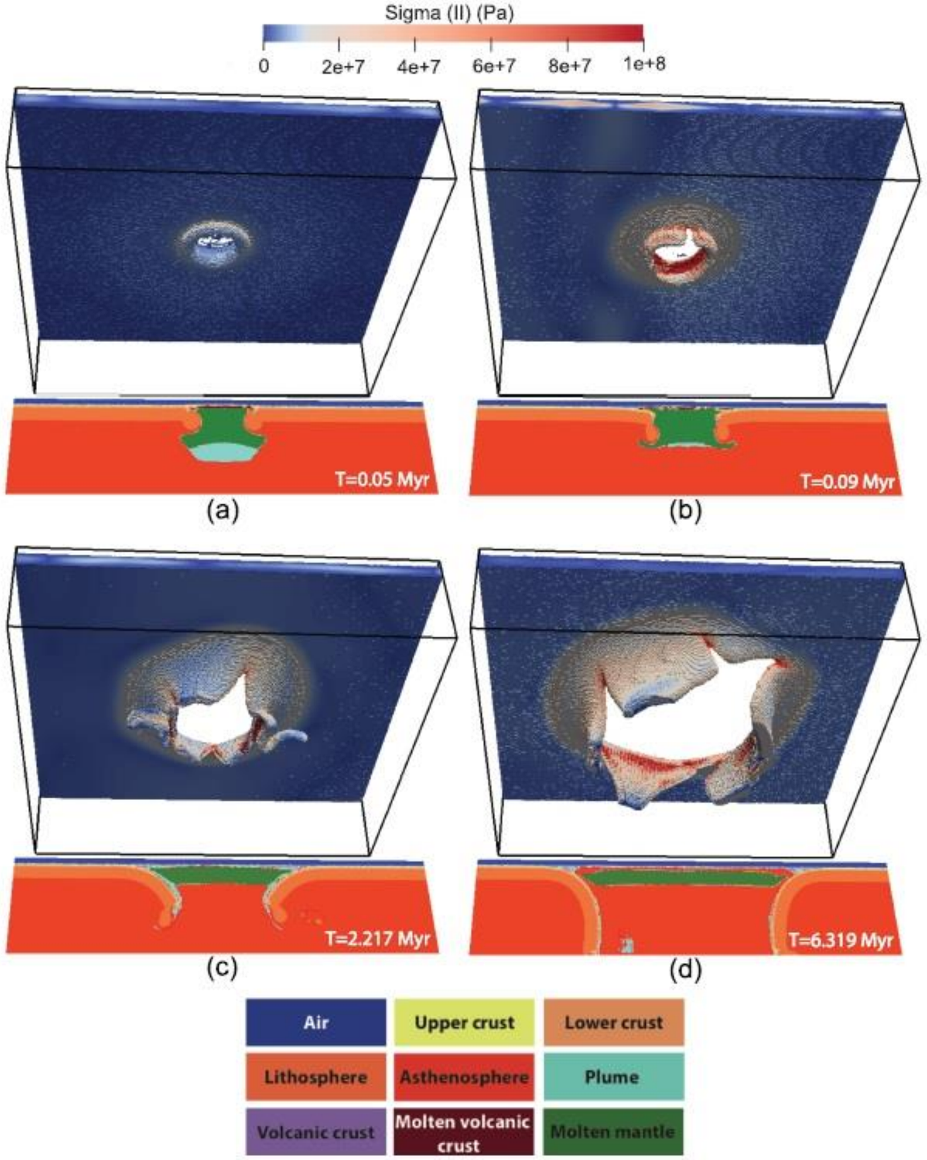
Four distinctly different lithospheric deformation patterns:

(a) multi-slab subduction initiation, (b) single-slab subduction initiation, (c) plateau formation without subduction initiation, and (d) episodic short-lived circular subduction initiation.

Crustal Thickness Effect

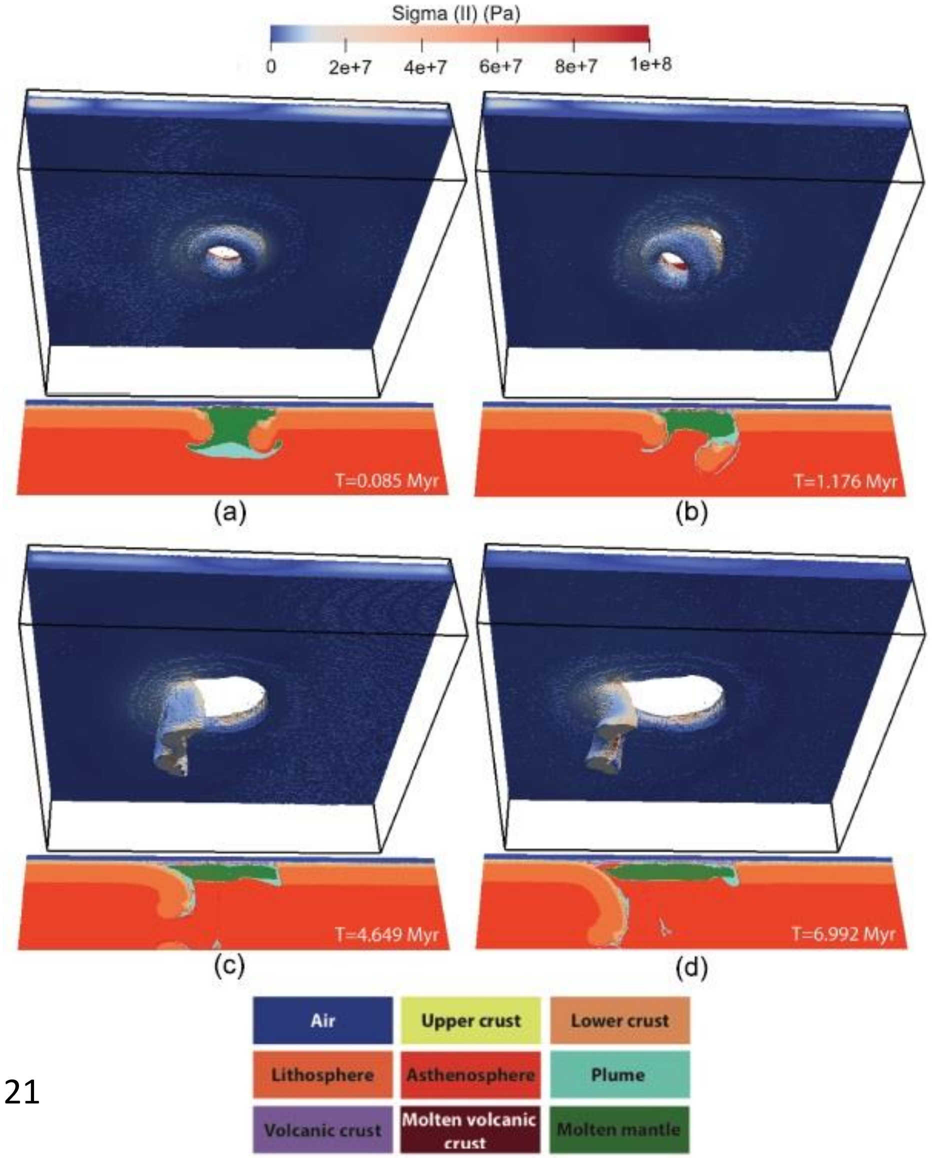
Multi-slab subduction initiation

Thin crust (8 km) and young oceanic lithosphere (age of 20 Myr)



Single-slab subduction

Thicker crust (12 km) and older oceanic lithosphere (age of 40 Myr)

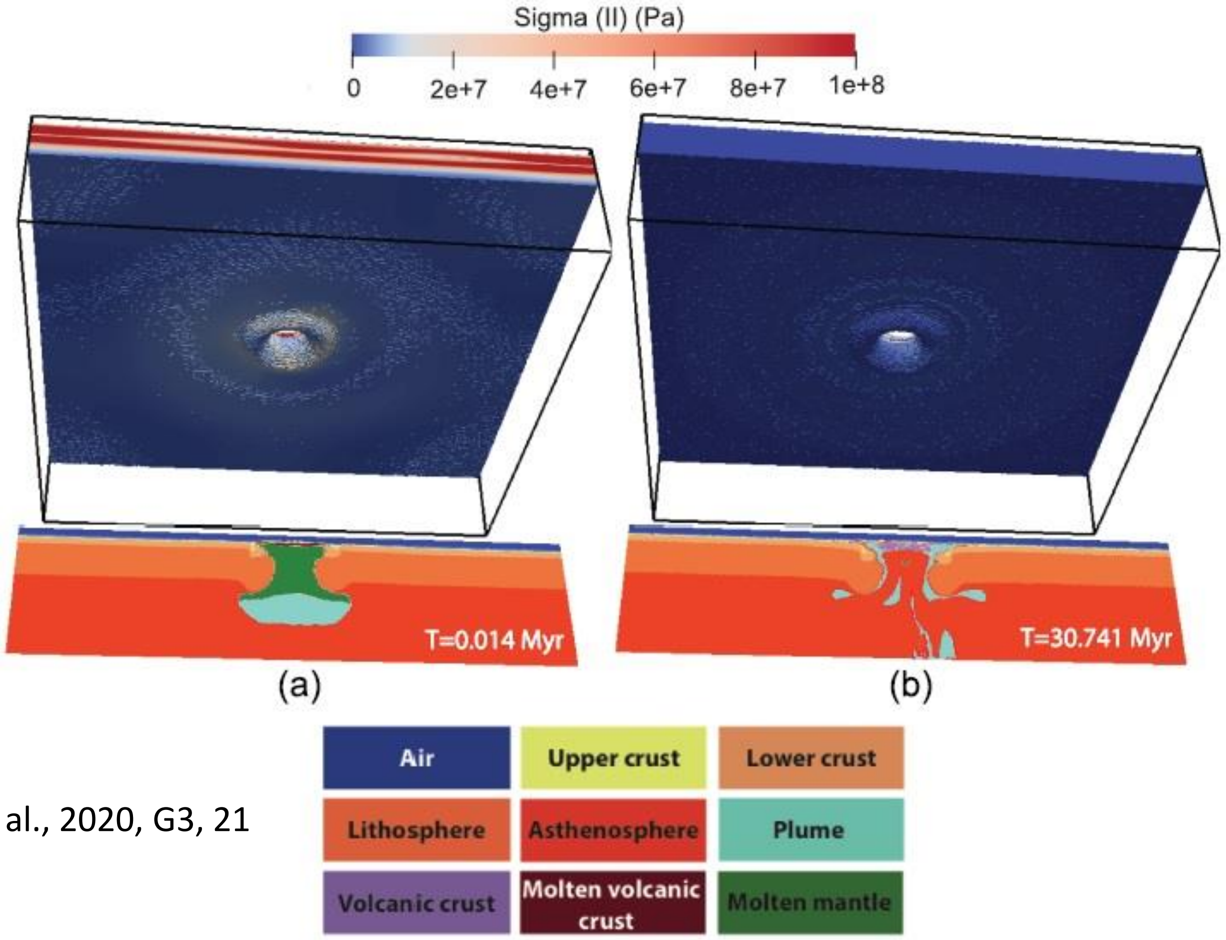


Baes et al., 2020, G3, 21

Crustal Thickness Effect

Plateau without subduction initiation

Thicker crust (> 12 km) and older oceanic lithosphere (age > 50 Myr)



Baes et al., 2020, G3, 21

- With the thickening of the crust and age of the lithosphere, the resistance of the lithosphere to descending increases, and thus the deformation regime, following plume-lithosphere interaction, changes progressively from multi-slab subduction initiation to plateau formation without subduction initiation.

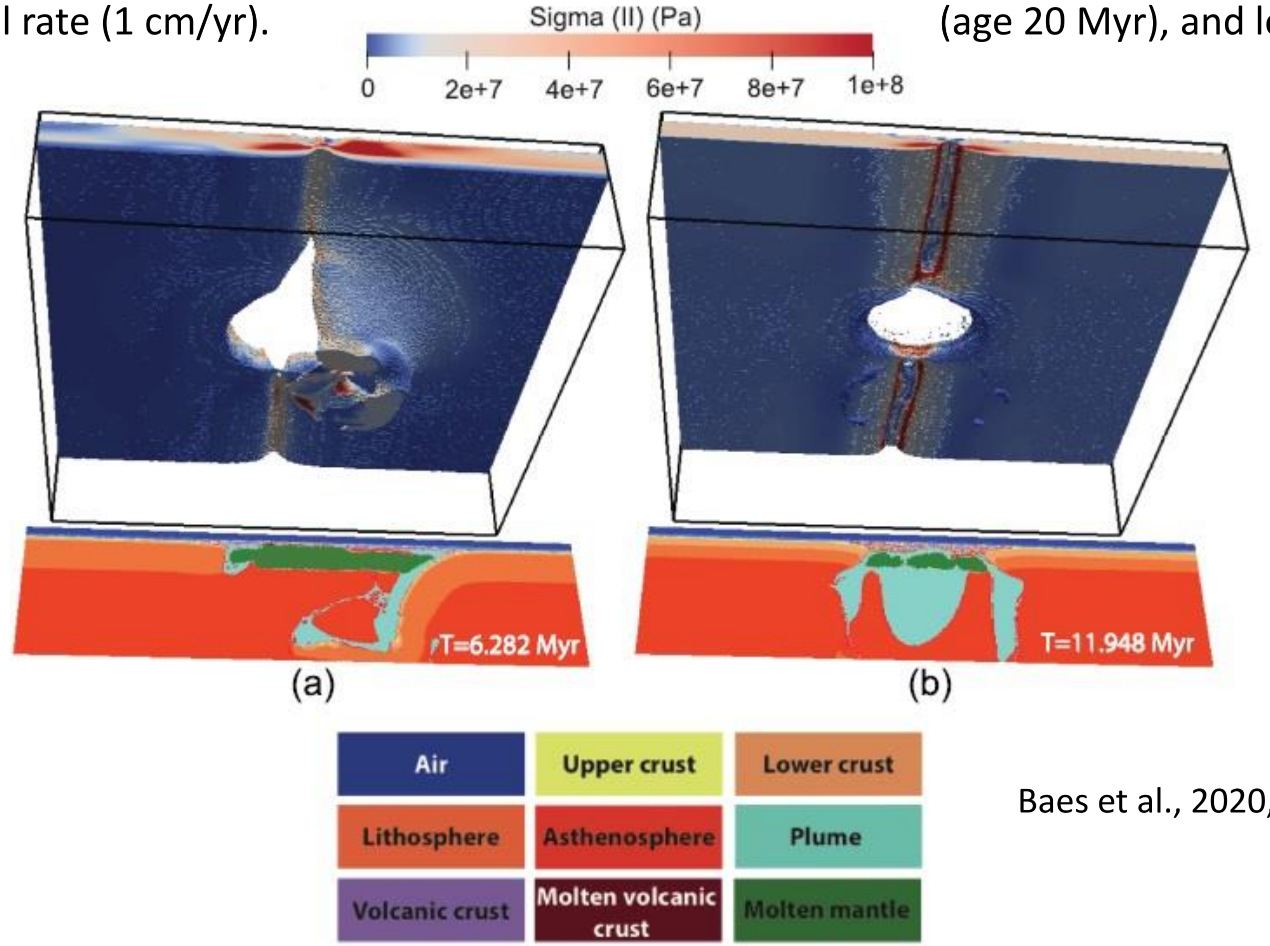
Extensional Regime Effect

Single-slab subduction

Thin crust (8 km), old oceanic lithosphere (age of 50 Myr), and high extensional rate (1 cm/yr).

Plateau without subduction initiation

Thicker crust (20 km), young oceanic lithosphere (age 20 Myr), and low extensional rate (0.5 cm/yr).



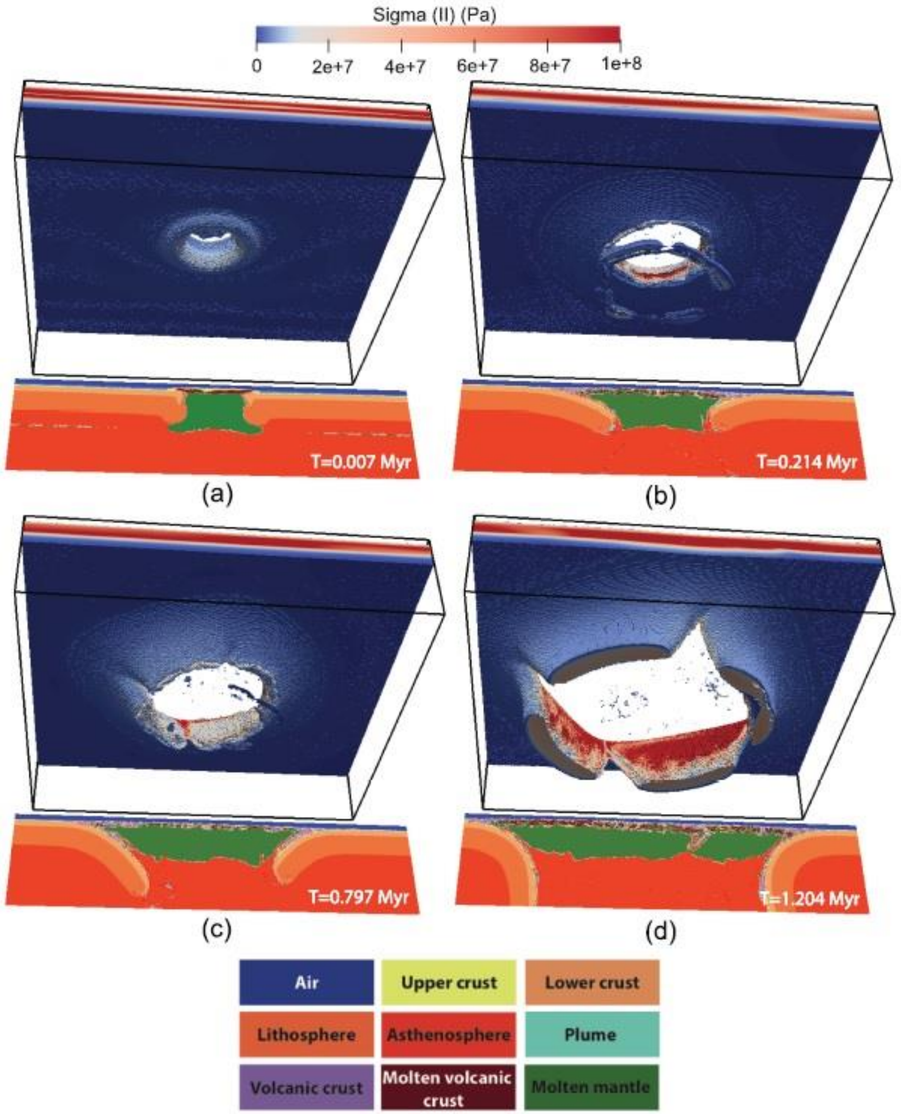
Baes et al., 2020, G3, 21

- In a low-rate extensional regime (0.5 cm/yr) single-slab subduction of a lithosphere with a typical crustal thickness of 8 km occurs only if the age of lithosphere is less than 50 Myr.
- Single-slab subduction of lithospheres with thick plateau of 20 km is possible when the age of lithosphere is between 30 and 40 Myr.
- Higher extensional regimes facilitate subduction of older lithospheres.

Mantle Temperature Effect

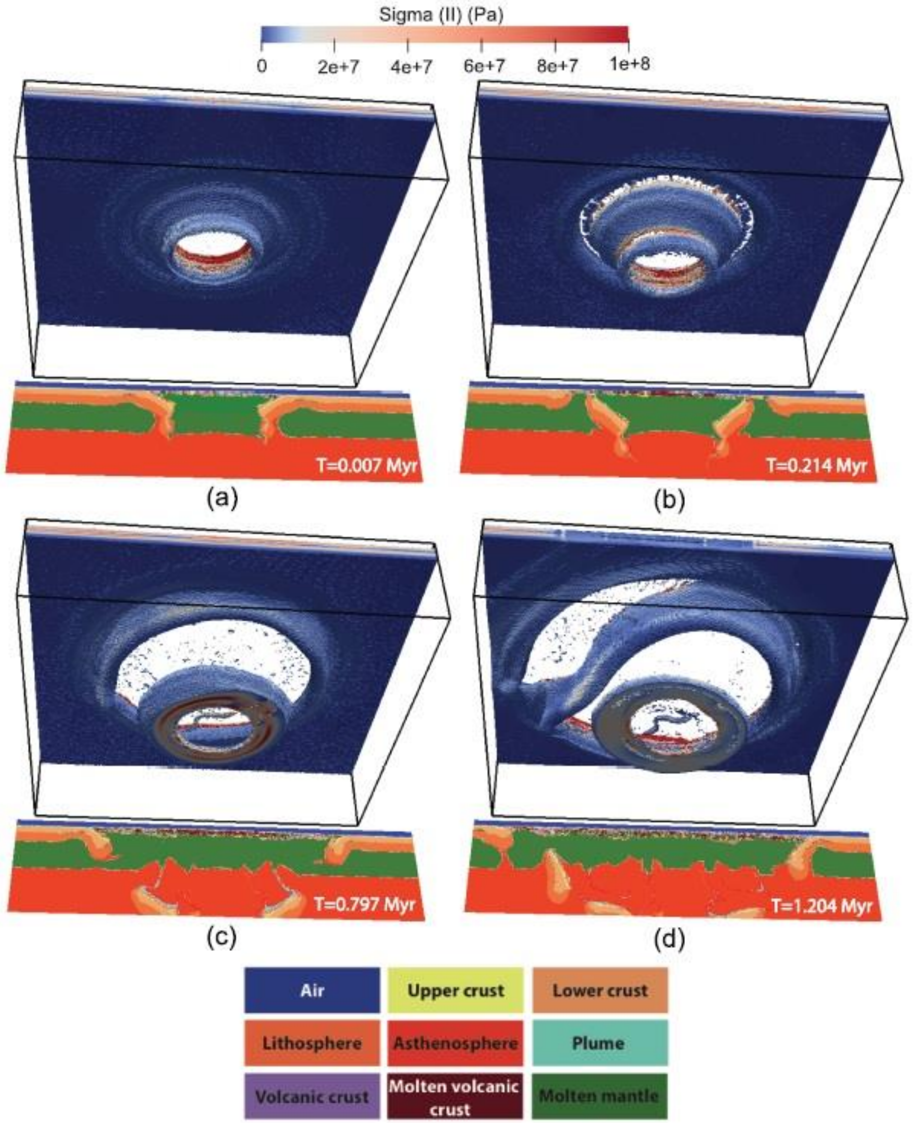
Multi-slab subduction initiation

Thick crust (20-30 km), old oceanic lithosphere (age > 50 Myr), > 200 K hotter mantle (Archean conditions)



Short-lived circular subdution zone

Thick crust (20-30 km), young oceanic lithosphere (age < 50 Myr), > 200 K hotter mantle (Archean conditions)

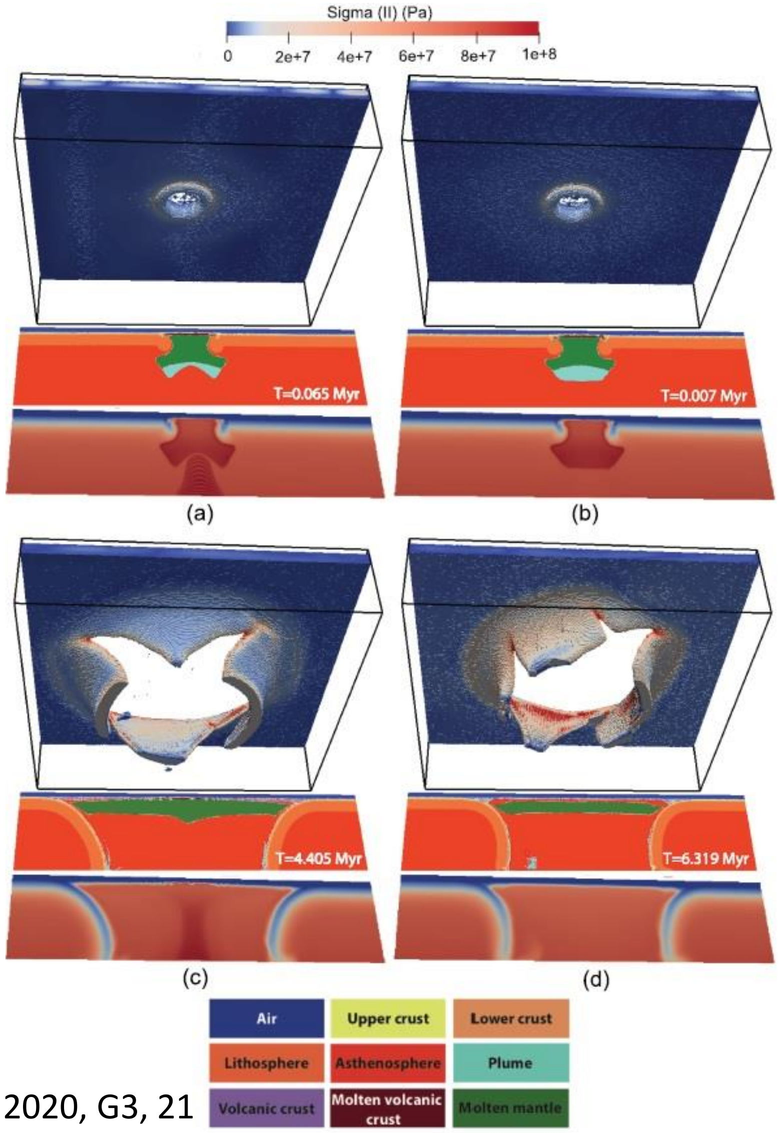


Baes et al., 2020, G3, 21

Uprising of Hot Mantle Material
(Plume with a Tail)

Multi-slab subduction initiation

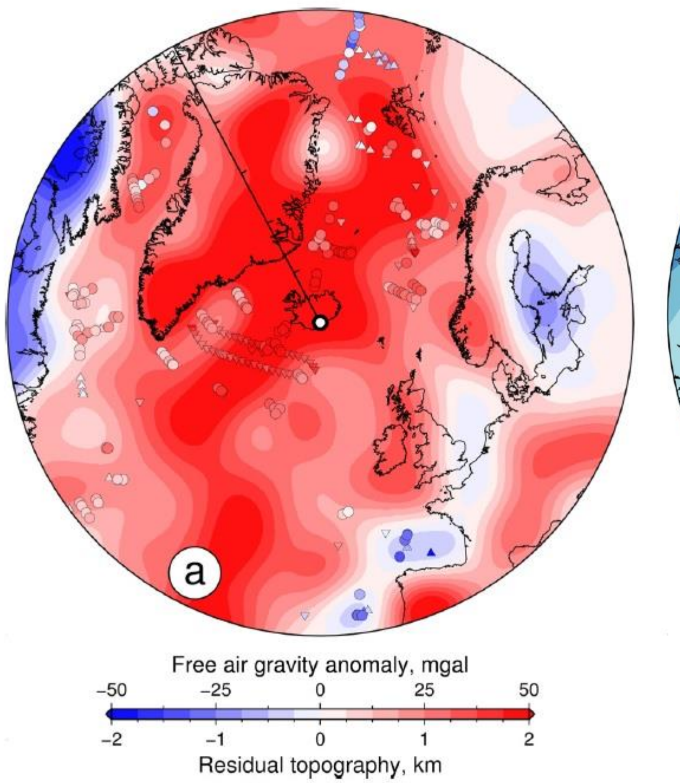
Thin crust (8 km), young oceanic lithosphere (age: 20 Myr), and 200 K higher temperature at the lower model boundary.

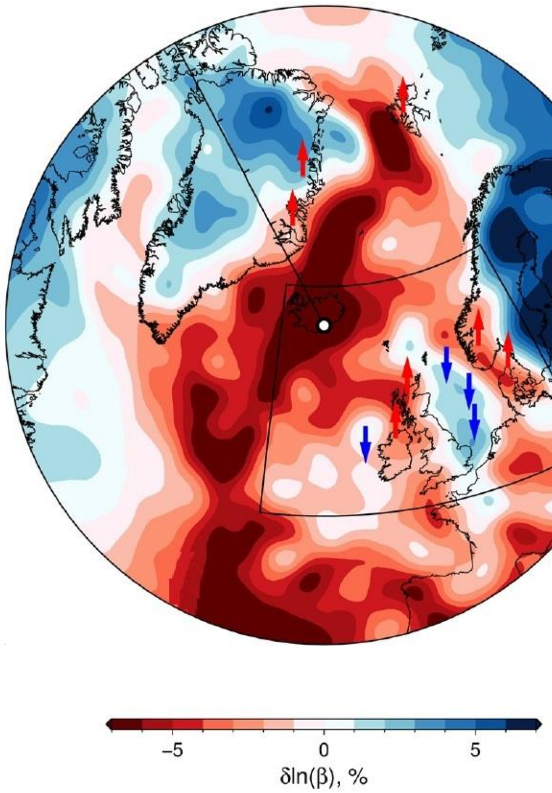


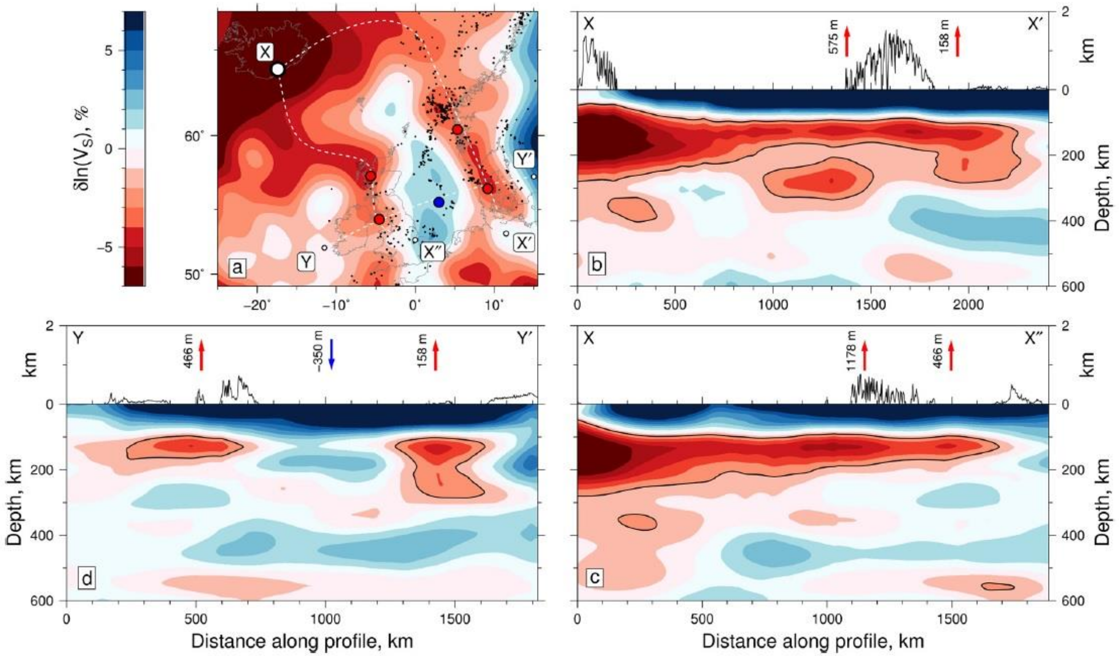
Baes et al., 2020, G3, 21

- Outcomes of experiments with different lithospheric age, extension rate, and thickness of the crust, show that continuous rising of hot mantel materials towards the surface facilitates subduction initiation, especially when a mantle plume interacts with a lithosphere containing a thick plateau.
- Breaking of the lithosphere depends on several parameters such as plume volume, plume buoyancy and lithospheric thickness.
- **Subductability of the lithosphere is related to the strength and buoyancy of the lithosphere:** lowering the strength of the lithosphere can facilitate subduction initiation. Negative buoyancy of the lithosphere increases with decreasing of the crustal thickness and increasing of the lithospheric age.
- Plume-lithosphere interaction may lead to formation of plateau without subduction initiation if the lithosphere is older than 50.
- This indicates that although the aging of the lithosphere increases the negative buoyancy of the plate, its strength growth acts as a resistive force in subduction initiation process.
- The formation of single-slab or multi-slab subduction depends on the strength of the lithosphere.
- During Archean times, multi-slab subduction initiation could occur if the lithosphere is older than 50 Myr, while at the present-day, Earth, it can occur only if a plume interacts with a relatively young oceanic lithosphere.

Radial viscous fingering generated by plumes



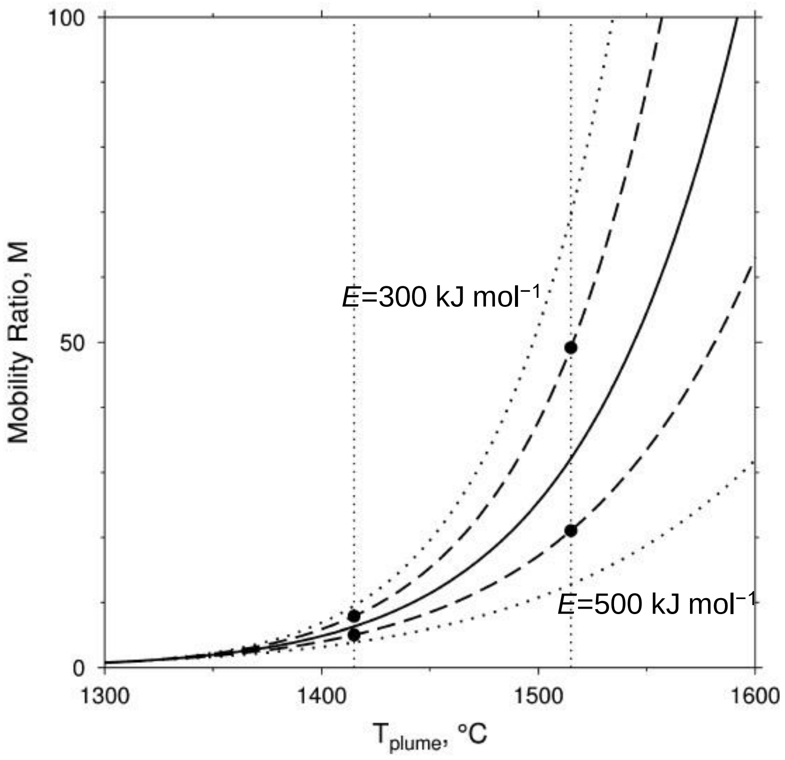




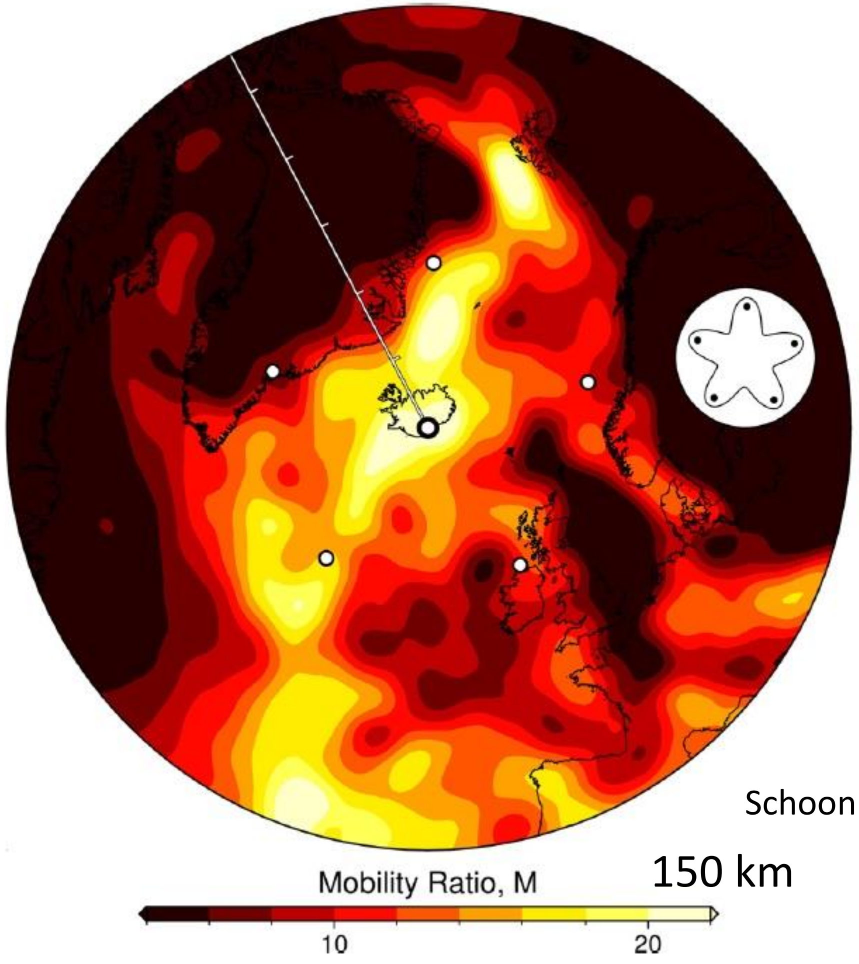
Schoonman et al., 2017, EPSL 468

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



Vertical dotted lines = range of plume temperatures



Schoonman et al., 2017, EPSL 468

150 km

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

η_r= mantle viscosity

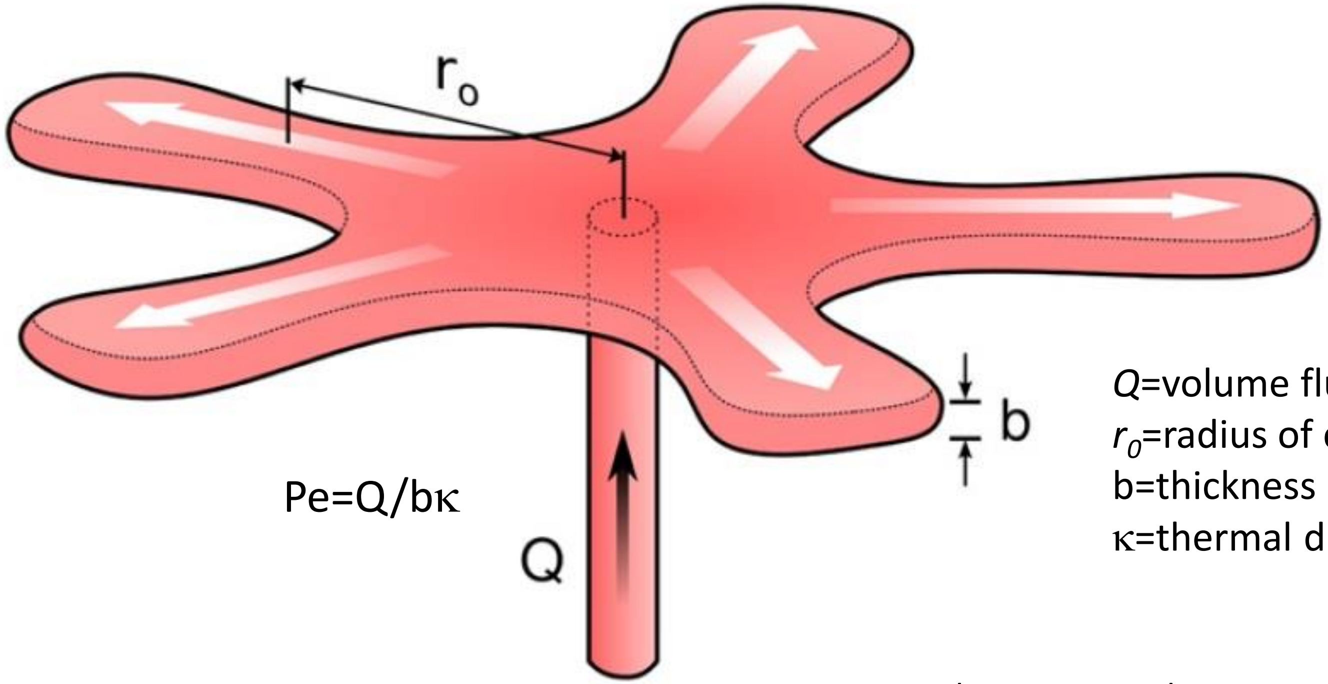
η= plume viscosity

T_r=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 (m^3/s)
r_0=radius of central portion of plume
b=thickness of the layer (~100 km)
κ=thermal diffusivity (m^2/s)

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 convective and diffusive transport rates ~$R_a^{2/3}$), and by the thickness of the horizontal layer into which fluid is injected (b~100 km).
- 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.

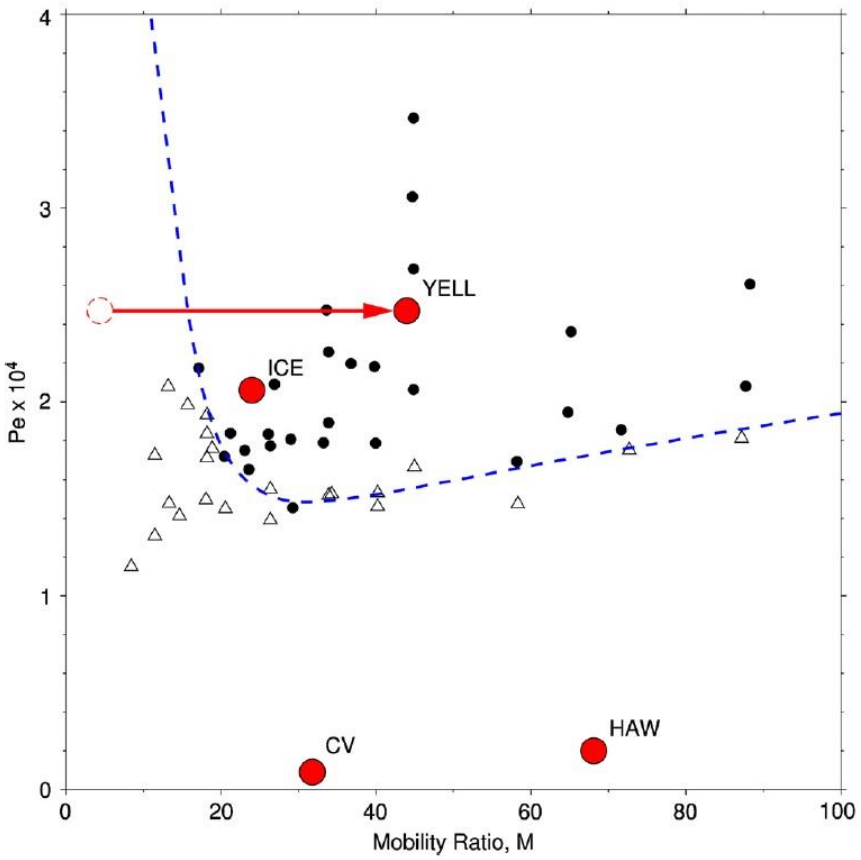
Radial viscous fingering generated by plumes
(experimental analysis)

B (buoyancy flux)=$g\Delta\rho\pi r^2 u$
B (buoyancy flux)=$\Delta\rho WEu_P$

WE=cross-sectional area of the swell
u=plume velocity
u_P=pate velocity

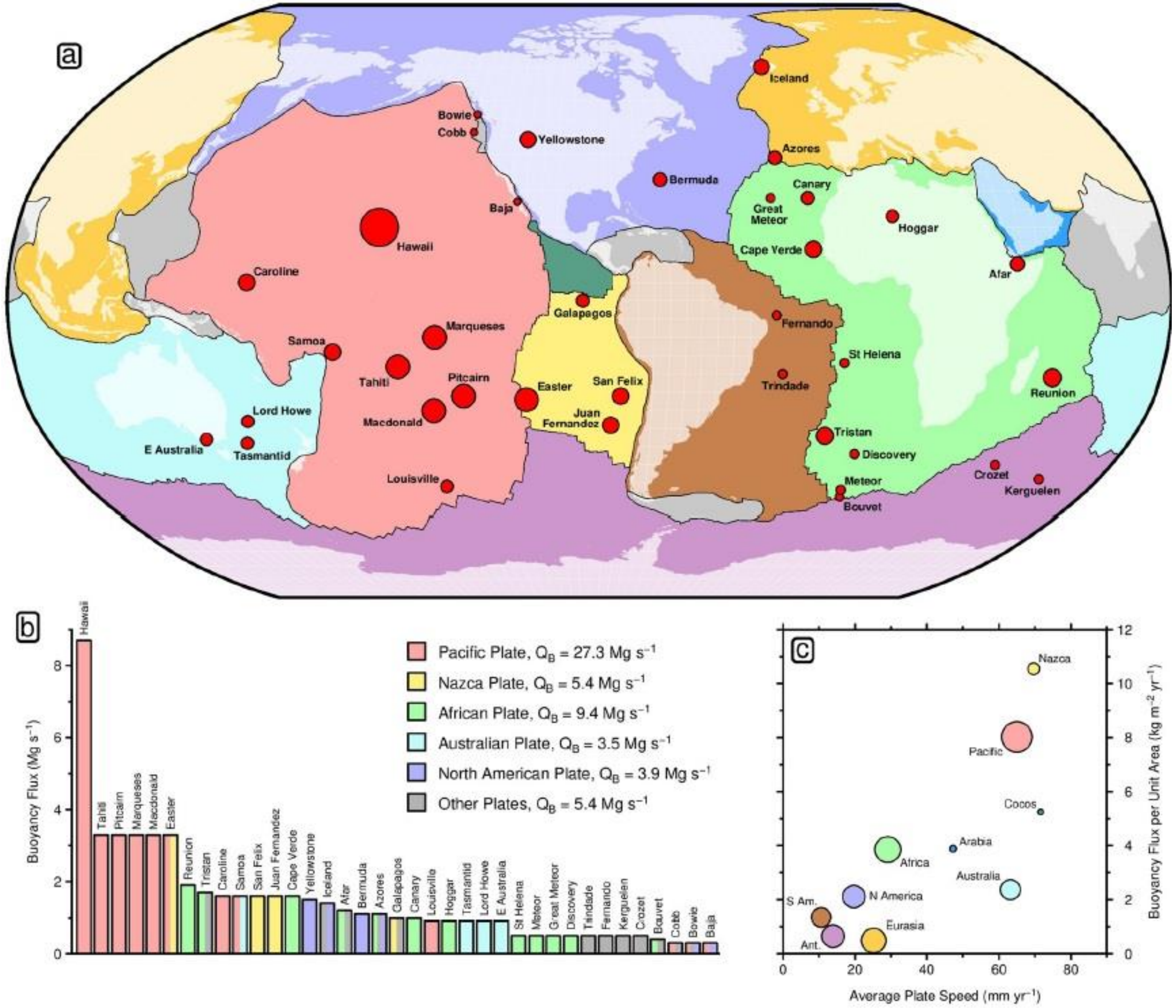
Plume	Radius (km)	ΔT (°C)	B (Mg s^{-1})	b (km)	M	$Pe \times 10^4$
Iceland	1200	150	27	100	24	2.06
Hawaii	500	250	5	120	68	0.20
Cape Verde	390	190	2	130	31	0.09
Yellowstone	1000	80	17	100	4.4	2.44

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).

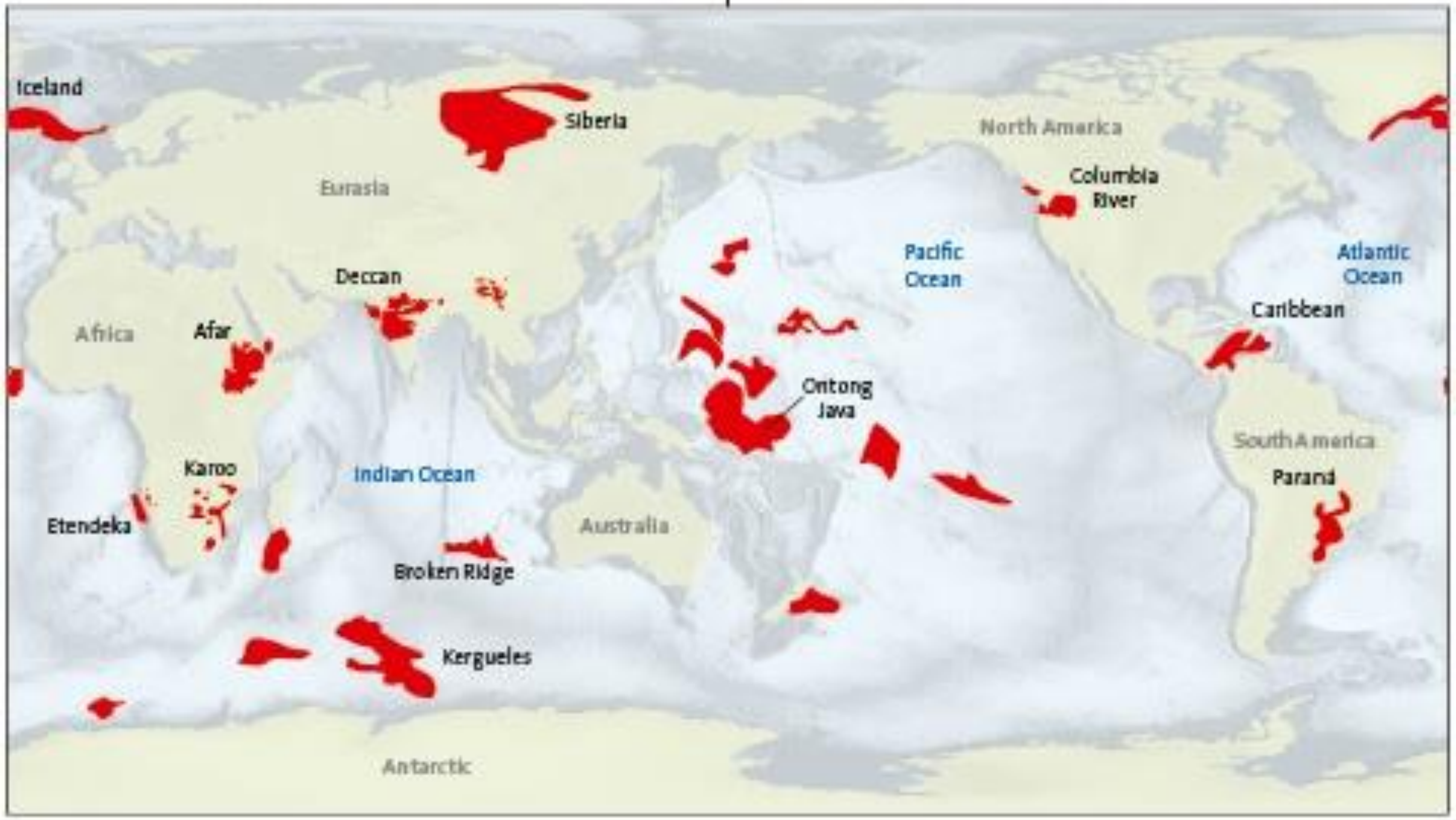
Buoyancy Flux and Plate Velocity



Hoggard et al., 2020, EPSL, 542

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

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

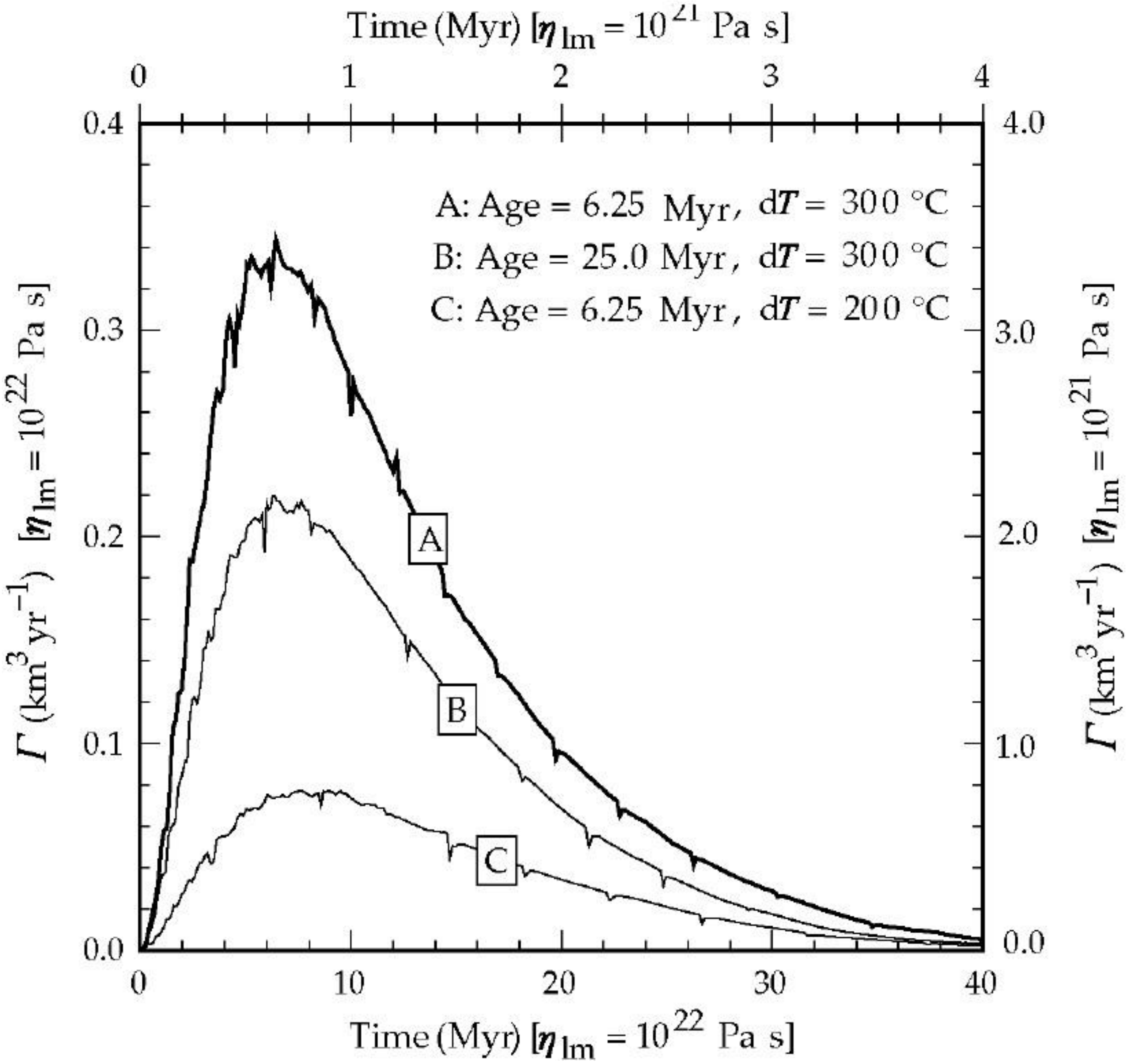


They extend up to 2000 km across, several km thick, 10 million km^3 of volcanic products

- 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

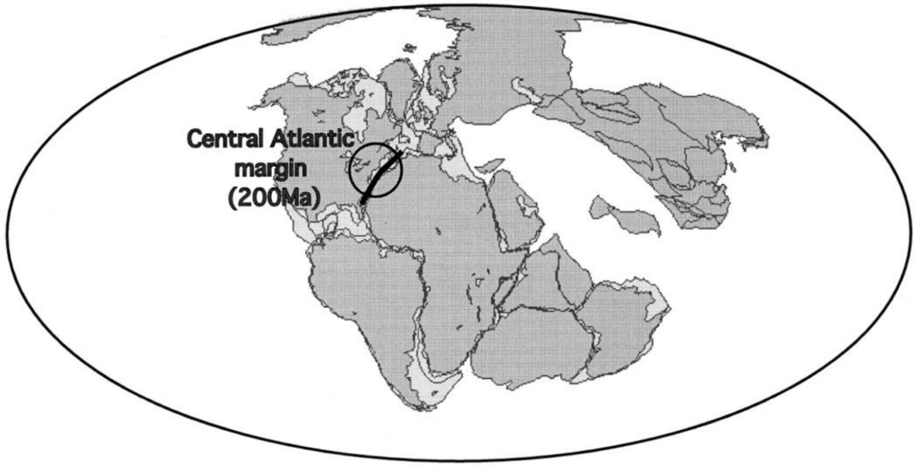
LIPs (Large igneous provinces)

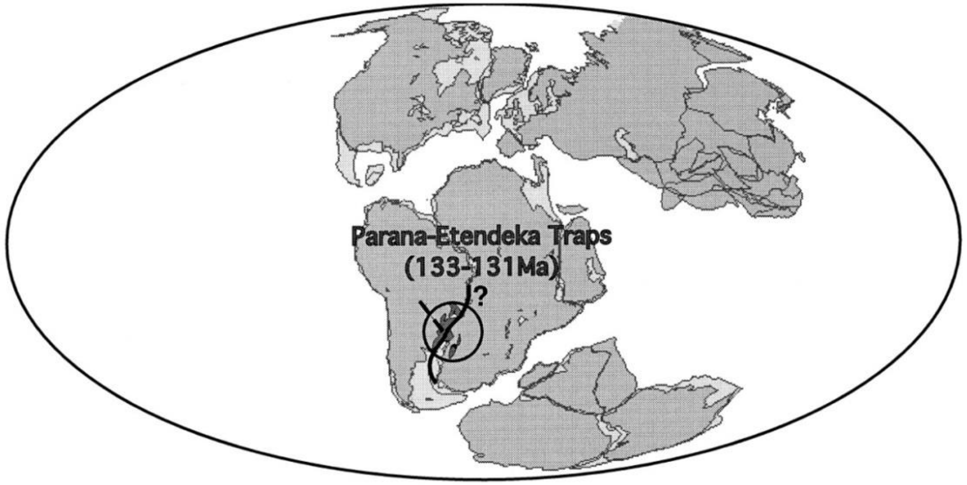
Rates of magma generation, Γ, for a plume head that includes 15% additional basaltic component

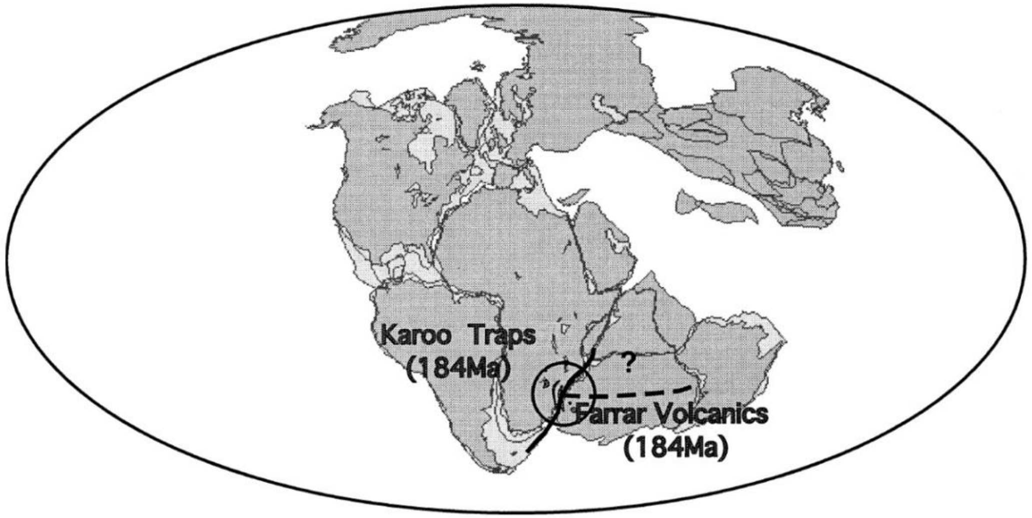


dT=plume temperature excesses
η=mantle viscosity

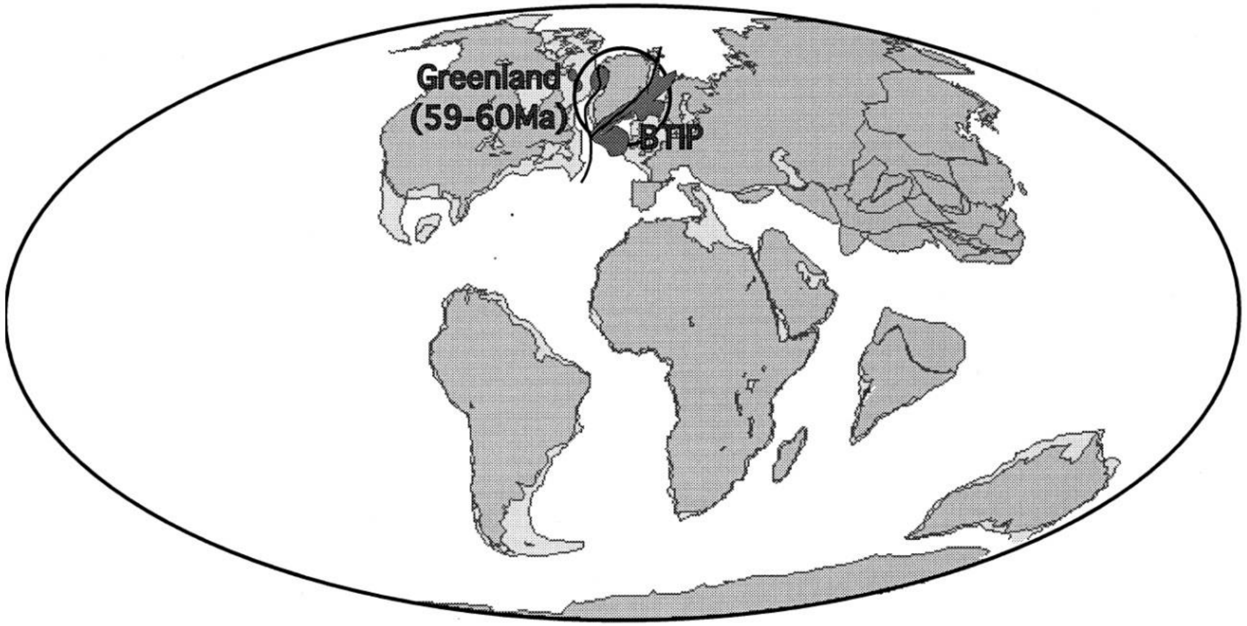
LIPs often predate continental break-up

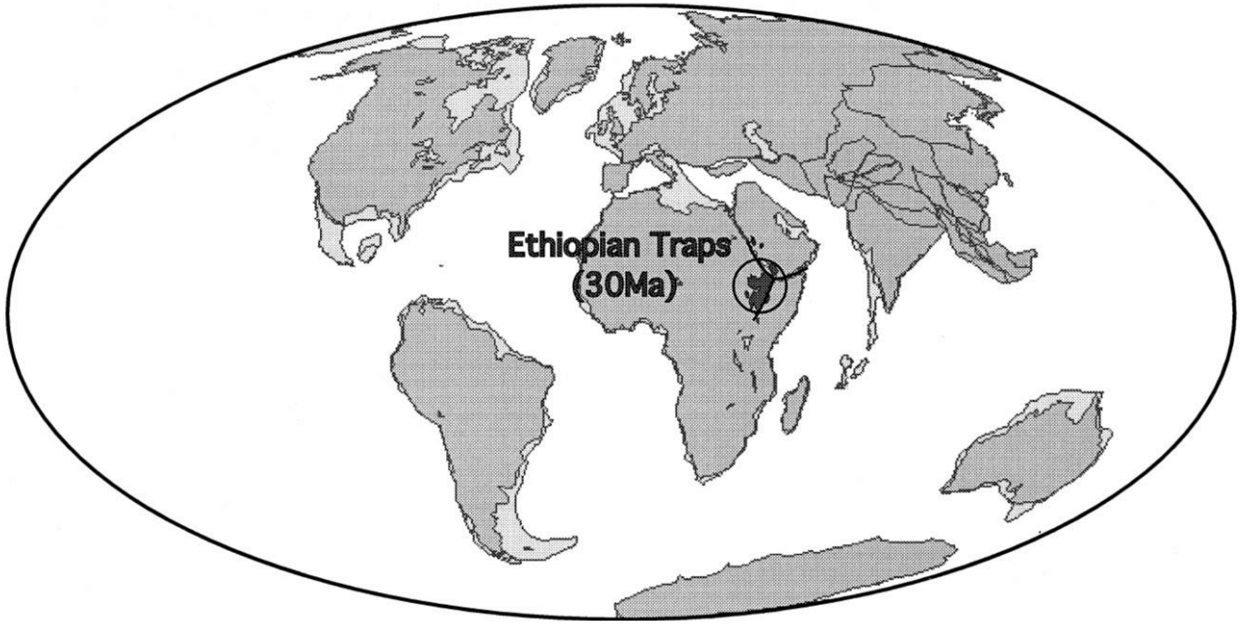




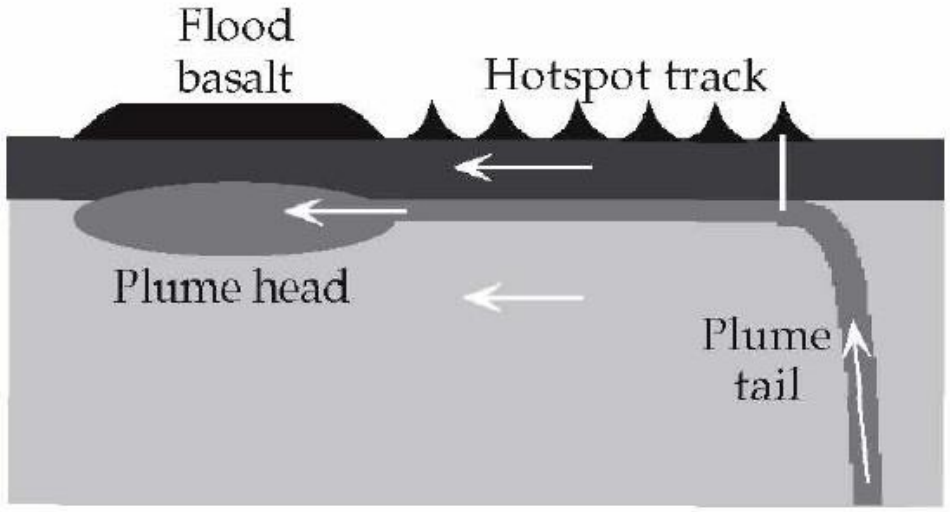


LIPs often predate continental break-up





Connection of LIPs to mantle plume



At least six examples have strong geographical, geochronological, and geochemical connections between hot-spot volcanism and flood basalt provinces: (1) Iceland and the North Atlantic Volcanic Province; (2) Kerguelen, and Bunbury, Naturaliste, Rajmahal (E. India), Broken Ridge, and Ninetyeast Ridge; (3) Reunion and Deccan, W. Indian, Chagos–Laccadive, Mascarenas, Mauritius; (4) Marion and Madagascar (Storey *et al*., 1997); (5) Tristan da Cunha and Paranà, Etendeka, Rio Grande, Walvis Ridge; (6) Galapagos and Caribbean.

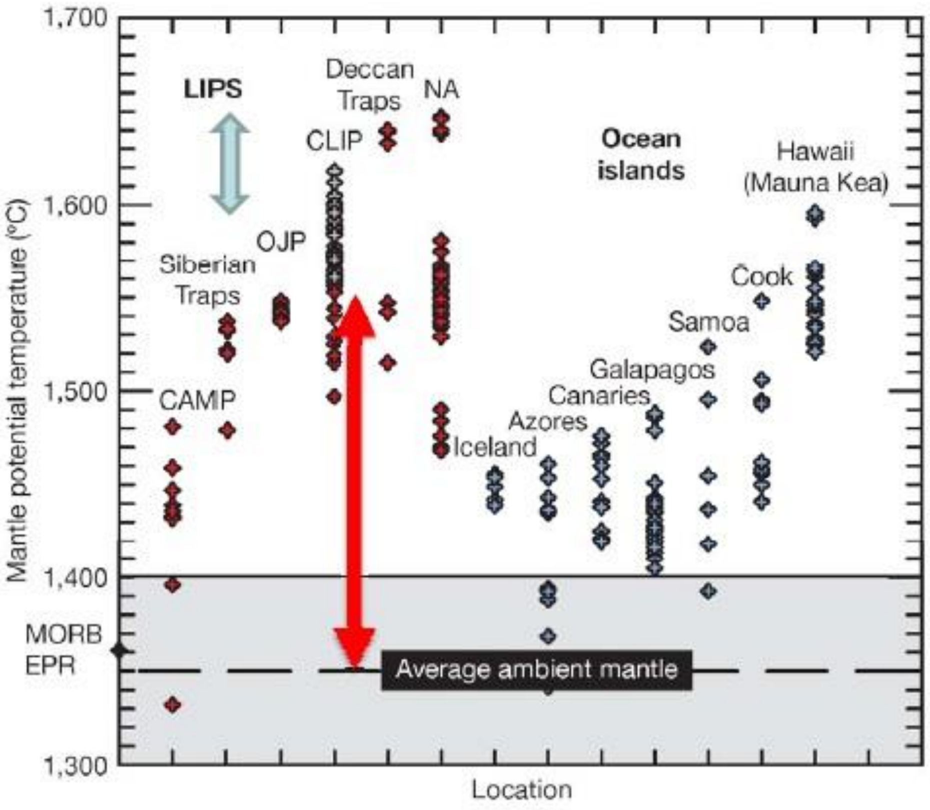
- **The observation of large plume heads followed by thin tails in fluid dynamical experiments has traditionally been used to explain the LIP-hotspot connection**: the hottest material of rising plume heads will erupt first, which explains high MgO basalts early in the LIP record. Furthermore, the arrival of the plume at the surface will lead to uplift and extension, often observed in the geological record of LIPs.
- The viscosity change can also completely break apart a starting mantle-plume head into multiple plumes, perhaps contributing to multiple flood basalt episodes.

Alternative hypothesis to the plume origin of the LIPs:

(1) Excess heat can build below continents during tectonic quiescence and/or supercontinent formation, causing the massive eruptions during continental breakup (it addresses the correlation between LIPs and continental breakup and the lack of connection of some continental LIPs to hot-spot trails).
(2) Delamination of continental lithosphere and secondary convection at rifted margins have been forwarded to generate LIPs near continents (it addresses the lack of connection of some continental LIPs to hot-spot trails).
(3) Compositional, rather than thermal effects cause excess melting, for example, more fertile mantle such as eclogite and/or water in the source (it addresses the lack of uplift of some LIPs, like OJP and the lack of connection of some continental LIPs to hot-spot trails).
(4) Meteorite impact could be responsible for the emplacement of LIPs, since the decompression of mantle following impact may generate extensive melting with less uplift than expected from a hot plume head and without a connection to a hot-spot track (but no evidence).
(5) Oscillatory instabilities in starting plumes can be caused by the competing effects of thermal and chemical buoyancy.

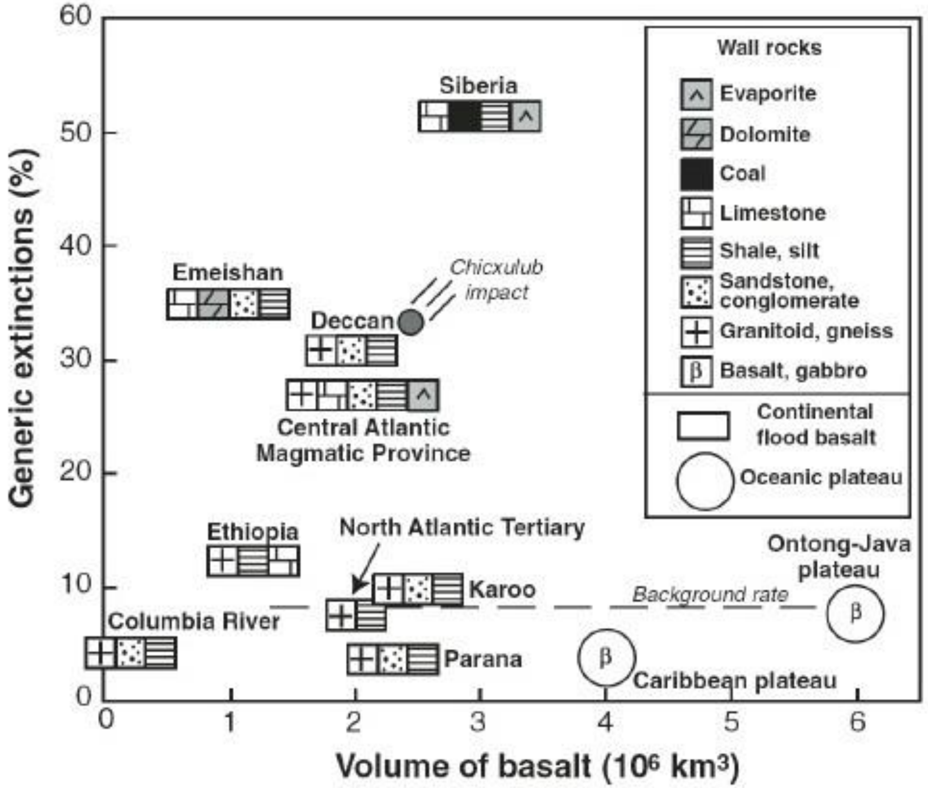
LIPs Main Characteristics

LIPs source has high temperature



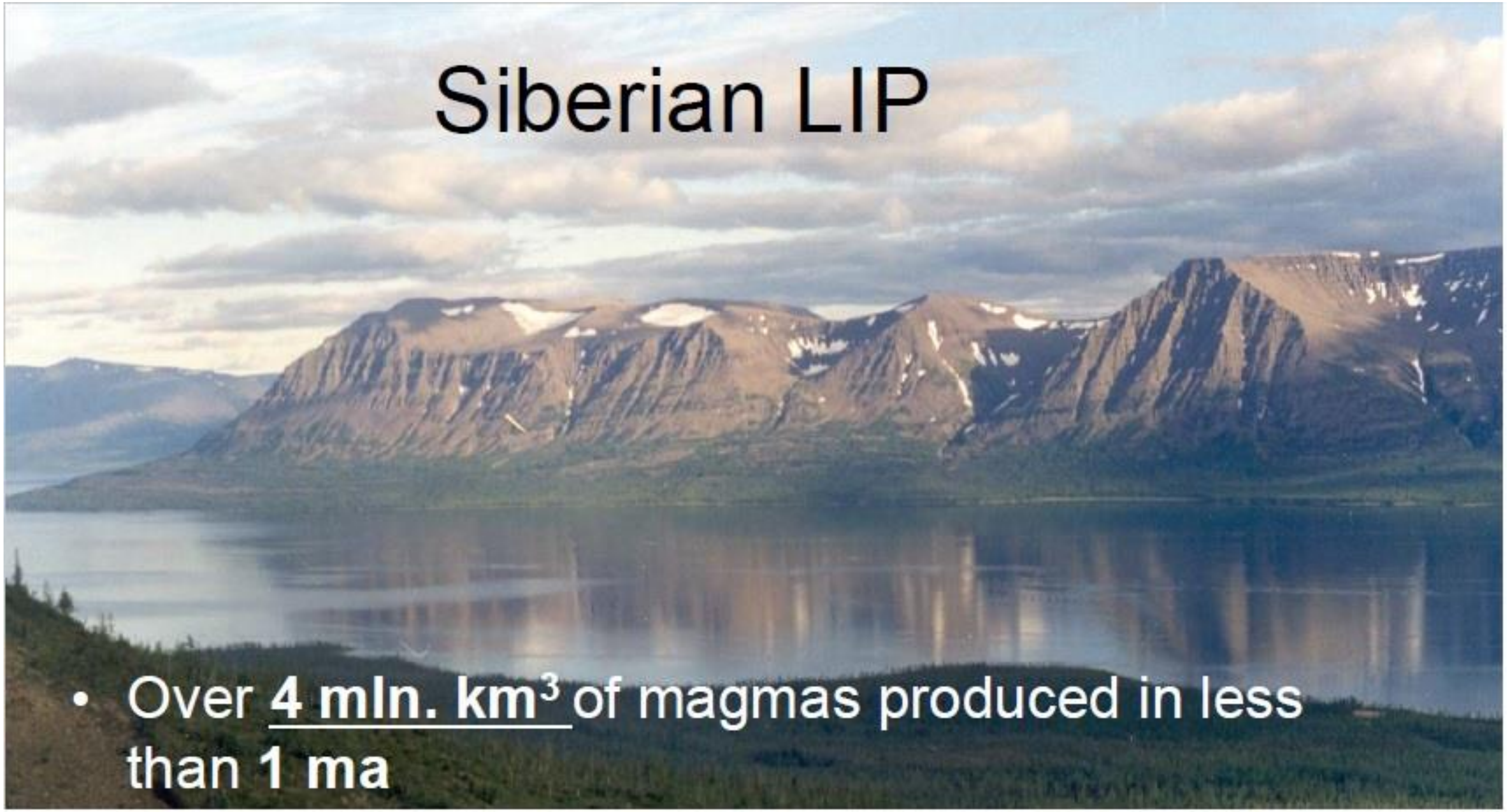
Herzberg and Gazel, 2009, Nature

LIPs correlate with mass extinction events

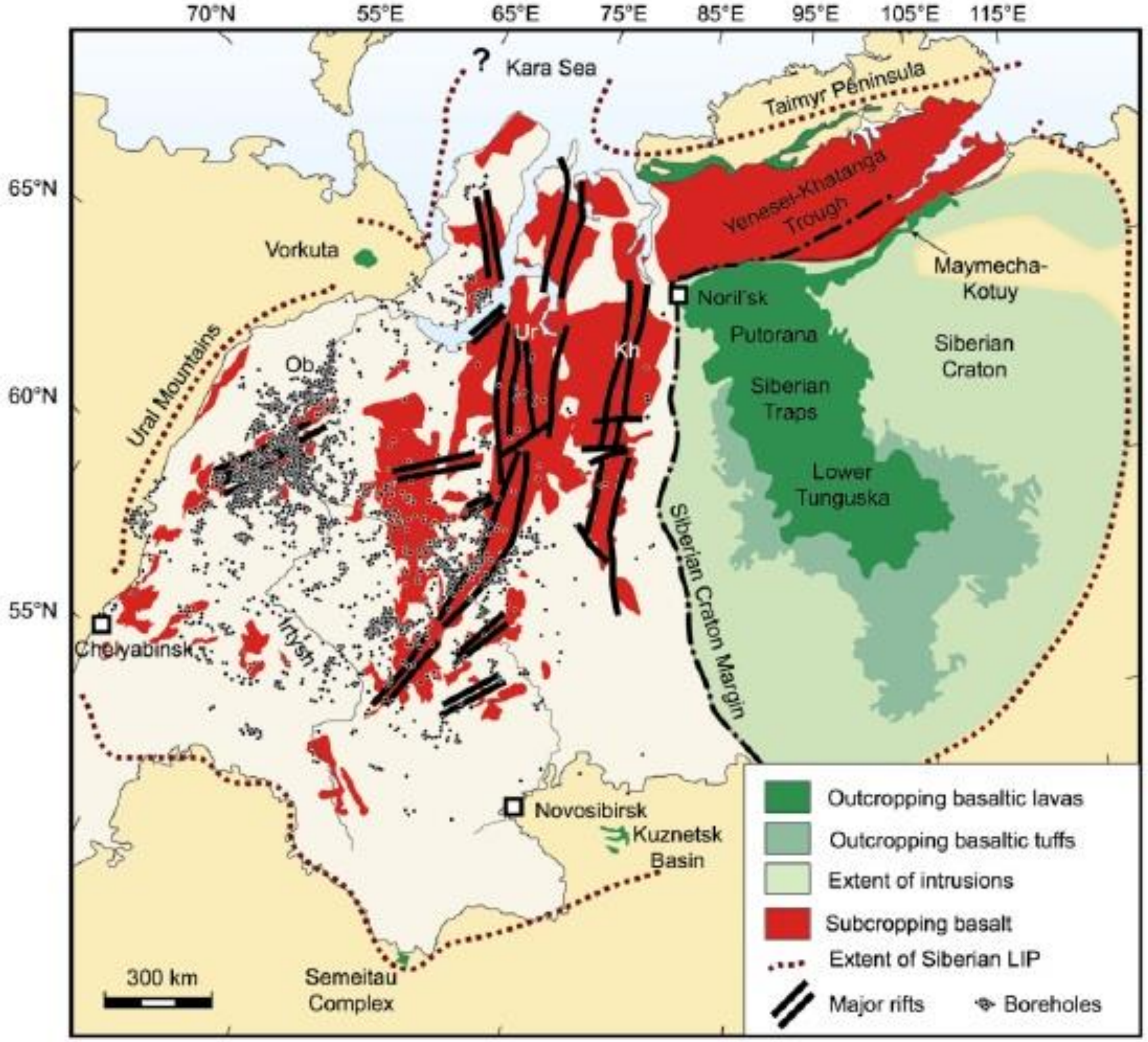


No Correlation with LIPs volume

Siberian Traps



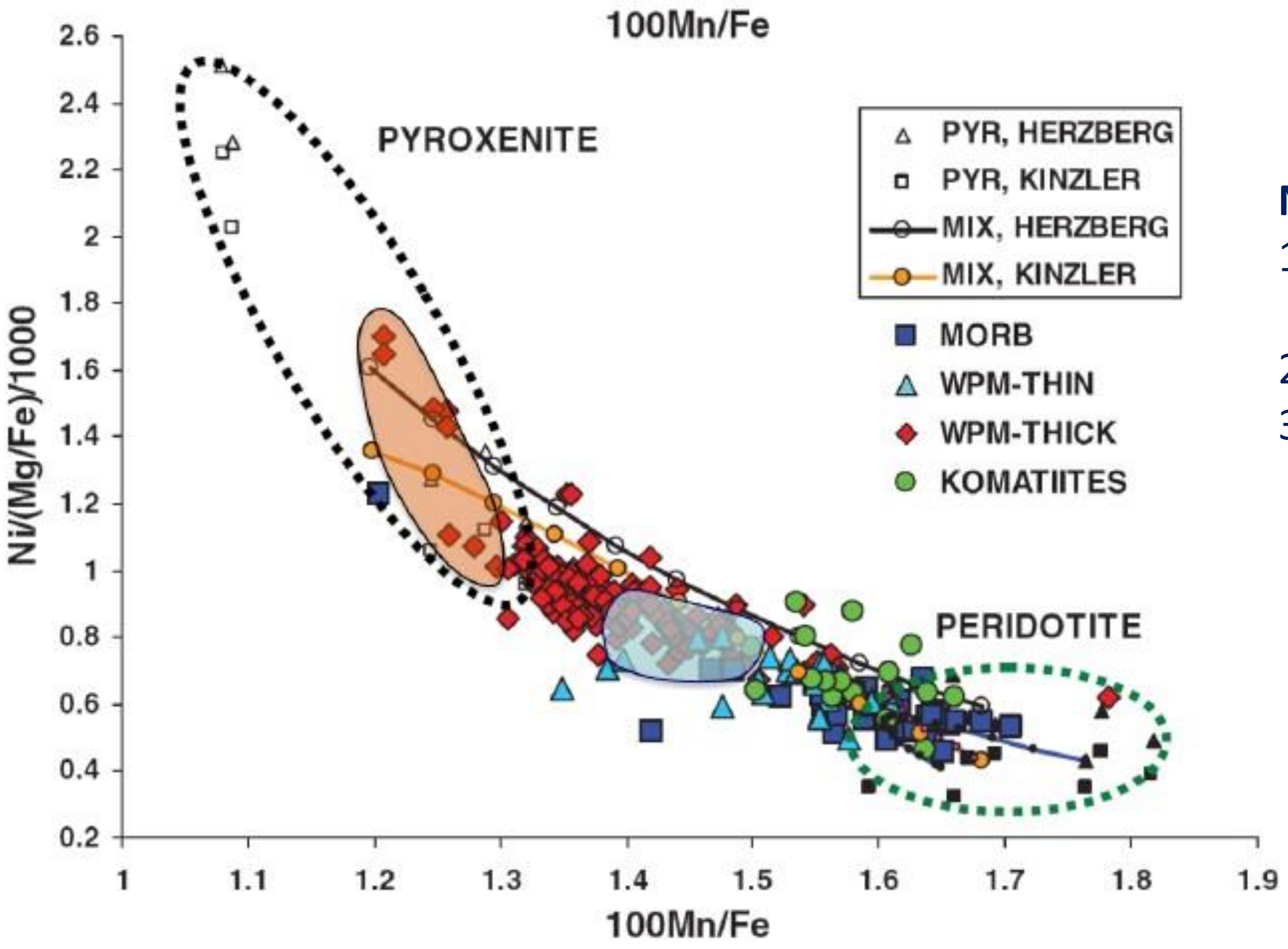
Siberian Traps



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



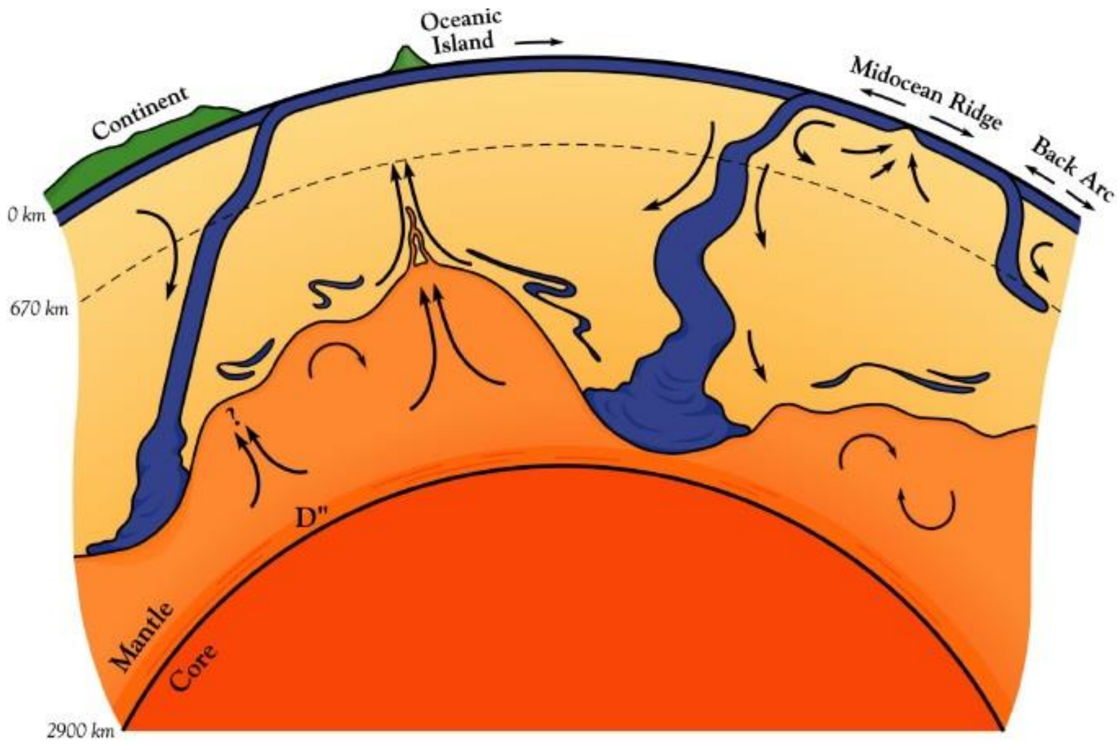
Sobolev et al, *Science*, 2007

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.

Piroxenite formation

Crustal recycling

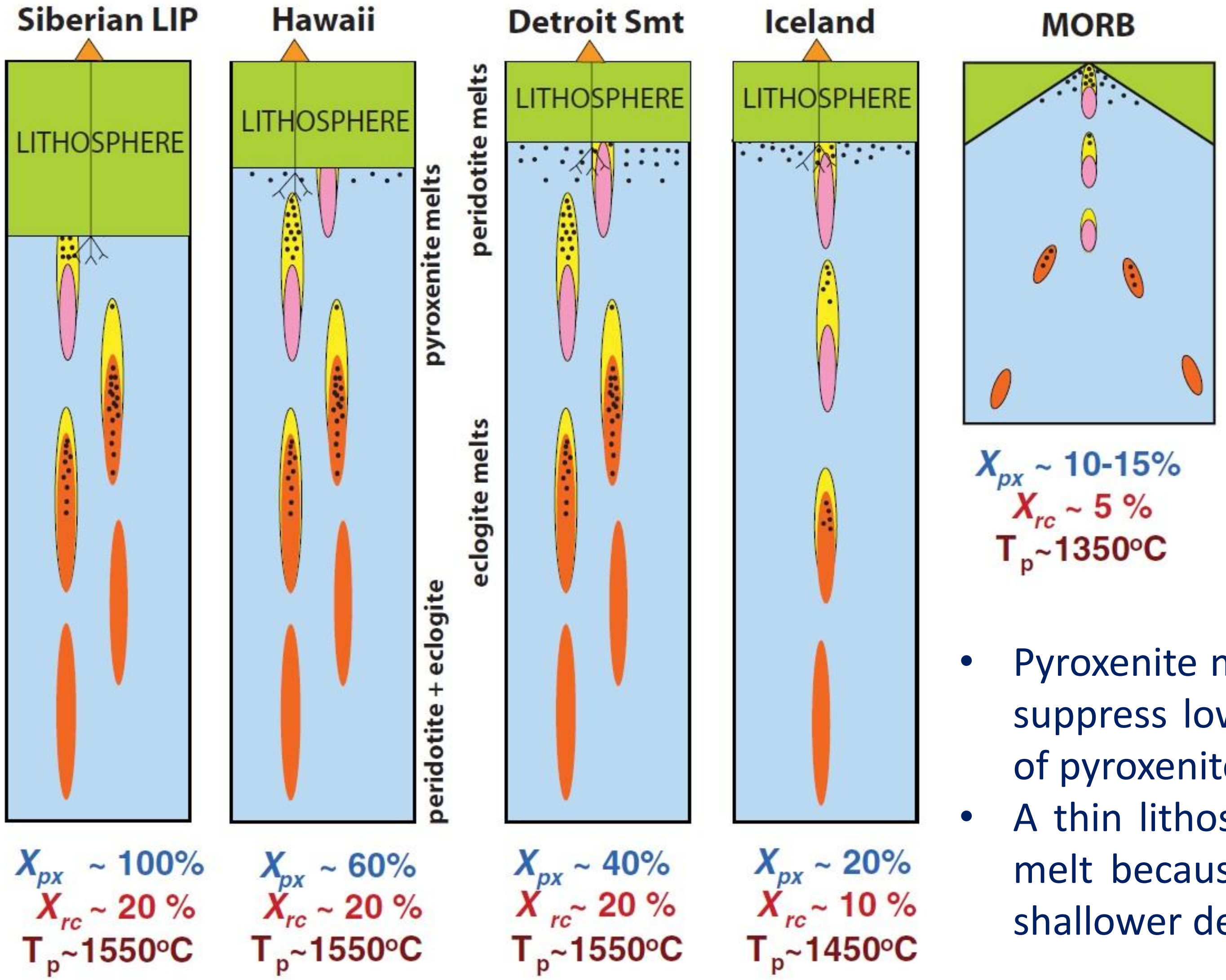


Kellogg et al., 1999

- 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
X_{px}=amount of pyroxenite derived melt
X_{rc}=amount of recycled oceanic crust in the peridotitic mantle

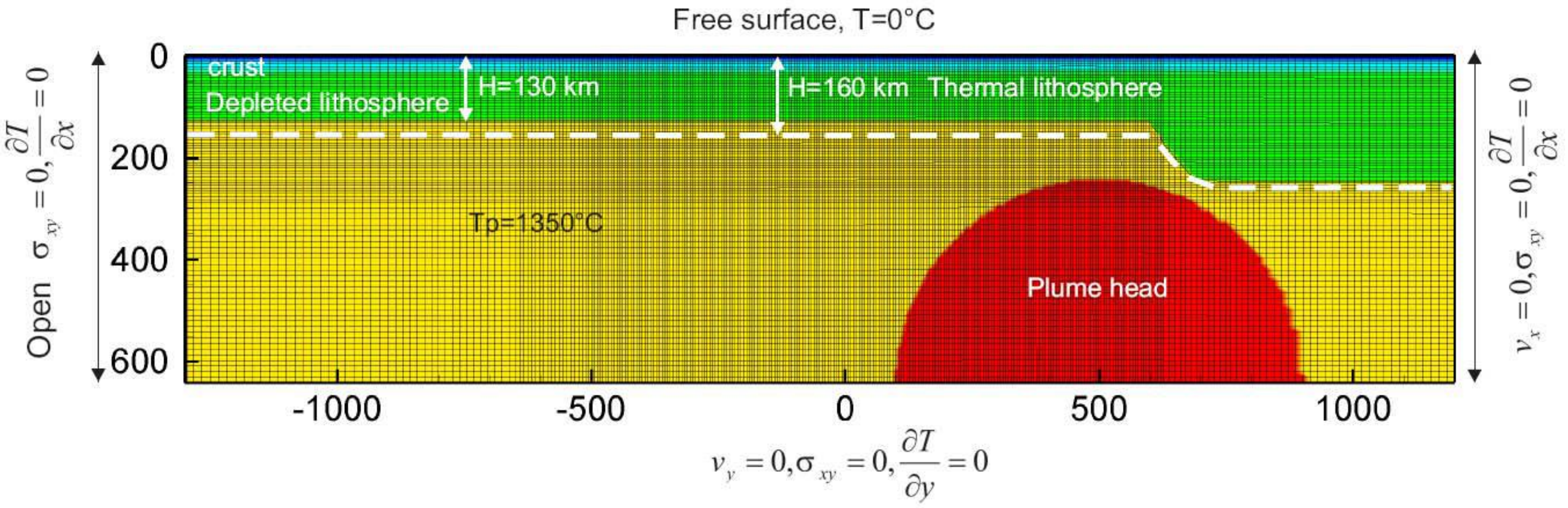
- Pyroxenite melts at higher *P* than peridotite, a thick lithosphere will suppress low-depth peridotite melting and favor a high proportion of pyroxenite derived melts.
- A thin lithosphere favors a higher proportion of peridotite-derived melt because of the increasing degree of melting of peridotite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

Sobolev et al, *Science*, 2007

Plume numerical models

Model Setup

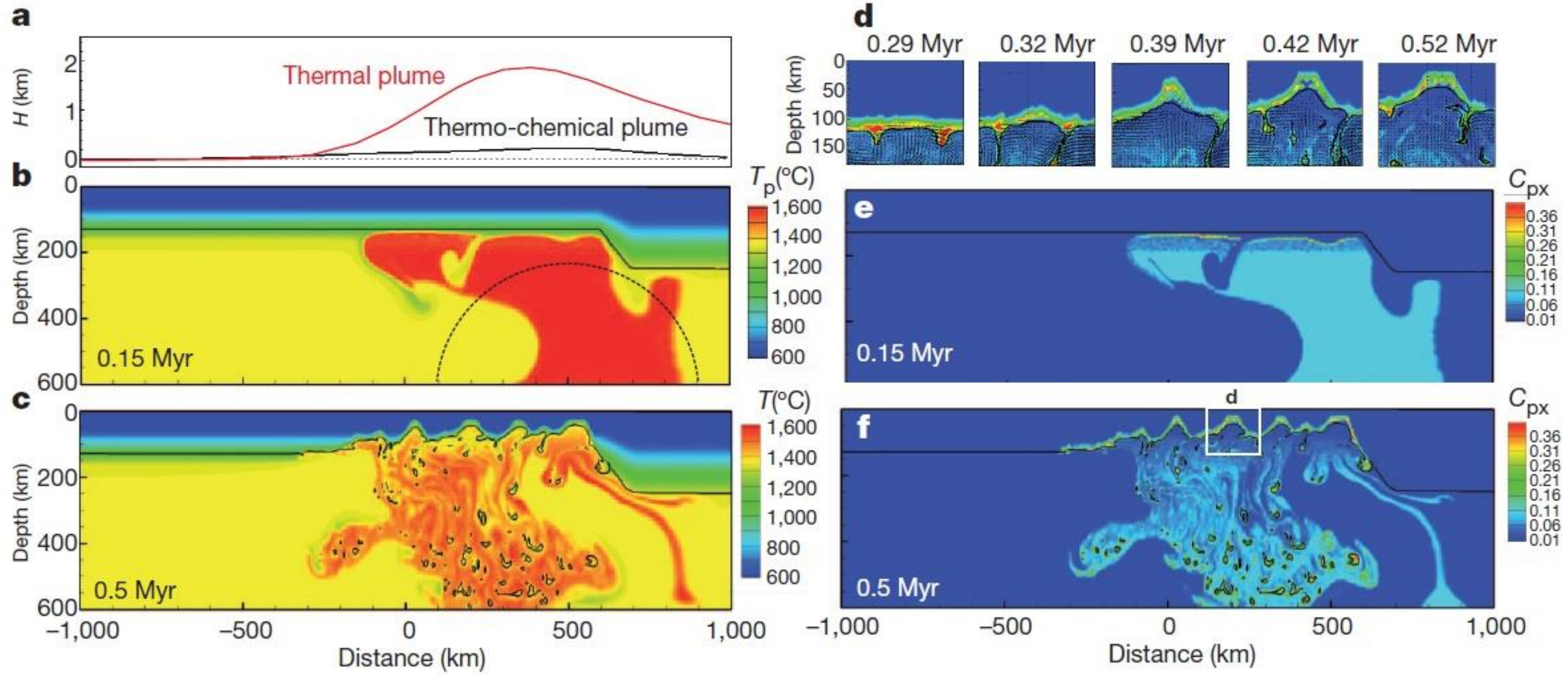
- Plume potential temperature Tp=1600°C
- Eclogite content in plume 10-20wt% (15wt%)
- Initial lithospheric thickness = 130 km



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

Sobolev et al., 2011, Nature

Plume numerical models



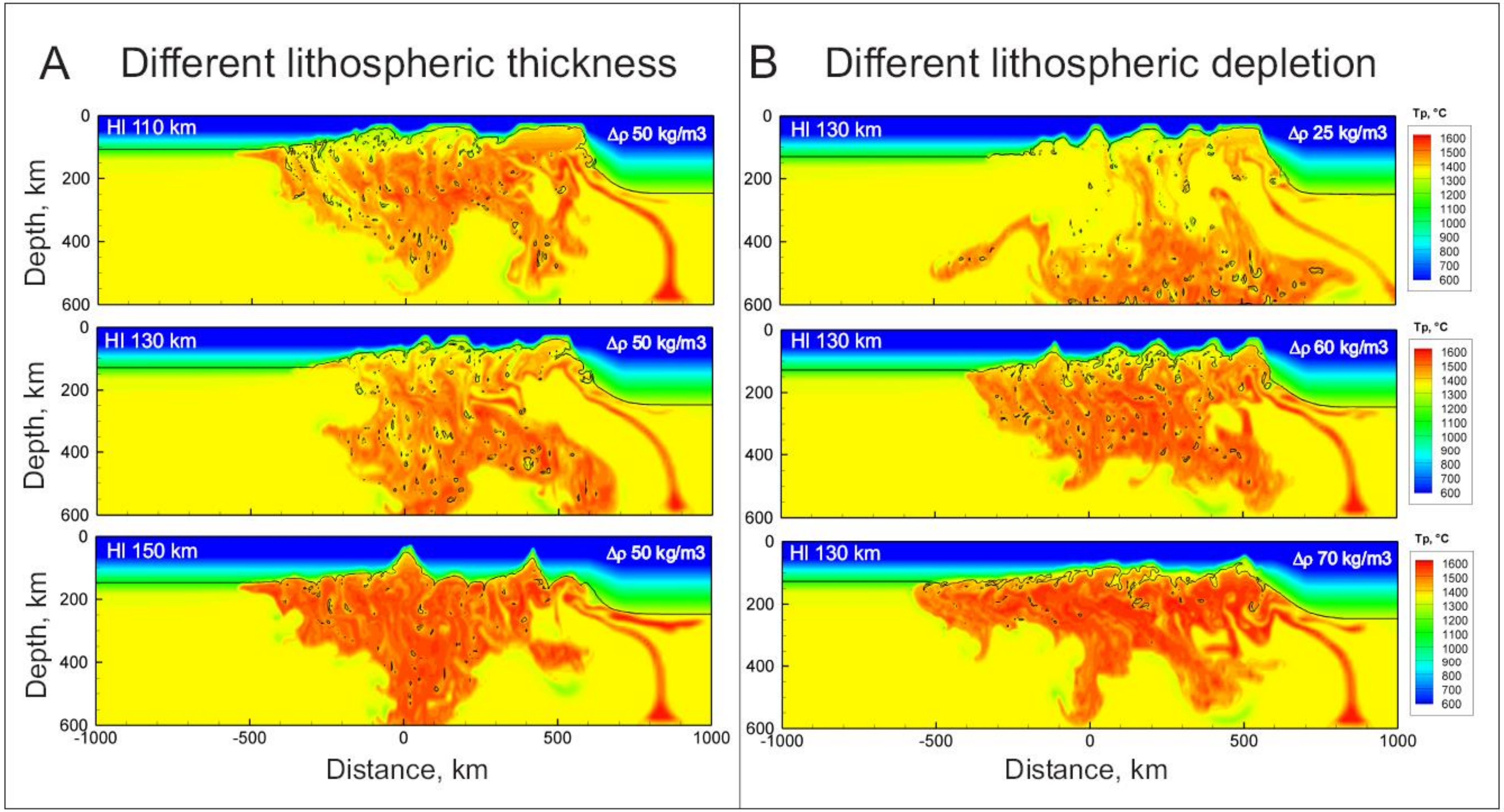
Sobolev et al., 2011, Nature

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

Large fraction (15 wt%) of dense recycled material is present within the plume, which strongly decreases its buoyancy, causing 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

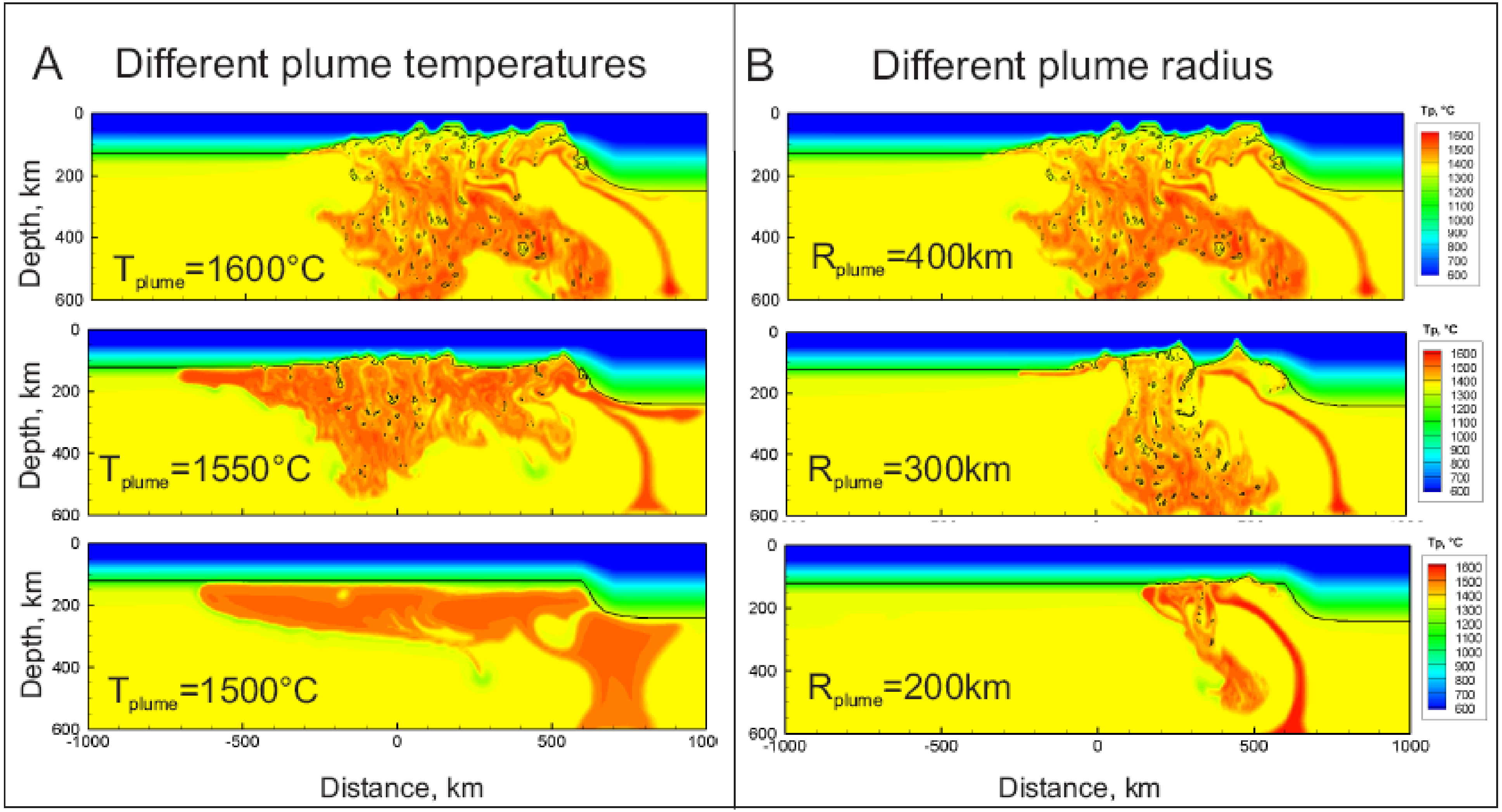


Hl= thickness of the depleted lithosphere

model time 1.0 Myr

Sobolev et al., 2011, Nature

Effect of plume on the intensity of the lithospheric destruction

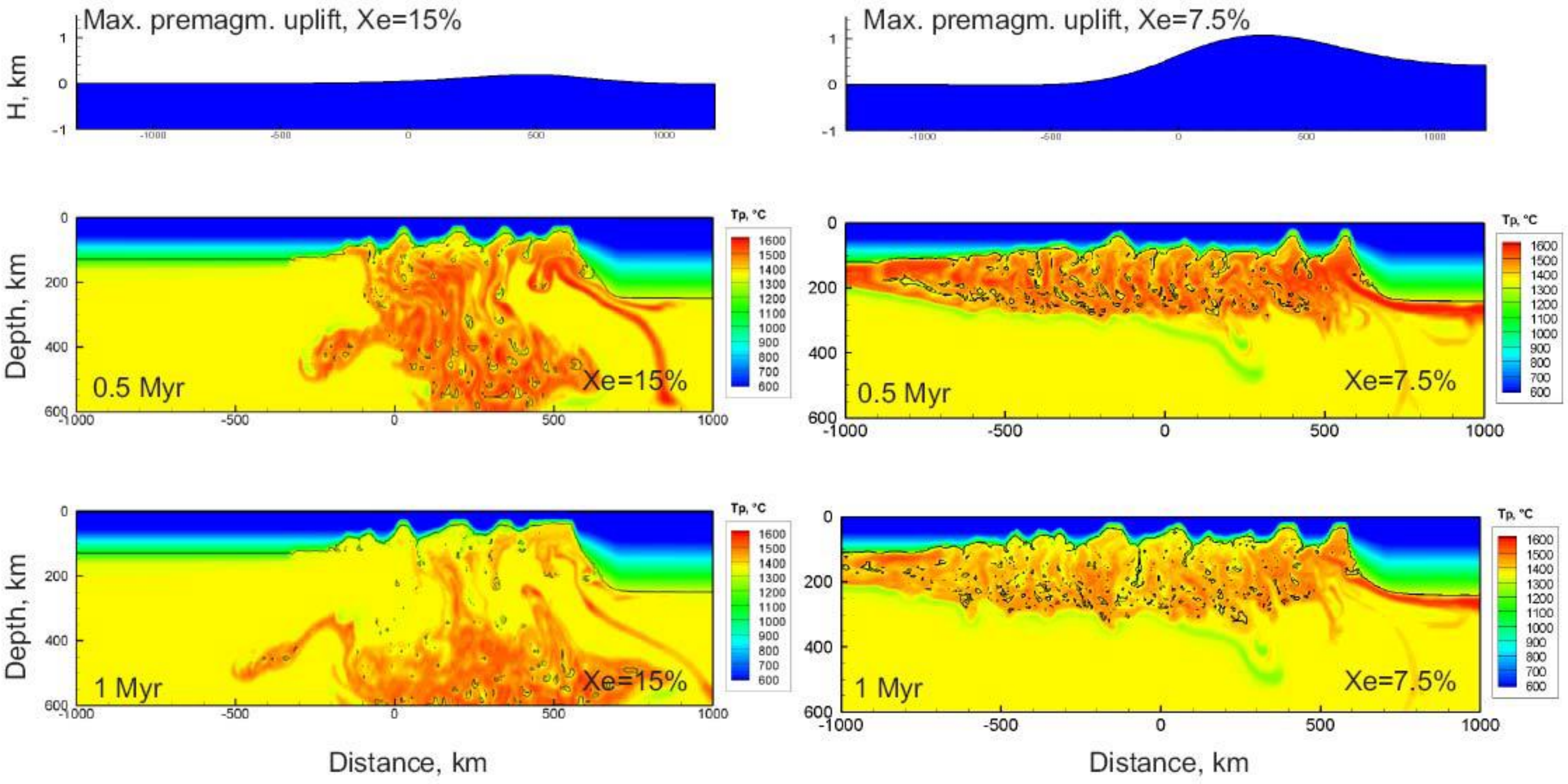


model time 1.0 Myr

Sobolev et al., 2011, Nature

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

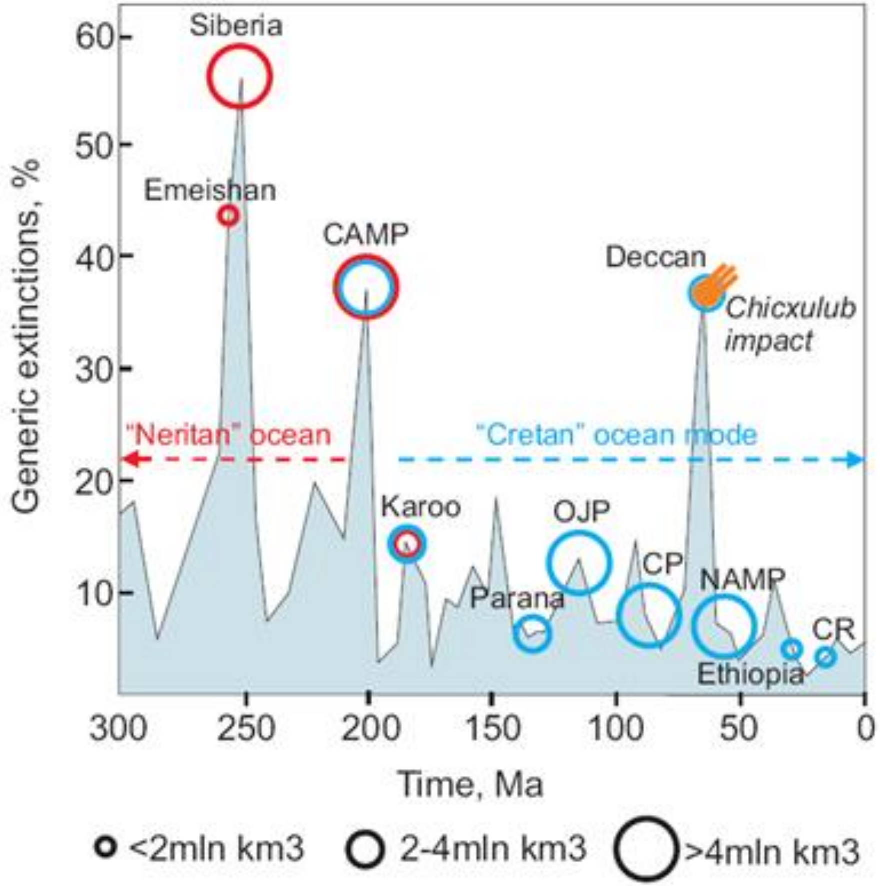
Different plume composition



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 Atlantic Magmatic Provinces, OJP, Ontong Java; CP, Caribbean Plateaux; CR, Columbian River basalts.

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