Course of Geodynamics

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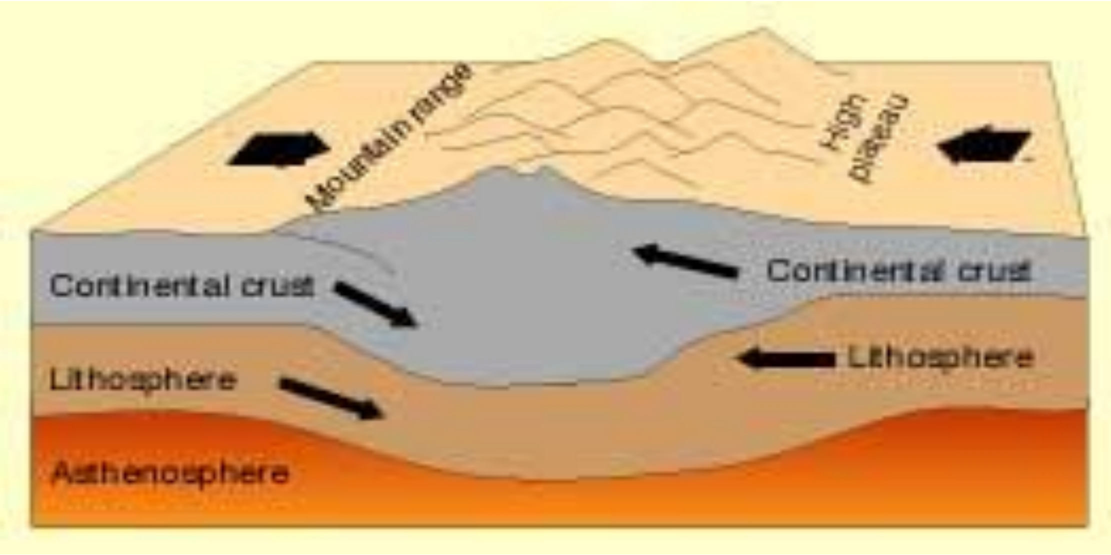
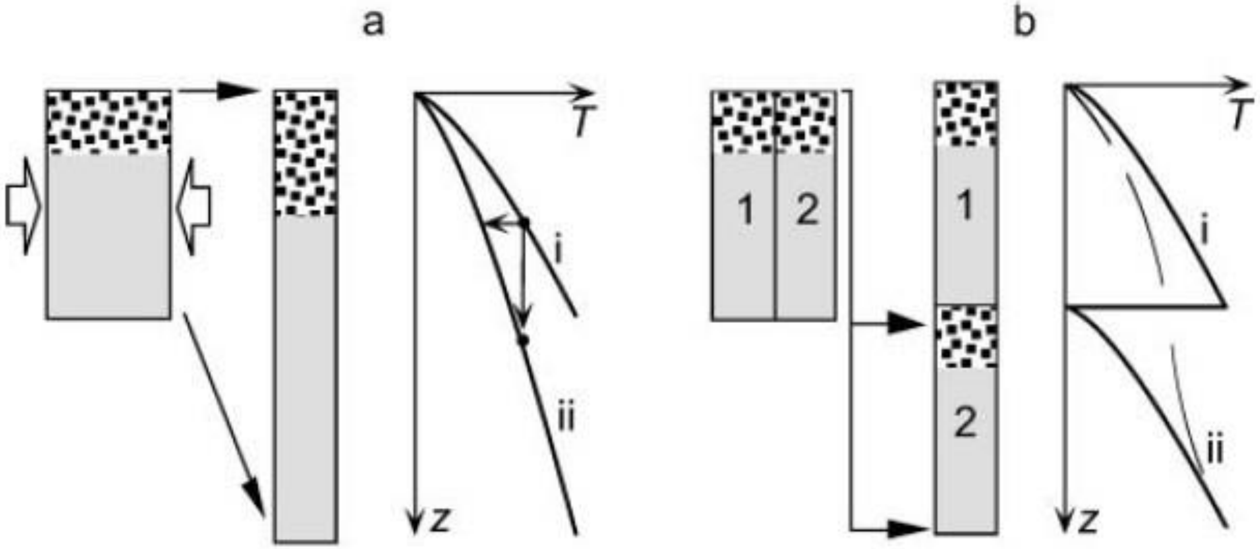
Course Outline:

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. Strength and effective elastic thickness 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

Continental Collision



Geotherms after thickening (before re-equilibration)

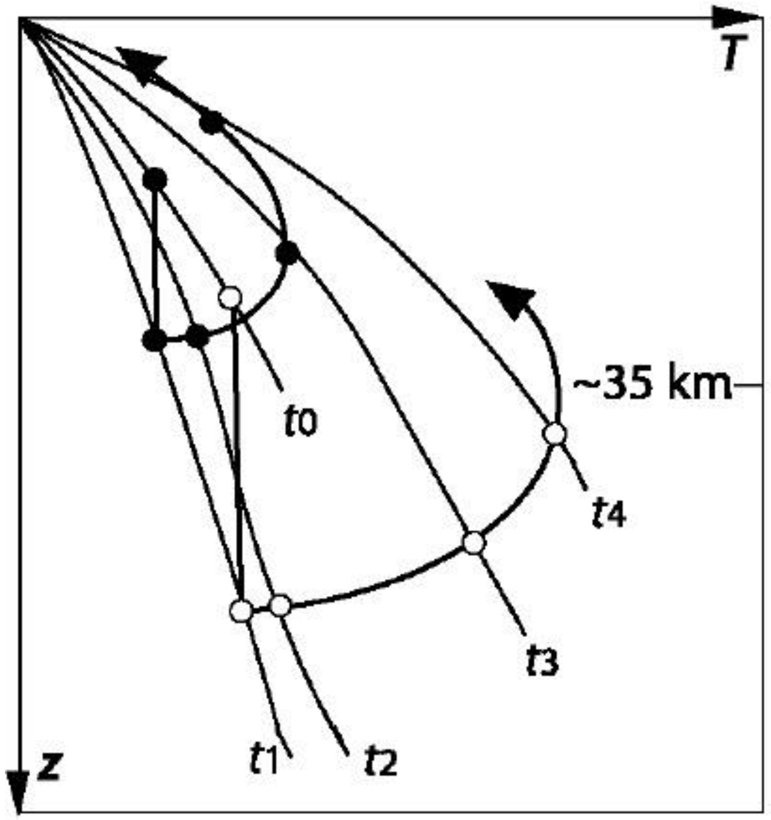


Collision of two continental plates leads to intense deformation and interfingering of both plates (different thickness, strength, and density with respect ot the oceanic plates)

Heating and collision do not occur simultaneously:

- Thickening of the crust is more rapid (10^{-14} s^{-1},indeed crustal thickening events last only few Myr, while orogenic cycles last about 10-100 Myr) than thermal equilibration (few tens of Myr), which increases quadratically with the thickness of the crust.
- Thickening of the crust causes an increase of the amount of radiogenic elements and thus its heating.

Continental Collision



- Extension and erosion processes following mountain building lead to cooling processes.
- The thermal evolution of rocks in a given orogen depends on the interplay and competition of the heating and cooling mechanisms.
- After the onset of denudation at the surface, the heating mechanisms wane and the influence of cooling mechanisms increases.
- Thermal evolution is shifted in time for different crustal levels: cooling of the upper crust commences at time t_3, while the lower crust heats at least until time t_4.
- There is a positive correlation between metamorphic grade and the time of metamorphism: The higher the metamorphic grade of a rock, the later its peak metamorphism occurred (while for contact metamorphic rocks this relationship is exactly the opposite).

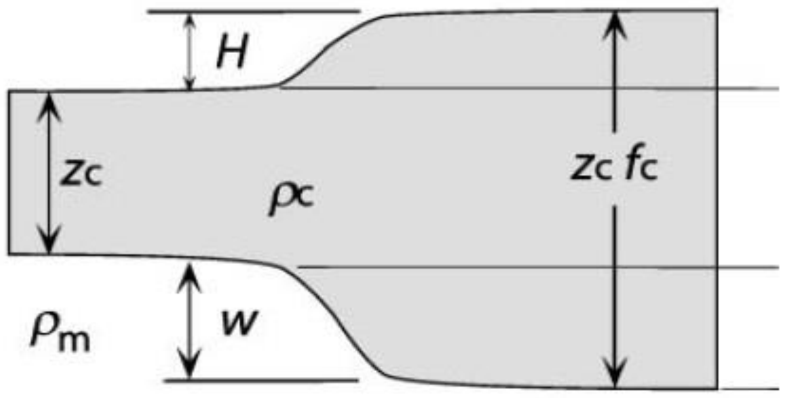
Continental Collision (mechanical description)

Force Balances in Orogens:

1) **Driving forces (F_d)**: Forces applied from the outside to an orogen (ridge push and slab pull).
2) **Internal forces (F_I)**: Forces internal to the lithosphere as inherent strength (vertically integrated strength).
3) **Potential energy (E_P)**: Forces resulting from the potential energy difference of an orogen relative to its surroundings (gravitational stresses or horizontal buoyancy forces). It grows with the square of the thickenss of the orogenic root and surface elevation.

We estimate E_p at depth z by summing up (i.e. integrate) the vertical stresses in the lithospheric column of interest between the surface and the depth of interest z (isostatic compensation depth).

$$E_{\mathrm{p}} = \int_0^z \sigma_{zz}\,\mathrm{d}z = \int_0^z \rho g z\,\mathrm{d}z = \frac{\rho g z^2}{2}$$



$$E_{\mathrm{p}}^{\mathrm{foreland}} = \rho_c g z_c^2/2$$

$$E_{\mathrm{p}}^{\mathrm{range}} = \rho_c g(H+z_c)^2/2 + \Delta\rho g w^2/2$$

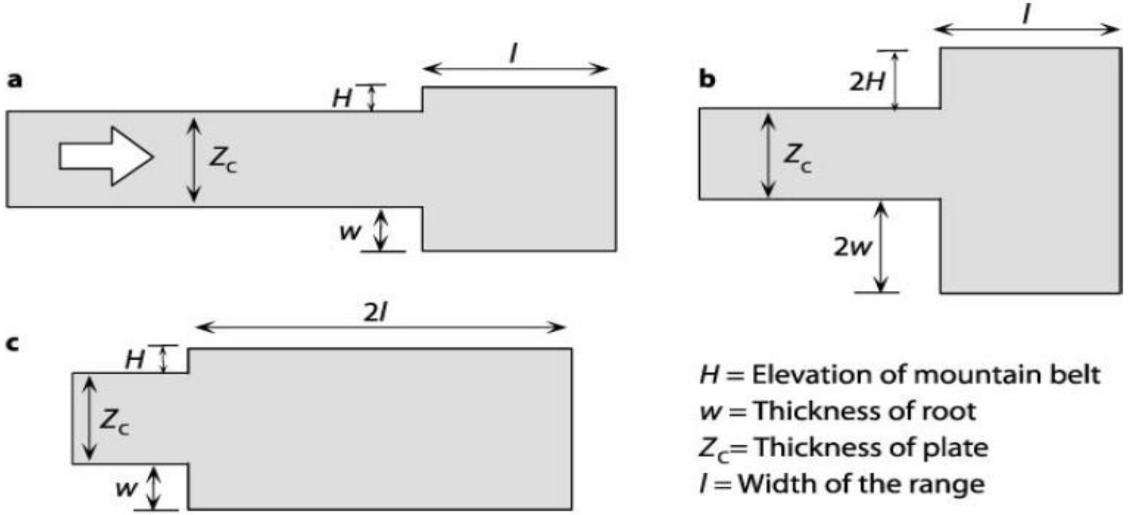
$$\Delta E_{\mathrm{p}} = F_{\mathrm{b}} = E_{\mathrm{p}}^{\mathrm{range}} - E_{\mathrm{p}}^{\mathrm{foreland}} = \rho_c g H^2/2 + \rho_c g H z_c + \Delta\rho g w^2/2$$

In isostatic conditions: $\Delta\rho w = H\rho_c$ $\quad \Delta E_{\mathrm{p}} = F_{\mathrm{b}} = \rho_c g H\,(H/2 + z_c + w/2)$

- It takes significantly more energy to increase the surface elevation of a high mountain range by one meter than it takes to increase the elevation of a low range by the same amount: **The height of a mountain range and the thickness of an orogenic root are limited, if the driving force is a constant**.

The limiting elevation of an orogen is reached when $\mathbf{F_d}$ (N/m)= $\mathbf{E_P}$ (J/m^2).

Continental Collision
(mechanical description)



H = Elevation of mountain belt
w = Thickness of root
Z_c = Thickness of plate
l = Width of the range

Potential energy per meter length of orogen:

$$\Delta E_{p,m^{-1}} = \rho_c g H l \left(H/2 + z_c + w/2\right)$$

If the crust inside the orogen is doubled in thickness:

$$\Delta E^{\mathrm{high}}_{p,m^{-1}} = 2\rho_c g H l \left(H + z_c + w\right)$$

If the growth of the mountain range is by doubling its width:

$$\Delta E^{\mathrm{wide}}_{p,m^{-1}} = 2\rho_c g H l \left(H/2 + z_c + w/2\right)$$

The difference of the potential energy increases between the two deformation styles is given by the difference between the two:

$$\Delta E^{\mathrm{high}}_{p,m^{-1}} - \Delta E^{\mathrm{wide}}_{p,m^{-1}} = \rho_c g H l \left(H + w\right) = \left(\frac{\rho_c \rho_m}{\rho_m - \rho_c}\right) g l H^2$$

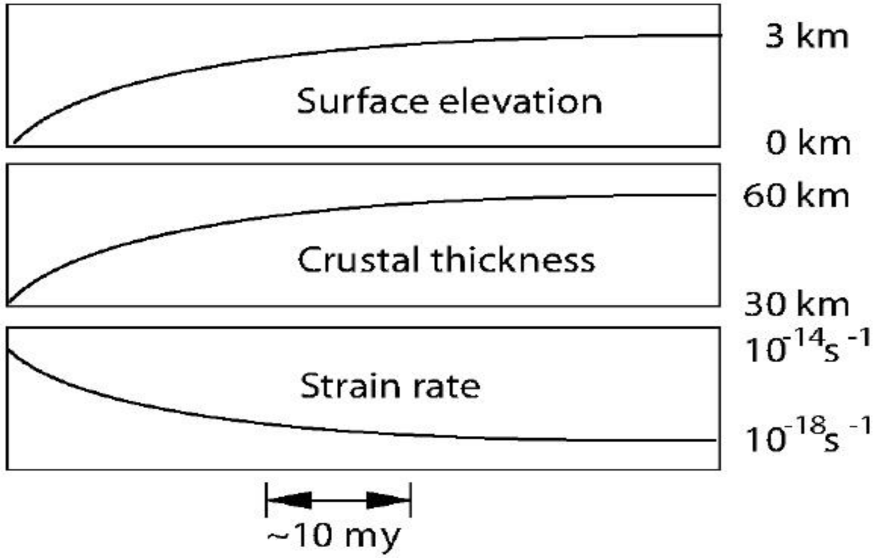
(in isostatic conditions: $\Delta\rho w = H\rho_c$)

- **When the convergence cannot be compensated by vertical growth of the range, it must be compensated by lateral growth of the range.**

Effective driving force (F_{eff}) = $F_d - F_b = F_l$ (when the orogen is deforming) **If $F_{eff} < F_l$ there is no deformation**

F_d=tectonic driving force per meter length of orogen
F_d=gravitational stress (horizontal buoyancy force) times the thickness of the lithosphere
F_l=vertically integrated strength

Continental Collision (mechanical description)



If the tectonic driving force remains constant to time, surface elevation and crustal thickness converge to a steady state when the magnitude of the horizontal buoyancy force approaches the tectonic driving force: the convergent strain rate goes towards zero (steady state conditions).

$$F_b = F_d \text{ and } F_{eff} = F_l = 0$$

Collisional orogen are self-limiting

$$F_d = F_b + F_l$$

If F_l decreases (e.g., T >>), F_b increases at costant F_d
Consequently, deformation is possible without changes of F_d

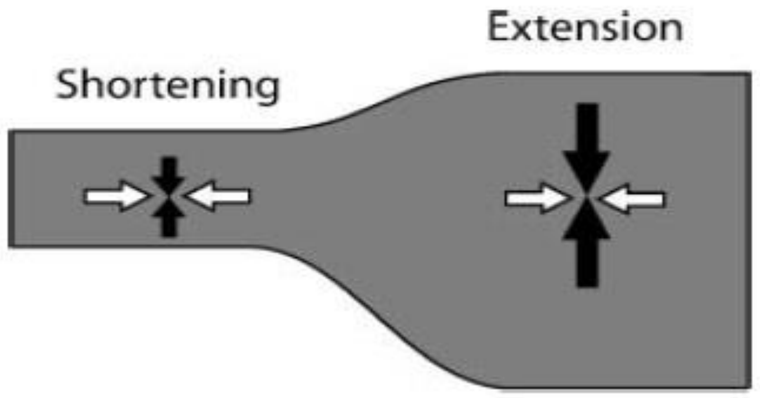
$F_b = - F_l$ This is the case in which a mountain range collapses under its own weight, $F_d = 0$

$F_d = F_l$ At the start of the growth of a new mountain belt, $F_b = 0$

Continental Collision

- **Changes in the stress field in collisional orogens (increase of potential energy of the mountain range)**

Horizontal stress in an orogen is constant at any one depth (mountain ranges and plateaus transmit horizontal forces without changing their magnitude), vertical stress is not, consequently there are some regions in extension and some others in compression (e.g., Tibetan Plateau is extending laterally, while there is thrust tectonics in the surrounding regions).

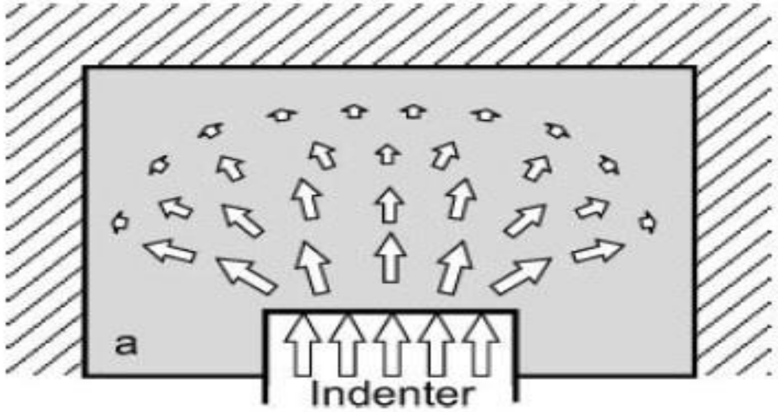


- **Changes in the stress field during the aging of plates (increase/decrease of the mean potential energy of the entire plate)**

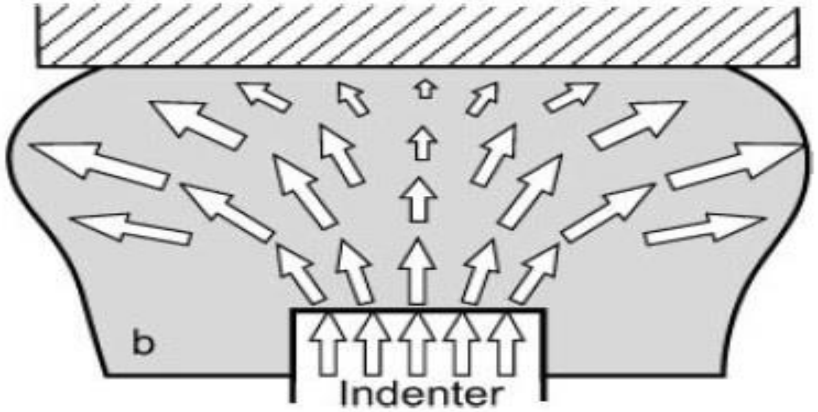
Orogen Parallel Extension

Orogen parallel extension in convergent orogens can occur when:

1. Unconstrained boundaries
2. Decrease in the convergence rate between two plates, if F_d<, F_b>, for F_l = constant: Post-orogenic collapse
3. Changes in the rheology of the plate, if F_l>, F_b< for F_d = constant).
4. External addition of potential energy to the plate (e.g., mantle delamination).



Constrained boundaries



Unconstrained boundaries

- **Argand number *Ar* is a measure for the ease with which the lithosphere deforms in response to gravitational stresses:** It tells if an orogen is likely to flow apart at the same rate it is being built, or if significant amounts of potential energy may be stored within it before it would collapse slowly under the influence of gravitational stress.

Argand number (A_r) may be interpreted as dimensionless ratio of the additional pressure $P_{(L)}$, that arises because of the thickness difference *L* between two plates and the stress τ_0, that is necessary to deform a plate with a significant strain rate

$$\dot{\epsilon}_0 = U_0/L \qquad \bar{B} = A^{(-1/n)} e^{Q/nRT}$$

U_0= collisional velocity between two plates

B summarizes all temperature dependent terms of the power law

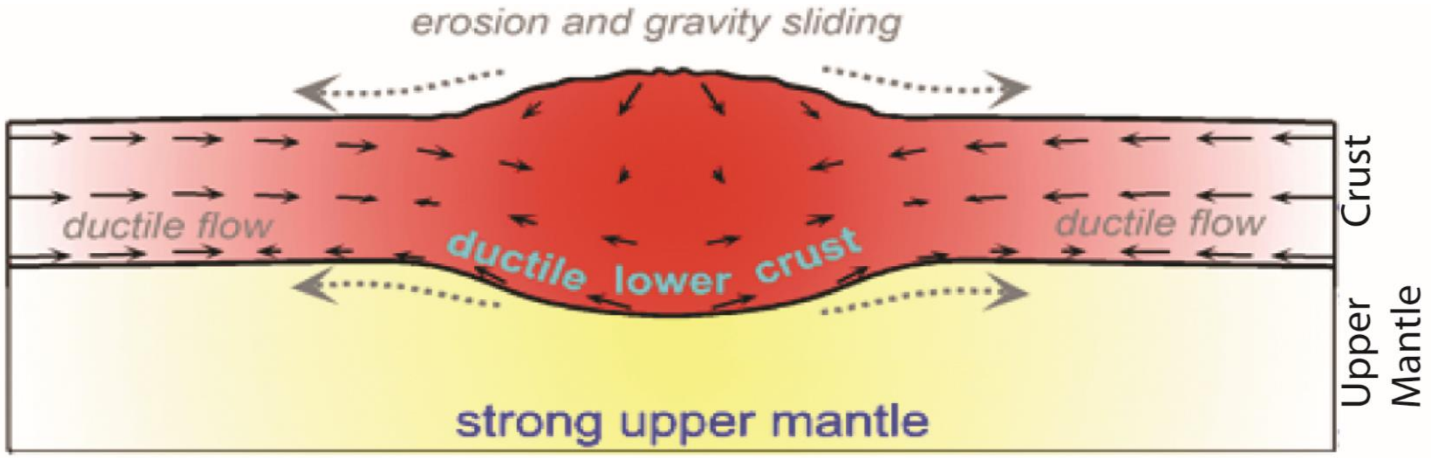
$$Ar = \frac{\rho_c g L(1-\rho_c/\rho_m)}{B(U_0/L)^{1/n}} = \frac{P_{(L)}}{\tau_0}$$

A_r>>1 the crust tend to flow, since viscosity of the orogen is small

How much can thicken the crust?

Max 70-80 km, further thickening cause Gravitational Collapse and Denudation

- **Burial during collision causes heating and weakening of rocks.**
- **Lateral deep crustal flow is driven by gravity forces (gravitational collapse).**
- **Upper crust thins and extends tectonically (denudation).**



Younger extensional structures overprint older compressional ones

Other processes related to collision: Lateral Escape of crustal blocks towards free space (e.g., Himalaya) or Lateral Tectonic Extrusion if accompanied by gravitational collapse (e.g., Eastern Alps).

Orogenesis

- Orogenesis occurs at convergent plate margins and involves (1) intra-plate shortening, (2) crustal thickening (up to 70 km), (4) deformation, (4) metamorphism, and (5) topographic uplift.
- Processes that change the strength of continental lithosphere during orogenesis commonly include magmatism, metamorphism, crustal melting, crustal thickening, sedimentation, and erosion.

Orogens related to oceanic subduction

Non-collisional Andean Type orogens: ocean-continent convergence, causing oceanic lithosphere subduction beneath continental margins, with collision of islands arcs, oceanic plateaus, microcontinents, and compression within the override plate.

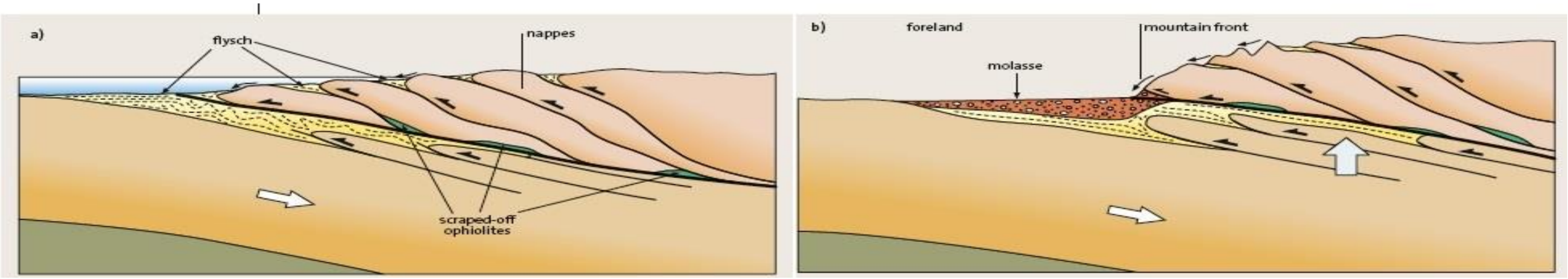
Collisional orogens: develop where a continent or island arc collides with a continental margin, following the closure of an ocean (it completes a Wilson cycle). The two continental margins become overthrust and underthrust, folded and metamorphosed.
Example: Belts in Taiwan and Himalayan Orogen.

Orogens not related to oceanic subduction

Accretionary Orogens: Orogens that have grown by collision of distinctive assemblages of crustal material (terranes) and through magma addition, sedimentation, and the creation and destruction of extensional basins.
Example: Western Cordillera of North America.

Alpine-type continent-continent collision

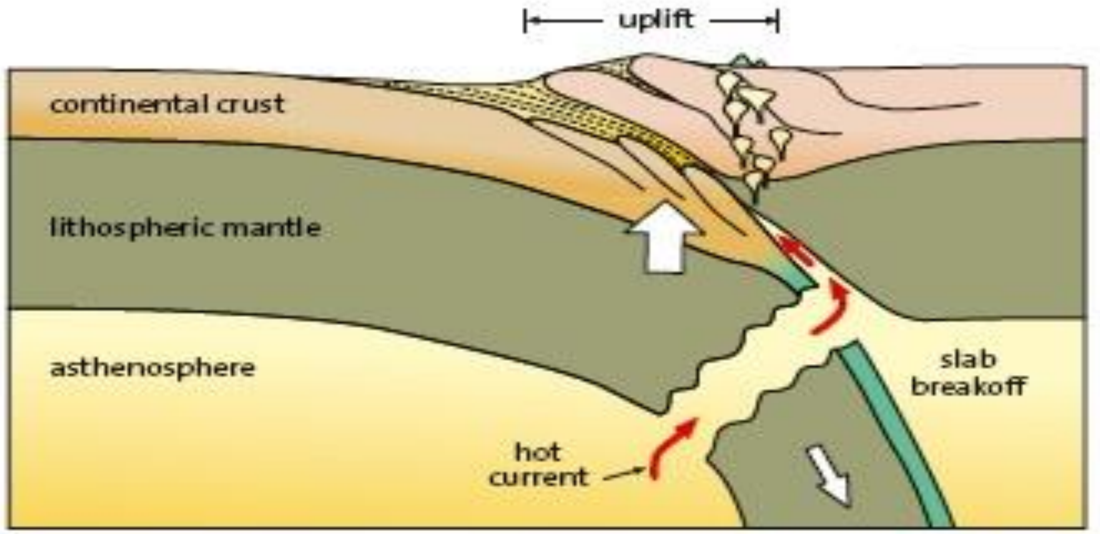
Main features: (a) large nappes, (b) broad belts of deformation and regional metamorphism, (c) the occurrence of ophiolitic sutures (remnants of the ocean floor), (d) island-arc magmatism (limited by the duration of subduction).



How does uplift occur?

First Phase (Pre-uplift): (1) the subducting continental margin is pulled downward by the attached oceanic slab; (2) The dense oceanic slab forms a counterweight against the thickened and buoyant continental crust.

Second Phase (uplift): (1) the resistance against compression in the collision zone becomes strong enough, so that the dense, heavy subducting oceanic lithosphere breaks off; (2) hot asthenospheric mantle ascends into the newly created space and to cause partial melting and loss of the heavy counterweight.



What is influencing uplift of mountain ranges?

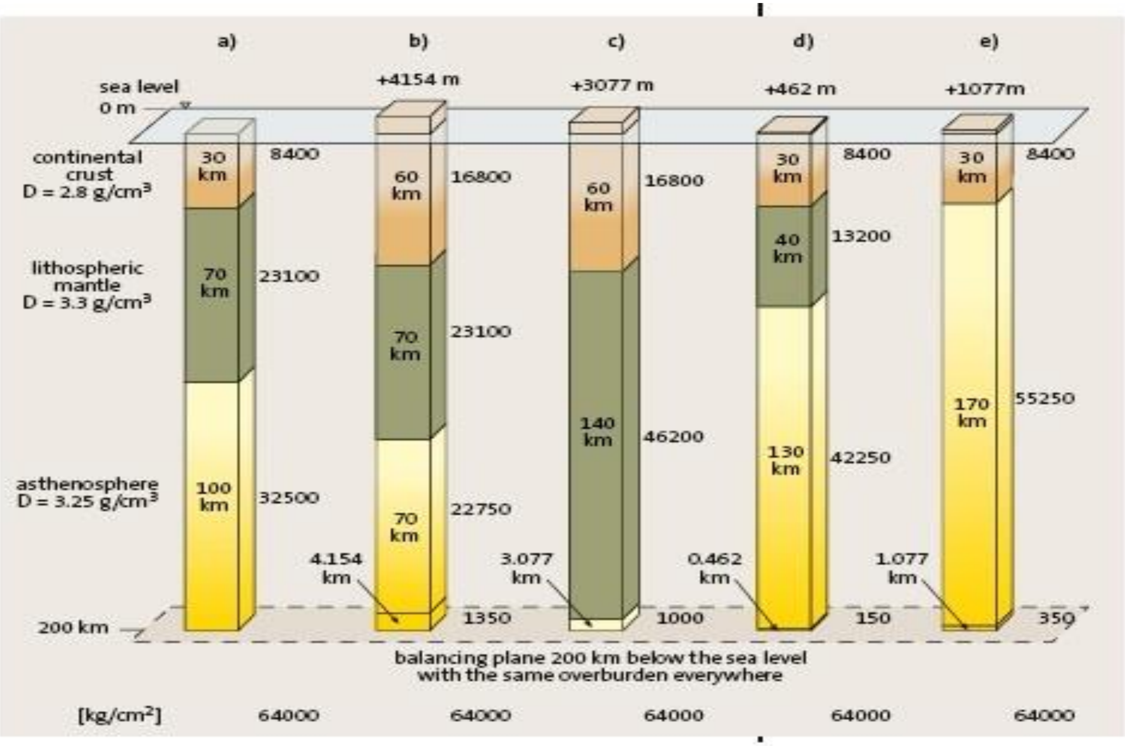
Tectonics Effects

Climatic Effects

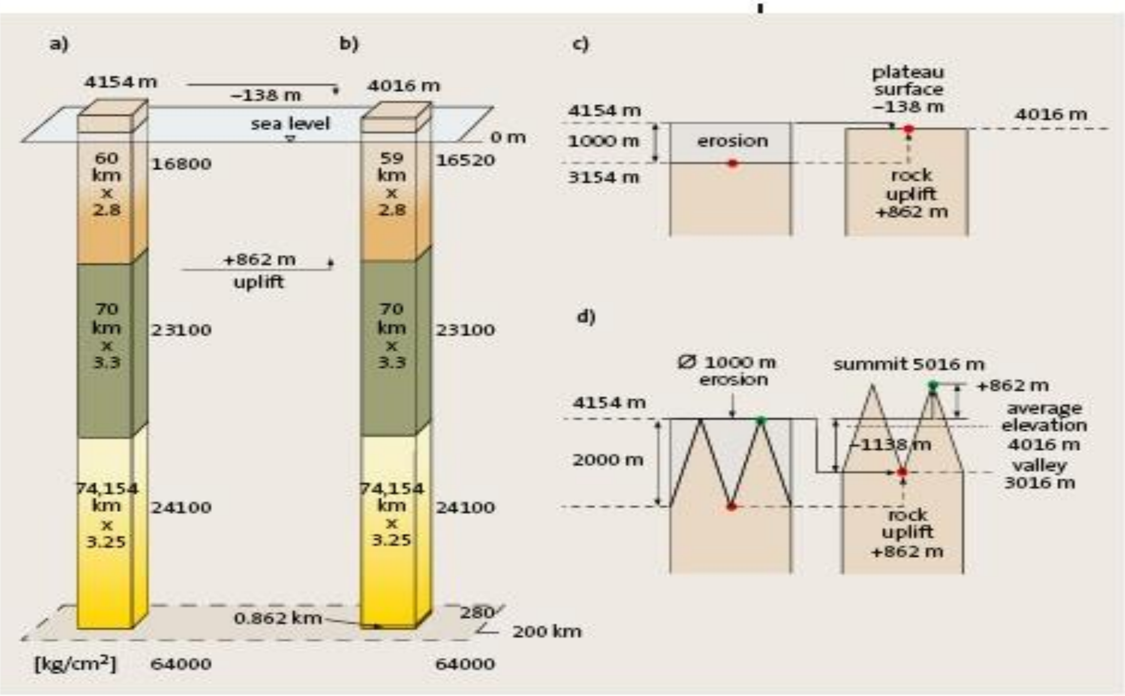
(explain morphological difference between Himalaya and Tibetan Plateau)

Thickening

Delamination

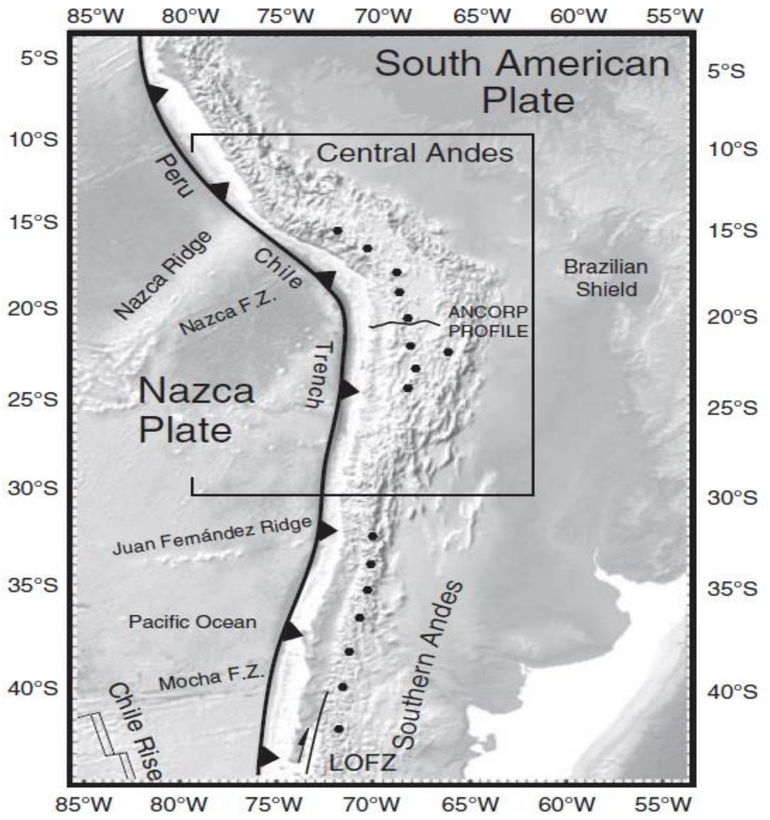


Erosion

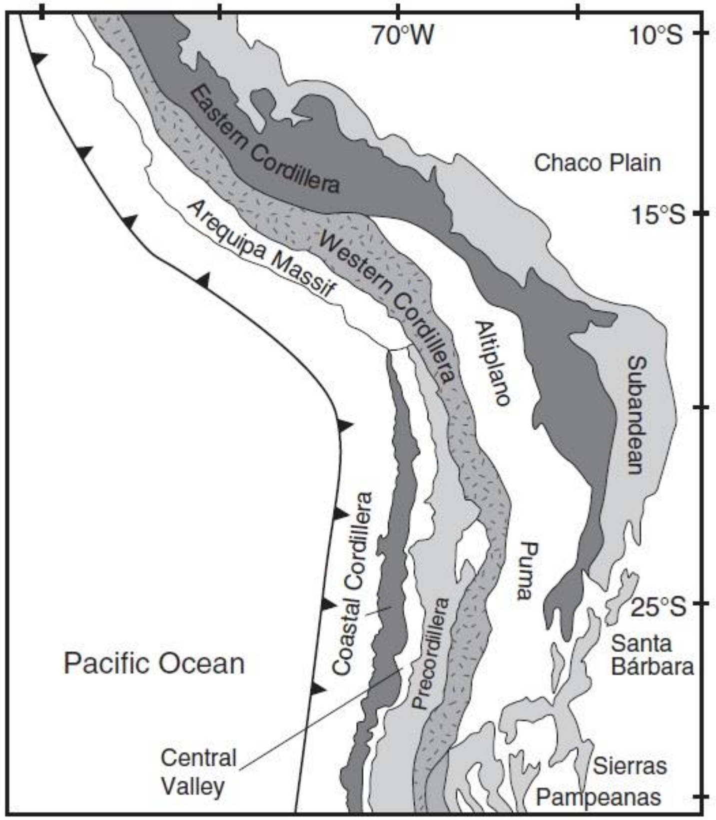


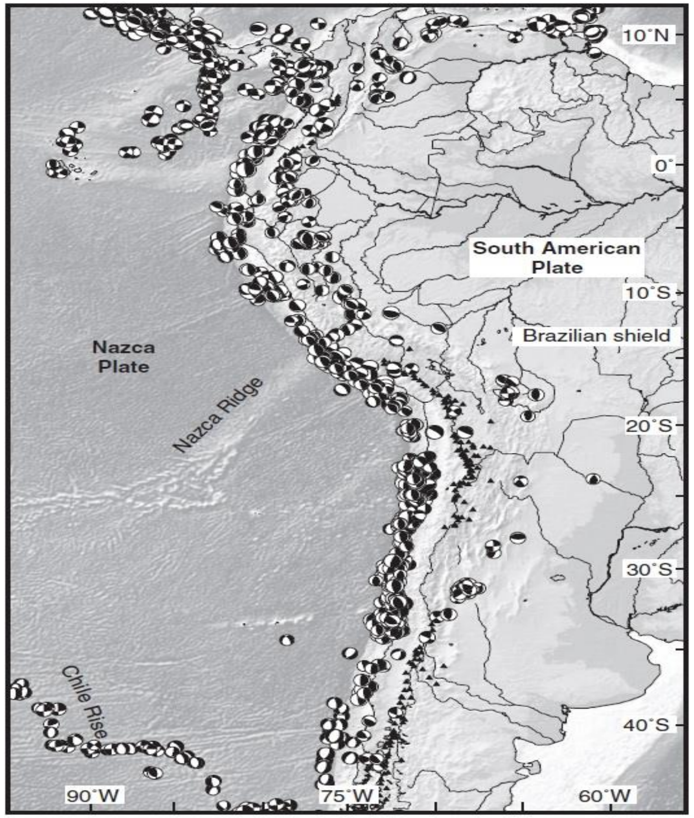
Andes

- 7500 km from Venezuela and Colombia in the north to Tierra de Fuego in the south.
- Geodetic data suggest that convergence velocities with respect to South America are 66–74 mm yr^{-1} at the trench (likely decelerated from a peak of ~150 $mmyr^{-1}$ occurred 25 Myr ago).
- Compressional regime started in the Early Mesozoic and continued in the Cenozoic due to (1) trenchward acceleration of the South American plate and (2) coupling between subducting oceanic lithosphere and overriding continent.
- Significant shortening and crustal thickening initiated at 50-45 Myr to the north of 25-30°S and 20 Myr to the south.
- Focal mechanism solutions show compression and strike-slip motions.
- Altiplano-Puna plateau: broad plateau, 3.8–4.5 km high, 1800 km long, and 350–400 km wide, starting uplift in the Miocene.

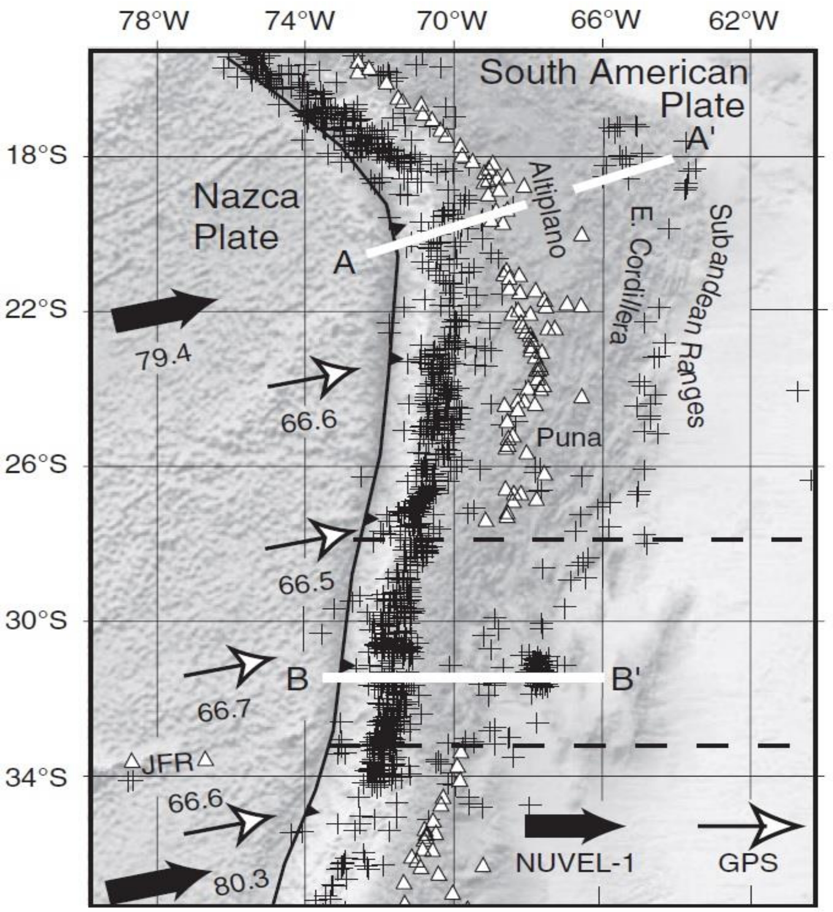


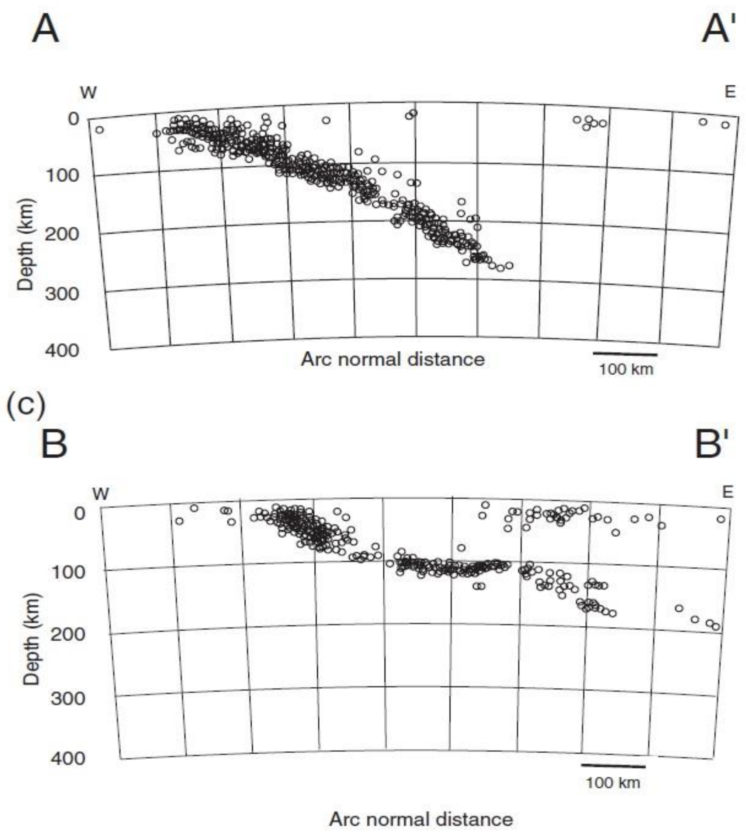
Black dots show active volcanoes





Alternation of flat and steep subduction segments





Beneath southern Peru and Bolivia, the Benioff zone dips about 30°.

Beneath north-central Chile, it initially forms an angle of 30° to a depth of ~100 km and then dips at angles of 0–10° for several hundred km (shallow angle likely developped after the subduction of an oceanic ridge) .

Different dips can be explained by (1) lithospheric tear or (2) distorted down-going plate

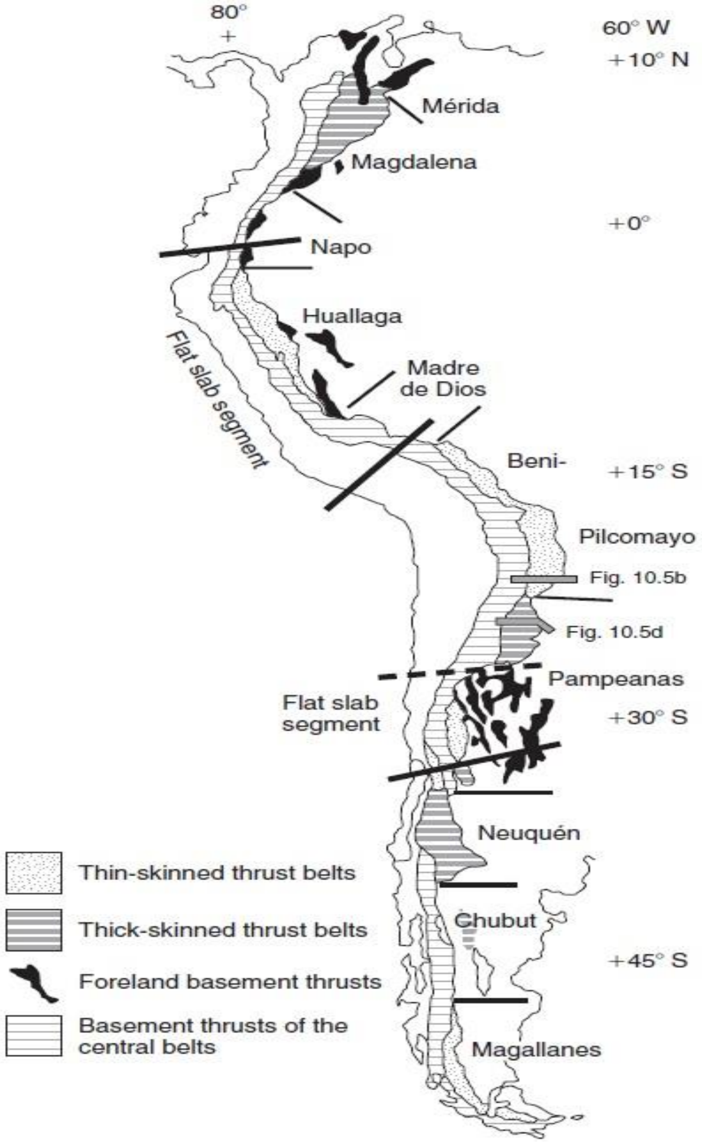
Above zones of flat subduction, the seismic energy released is 3-5 times larger and shallow seismicity is more abundant and broadly distributed than over neighboring steep segments.

Flat segments are strongly coupled to the overriding continental plate:

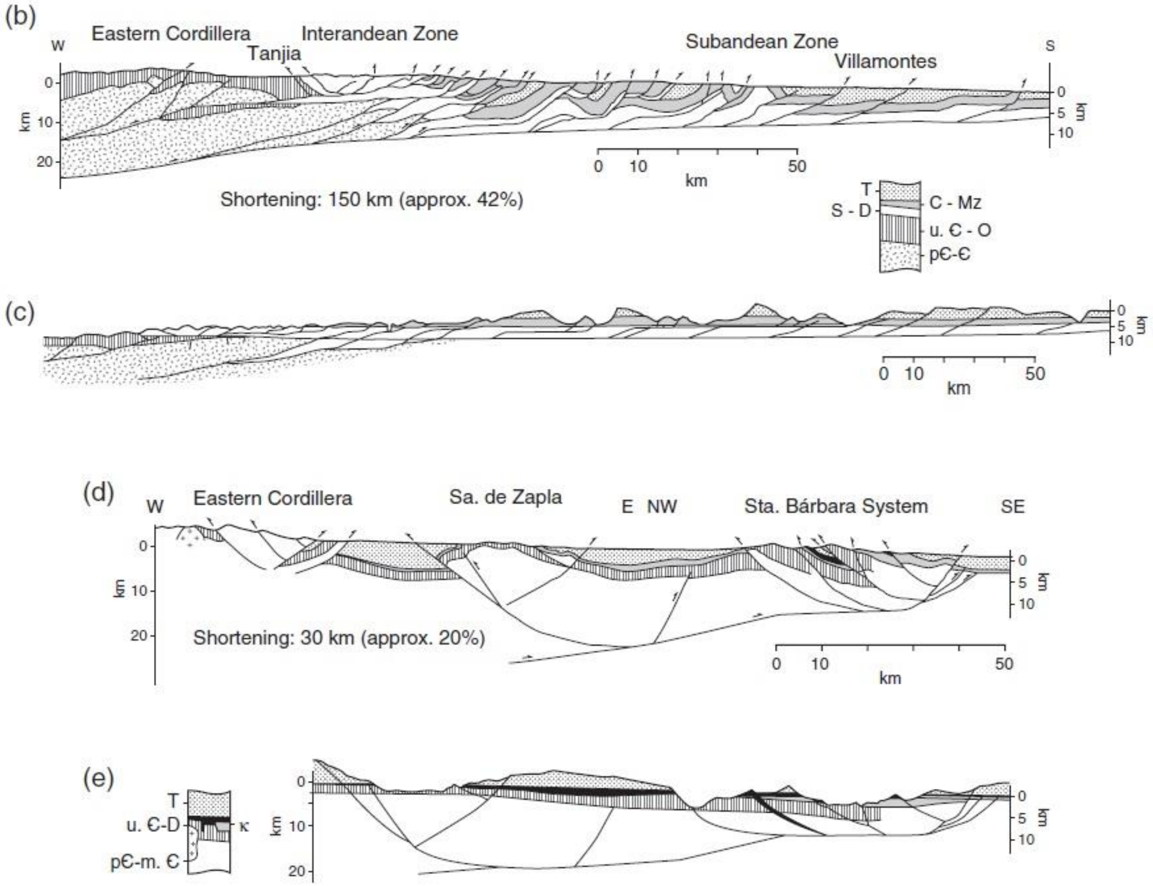
Cool slab at shallow depths beneath the continental lithosphere enables (1) to transmit stress at long distances = more shallow seismicity and (2) enables to eliminate asthenospheric wedge = no magmatism (active volcanic arc where subduction plate deeply steeps).

Deformation style of South American Cordillera

Alternations among different styles of shortening along the strike of the orogen cause segmentation of the Andean foreland (thin and thick-skinned tectonics). The development of thin- or thick-skinned styles of shortening commonly is controlled by the presence of inherited stratigraphic and structural heterogeneities in the crust.



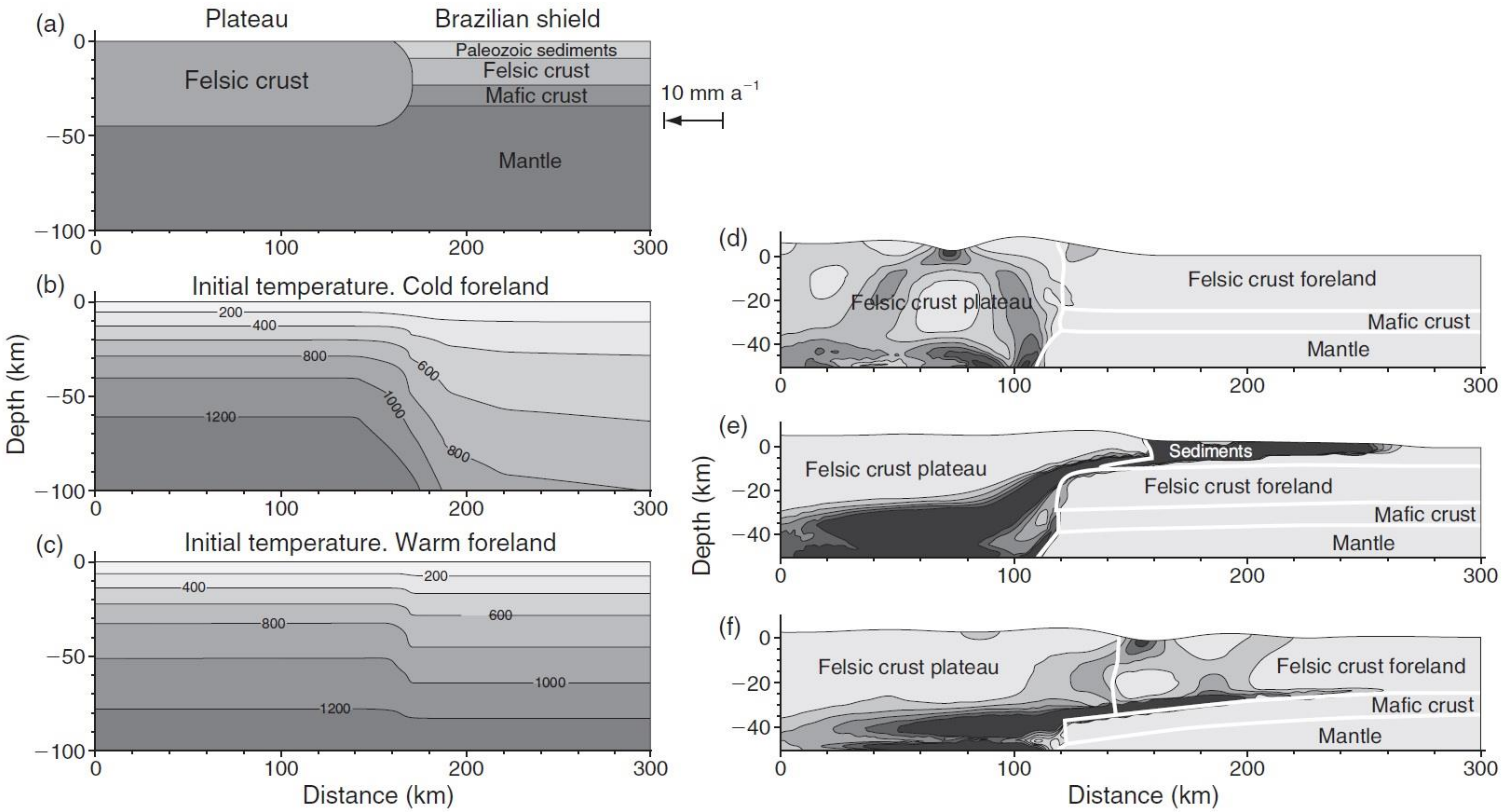
- In *thin-skinned* thrust belts, the lowermost, or basal, décollement separates a laterally displaced sedimentary cover from an underlying basement that is still in its original position.
- In *thick-skinned* styles, the décollement surface cuts down through and involves the crystalline basement.



Modes of shortening in foreland fold-thrust belts

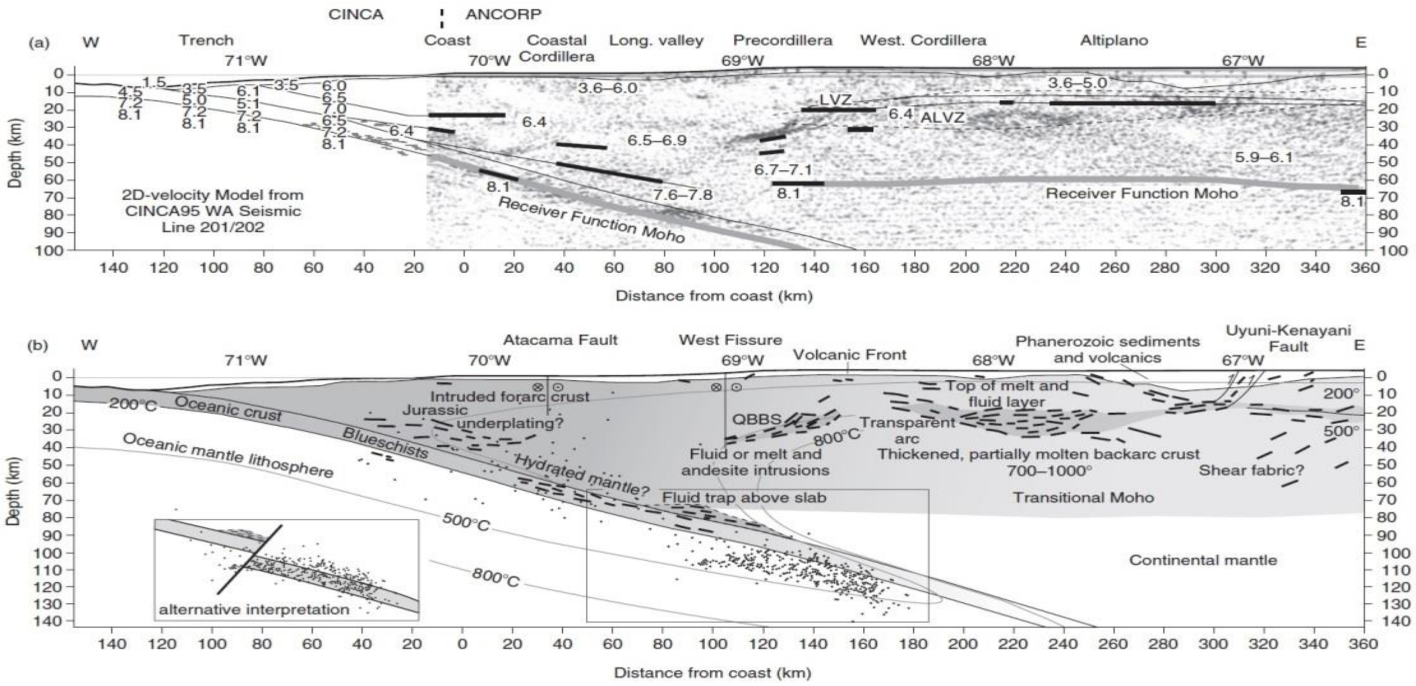
- **Thin-skinned styles** usually occur in regions that have accumulated >3 km of sediment, where the low mechanical strength of the sequences localizes deformation above crystalline basement. If the upper crust is weak and the deep crust is cool and strong, then shortening leads to a mechanical failure of upper crustal sequences and the orogen grows laterally by thin-skinned deformation (thrust wedges show high tapers, asymmetric styles, and rapid lateral growth).
- **Thick-skinned styles** tend to occur in regions where Mesozoic extensional basins have inverted. If the upper crust is strong and the deep crust relatively hot and weak, then shortening may localize into narrow zones and thick-skinned styles of deformation result. A weak middle and lower crust promotes ductile flow and inhibits the lateral growth of the thrust wedge (that usually show low tapers and symmetric style).

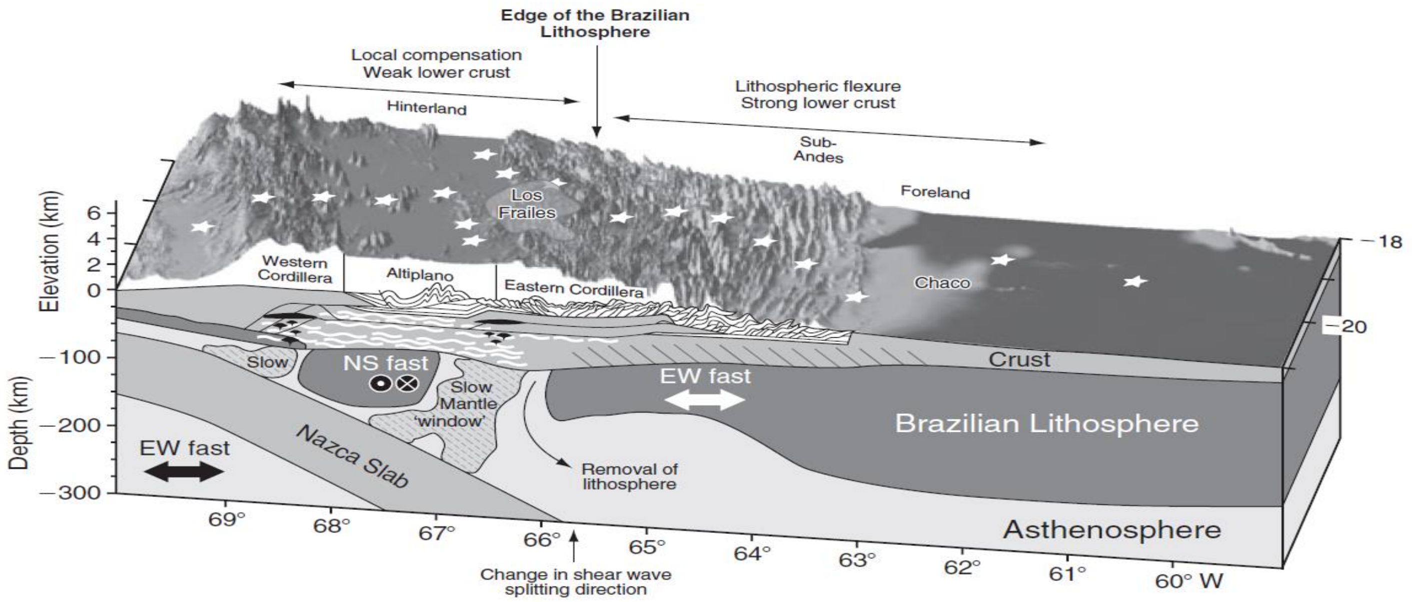
Cold, rigid lithosphere of the Brazilian Shield indented into the adjacent Altiplano-Puna.



a-d) Where the Paleozoic sediments are strong (or absent) and lie on top of a cold strong Brazilian Shield, the crust and mantle deform together homogeneously in pure shear mode.
b-e) Where the Paleozoic sediments are weak and the foreland cold and strong, the foreland displays a simple shear thin-skinned mode of deformation. Underthrusting of the shield is accompanied by the eastward propagation of the thin-skinned thrust belt above a shallow décollement (8-14 depth).
c-f) Where the Paleozoic sediments are weak and the foreland warm and weak, deformation in the foreland is thick-skinned with a deep décollement (~25 km depth).

Deep structure of the central Andes

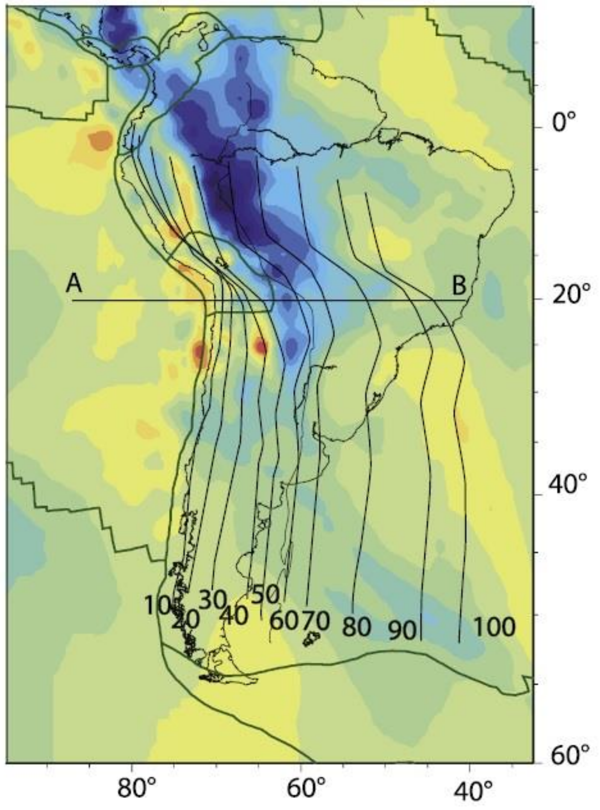


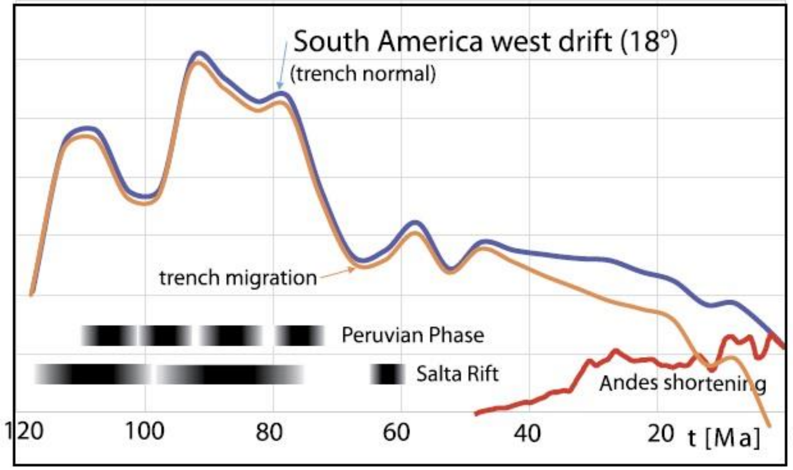


- Parallel to the slab, highly reflective zones indicate the presence of trapped fluids and sheared, hydrated mantle at the top of descending slab.
- Crustal thickness increases from ~35 km to ~70 km beneath the Western Cordillera and Altiplano, where it reaches a maximum of 75 km under its northern part, and decreases to 50 km under the Puna Plateau.
- Within the crust, seismic velocities indicate the presence of a 15- to 20-km-thick zone of low seismic wave speeds at depths of 14–30 km beneath the Western Cordillera and Altiplano-Puna.
- Average crustal *Vp/Vs* ratios of 1.77 beneath the plateau and max values of 1.80–1.85 beneath the active volcanic arc suggest the presence of high crustal temperatures and widespread intra-crustal melting (weak crust).
- The lithosphere is 100–150 km thick below the Altiplano and several tens of km thinner beneath the Puna plateau (explaining the high elevation).

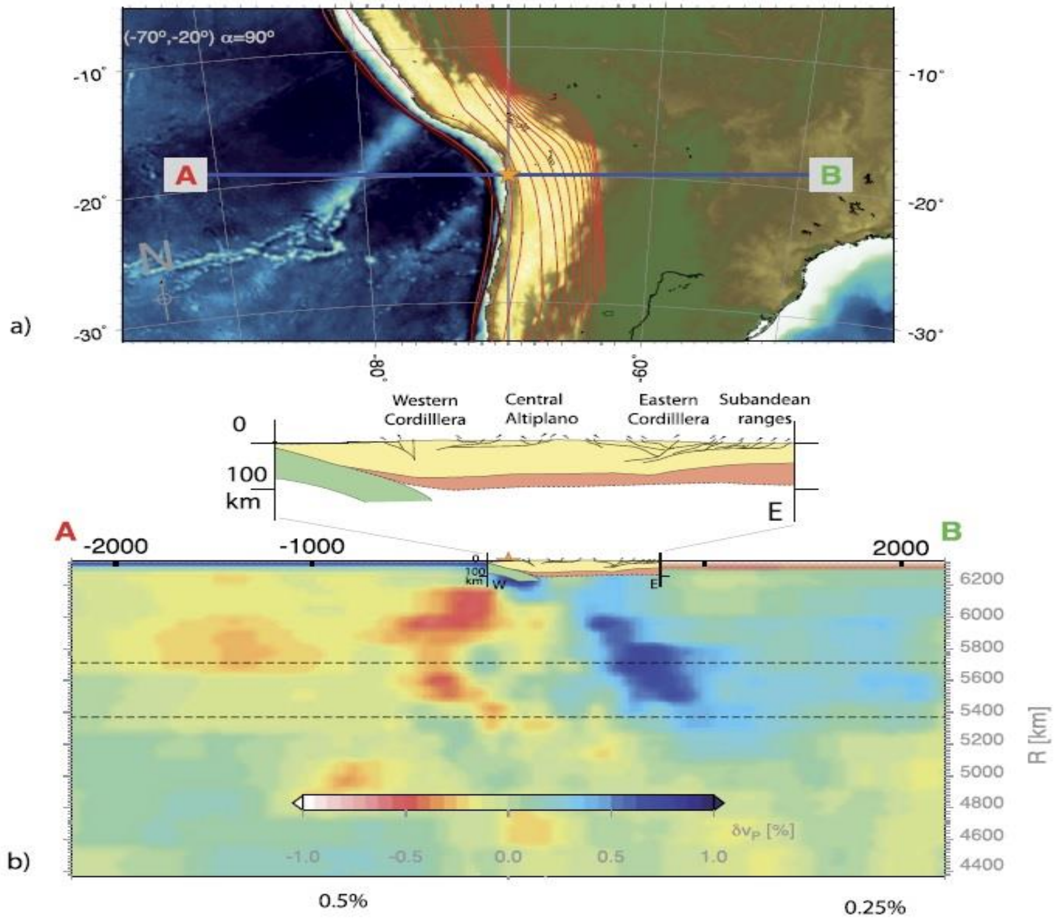
Seismic Tomography and Numerical Modelling

900 Km MIT08 Model





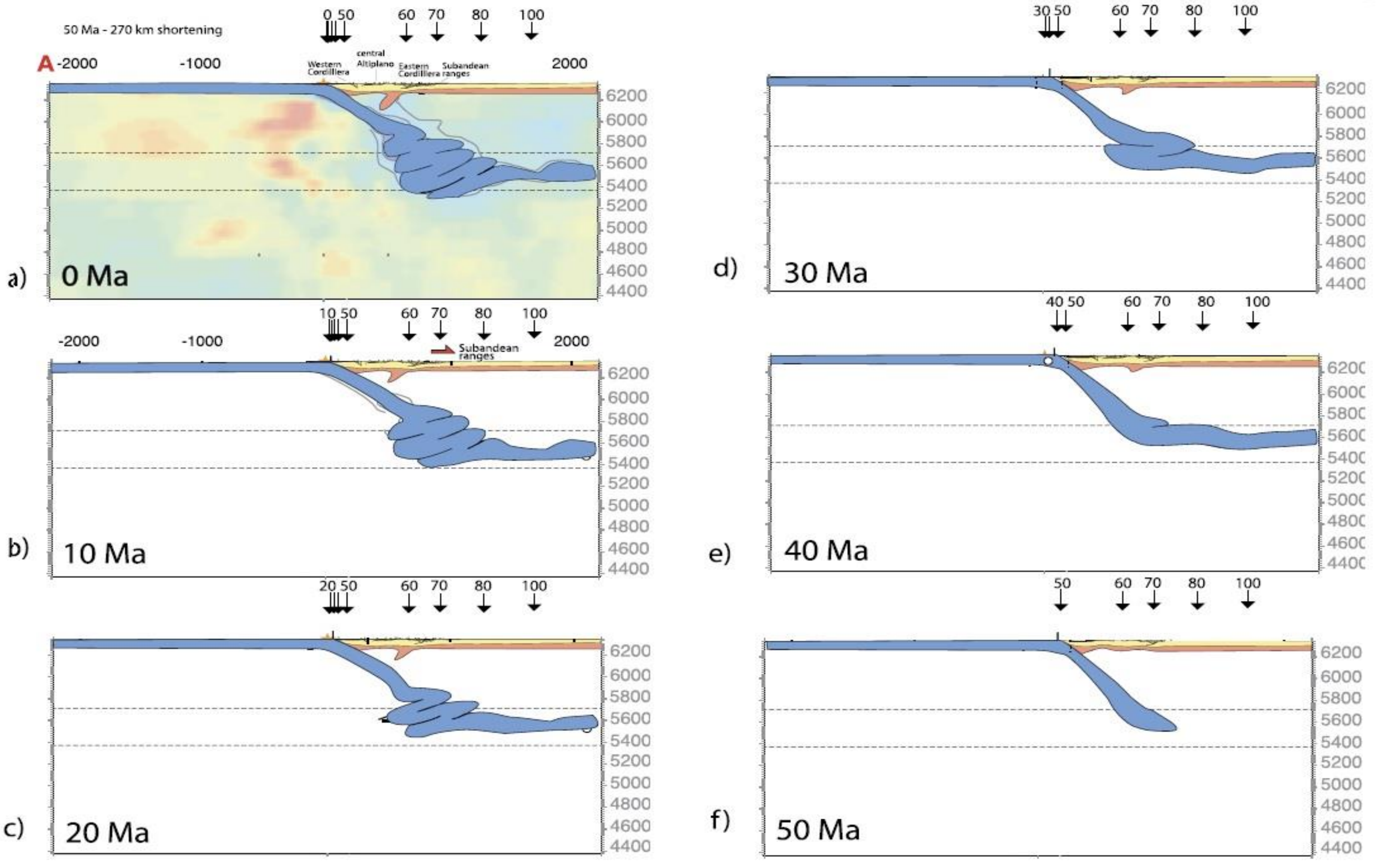
Faccenna et al., 2017, EPSL, 463



- The westward motion of South America sharply decreased at ~75–80 Myr from ~5 to ~2.5–3 cm/yr, and then from 40 Myr progressively decreased down to the present-day rate of ~1cm/yr.

Model of subduction system

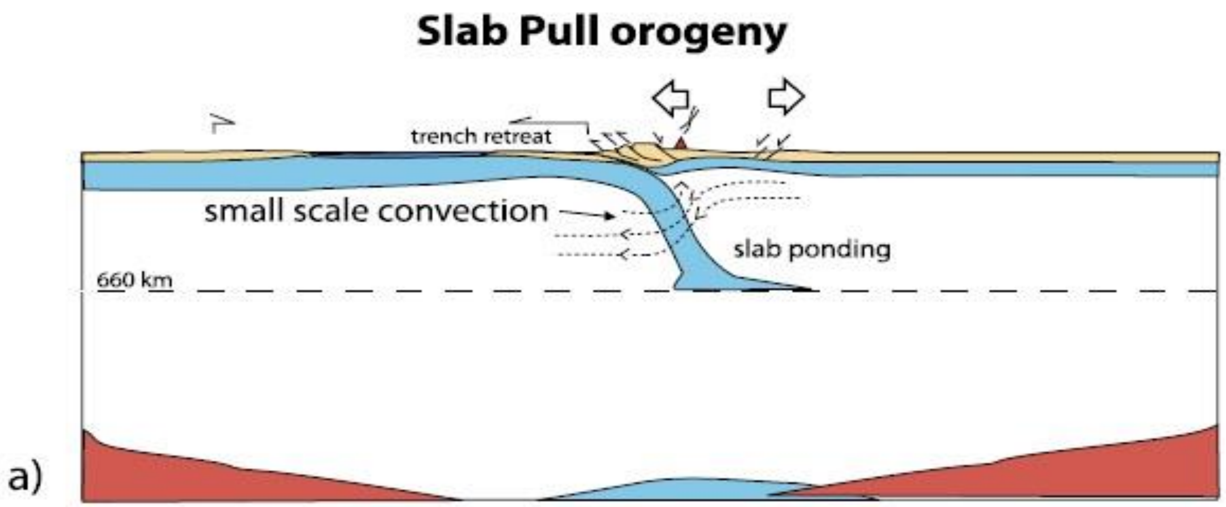
Reconstruction of subduction system along a 20°S section



Black arrows show trench position in time

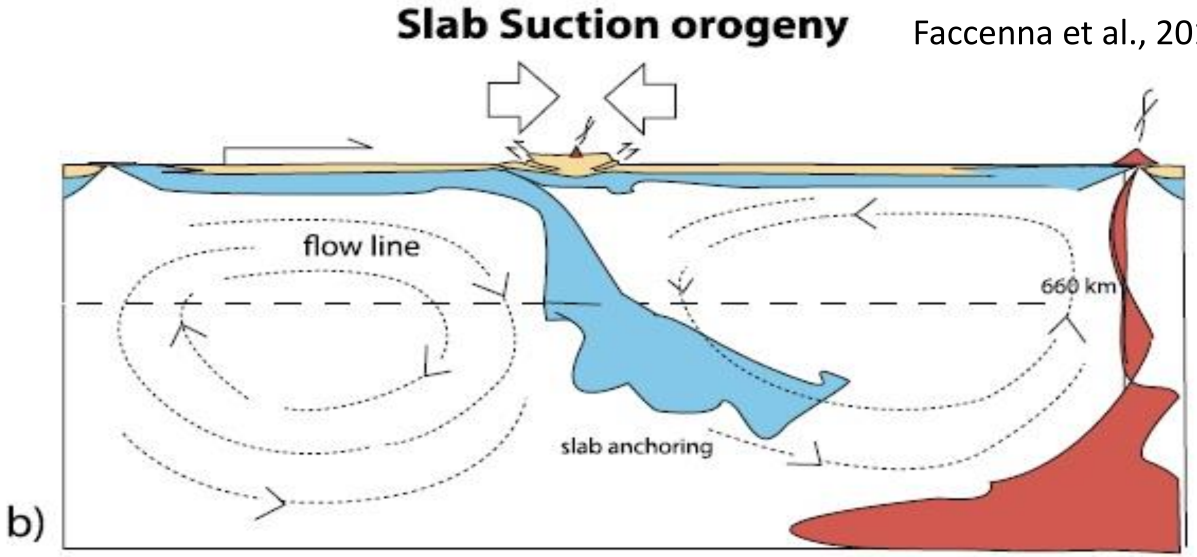
Faccenna et al., 2017, EPSL, 463

Model of orogenic buildup



(a) orogenic wedge grows asymmetrically due to the accretion of crustal slices. Subduction is confined in the upper mantle favoring trench rollback and extension of the upper plate (no crustal thickening).

Faccenna et al., 2017, EPSL, 463



(b) orogenic growth with crustal thickening formed by advancing or stationary trench against a stationary or advancing upper plate. Slab downwelling into the lower mantle generates a suction force that drag plates in collision.

The transition between those two styles of orogeny is ultimately controlled by the capability of the slab to penetrate into the lower mantle. The main orogenic phase (onset of shortening) started at ~50 Myr, after a long phase of westward drift of South America.

Mechanisms of formation of non-collisional orogens

- **Orogenesis at ocean–continent convergent margins initiates when: (i) the upper continental plate is thrown into compression and (ii) the converging plates are sufficiently coupled to allow compressional stresses to be transmitted into the interior of the upper plate:** High convergence rates and the underthrusting of young, thick, and/or buoyant lithosphere causes (1) compression, (2) decrease slab dips, and (3) enhance the transfer of compressional stress.
- However, from the Altiplano region northward and southward, there is a decrease in the total amount of crustal shortening and thickening with no direct correspondence to either the slab age or the convergence rate

Other factors control shortening and thickening variations of the Andes:

1. Strength of inter-plate coupling at the trench
2. Internal structure and rheology of the continental plate

In the backarc-foreland domain, deformation is controlled by the absolute velocity of the continental plate, its rheology, and the strength of inter-plate coupling at the trench.

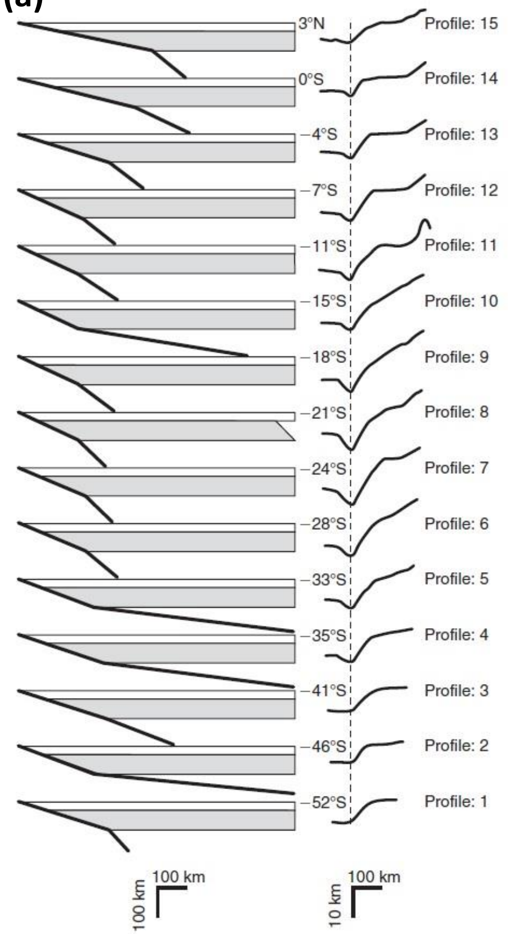
- The strength of inter-plate coupling along the Peru–Chile Trench is controlled by the rate and age of subducting lithosphere and the amount of surface erosion and deposition: A dry, sediment-starved trench may result in a high degree of friction along the Nazca–South American plate interface, increasing shear stress, leads to increased compression and uplift in the central Andes, while the presence of weak sediments in the Southern Andes produce opposite conditions.

Strength of inter-plate coupling at the trench

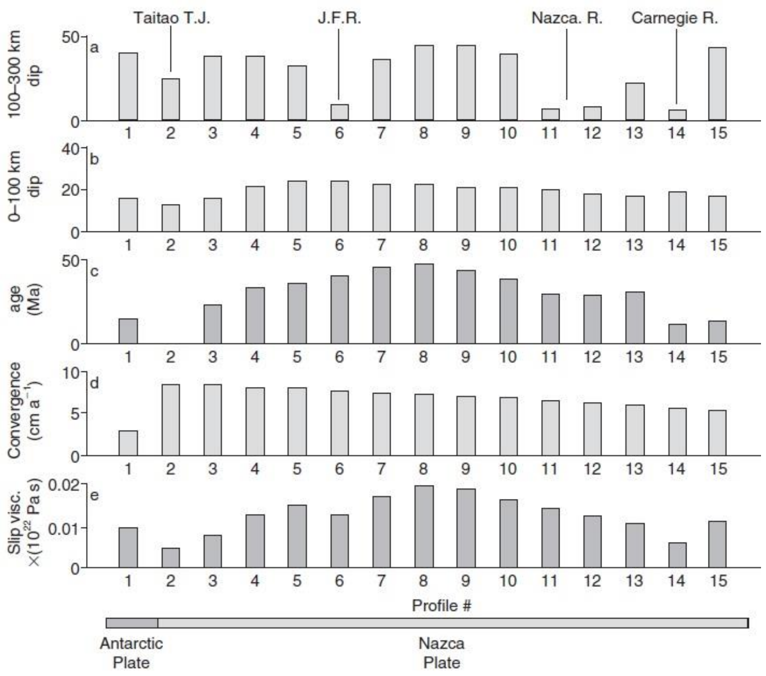
Empirical relationship between trench topography and the degree of coupling across the slipping interface was derived using along strike variations in the shape of the inner trench slope:

- Buoyancy forces associated with continental crust dominate the force balance if the strength of the plate interface is low, resulting in an upward movement of the forearc.
- Tectonic forces associated with the sinking of oceanic lithosphere dominate if the strength of the plate interface is high, causing downward movement of the forearc.

(a)



(b)



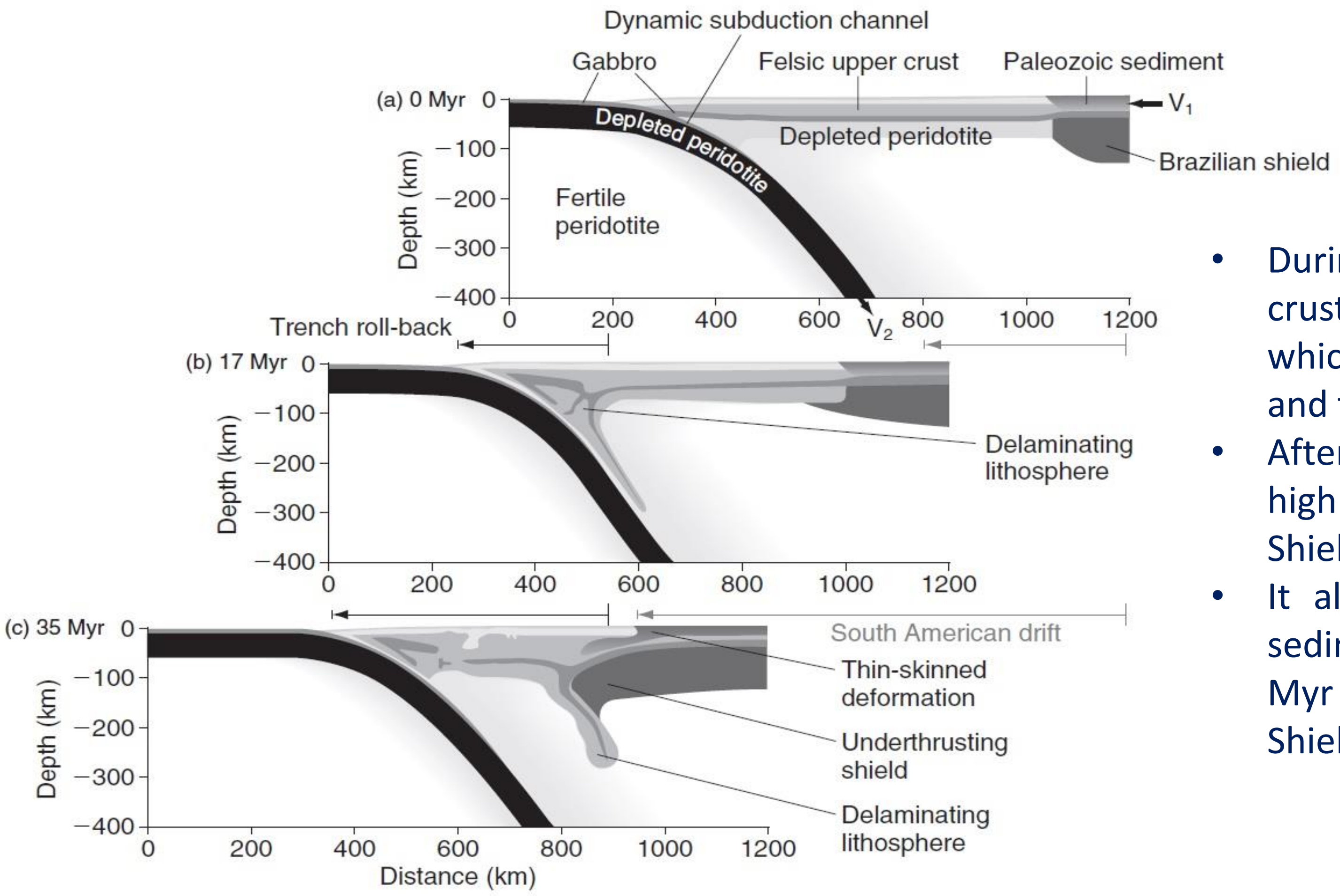
- Strongest inter-plate coupling occurs in the central Andes near latitude 21°S, where inner trench slopes are steepest and the age of subducted lithosphere is the oldest.
- Weak coupling occurs in the southern Andes south of 35°S, where opposite conditions are observed.

(a) Trench topography for 15 profiles of the Andes between 3°N and 56°S showing dip of the Benioff zone and bathymetric profiles (b) Model results showing slab dip under asthenospheric wedge, near-trench slab dip angle, age of subducting slab at the trench, convergence velocity, slip layer viscosity for a layer of 10 km thickness.

Strong coupling results in large amounts of compression in the backarc, which increases crustal shortening and thickening.

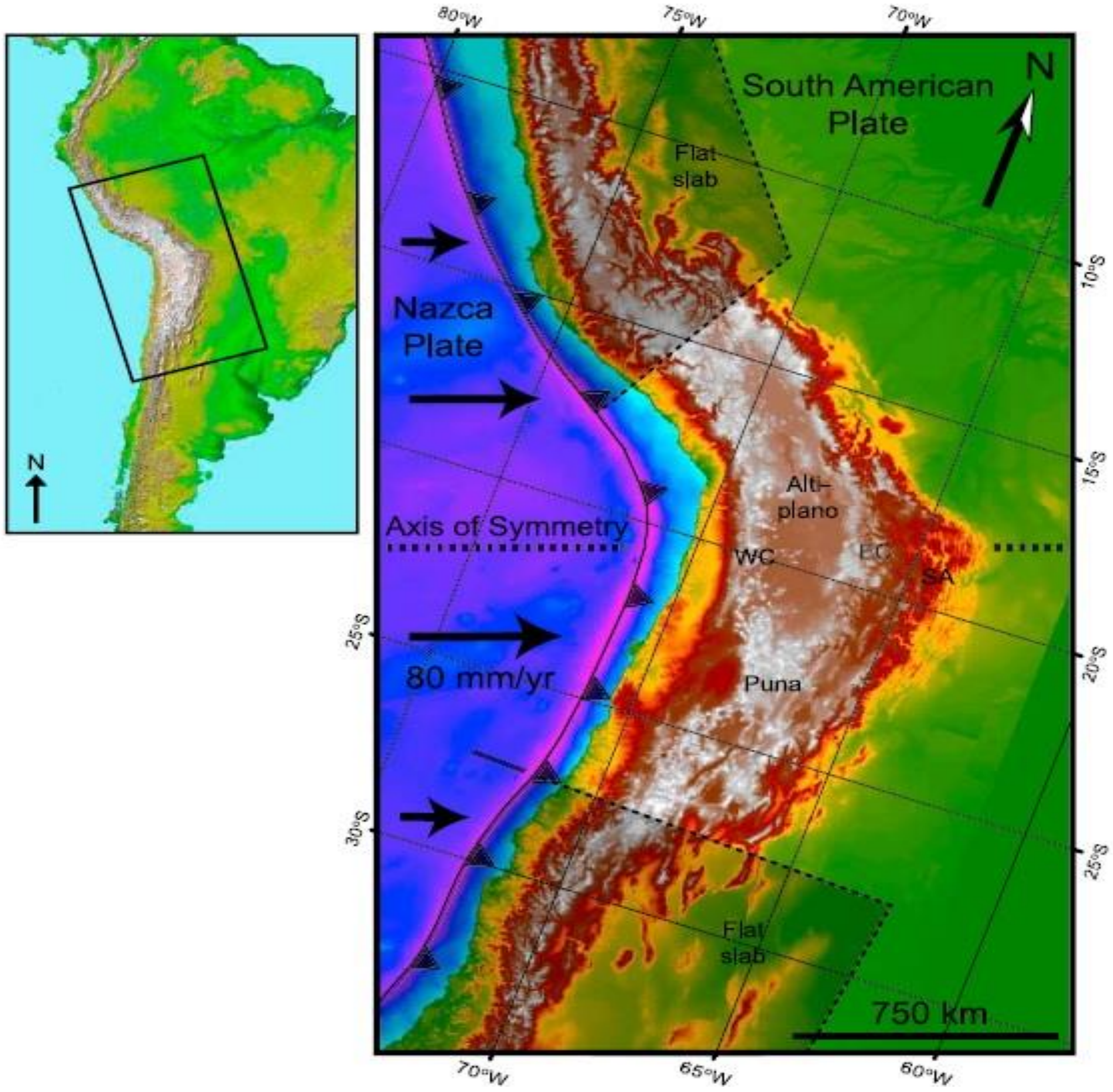
Structure and rheology of continental plate

Initial Conidions: (1) The central Andes involve a thick felsic upper crust, a thin gabbroic lower crust, and a total thickness of 40–45 km, already shortened prior to the start of deformation at 30–35 Myr; **(2)** The southern Andes consist of an upper and lower crust of equal thickness and a total crustal thickness of 35–40 km; **(3)** Subduction initially occurs at a low angle below a 100- to 130-km-thick continental lithosphere.



- During shortening the crustal thickness doubles while the lower crust and mantle lithosphere become thinner by delamination, which causes an increase of Moho *T*, leading to crustal wakening and flow.
- After 20–25 Myr in model time, tectonic shortening generates high topography between the magmatic arc and the Brazilian Shield.
- It also occurs mechanical failure of the wedge of Paleozoic sediments by thin skinned thrust faulting in the foreland at 25 Myr model time, followed by underthrusting of the Brazilian Shield under the plateau.

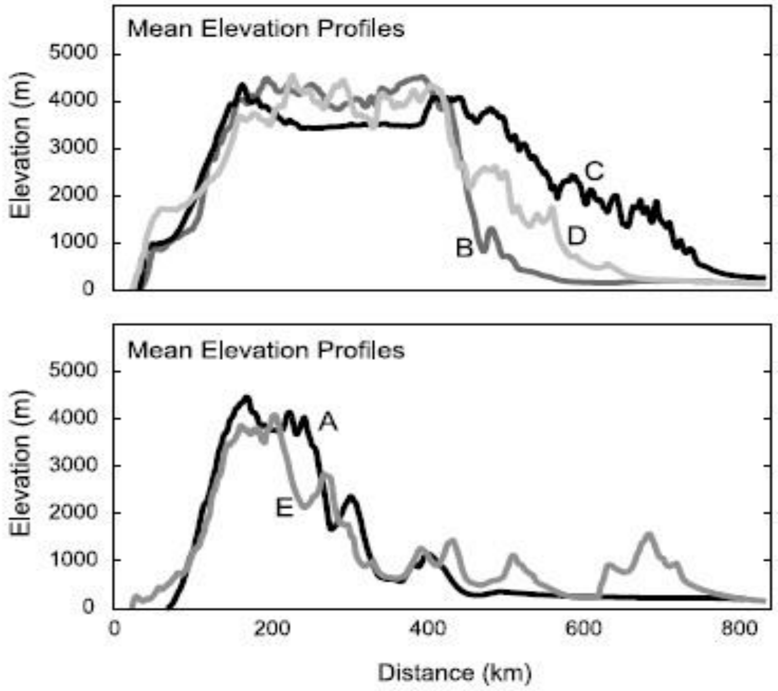
Topography of the Andes

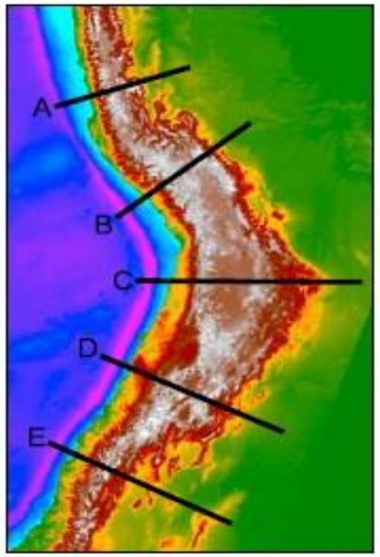


Ouimet and Cook, 2010, Tectonics, 29

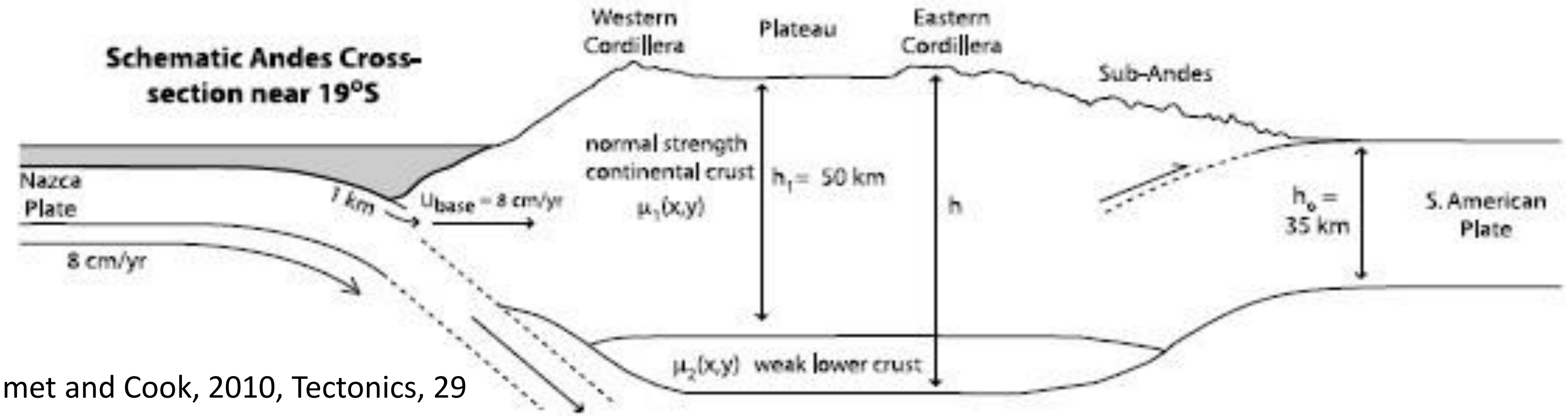
- The highest and widest parts of the orogen are in the central Andes (between 12°S and 30°S), with a topography of 3.5–4 km persisting over an area of ~600,000 km^2, a maximum width of the order of 600–700 km, and crustal thicknesses as large as 60–70 km.
- In the northern Altiplano (~14°S) and in the southern Puna (~28°S), the plateau and cordillera are less well defined, indicating a transition to different styles of deformation. This is likely a result from the presence of flat slab subduction beneath the northern and southern parts of the orogen.
- From 2°S to ~14°S and 28°S to 34°S, the Nazca plate subducts shallowly beneath the Andes (at <10°). This contrasts with subduction beneath the central Andes, where the Nazca plate subducts at a steeper angle (~30°).

Topography of the Andes



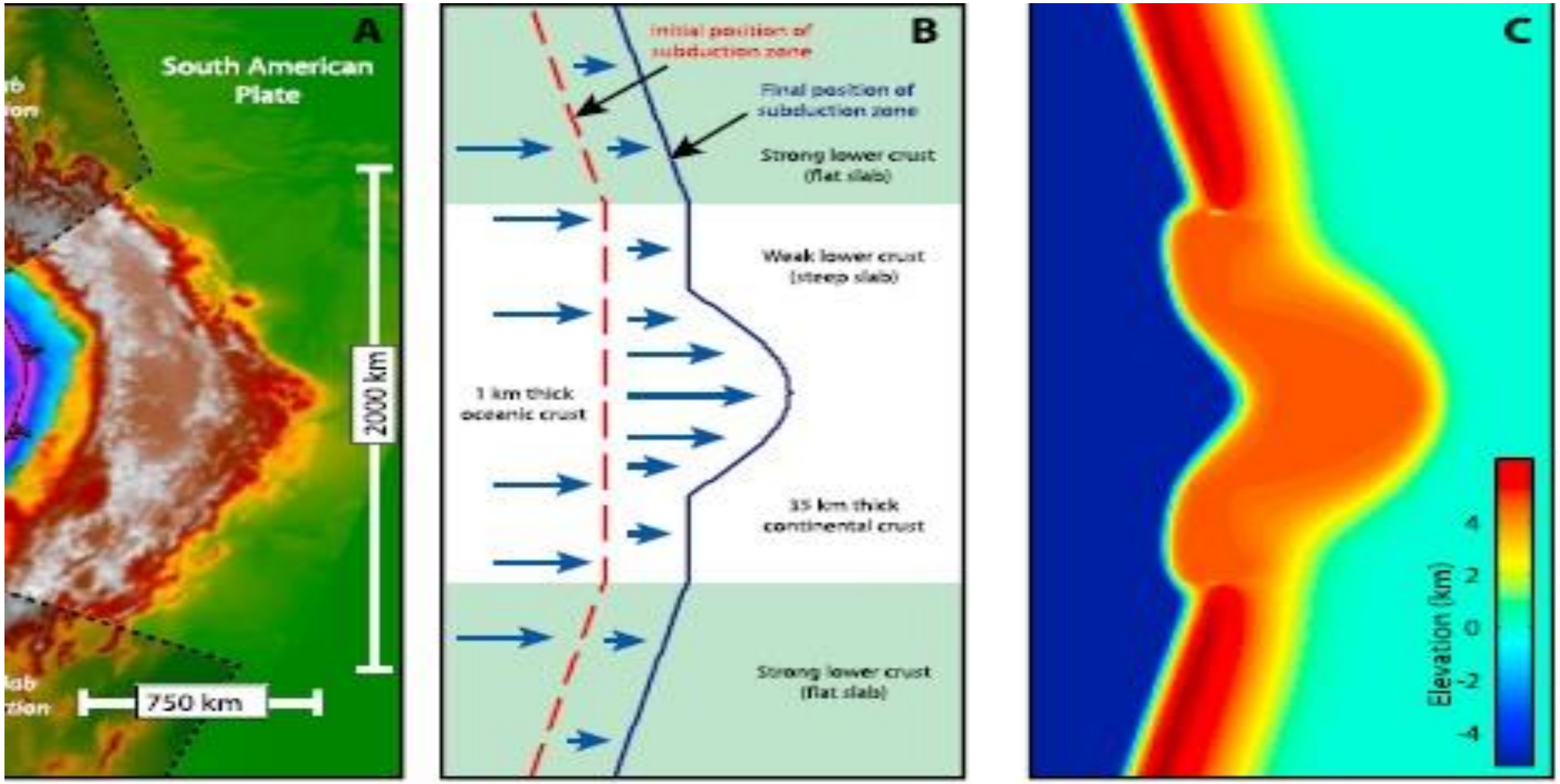


- In the orogenic segment underlain by a more steeply subducting slab, the viscosity of the lower crust begins to drop when the crust reaches a thickness of 50 km and arrives at its minimum value of 10^{17} Pa s when the crust reaches a thickness of 65 km.
- In the orogenic segments underlain by flat slab subduction (north of 14°S and south of 28°S) the lower crust is not allowed to weaken, and the viscosity remains uniform throughout the crust.
- In the center of the steep slab zone is a narrower oroclinal zone where there is more shortening within the South American plate than to the north and south (more continental lithosphere is incorporated into the orogen from the east).



Ouimet and Cook, 2010, Tectonics, 29

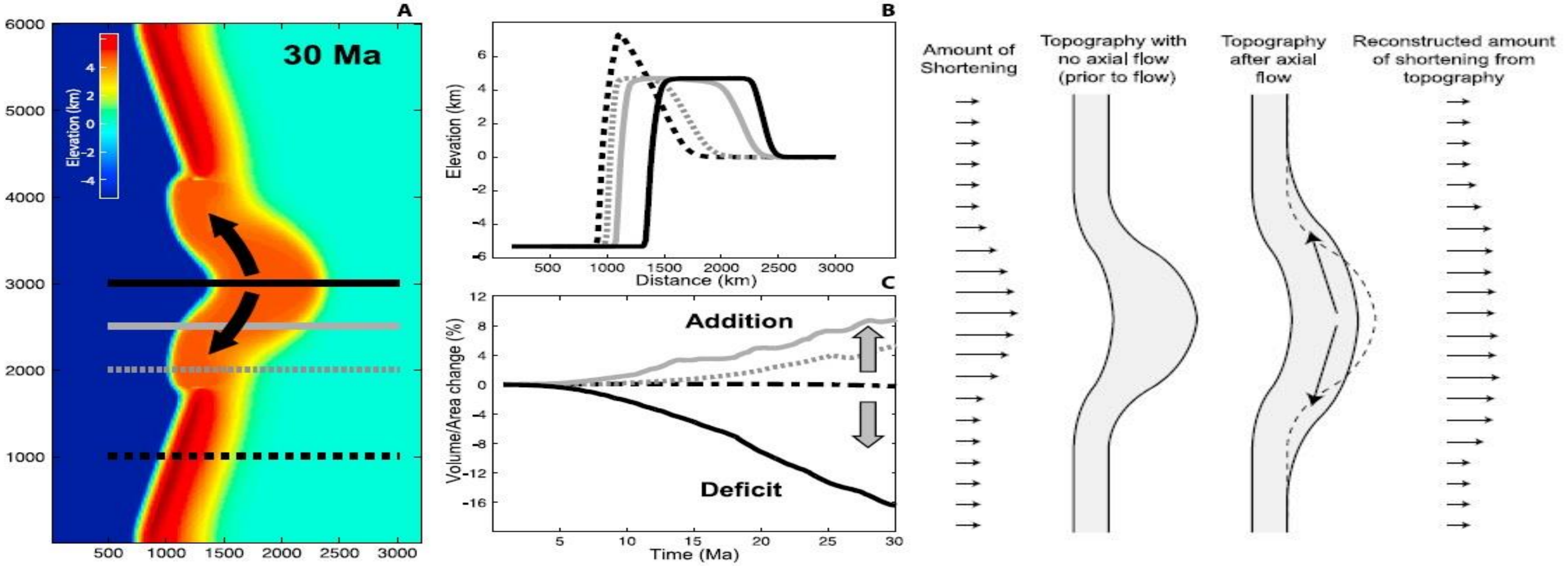
Numerical models boundary conditions



Ouimet and Cook, 2010, Tectonics, 29

- The model produces Andean-like topography provided that (1) the more steeply dipping slab segment beneath the central Andes is overlain by the weak lower or middle crust, (2) the flat slab subduction segments to the north and south of this zone are overlain by the strong middle and lower crust, and (3) the steeply dipping central slab segment is overlain by a narrow, centrally localized zone of increased crustal shortening.
- The axial lower crustal flow that thickens crust north and south of the orocline zone and stops when it reaches the strong flat slab segments, occurs in the later stages of orogen evolution, after significant crustal thickening in the center.

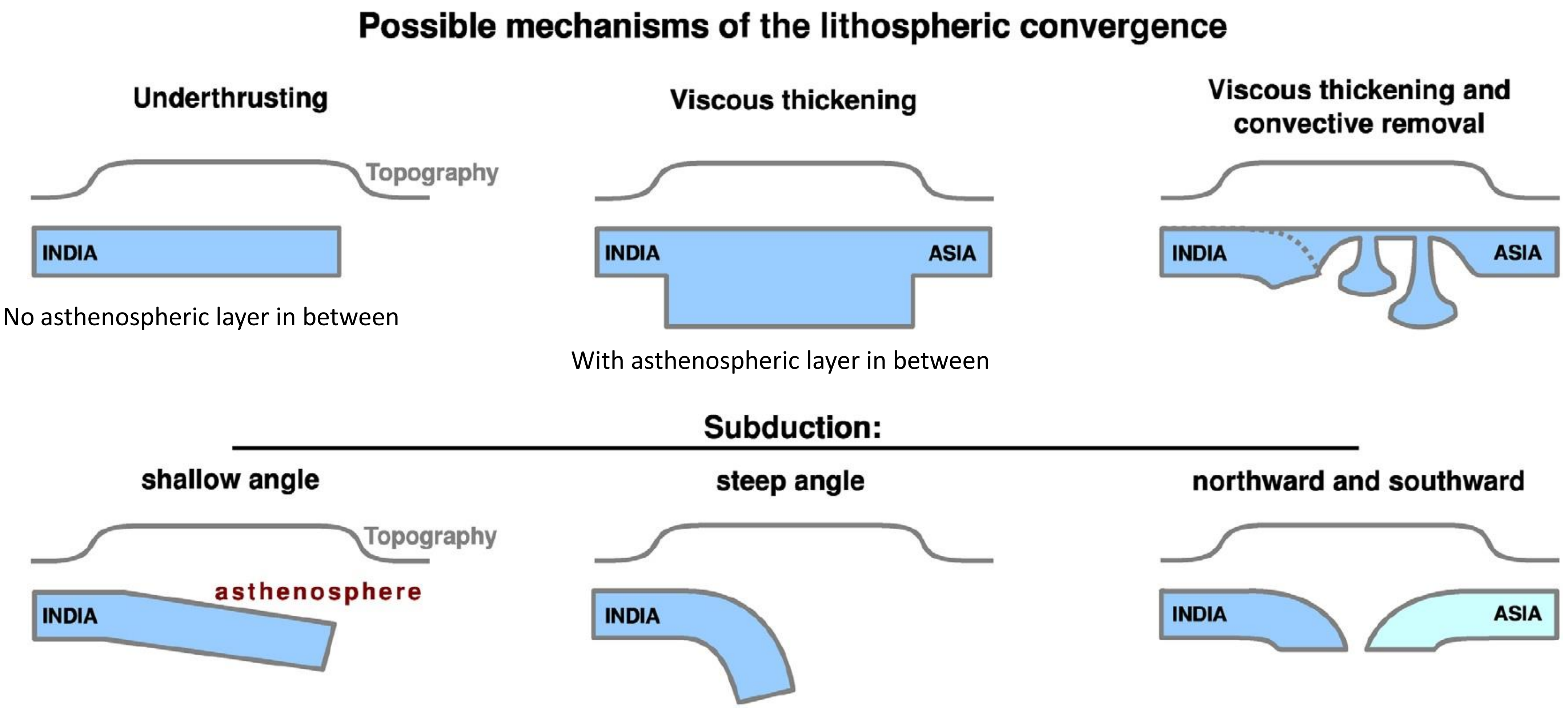
Numerical models results



Ouimet and Cook, 2010, Tectonics, 29

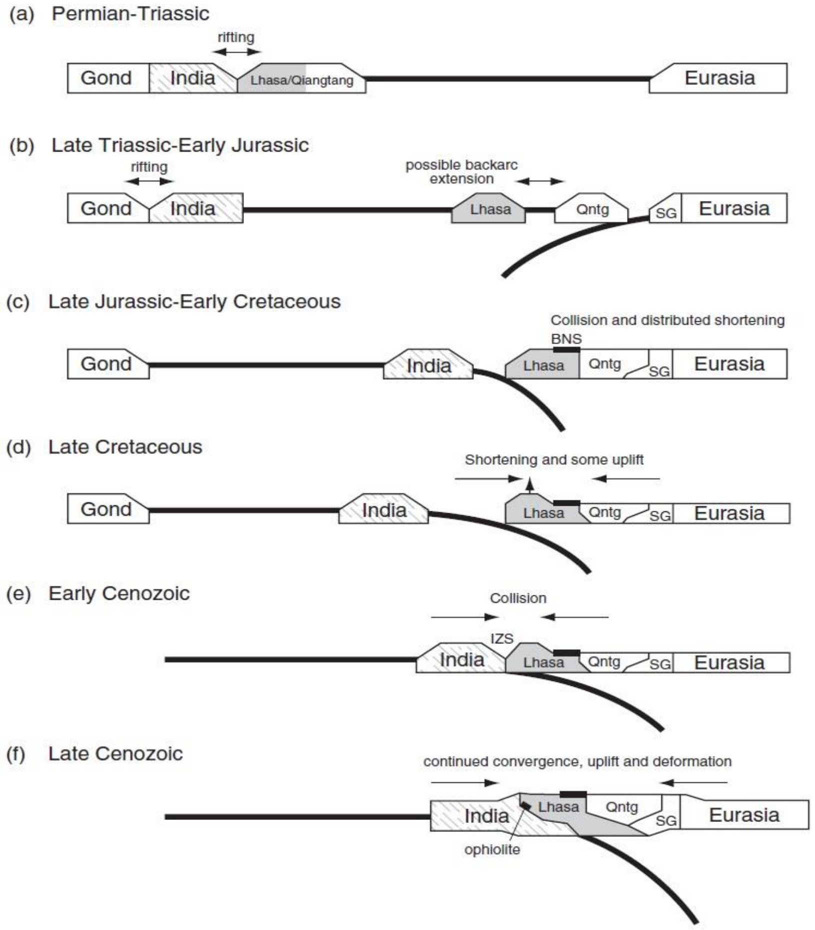
- As lower crust above the more steeply dipping segment of the subducting slab becomes hot and weak, increased shortening in the center of the Andes drives longitudinal flow of lower crust along strike to the north and south.
- Longitudinal flow does not penetrate into the flat slab regions, presumably because the crust there is stronger and the lower crust is unable to flow.

Continent-Continent Collision



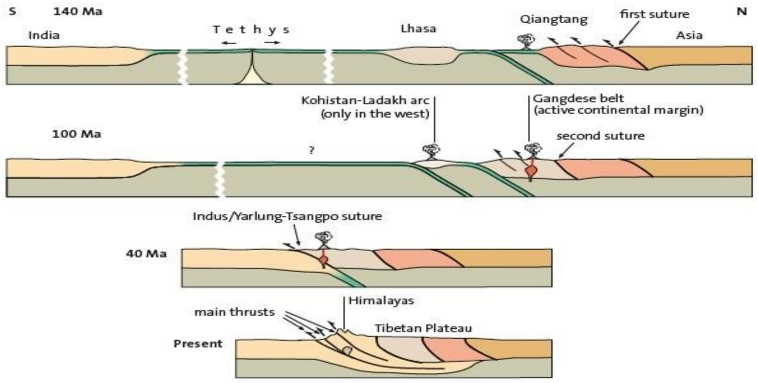
Agius and Lebedev, 2013, G3, 14

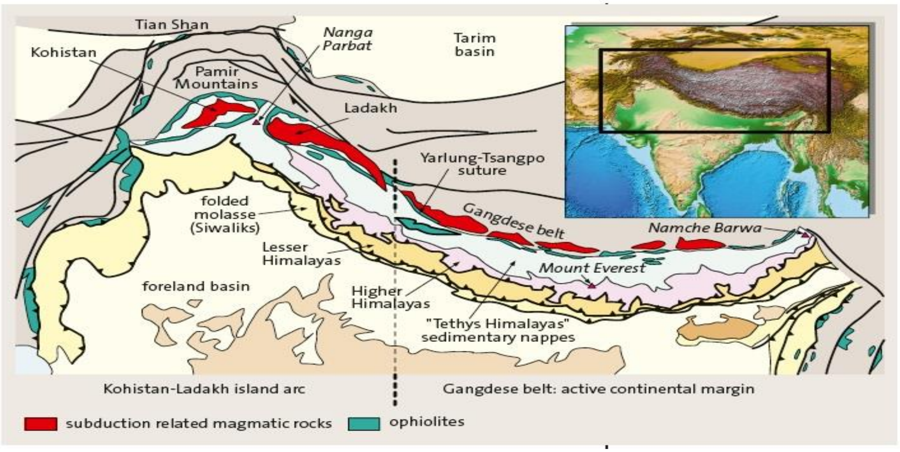
India-Eurasia Collision



Collisional history

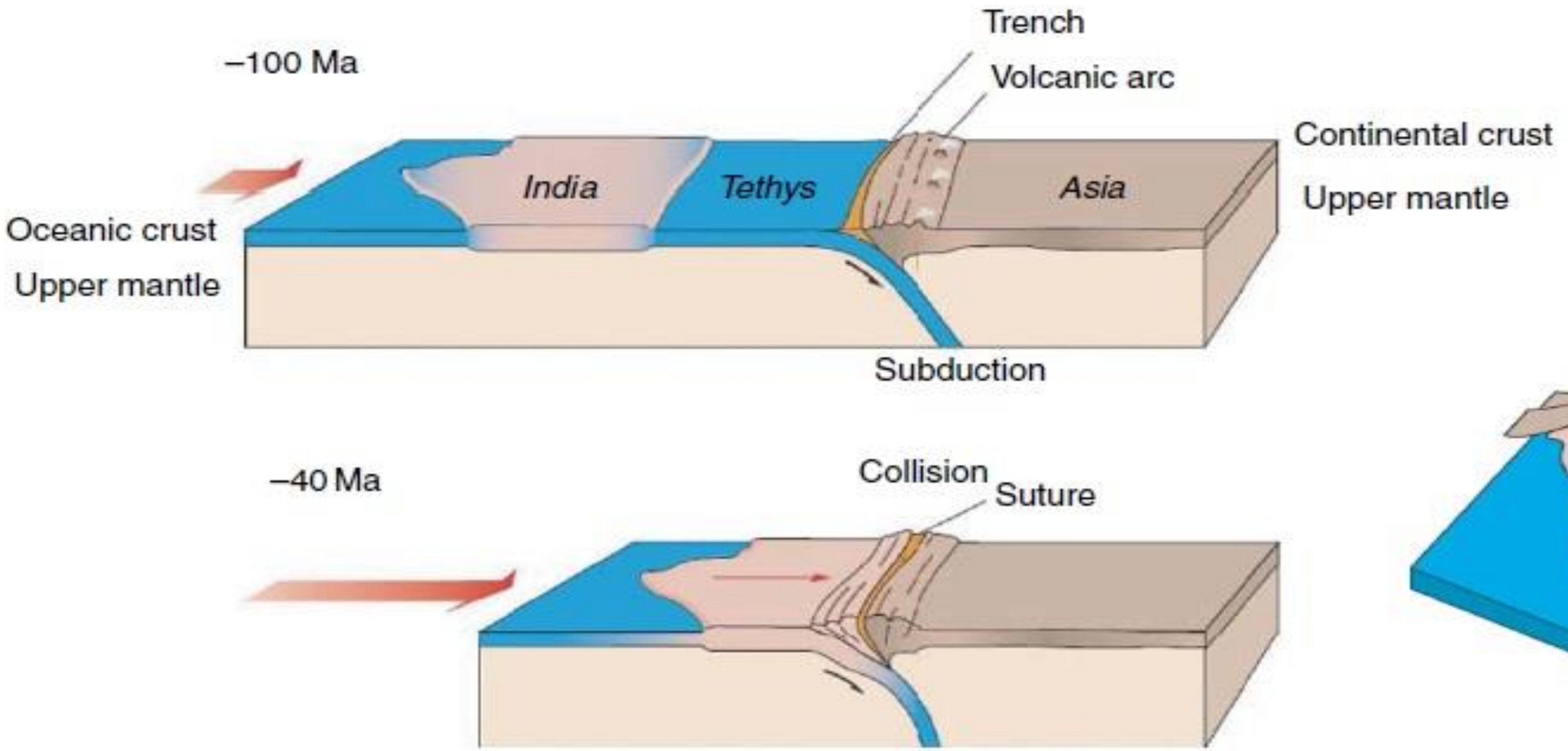
- Long history of subduction, arc magmatism, terrane accretion, and crustal thickening along the southern margin of Eurasia weakened its lithosphere, allowing deformation.
- Initial contact between some parts of India and Asia could have occurred as early as 70 Myr, all Tethyan oceanic lithosphere had disappeared by 45 Myr and at ~36 Myr.
- Strong Indian plate (Precambrian Shield) resisted shortening during collision and favored its underthrusting. Sediments along the north Indian margin were scraped off forming the Himalaya.
- The Himalayan–Tibetan orogen is built upon a collage of exotic material that became welded to the Eurasian Plate before the main India–Eurasia collision (accretionary orogen).

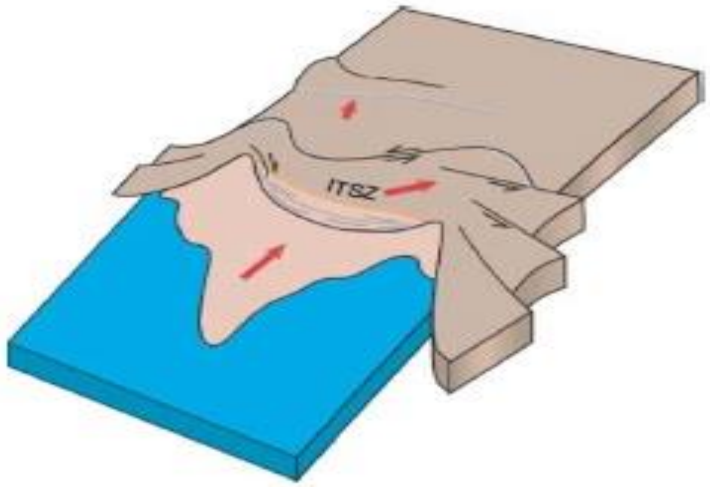




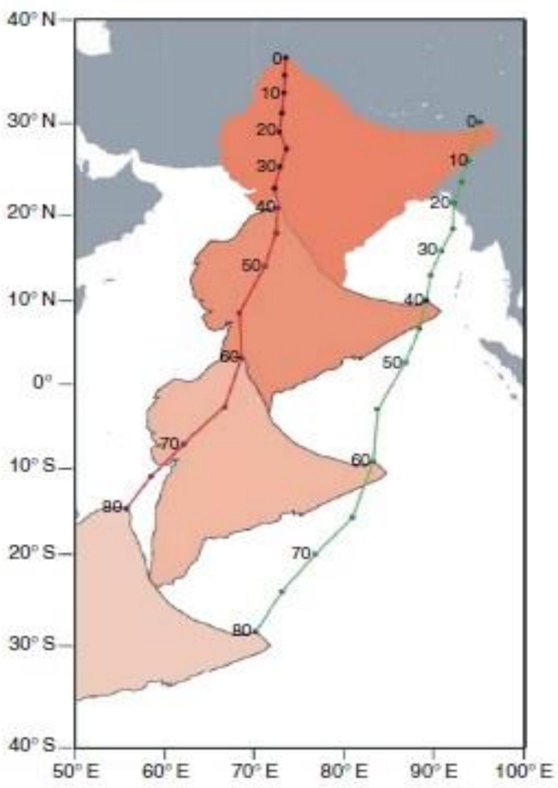
India-Eurasia Collision

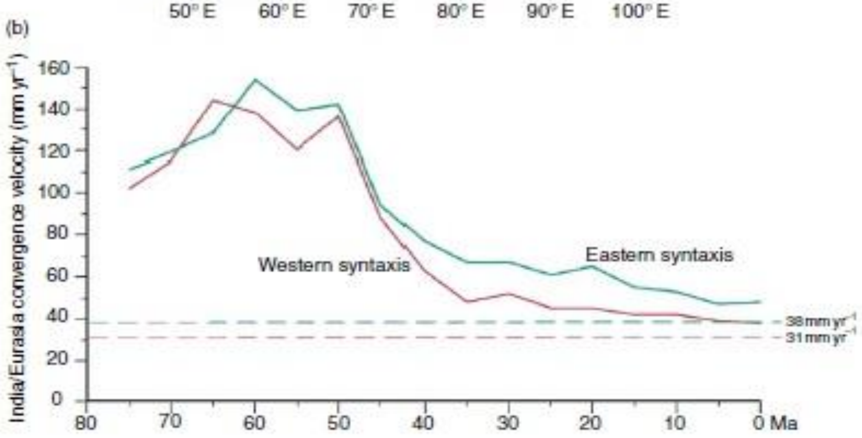
- The Himalayan-Tibetan orogen was created over the past 50-70 Myr.
- Collision reduced the speed from 100mm/yr to 50 mm/yr in the last 40 Myr.
- The motion caused the indentation of India into Asia of about 2000 km and a zone of active deformation, stretching 3000 km north of the Himalaya.

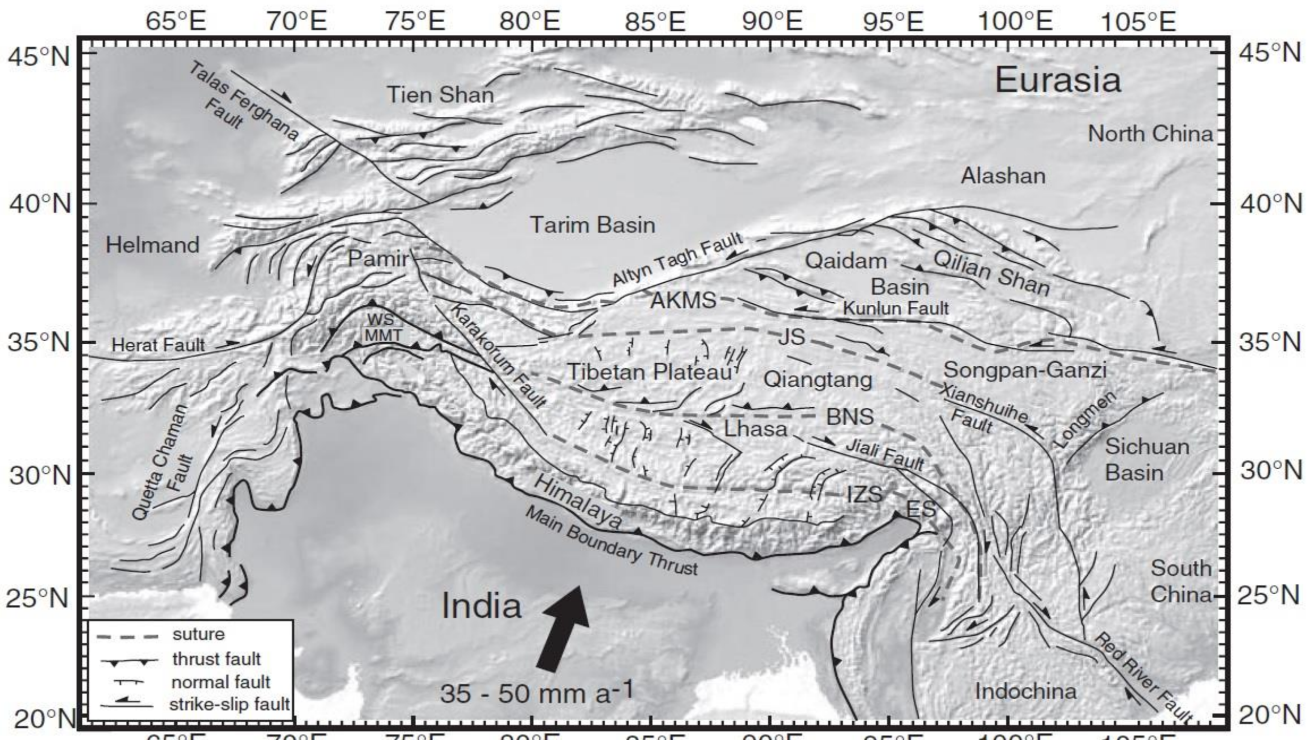




India-Eurasia Collision





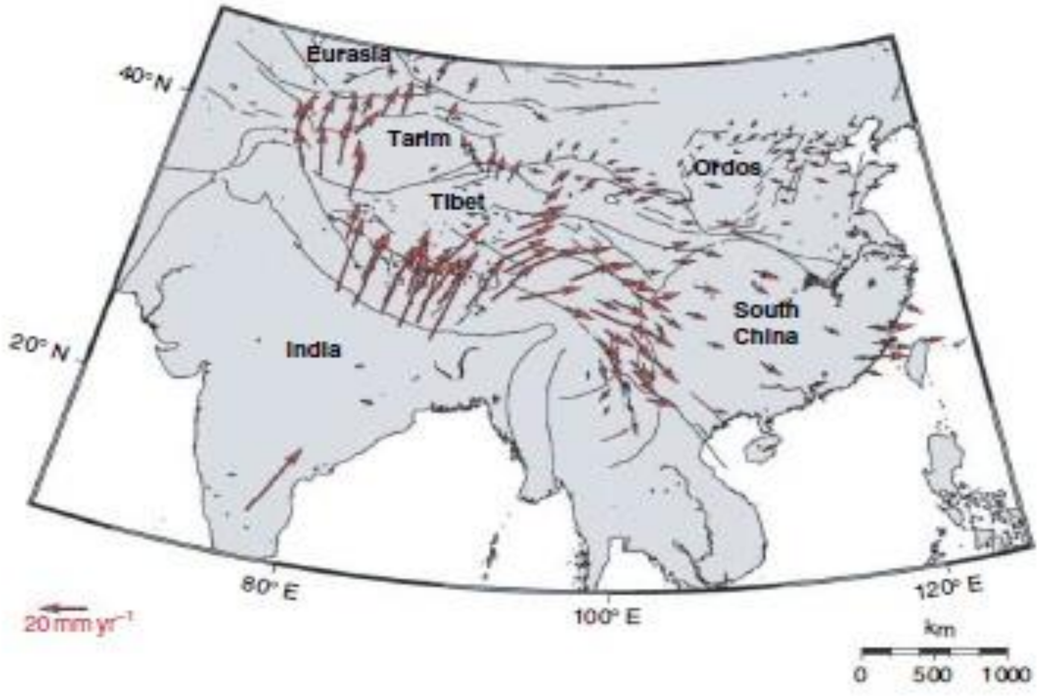


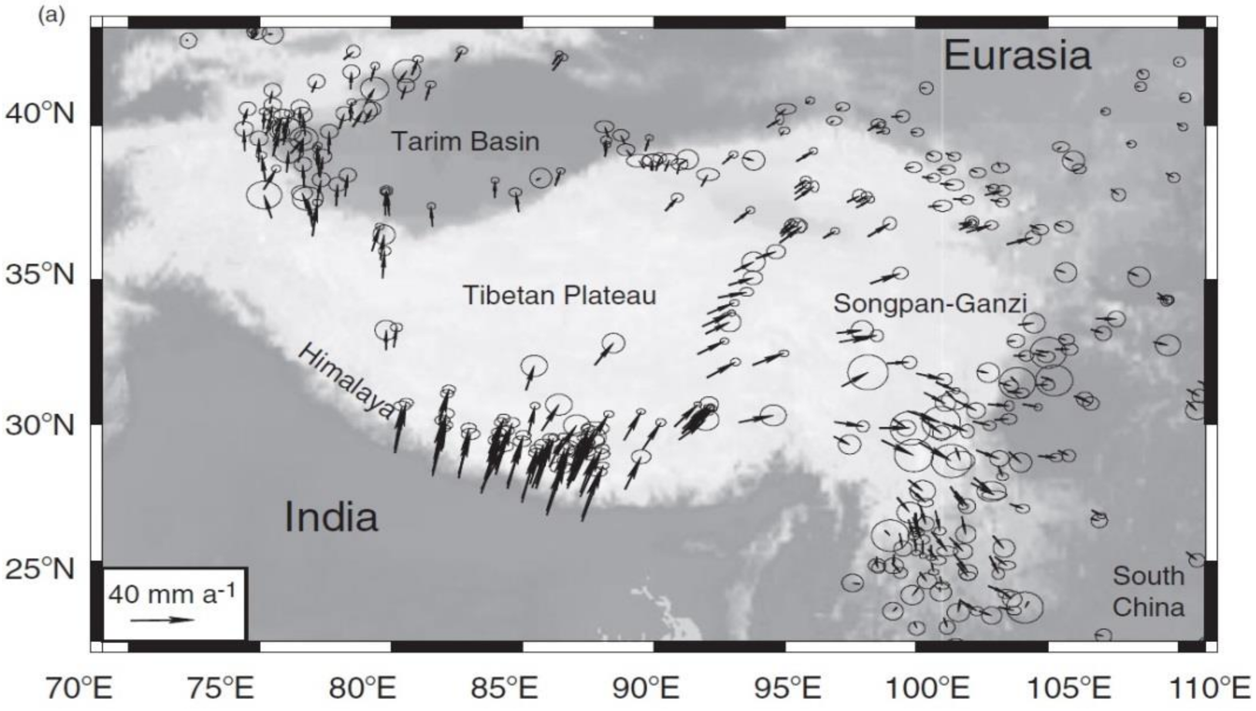
WS, Western Himalayan Syntaxis; ES, Eastern Himalayan Syntaxis; MMT, Main Mantle Thrust; AKMS, Ayimaqin–Kunlun–Mutztagh suture; JS, Jinsha suture; BNS, Bangong–Nujiang suture; IZS, Indus–Zangbo suture.

- India underwent a counterclockwise rotation to close the remaining part of the Neotethys
- The spreading of the mid-ocean ridge laying between India and Australia stopped and the two continent fused into a single plate.

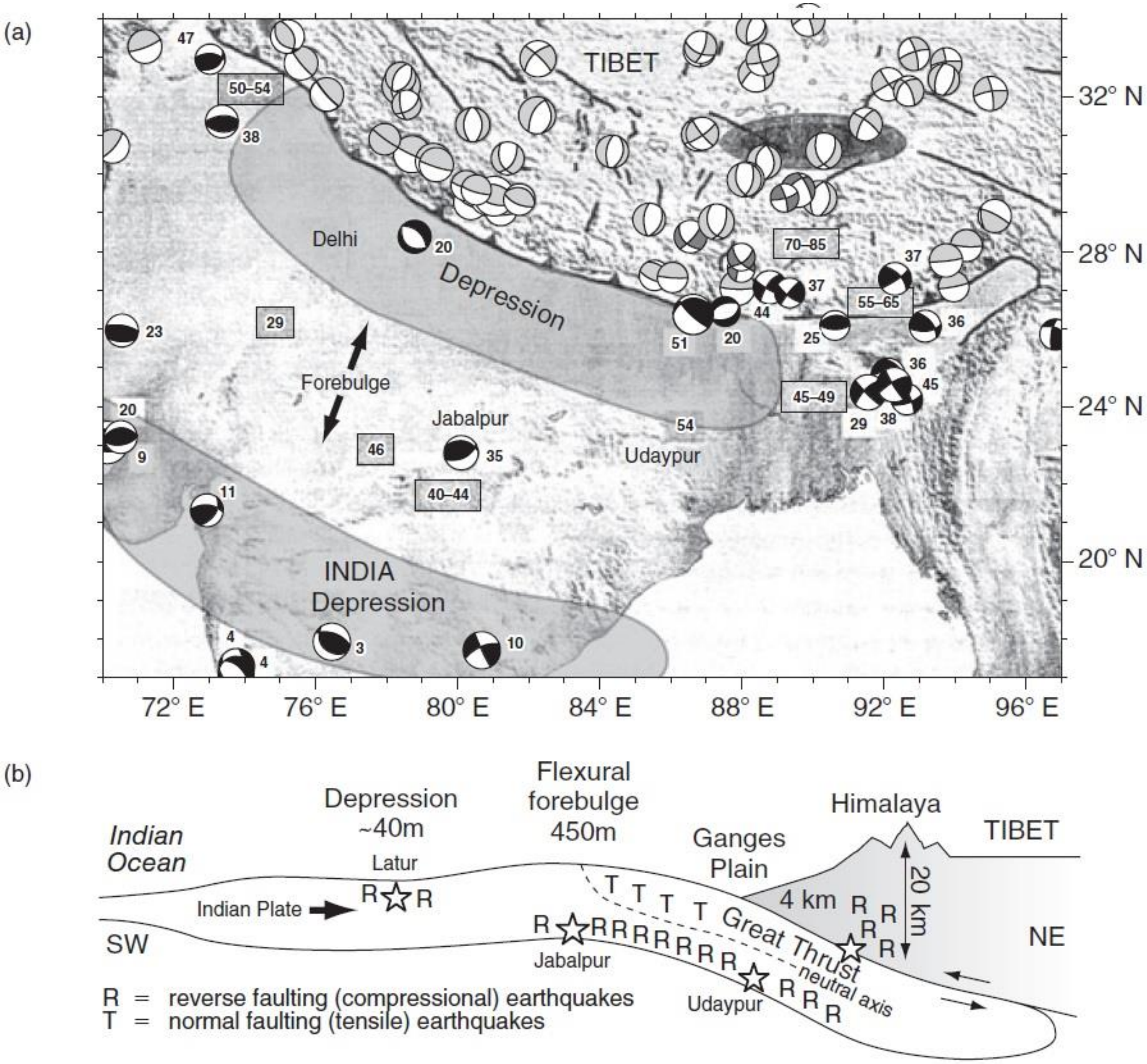
GPS velocities and seismic anisotropy

- India moves to the northeast with a rate of 35-38 mmm/yr relative to Siberia.
- Tibetan Plateau absorbs more than 90 % of the relative motion and is broadly distributed (GPS velocity mostly linear parallel to the direction of India-Eurasia collision, N21°E).
- Additional component of shortening is accommodated in Pamir, Tien Shan, and Qilian Shan.
- The mostly linear trend of GPS velocity suggests that the shortening across the plateau is broadly distributed (no deviations across individual fault zones).
- North China and South China are moving ESE at rates of 2-8 mm/yr and 6-11 mm/yr relative to Eurasia.
- South of the Kunlun Fault the Tibetan Plateau extrudes eastward relative to both India and Asia (lateral escape).





Lithospheric Structure of the Indo-Eurasia collision zone

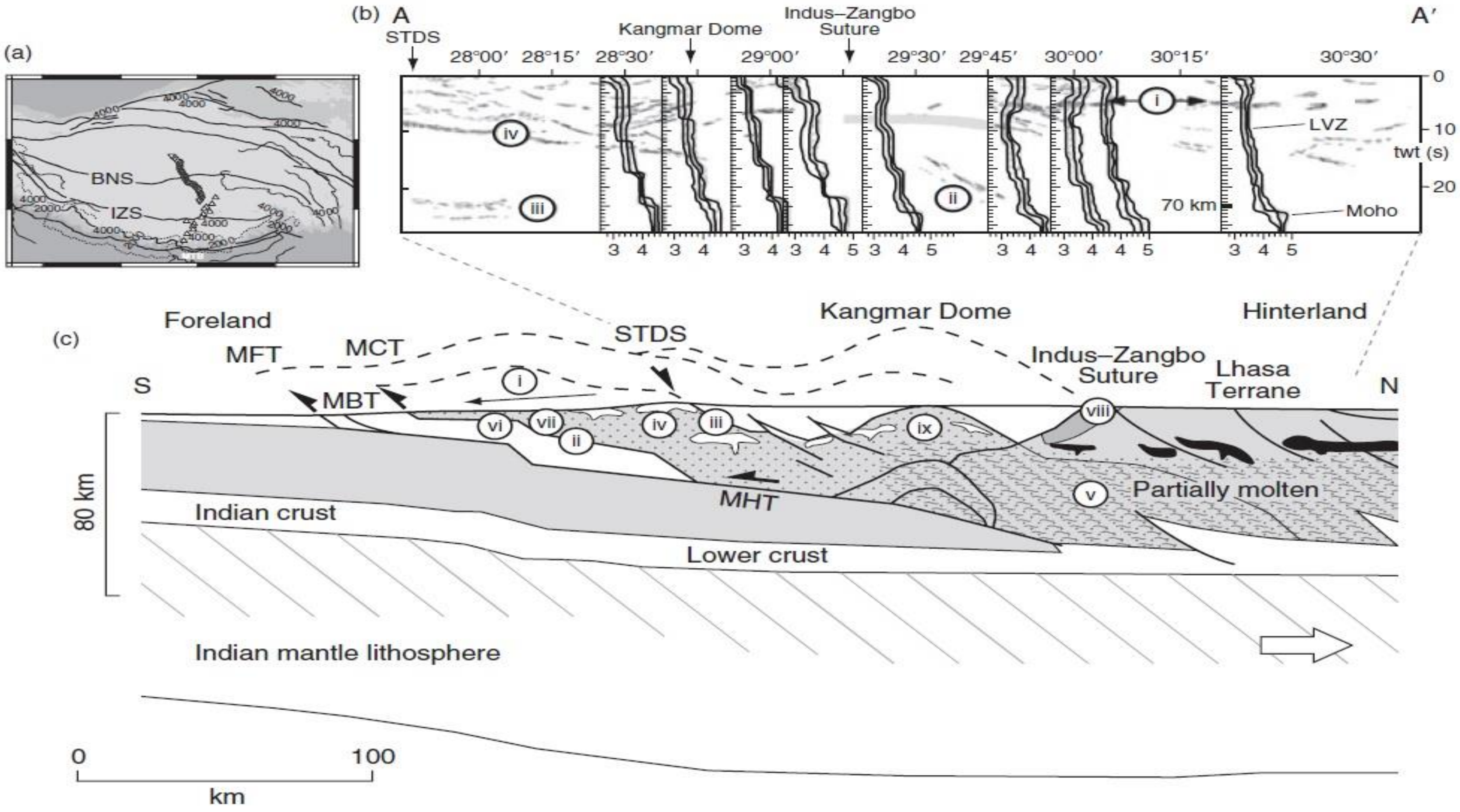


- The overall pattern of the deformation is similar to that occurring at ocean–continent convergence zones where an oceanic plate flexes downward into a subduction zone.
- North of the Himalaya, normal faulting and east–west extension dominate southern and central Tibet.
- Strike-slip faulting dominates a region some 1500 km wide north of the Himalaya and extending eastward into Indo-China.
- Convergence between India and Eurasia is accommodated by combinations of shortening, east–west extension, strike-slip faulting, lateral escape, and clockwise rotations, and uplift of the Tibetan Plateau, which started by Miocene time (at present-day rates are between 0.5 and 4 mm yr^{-1}).

Numbers represent depths. Black solutions are from events that occur within the Indian craton, light gray solutions are at depths of 10–15 km. Depths highlighted by a box are Moho depths from receiver function studies.

Himalaya structure

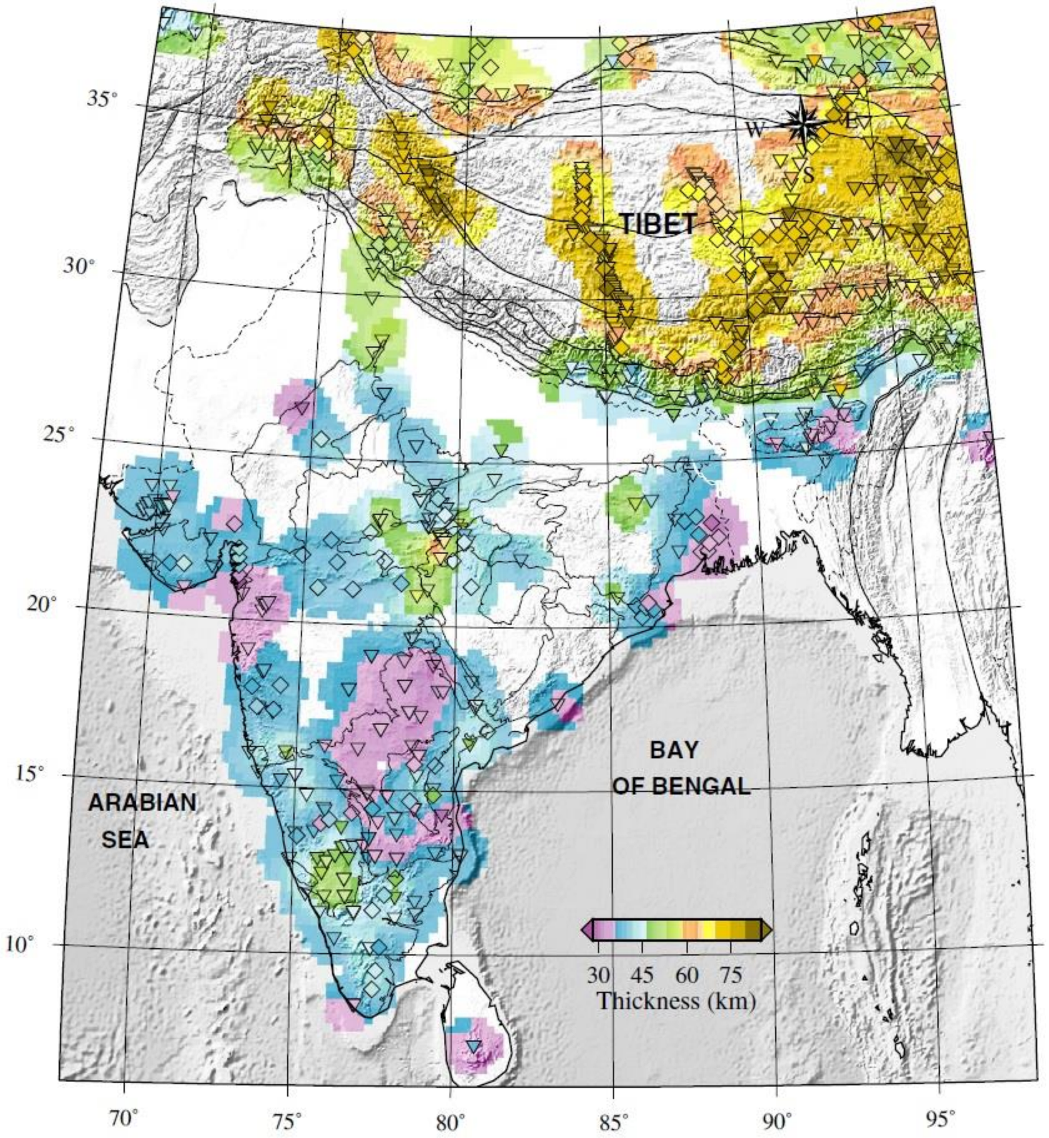
- Himalaya is composed of three imbricated thrust slices (250-350 km), separated by four major fault systems, accommodating almost the half of the ~ 2000 km of post-collisional shortening.
- Progressive decrease in the age of thrusting from north to south defines a foreland-propagating fold-thrust system.
- Each of three main thrusts merges downward into a common decollement (Main Himalayan Thrust).
- A crustal décollement surface above the Moho dips northward from 8 km below the Sub-Himalaya to a midcrustal depth of 20 km beneath the Greater Himalaya.



Moho Structure of the Indo-Eurasia collision zone

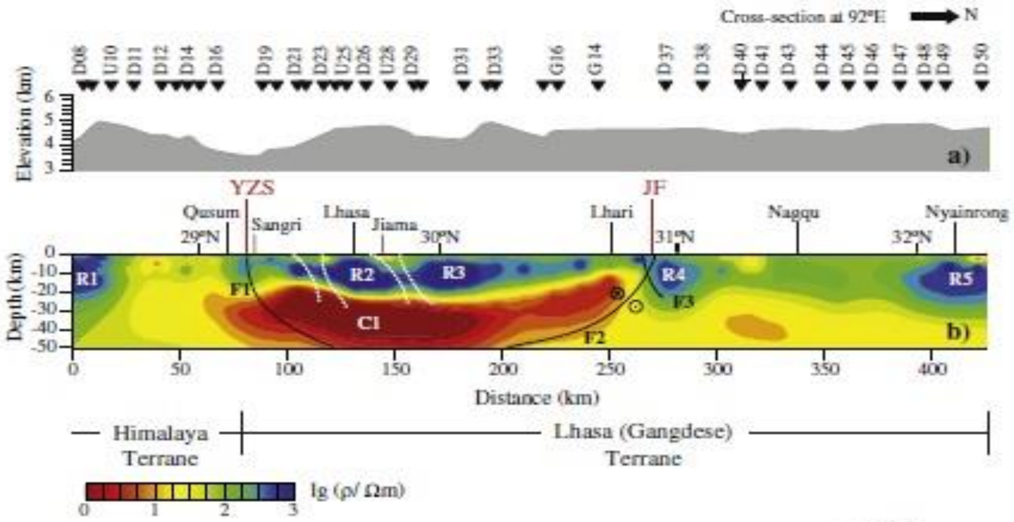
Moho Depth

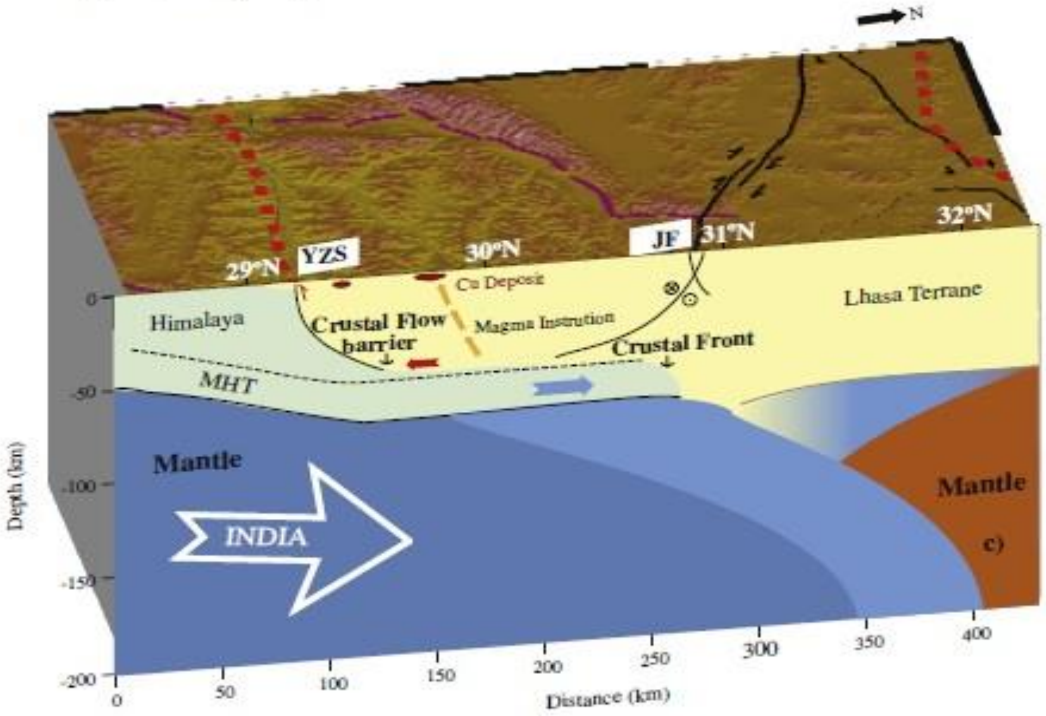
(from receiver functions and deep seismic sounding)



Singh et al., 2015, Tectonophysics, 644

Lithospheric Structure of the Indo-Eurasia collision zone





- Results from magnetotelluric investigations show that the Lhasa terrane is characterized by two layers: a resistive layer above−20km, corresponding to the presence of volcanic products and a conductive layer below−20km, likely reflecting a zone of partial melt (possible conductor channel between the crust and mantle).
- Crustal southwards extrusion from Tibet is limited in the southernmost portion of the Lhasa terrane (stopped at YSZ).
- The north Lhasa terrane has a cold crust that caused the crustal northern subduction front of the Indian subcontinent to be limited in the northern region of the Lhasa terrane (~30.8°N).

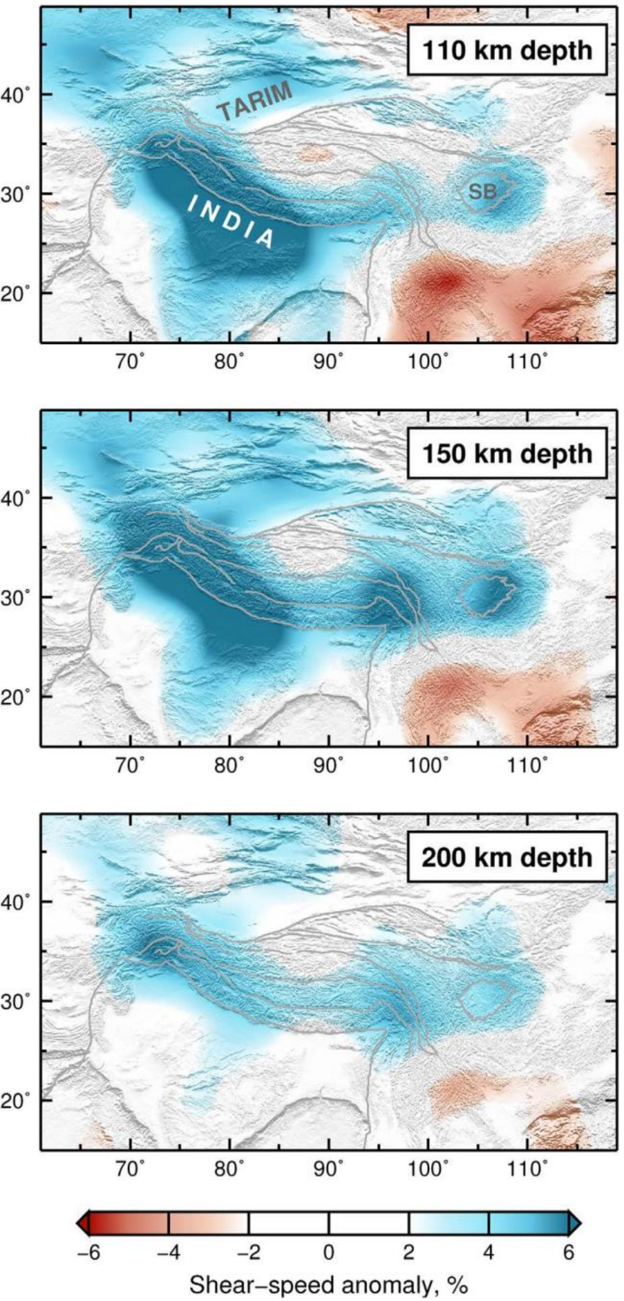
YZS: Yarlung–Zangbo sutures
JF: Jiali faults

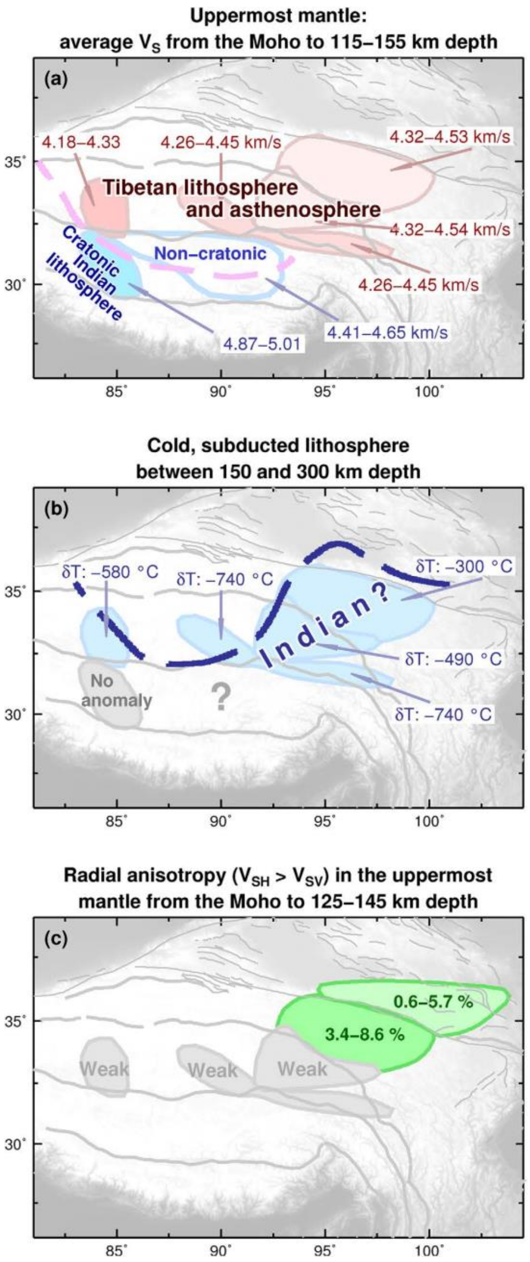
MHT= Main Himalayan thrust.

Xie et al., 2016, Tectonophysics, 675

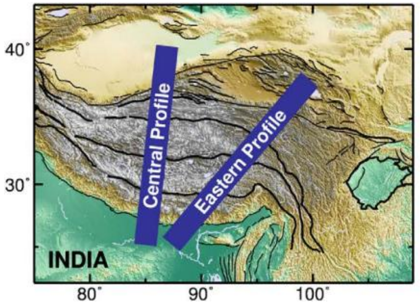
Lithospheric Structure of the Indo-Eurasia collision zone

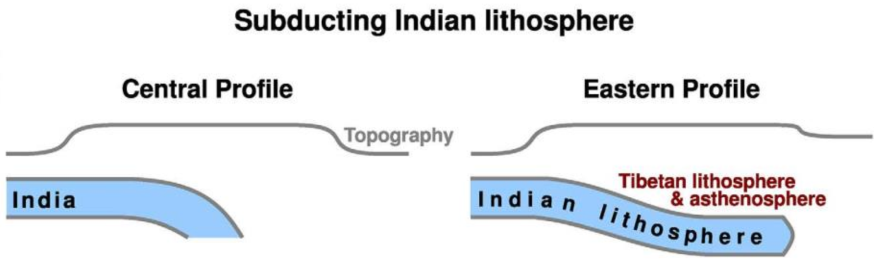
Agius and Lebedev, 2013, G3, 14





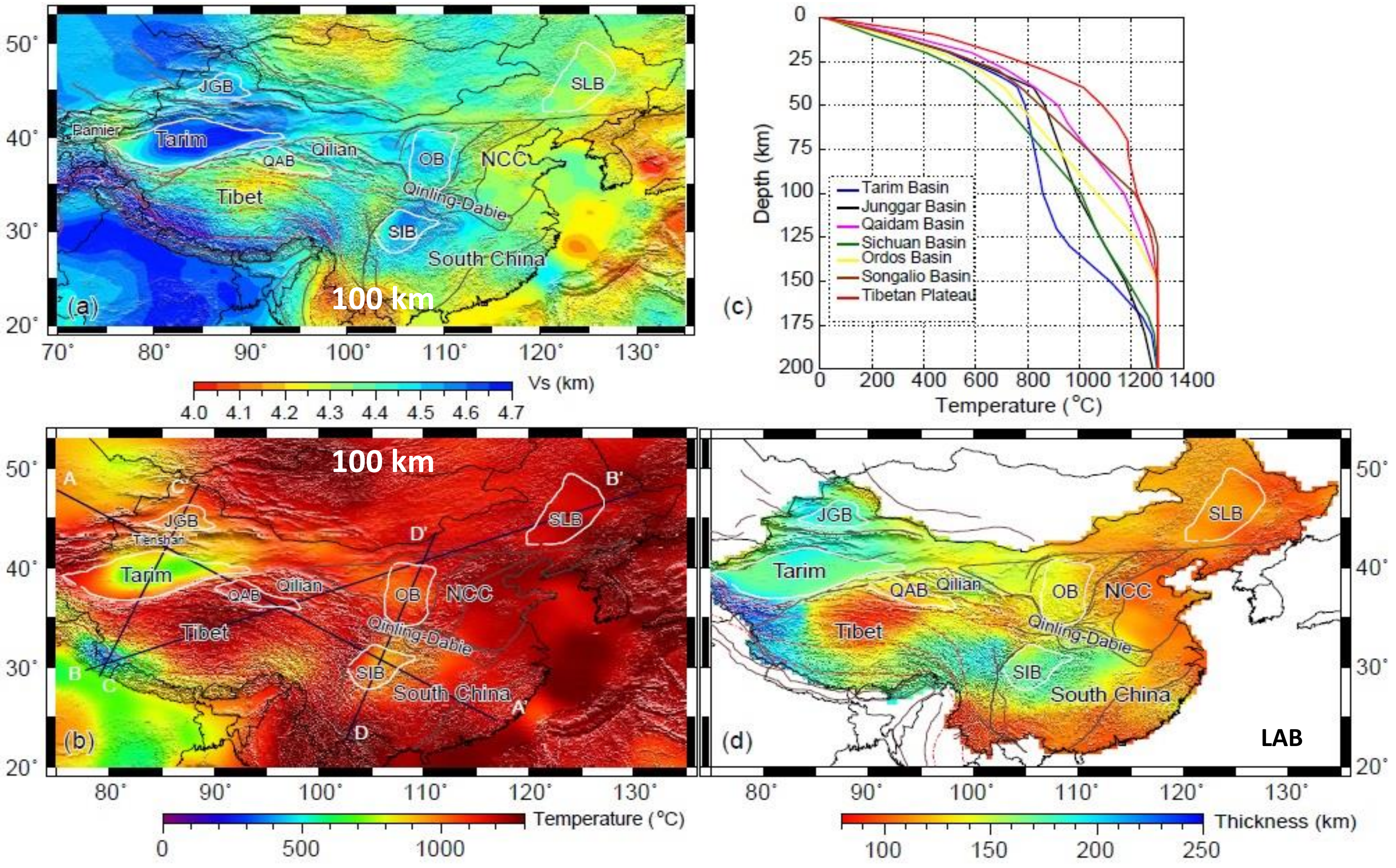
- In contrast to western and southwestern Tibet, where the mantle lithosphere correspond to that of the underthrusting Indian Plate, central and eastern Tibet are underlain by warm, Tibetan mantle lithosphere and asthenosphere.
- High velocity at 200 km depth may thus show the northern extent of the subducted Indian lithosphere that is now in the upper mantle beneath Tibet.
- Radial anisotropy in the Tibetan mantle lithosphere and asthenosphere is weak beneath the Qiangtang Terrane but strong beneath eastern Songpan-Ganzi, which is likely related to the asthenospheric flow above the subducting Indian slab.
- Downwelling accounts for the total amount of shortening in the Himalaya and Tibet and can explain the presence of warm shallow mantle (flow upward) .





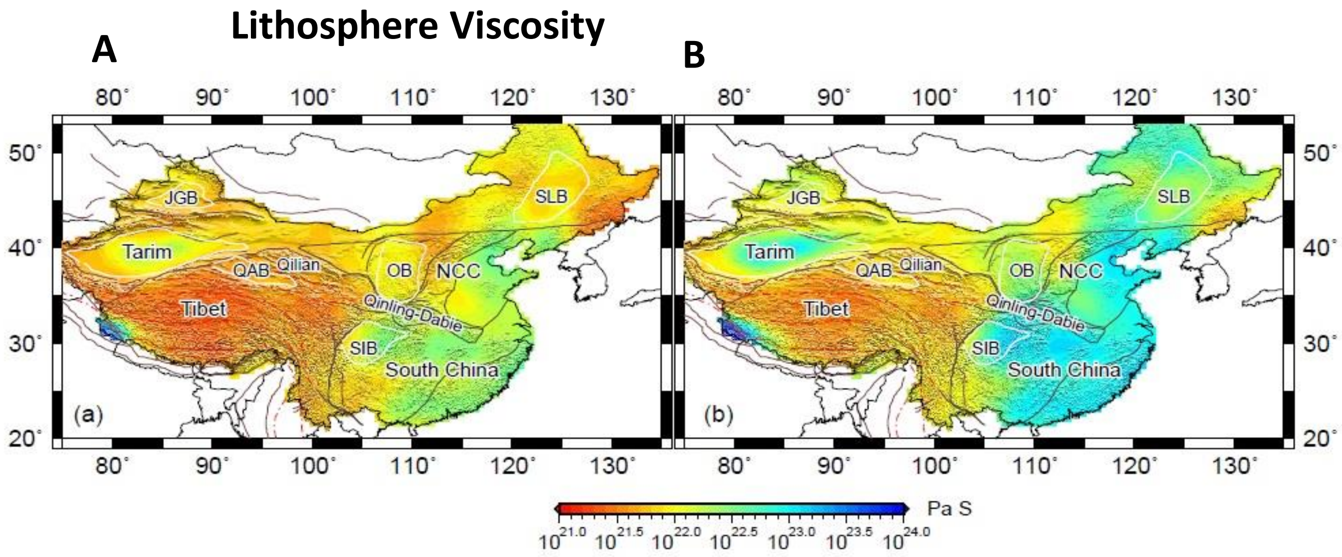
Thick, dashed blue line is + 2.5% anomaly contour at 200 km depth

Lithospheric Structure of the Indo-Eurasia collision zone

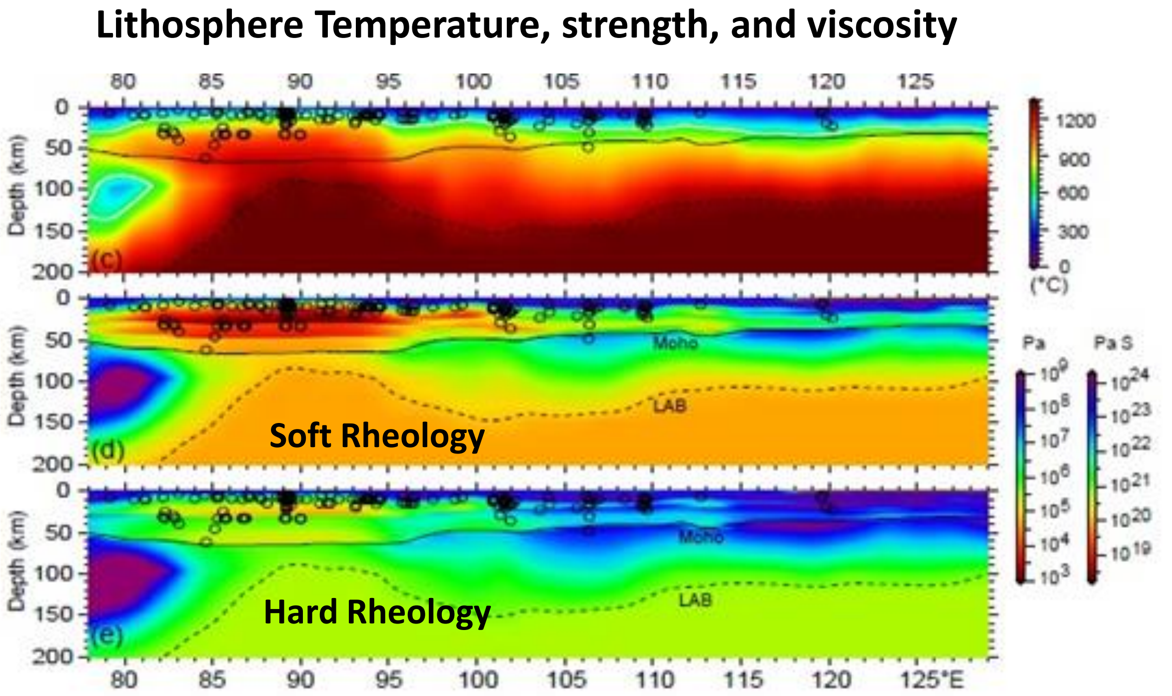


Deng and Tesauro et al., 2016, Tectonics, 35

Strength and viscosity of the Indo-Eurasia collision zone



Deng and Tesauro et al., 2016, Tectonics, 35



A=Soft Rheology: “wet quartzite,” “wet diorite,” “felsic granulite,” and “wet peridotite

B=Hard Rheology: “dry granite,” “dry diabase,” “mafic granulite,”and “dry peridotite”

- When the lower crust is relatively strong and resists flow, the crust tends to couple to the underlying mantle during shortening, crustal thickening is initially controlled by mantle subduction and results in a relatively narrow width, triangular shape orogen, and in the lack of a high orogenic Plateau.
- When the lower crust is relatively weak and flows easily, the crust decouples from the mantle, the mechanics of the orogen is dominated by identation and results in diffuse deformation.
- In Tibet low viscosity zones have developed in the deep crust during crustal thickening and wide, steep-sided plateaux have formed above the weak zones.

Mechanisms of continental collision

Precollisional history

- Millions of years of subduction, arc magmatism, terrane accretion, and crustal thickening along the southern margin of Eurasia weakened its lithosphere.
- Unlike Eurasia, the relatively cool and deeply rooted Precambrian shield of India resulted in a relatively strong plate that resisted shortening during collision (except for the sediments deposited on the passive continental margin of northern India).

Continental underthrusting

- The rheology of the two plates and the degree of mechanical coupling between them control shortening and the evolution of stresses within the overriding plate.
- Underthrusting of the Indian plate generates shortening and thus large crustal thickness (70-80 km) and uplift of the Tibetan Plateau.
- Removal or displacement of Asian lithosphere from Tibet may occur through different mechanisms: (1) Delamination of the lithosphere mantle beneath the Tibetan Plateau, (2) Southward subduction of the Asian mantle, (3) Removal of Asian mantle by strike-slip faulting during lateral escape of Tibet.

Indentation, lateral escape, and gravitational collapse

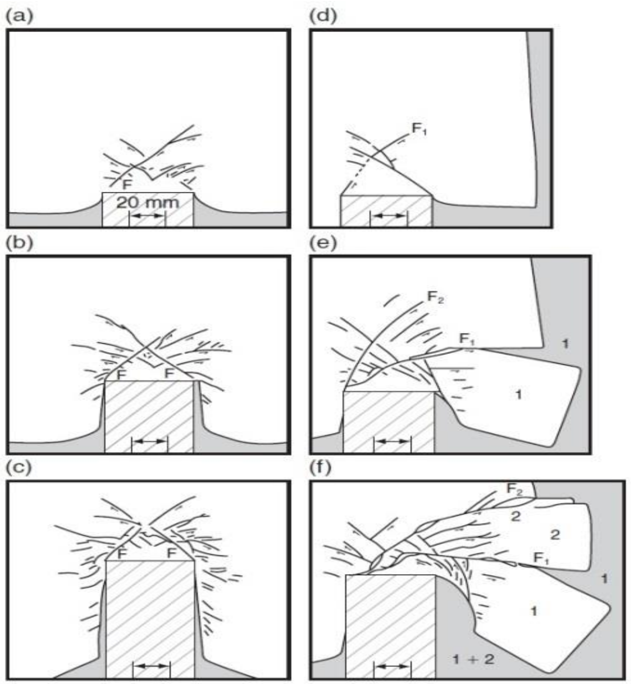
- Only part of the shortening has been accommodated by fold-thrust belt, since there is a shortening deficit ranging from 500 to over 1200 km: Hypotheses about the indentation of India into Asia and lateral escape of eastern Tibet.

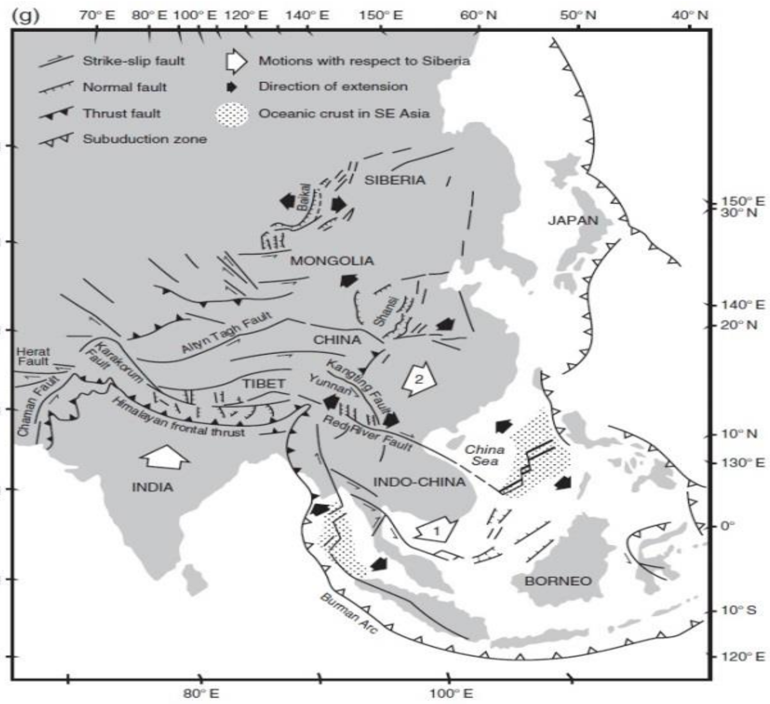
Mechanisms of continental collision

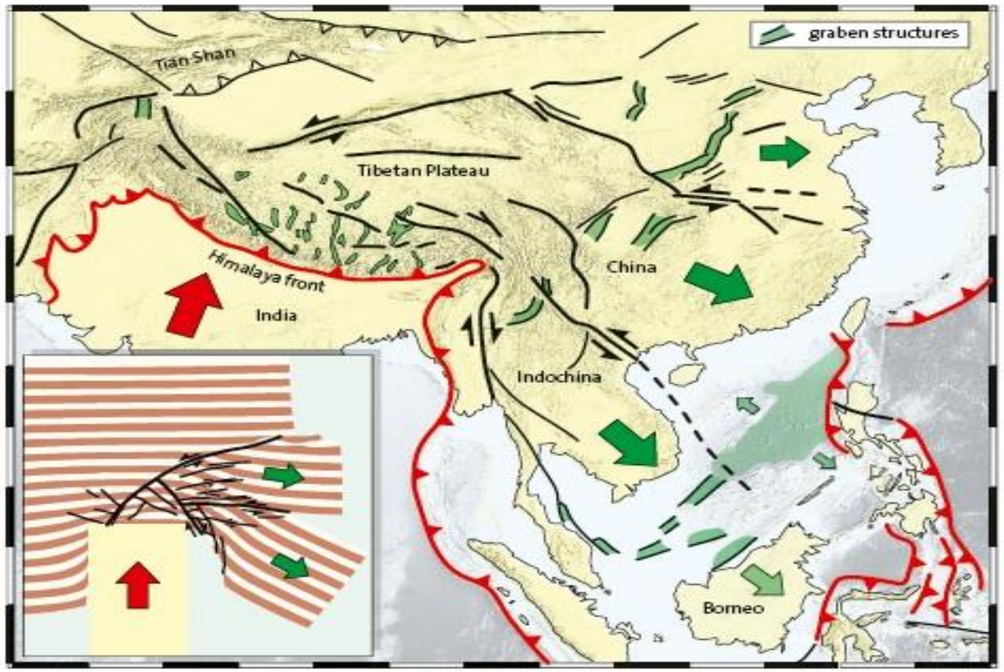
Indentation, lateral escape, and gravitational collapse

- Indentation: the process by which a rigid block presses into and deforms a softer block during convergence (lateral transmission of forces).

Bilaterally confined **Unilaterally confined**







Numbers associated with arrows show extrusion phases: 1 ~ 50–20 Myr; 2 ~ 20–0 Myr.

- Bilateral case produces a pattern of faults shaping a triangle welding the indenter
- Unilateral case generates an asymmetric pattern of faults, pull-apart basins and curvature of fault systems

Mechanisms of continental collision

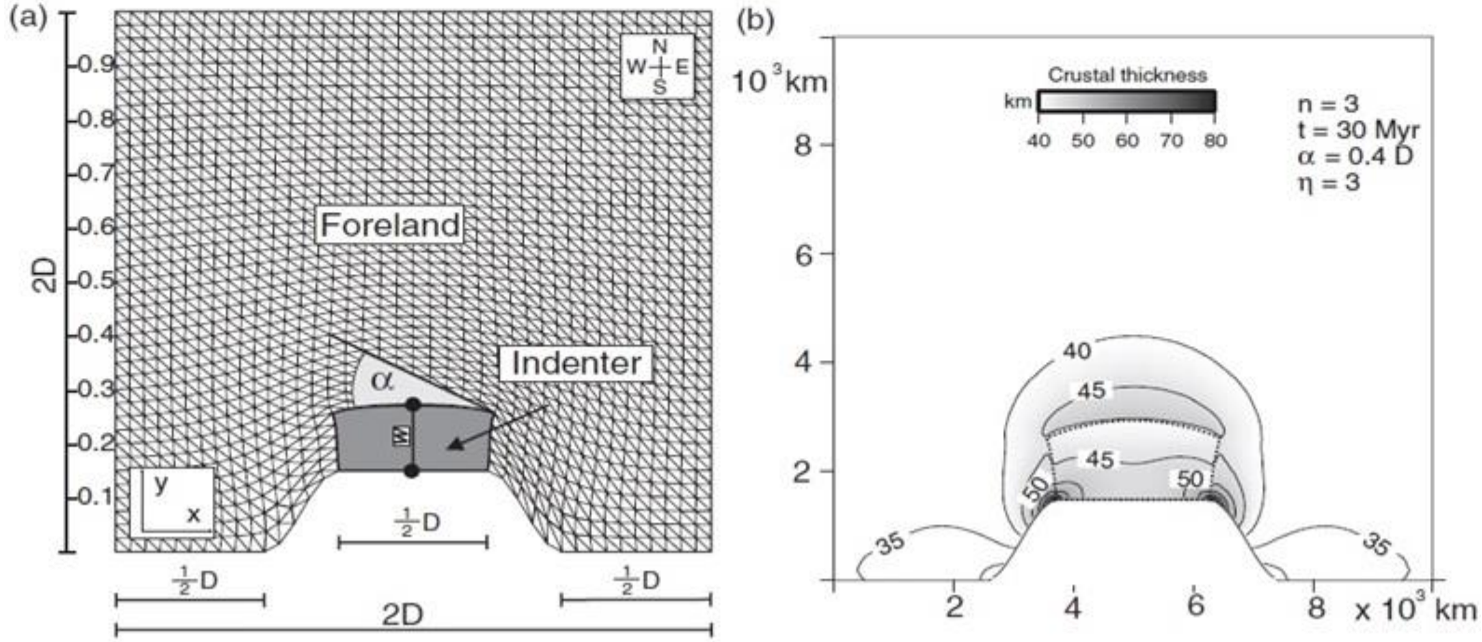
Limitations of the identation model:

- Identation predicts lateral displacements of hundreds to a thousand kms on the large strike-slip faults (not corresponding to the reality).
- Identation do not take into account the effects of variations in crustal thickness during deformation.
- Indentation explains strike slip faults in eastern Tibet, but not normal faults within Tibet.
- E-W extension may be the result of gravitational buoyancy forces associated with large thickness and high elevation of the plateau.
- Lateral gradients in gravitational potential energy may spread out the plateau (lateral extrusion).

Numerical models tested the effects of variation in the shape, convergence angle, and rheology of a continental indenter on both lateral and vertical strain patterns in Asia during lateral escape: How buoyancy forces arising from crustal thickening are balanced by edge forces from indentation.

Viscous Sheet Model:

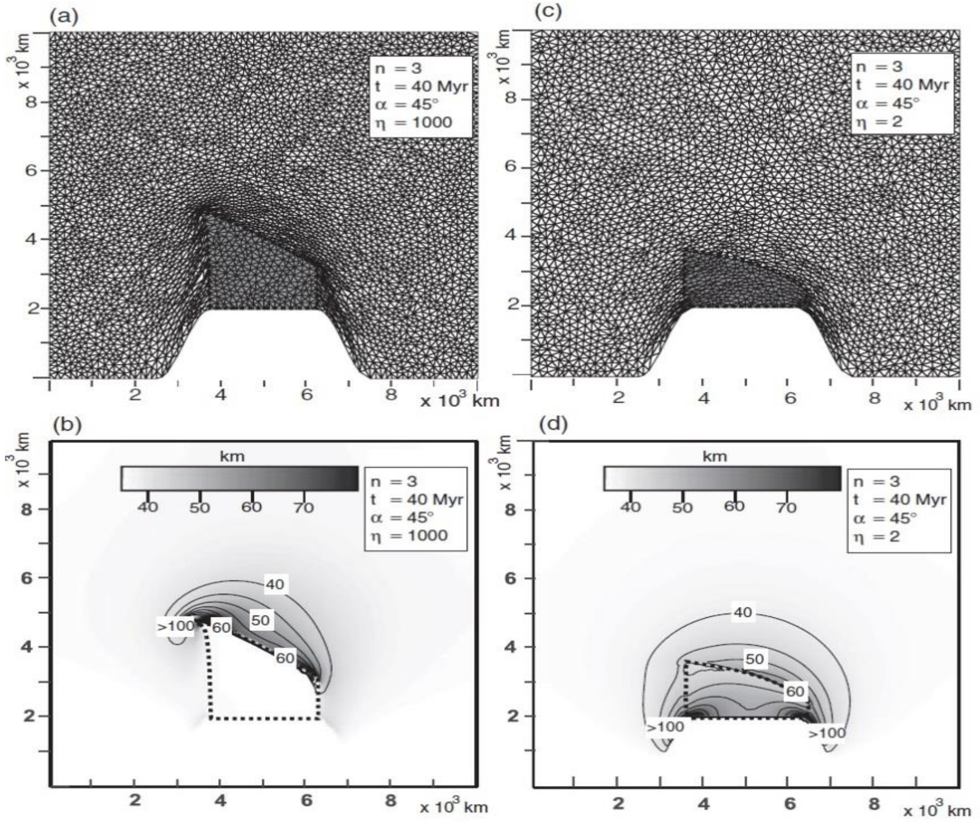
The motion of the identer results in deformation and thickening that is distributed between the indenter and the foreland to the north



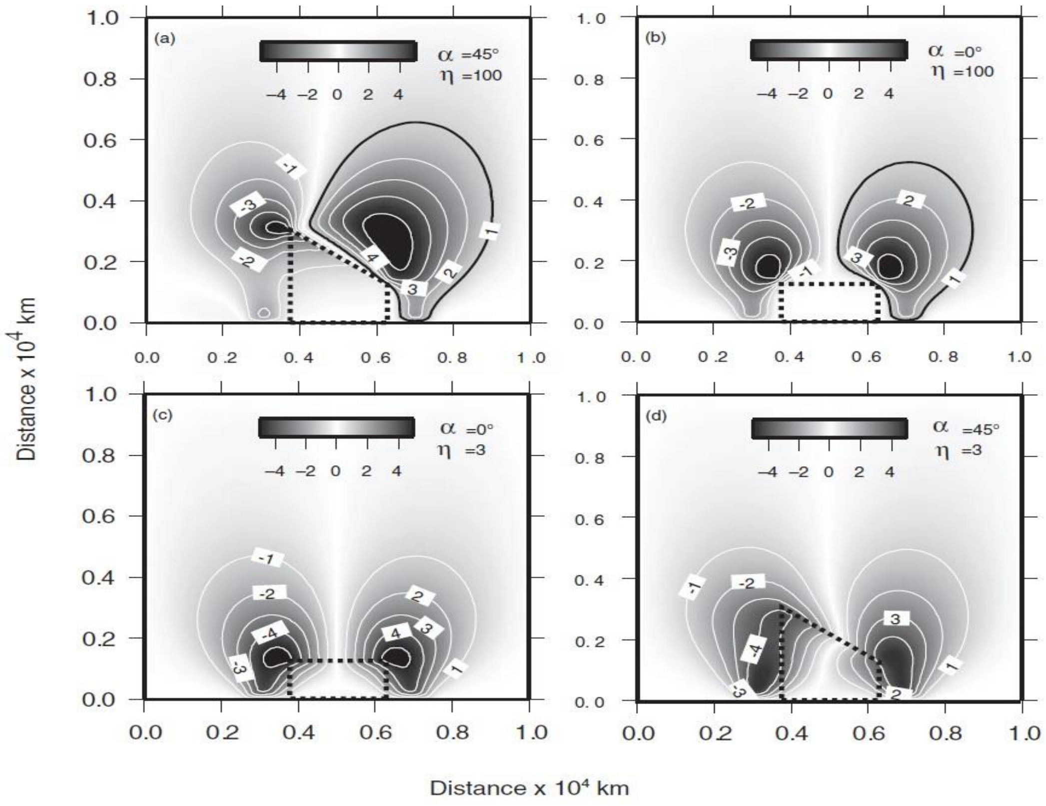
Variable: Viscosity contrast (η) between the indenter and the foreland and the angle (α) between the indenter front and the direction of indentation

Mechanisms of continental collision

- For a high viscosity identer (η), crustal thickening is at a maximum north of the western tip and slightly less in front of the northeastern edge: **lateral escape of the crust increases with indenter angle for relatively strong indenter rheologies.**
- **For a low viscosity indenter (η = 2 or 3), the indenter angle plays only a minor role** and accommodates most of the shortening and thickening, with the pattern becoming progressively more symmetric and delocalized through time.



Crustal Thickness

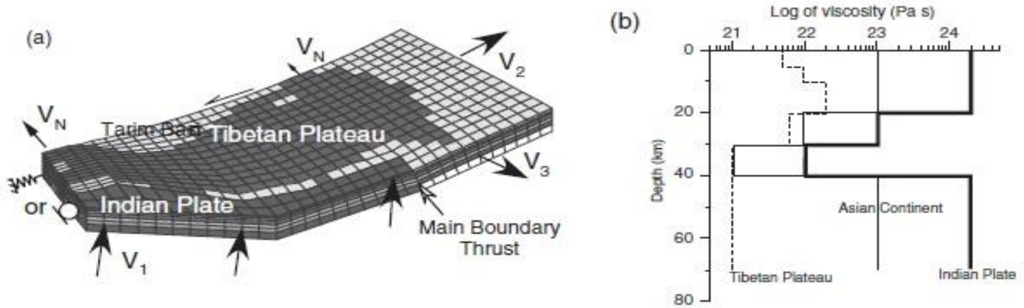


Horizontal displacement

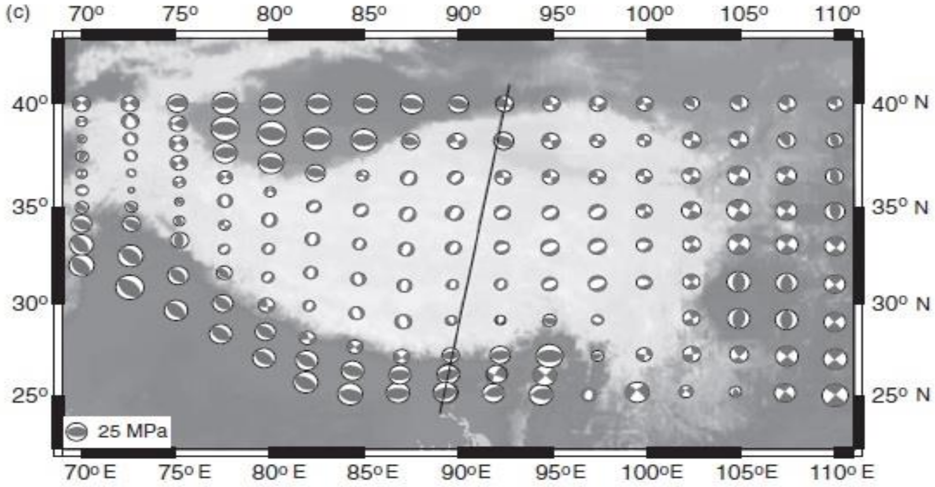
By decreasing the viscosity indenter, the indenter angle plays only a minor role

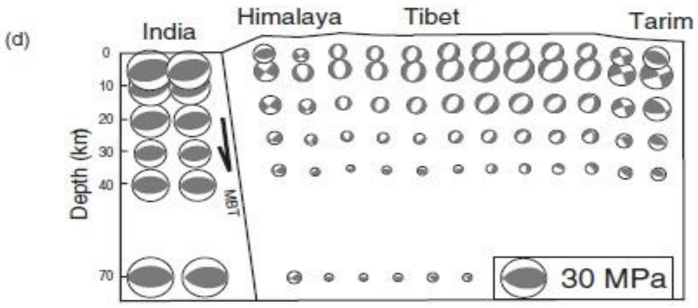
As the strength of Asia decreases, the magnitude and distribution of crustal thickening increase and gravitational buoyancy forces become more important.

Mechanisms of continental collision



Predicted Stress in the Upper Crust (10 km depth)





Predicted Stress with depth

Boundary Conditions: On the eastern and southeastern sides of the model, boundary conditions are assigned to simulate the lateral escape of the crust. On the west, the effects of a spring or roller simulate the lateral resisting force of a rigid block in Pamir. On the northern side of the model boundary conditions approximate the resistance to motion by the rigid Tarim Basin. An isostatic restoring force is applied to the bottom of the model.

Present state of stress in the Himalaya and Tibetan Plateau is considered due to: (1) a horizontal compressive force resulting from the collision of India with Asia; (2) buoyancy forces resulting from isostatically compensated topography; (3) basal drag on the Eurasia plate; (4) slab pull forces.

Results:

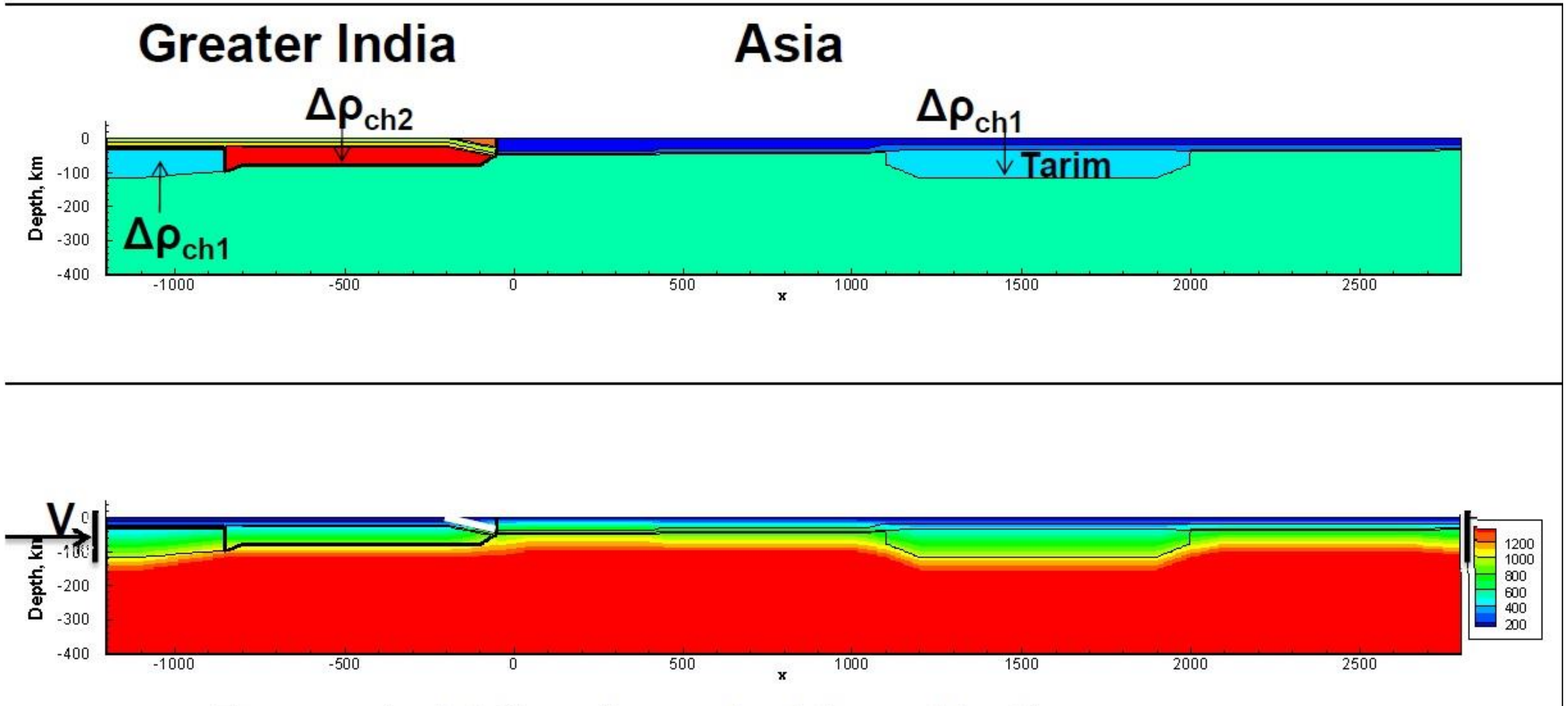
- Surface velocity field and regime of deformation in the orogen reflect a mechanical balance between gravitational buoyancy, the indenting Indian plate, and the specific geometry and the buoyancy conditions of the plateau.
- Difference between the weak rheology of the Tibetan Plateau and strong rheology of the Tarim basin enhances crustal thickening and topographic uplift.
- Extension and the high elevations of Tibet are obtained if the Tibetan crust is very weak.
- When the plateau is 50% lower than its present elevation (~5 km), strike-slip and reverse faulting dominate the plateau region. Significant crustal extension occurs when the plateau reaches 75% of its present elevation.
- Basal shear relieves the compressive (indentation) stresses that balance the buoyancy forces driving extension at the southern edge of Tibet (and thus it enhances extension).

Mechanisms of continental collision

From more than 1400 km of underthrusted India less than 700 km is located in shallow mantle

- What is mechanical condition for more than 1000 km underthrusting of India?
- What happened with the lithosphere of India during underthrusting and why?
- What happened with the lithosphere of Asia and what about delamination?

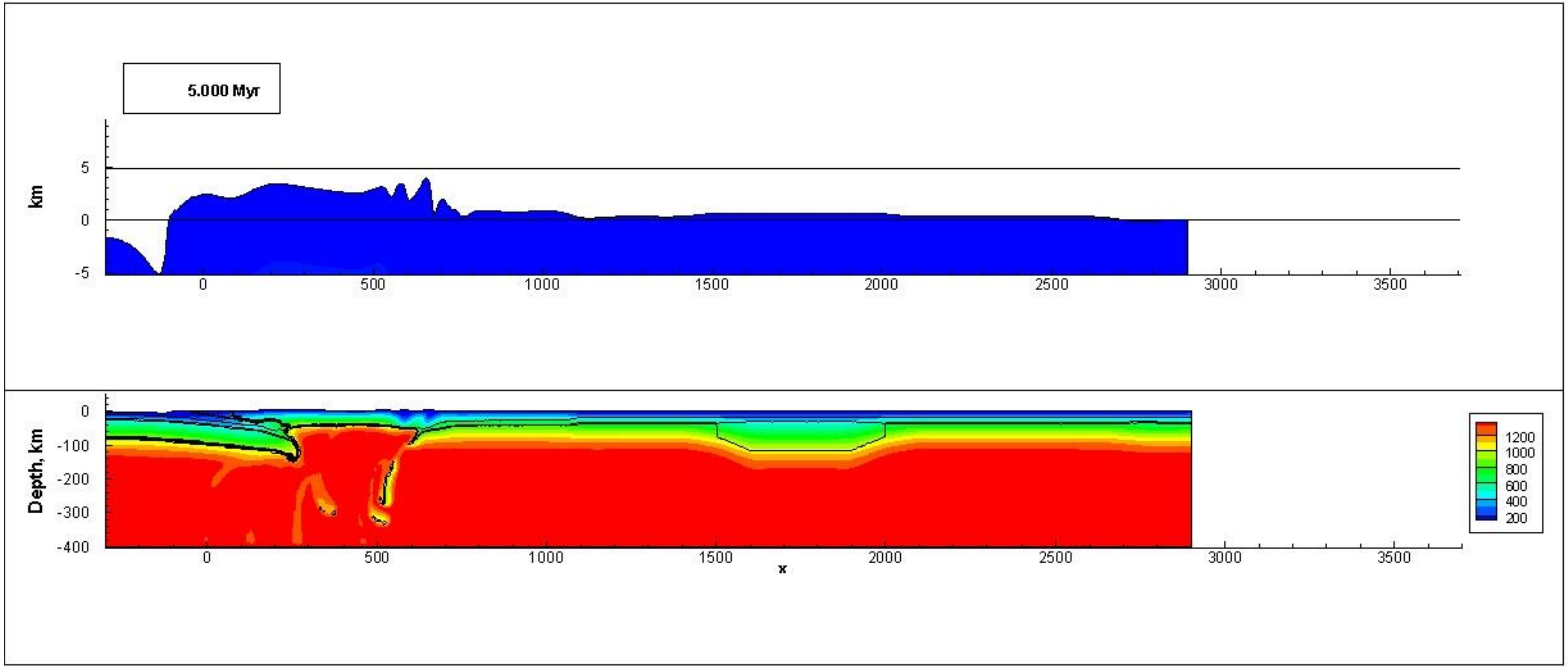
Model setup



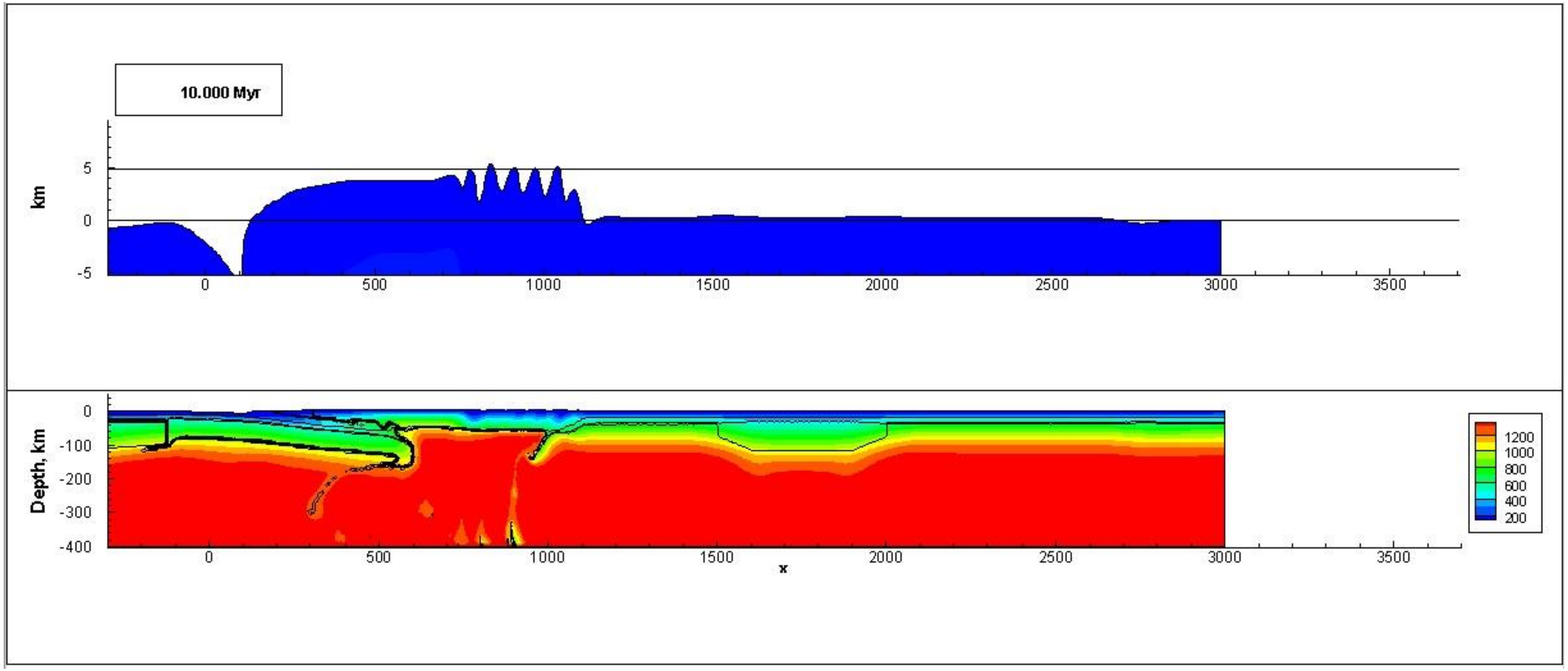
Free material flow through side and bottom boundaries

Mechanisms of continental collision

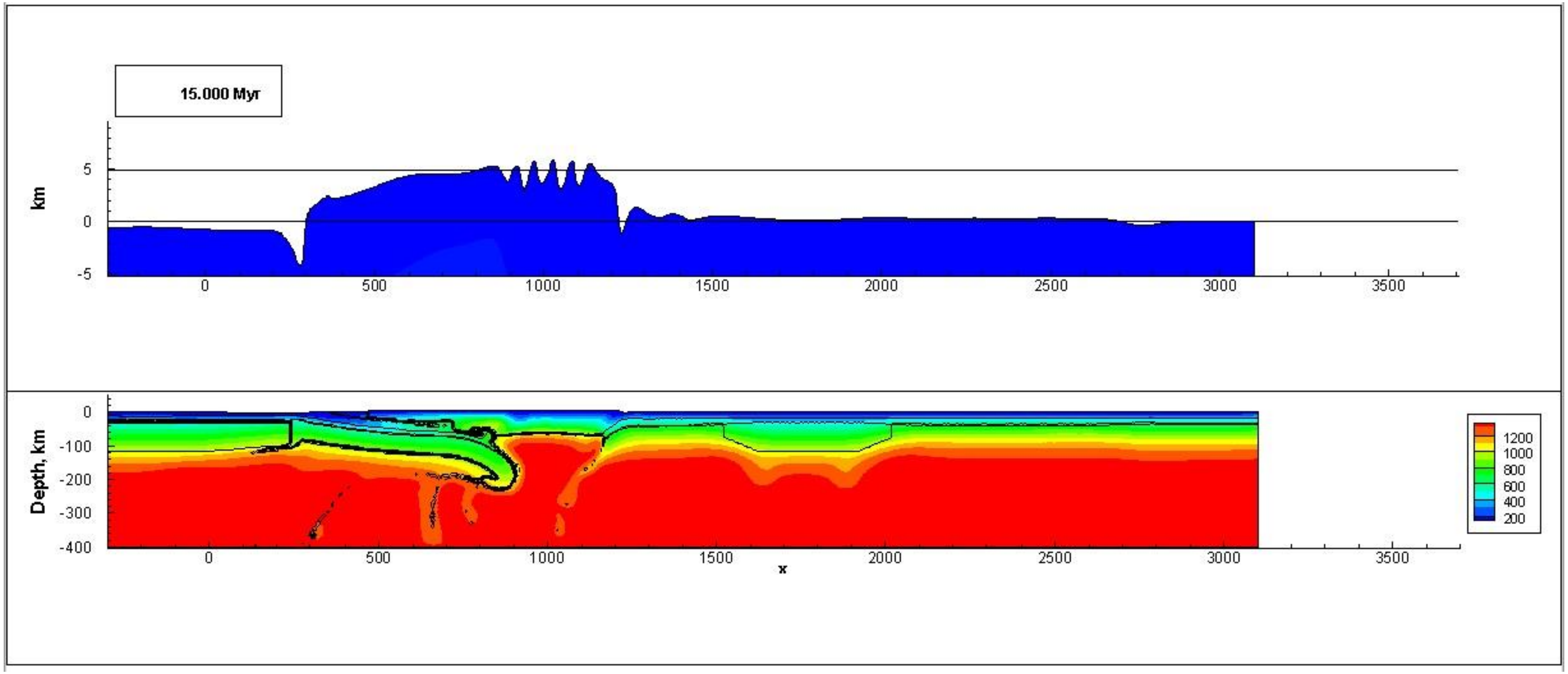
The model supports early high topography moving to the north together with the mantle-delamination front



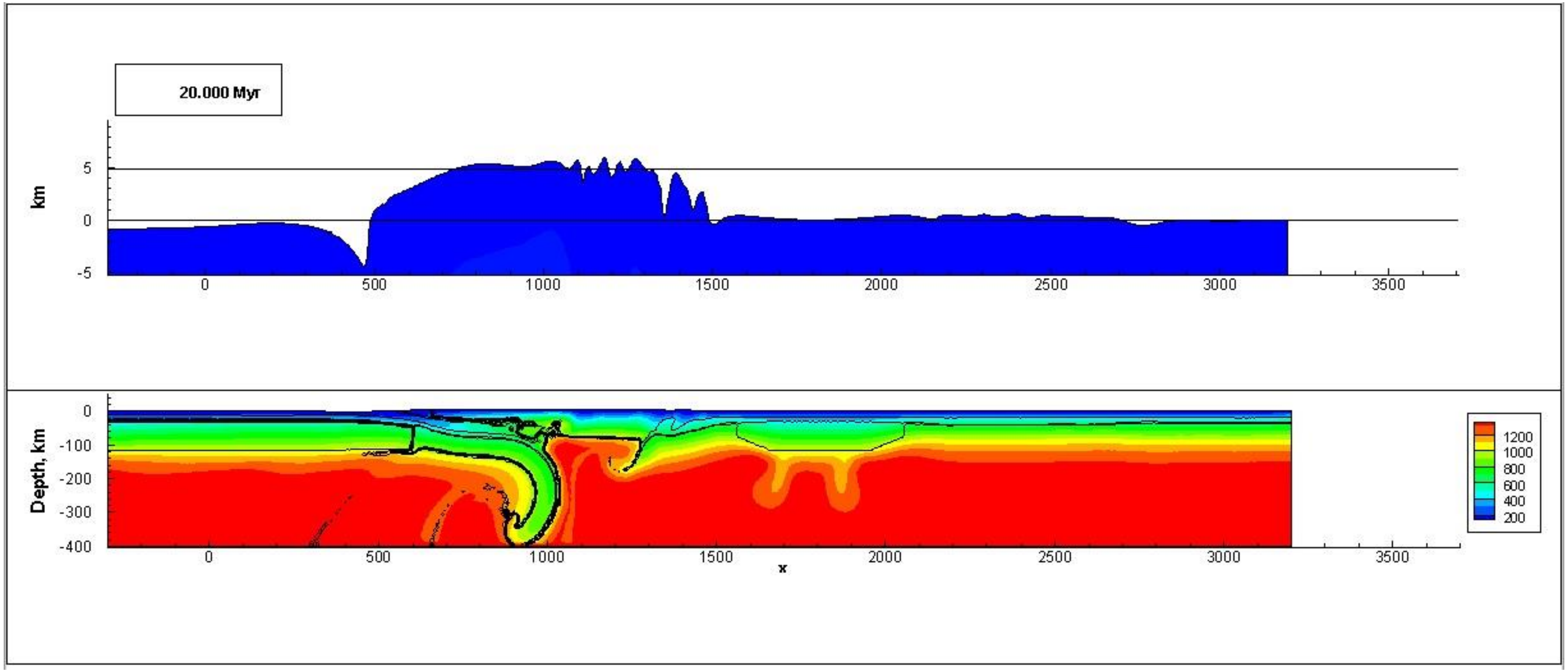
Mechanisms of continental collision



Mechanisms of continental collision



Mechanisms of continental collision



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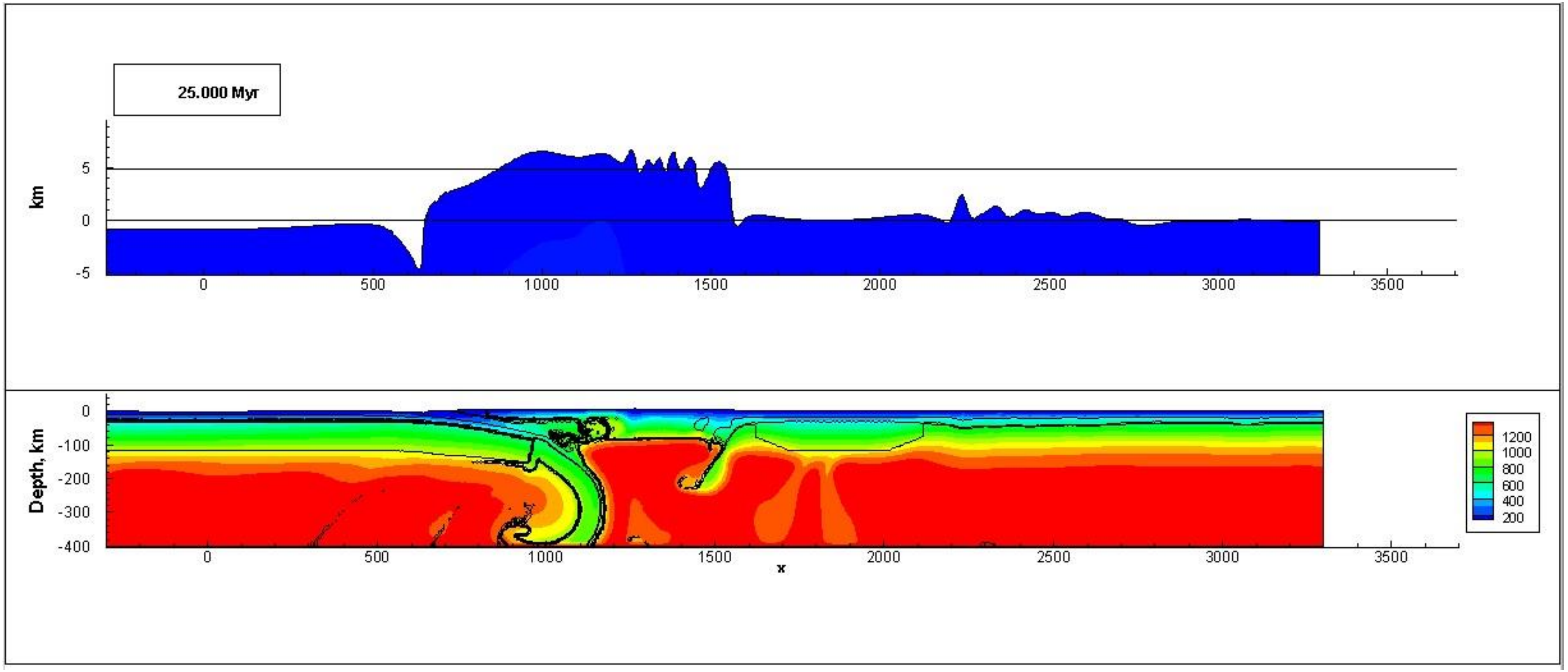

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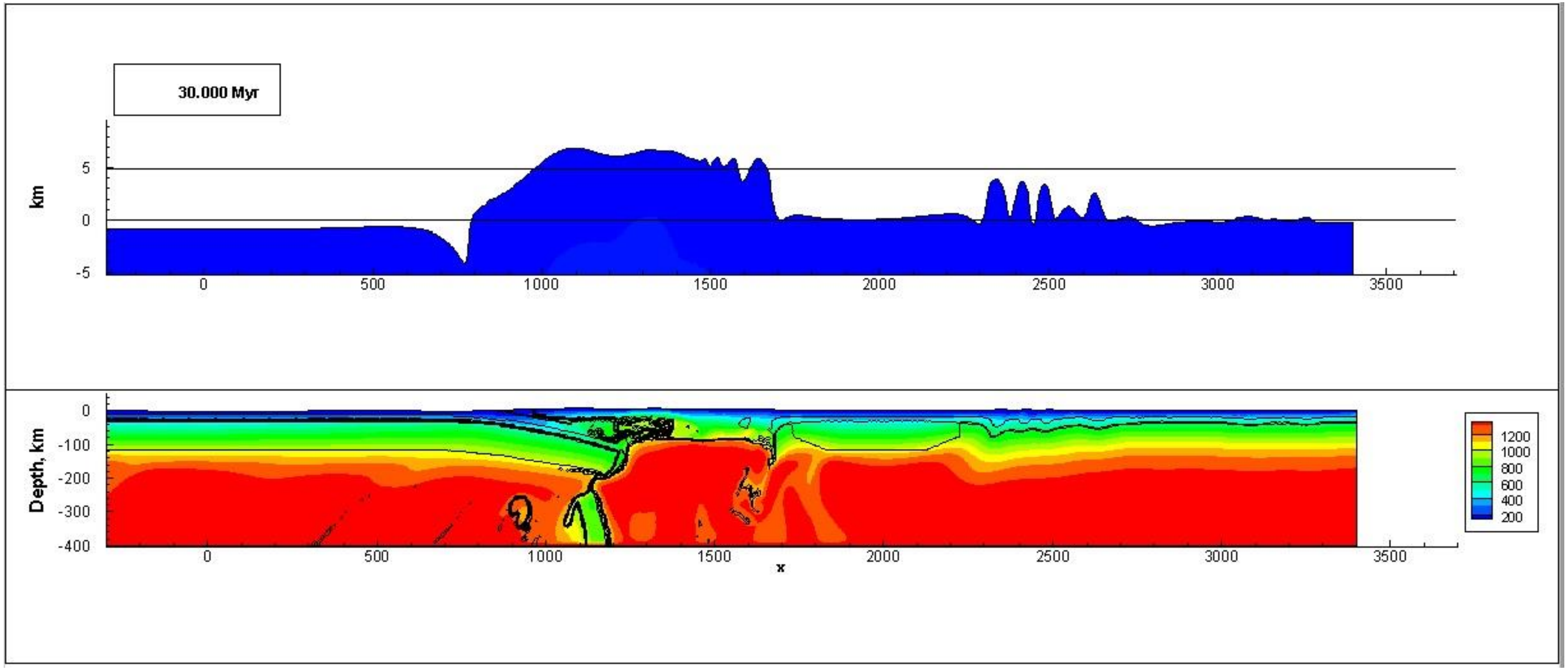
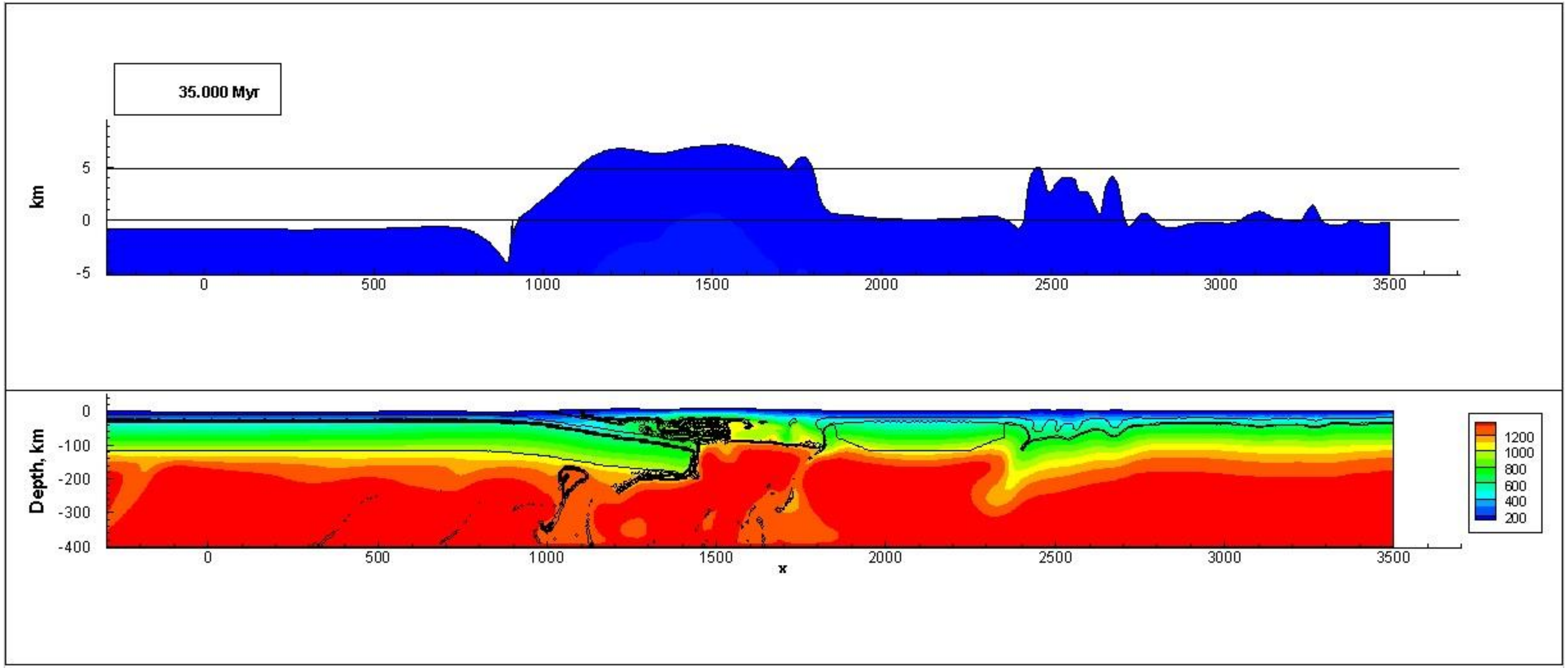
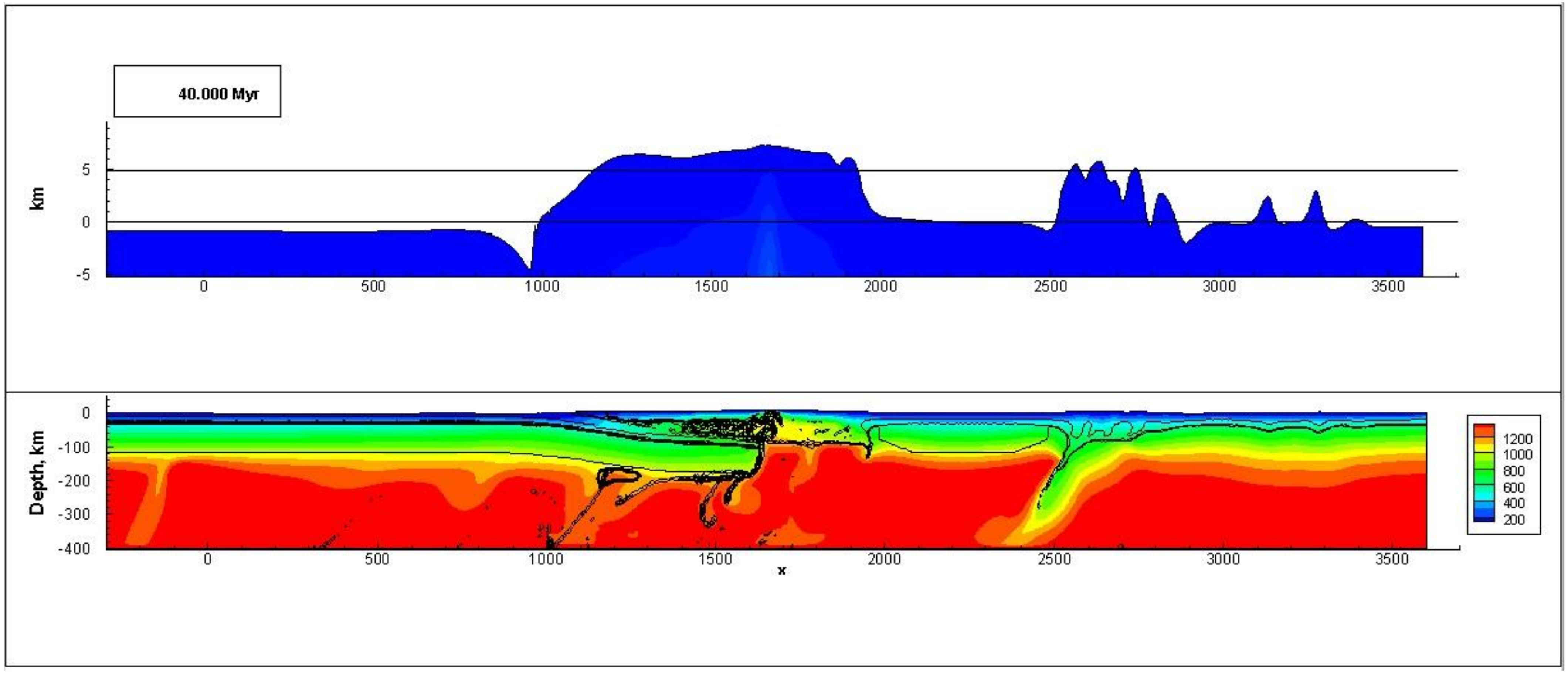


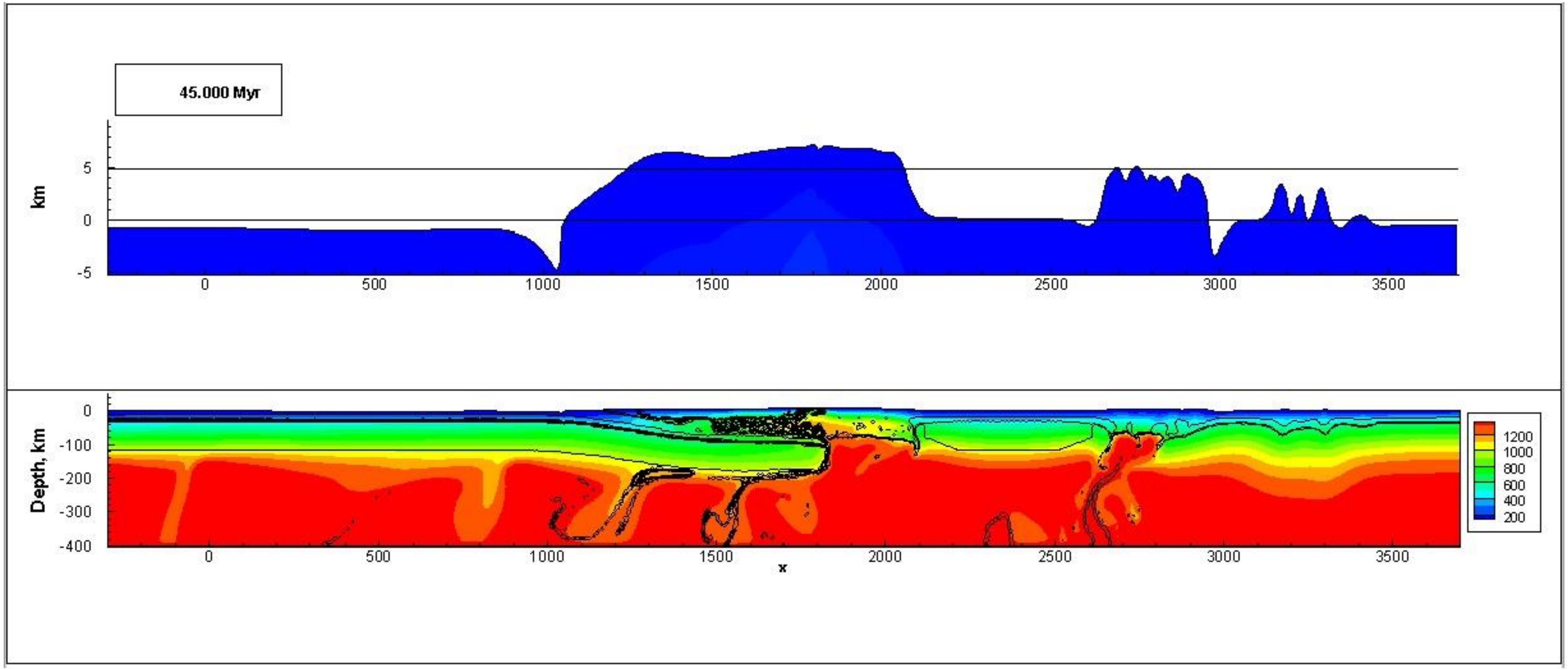
Mechanisms of continental collision



Mechanisms of continental collision

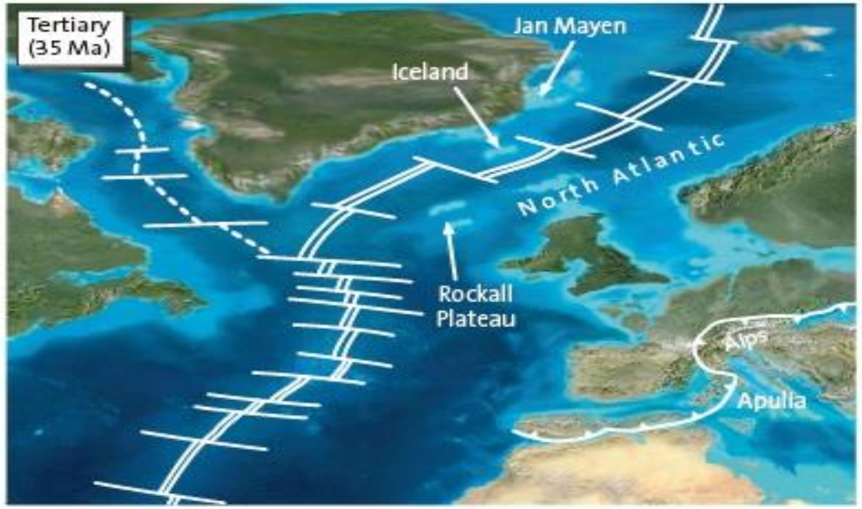


Mechanisms of continental collision



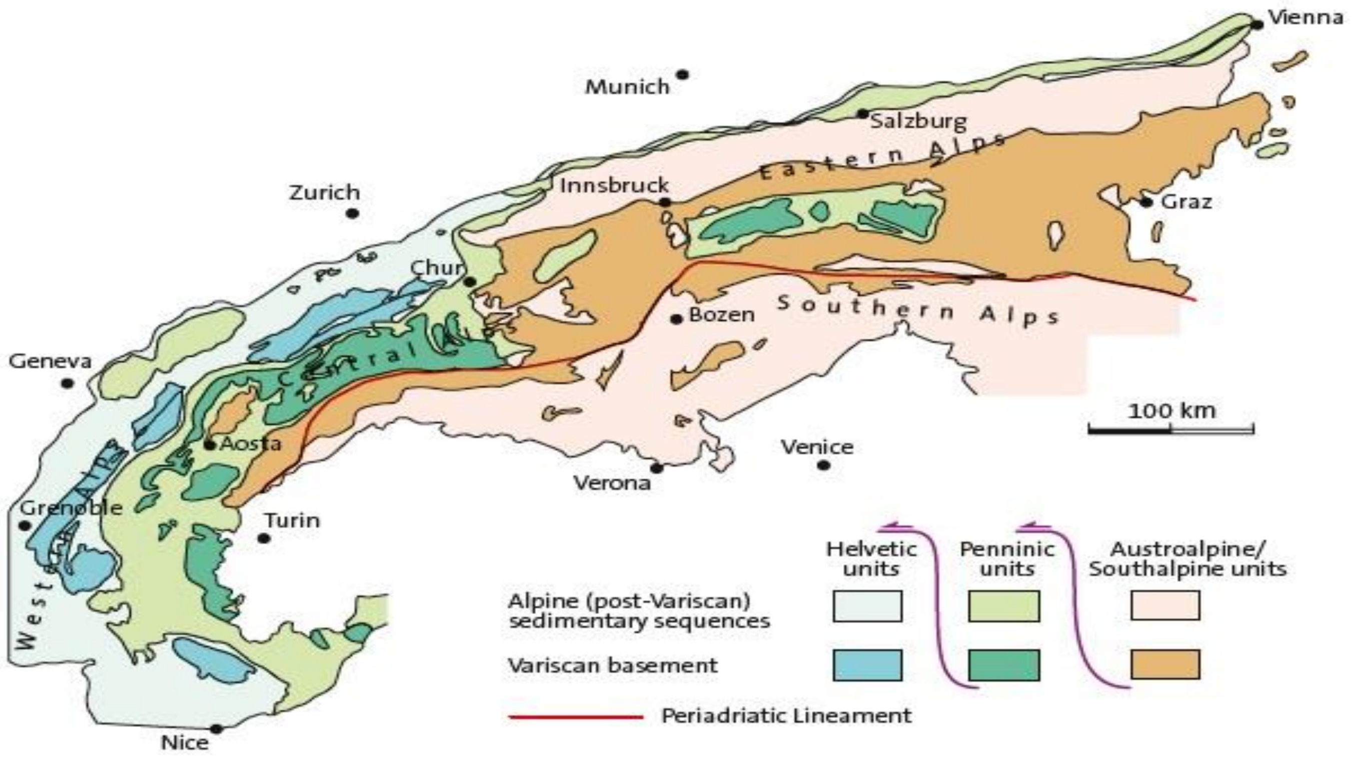
Alpine Chain

- The initial orogenic event occurred during the Middle Jurassic to Early Cretaceous (Eoalpine orogeny) with the formation of the Austroalpine nappe system (Eastern Alps).
- The Eoalpine orogeny continued into the Carpathians, Dinarides, and other mountain ranges further east (Cimmerian orogeny) in central Asia it marks the collision of the Lhasa Terrane with Eurasia in Tibet (partial closure of the Tethys ocean).
- Penninic-Ligurian Ocean opened in the Early Jurassic as an extension of the Atlantic Ocean and closed in the Paleogene (partially destroyed in the Alpine orogeny: different units subducted beneath the Austroalpine nappe), as the former African continental margin (Austroalpine unit) was thrust over the European continental margin (Helvetic unit).

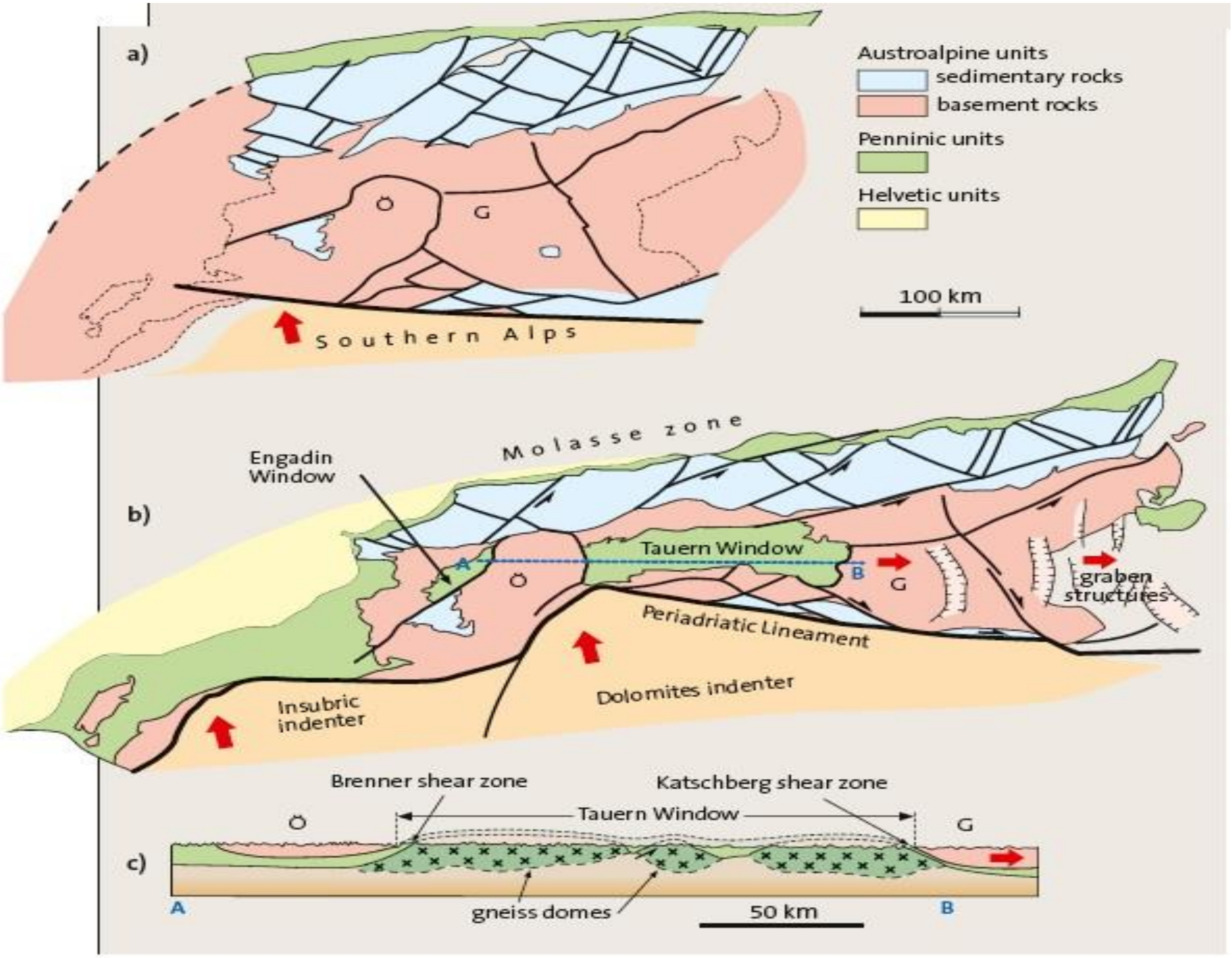


Alpine Chain

- The main orogenic phase lasted from Eocene to Oligocene and lead to the collision of the Austroalpine realm with the Middle Penninic continental mass, and, subsequently, with the European continental margin (the Helvetic realm).
- In the Early Oligocene, increasing volumes of sediment were transported from the uplifting orogen (consequently to the rapid stacking of numerous crustal sheets, and slab breakoff of the oceanic part of the subducting plate) into the foredeep, became rapidly filled.
- The basement of Western Alps is continuation of the Southern Central Massif and Black Forest and was thus part of Gondwana until its collision during the Variscan orogeny.

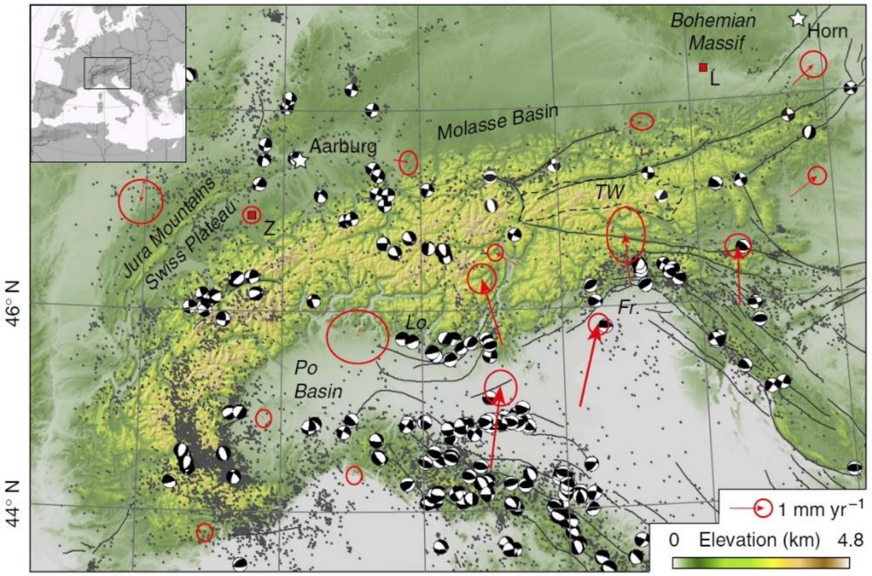


Alpine Chain

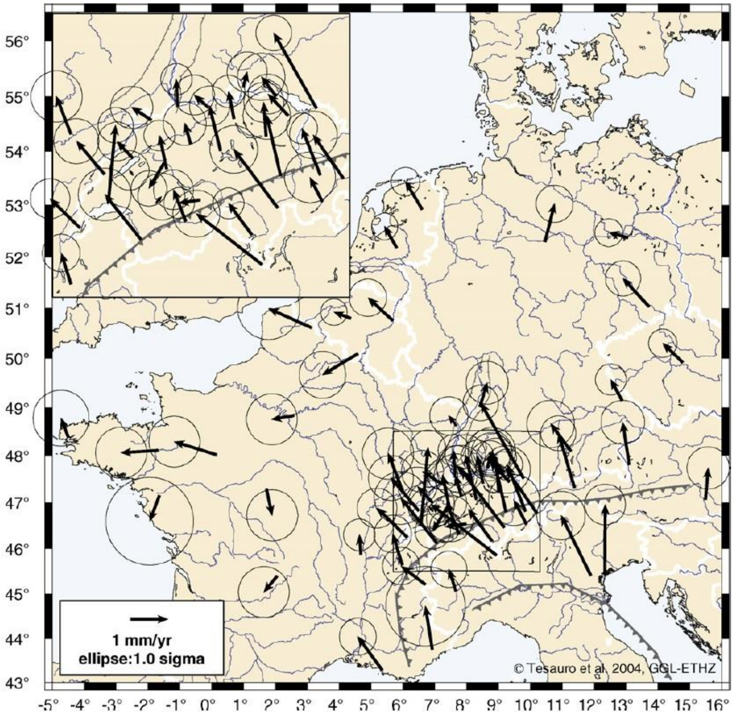


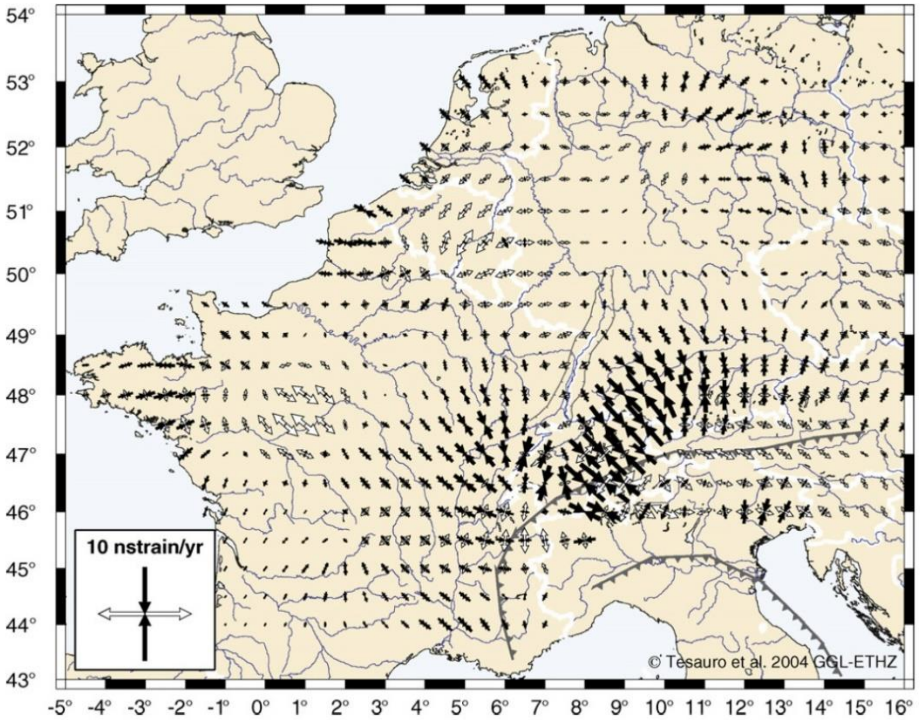
- The Southern Alps did not experience metamorphism and thus remained cool and strong and acted as an indenter (Insubric and Dolomites indenters) that pushed against the main body of the Alps to the north.
- As the strong, brittle masses pushed northward, they generated crustal stacks north of the Periadriatic Lineament (lateral tectonic extrusion) in the middle Miocene
- The Penninic rocks buried beneath the Austroalpine nappe system became rapidly exhumed and formed tectonic windows (Tauern Window (160 km-long and 30 km wide structure).
- The Tauern region consists of a brittle upper plate that comprises basement rocks that cooled following Cretaceous Eoalpine metamorphism and a lower plate of ductile, metamorphic Penninic rocks.
- Exhumation of the Tauern window occurred as a consequence of the rapid extension and unroofing in the upper brittle plate. In response, the lower ductile plate was deformed and extended along an E-W trend during the rapid uplift.
- Farther east of the Alps beneath the Neogene sedimentary cover of the Pannonian Basin, the Austroalpine unit is dissected by N-S-trending graben and horst structures, another indication of E-W stretching.

Alpine Chain



Mey et al., 2016, Nature Communications



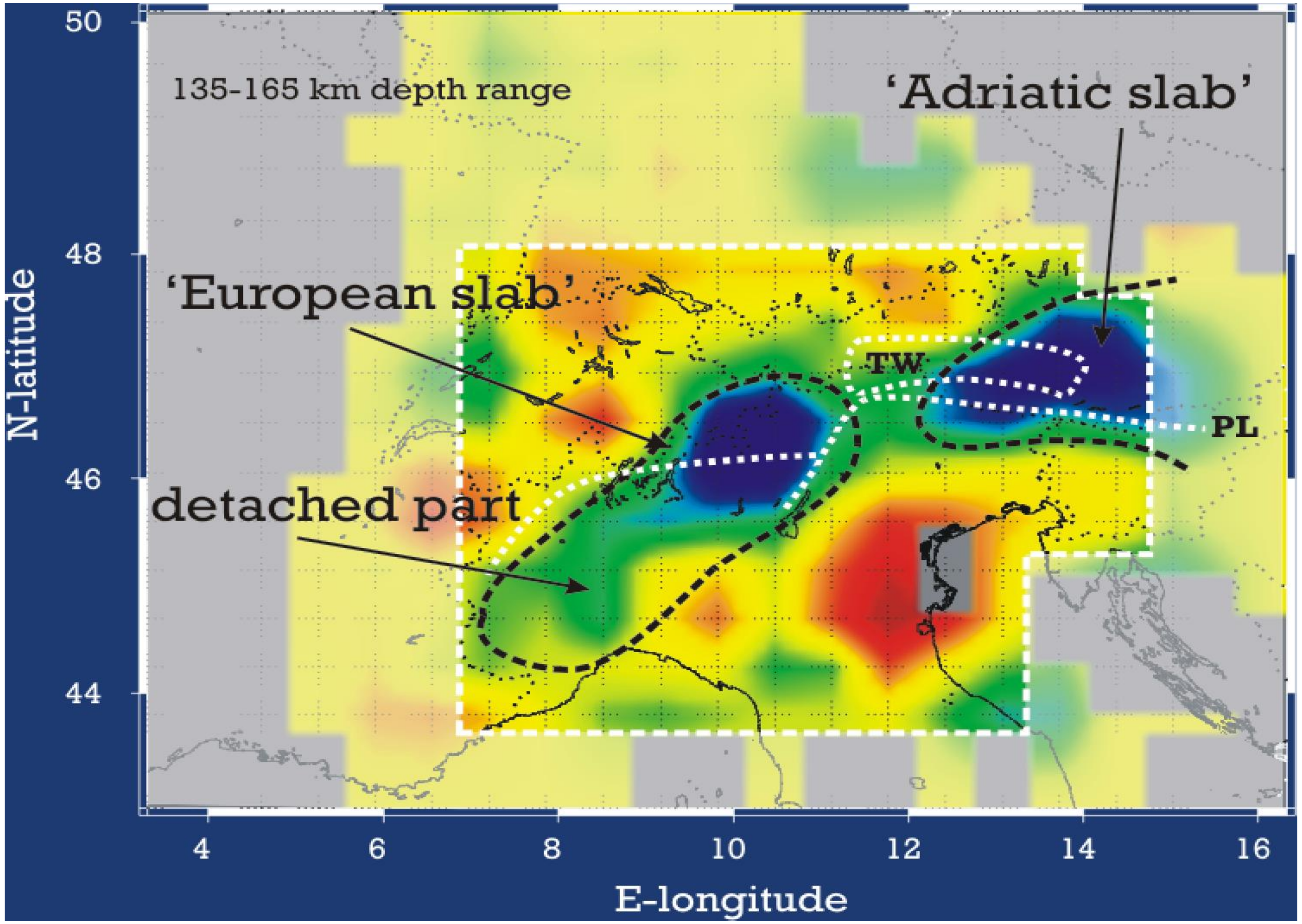


Tesauro et al., 2006, J. Geodyn. 42

- Ongoing crustal convergence of 1–2 mmyr across the Eastern Alps that is controlled by the counterclockwise rotation of the Adriatic plate
- In the Central and Western Alps only minor or no crustal shortening can be detected and earthquake focal plane solutions are dominated by extensional and strike-slip mechanisms.
- There is a change of orientation of the compressional strain rate axes from NW-SE to N-S, NNE-SSW along the Alpine chain.

Alpine Chain

Two lithospheric slabs beneath the Alps at 150 km depth

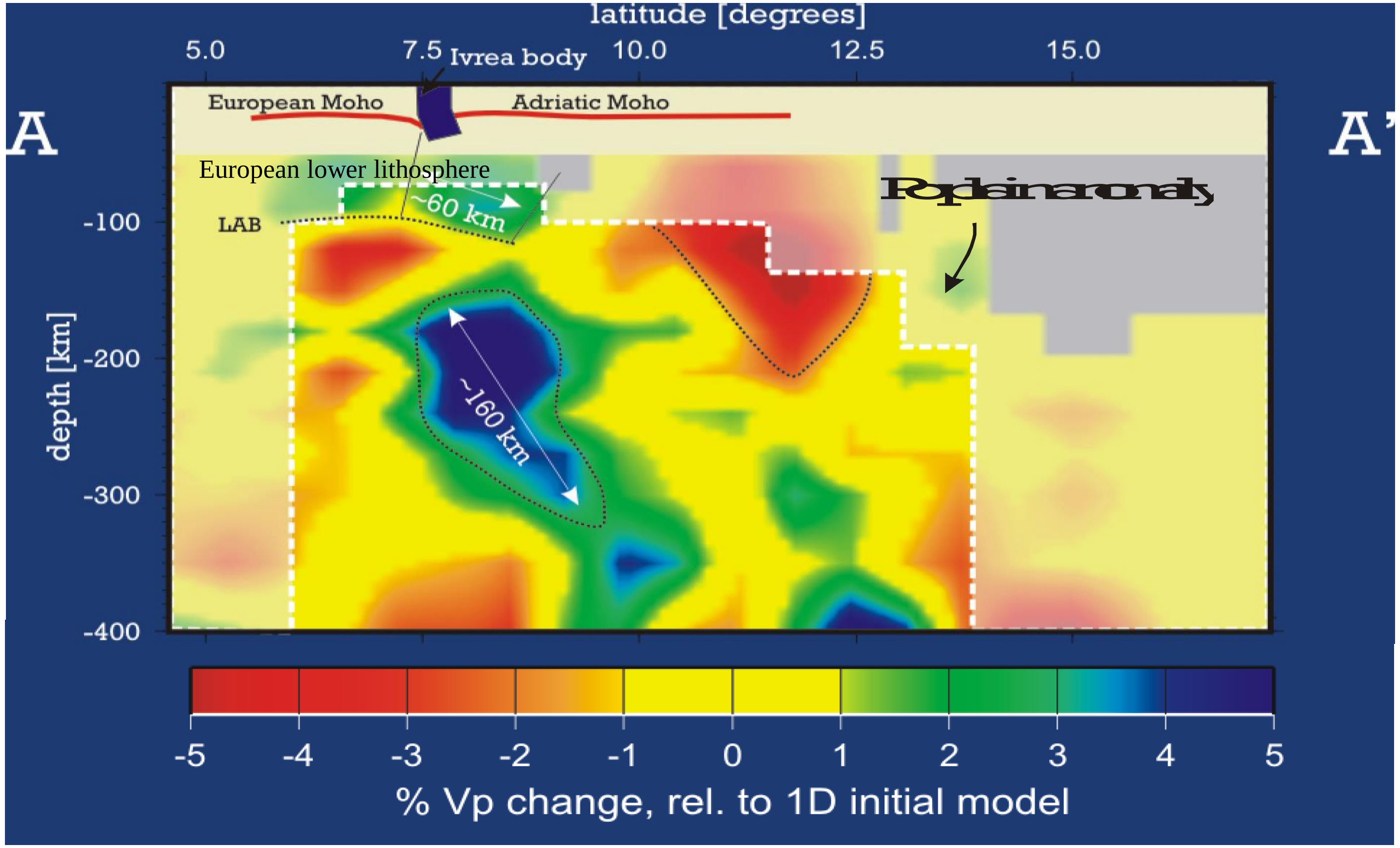


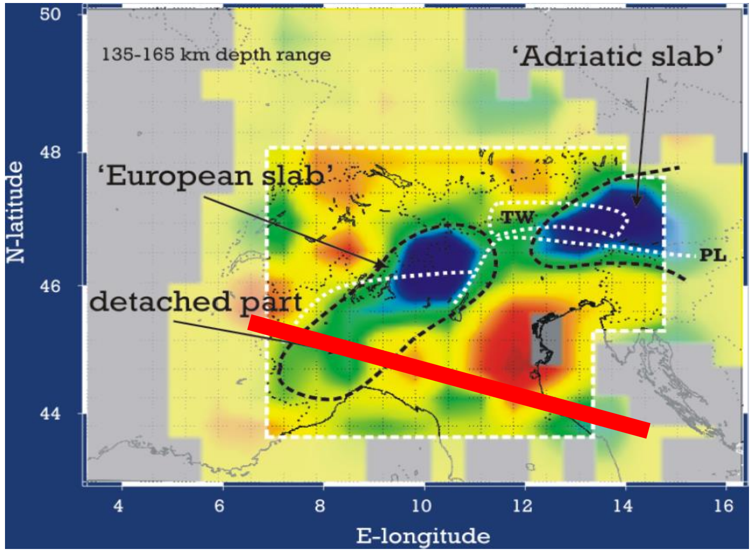
Lippitsch et al. 2003, JGR, 108

Alpine Chain

- At least 60 km of European lower lithosphere seem to be subducted beneath the Adriatic microplate: Before the actual collision in the late Eocene (40 Myr), at least two oceanic basins (Piemont- Ligurian and Valais) were subducted beneath the Adriatic microcontinent in the wider Alpine orogen.
- A clear gap appears between 110 and 150 km depth, which could indicate slab detachment from beneath the Ivrea body toward East, likely due to the Ivrea body, which acted as a buttress during the collision of the Adriatic and European plate, creating opposite buoyancy and forced a tear apart of the subducting slab.

Tear in E-Lithosphere beneath Western Alps?



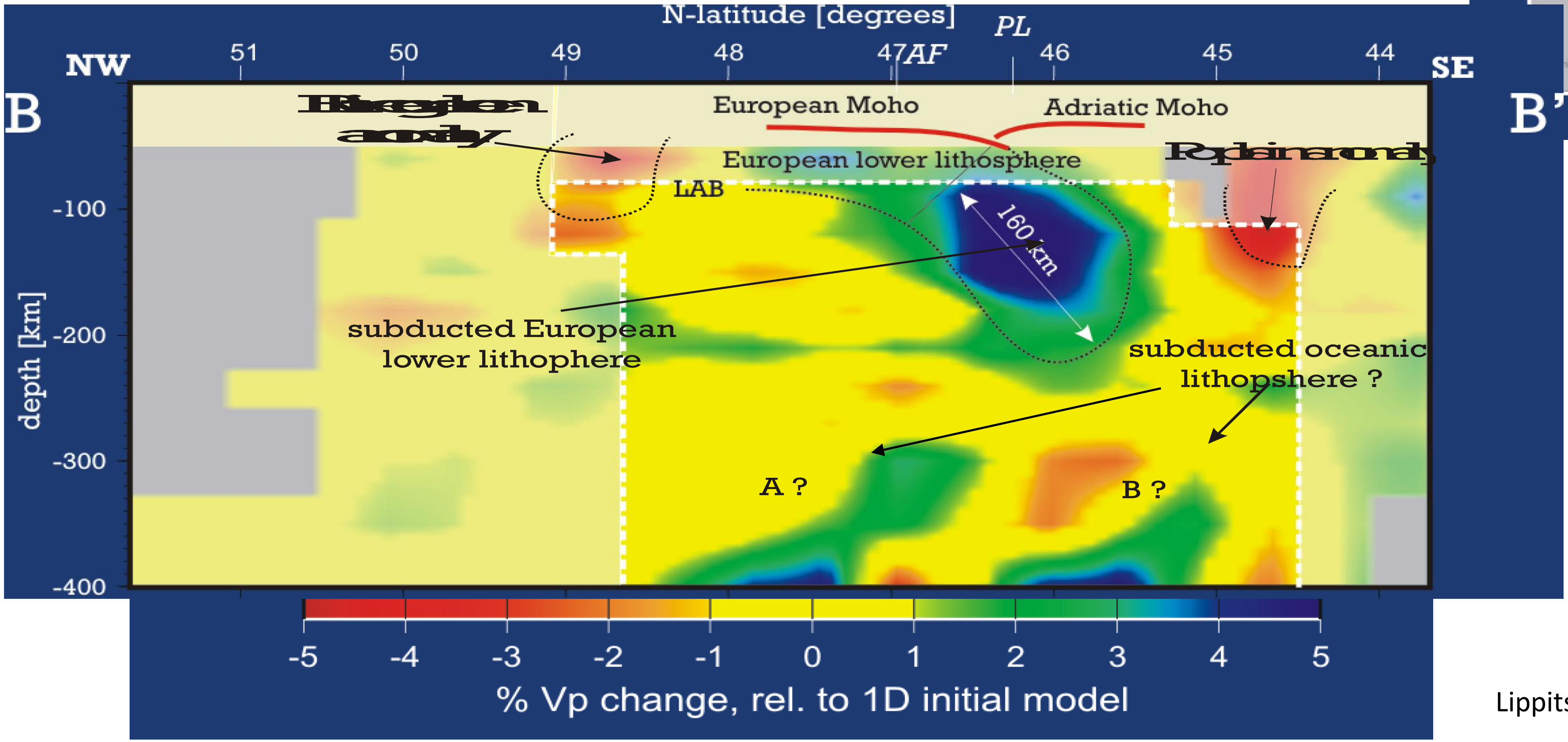


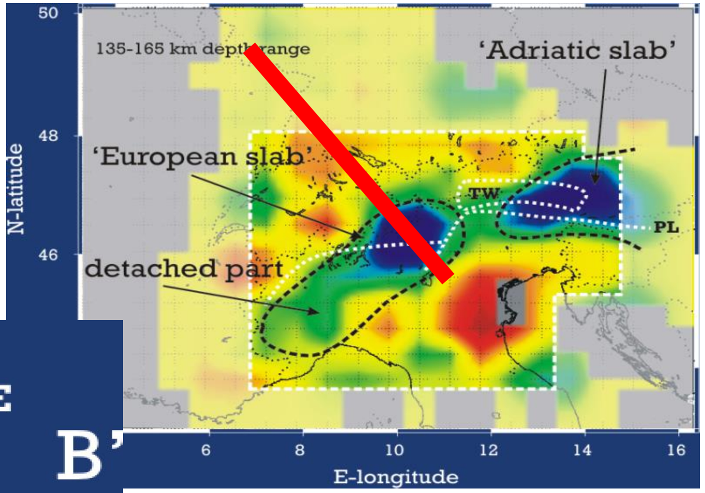
Lippitsch et al. 2003, JGR, 108

Alpine Chain

- In the central Alps the Adriatic lower crust is indenting European lithosphere, building an Adriatic lower crustal wedge, contrary to the western Alps where the Ivrea body acted as a buttress during the collision of Adriatic and European plates, preventing indentation from SE.
- From kinematic reconstructions, the estimated amount of postcollisional crustal shortening for the central Alps is about 164 km larger than in the western Alps and likely corresponds to 160 km lower lithospheric material that has subducted beneath the Adriatic microcontinent.

S-vergent subduction of E- mantle lithosphere beneath Central Alps



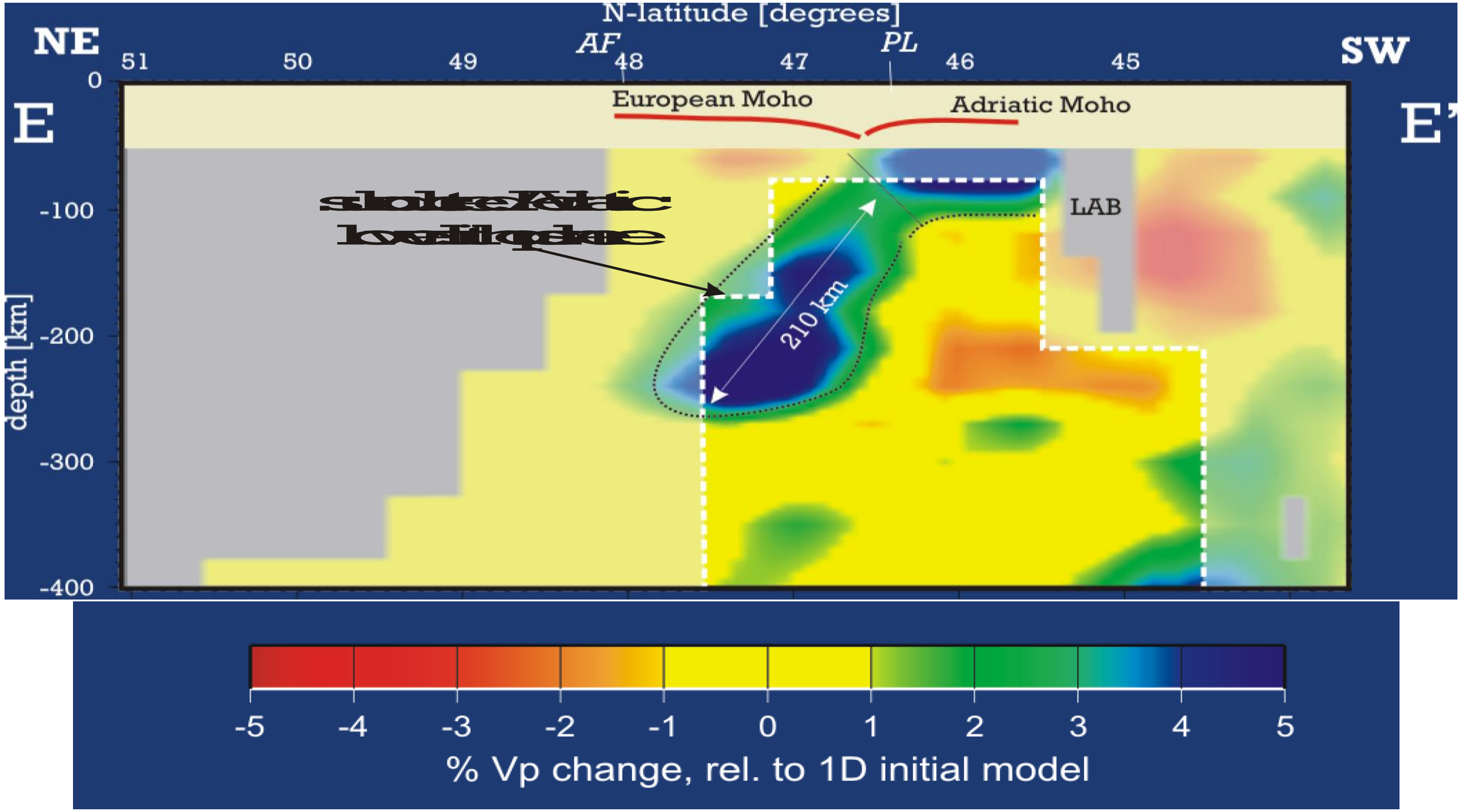


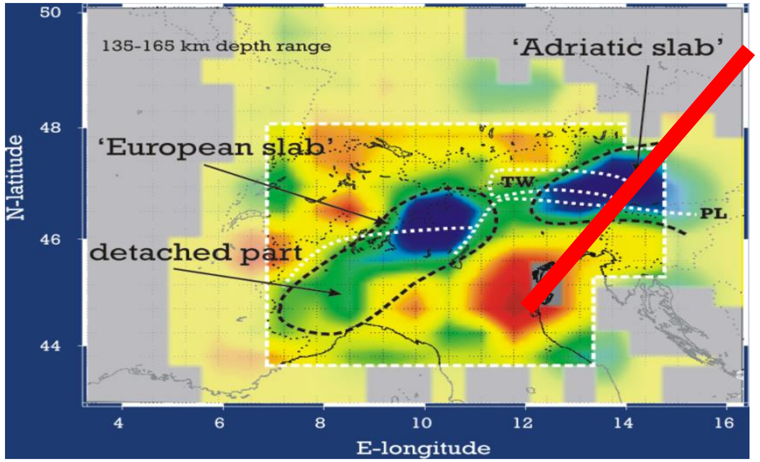
Lippitsch et al. 2003, JGR, 108

Alpine Chain

- Part of the Vardar ocean was subducted toward the north beneath the Austro- Alpine, forcing Adriatic continental lower lithosphere to subduct northeastward beneath the Austro-Alpine.
- Considering the rotation of the Adriatic microcontinent after collision, we must assume that the amount of subducted continental lithosphere increases toward the east.
- Crustal shortening of about 200 km occurred after collision at 40 Myr (100 km of shortening is estimated in the Friuli area in the last 20 Myr).

NE-vergent subduction of adriatic mantle lithosphere beneath Eastern Alps

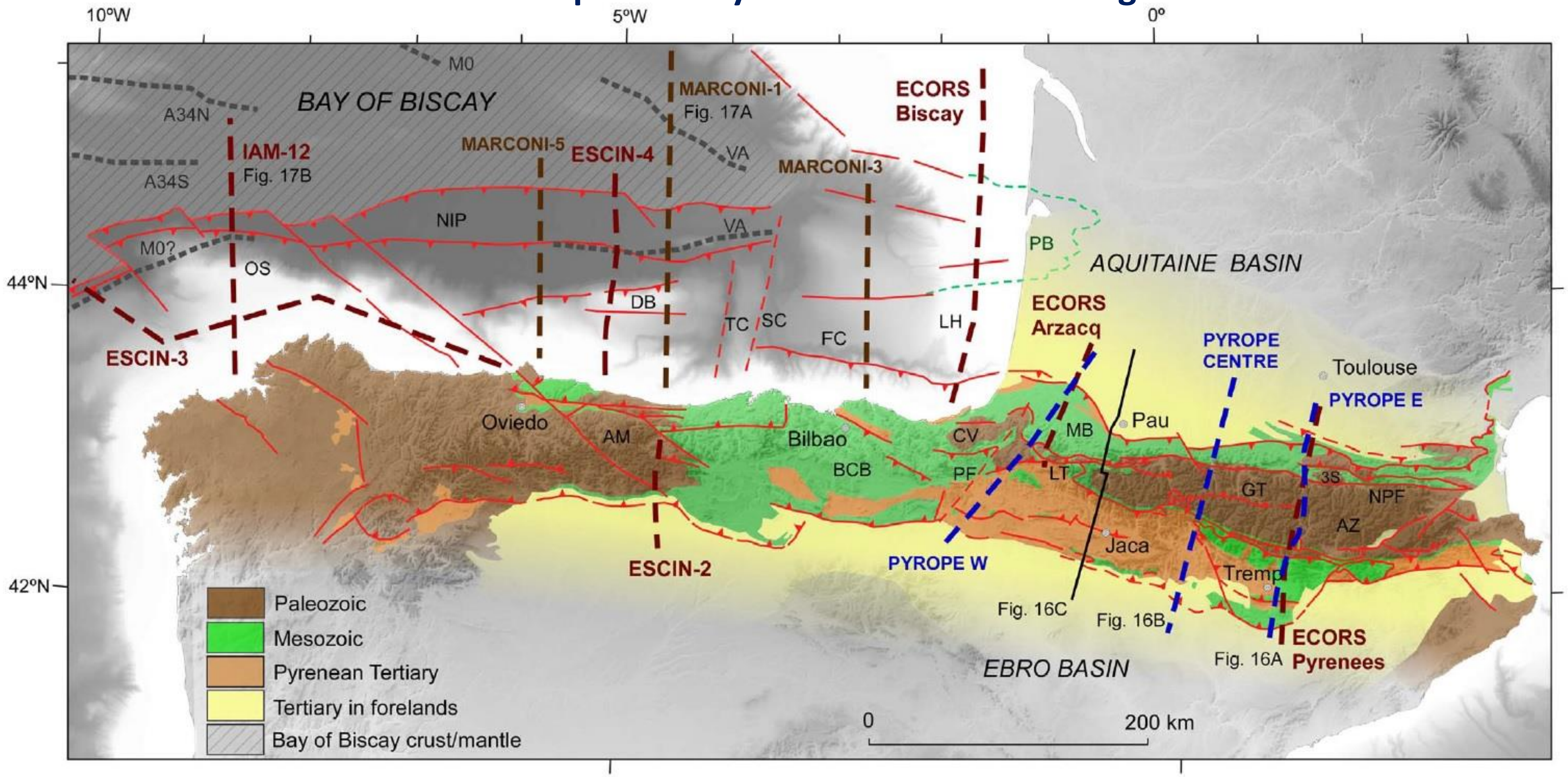




Lippitsch et al. 2003, JGR, 108

Pyrenean-Cantabrian orogenic belt

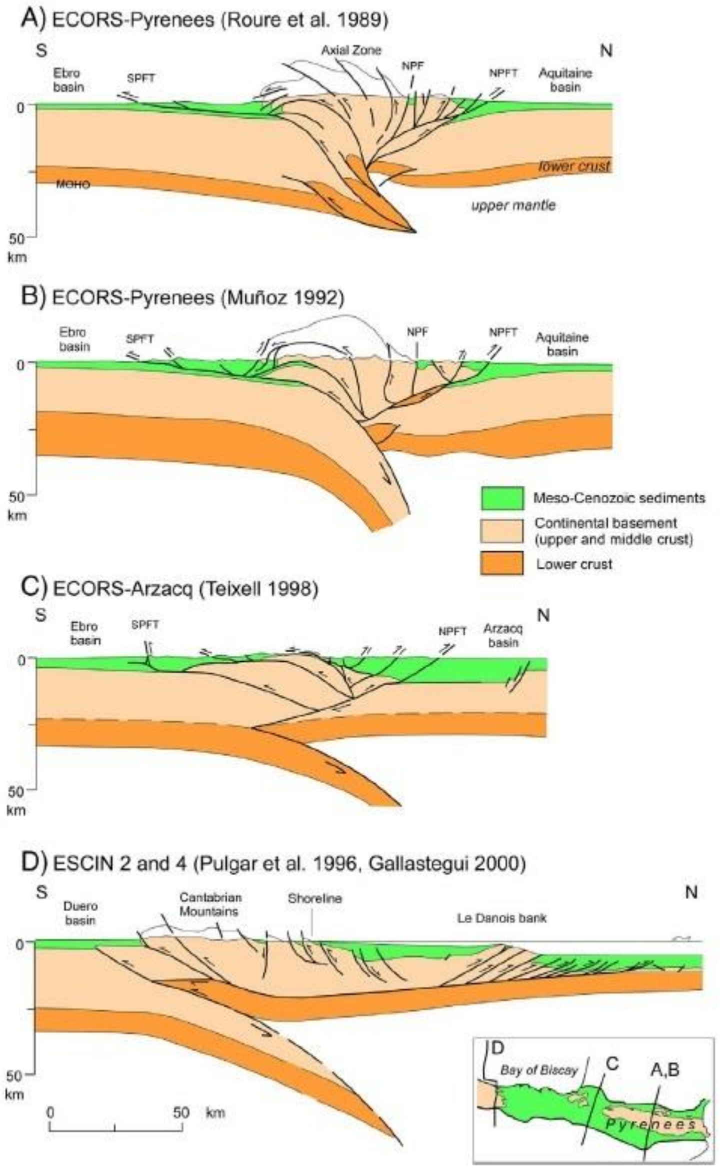
Tectonic map of the Pyrenean-Cantabrian orogenic belt

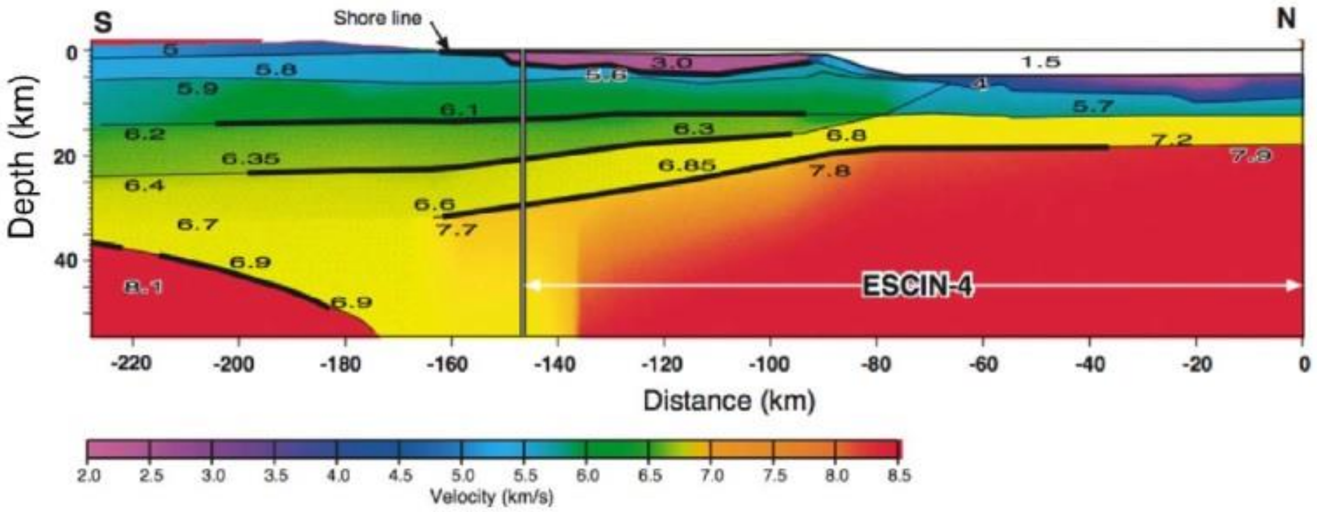


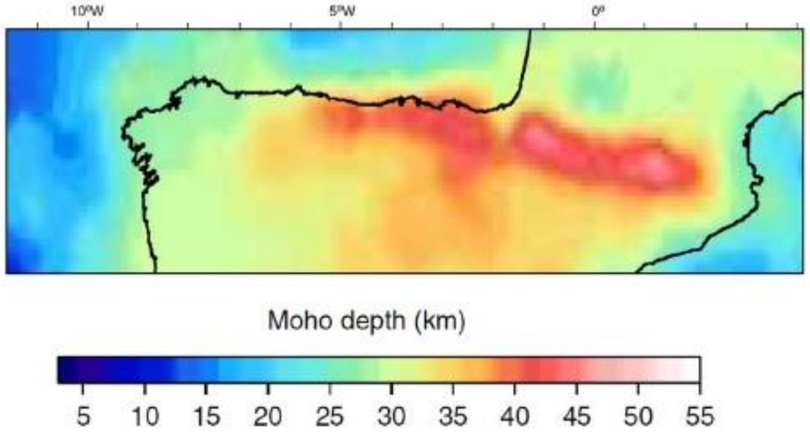
Teixell et al., 2018, Tectonophysics, 724-725

Brown: reflection profiles; blue: receiver function profiles; M0, A34N, A34S: labelled oceanic magnetic anomalies in the Bay of Biscay abyssal plain; VA: V-shaped magnetic anomaly of the eastern Bay of Biscay (traced after Sibuet et al., 2004); NIP: North Iberian thrust prism; OS: Ortegal spur; DB: Le Danois bank; TC, SC: Torrelavega and Santander canyons (transfer zone); FC: Cap Ferret canyon; PB: Parentis basin; LH: Landes high; AM: Asturian massif; BCB: Basque-Cantabrian basin; CV: Cinco Villas massif; PF: Pamplona fault (transfer zone); MB: Mauléon basin; LT: Lakora thrust; GT: Gavarnie thrust; AZ: Axial Zone; NPF: North Pyrenean fault; 3S: Trois-Seigneurs massif.

Pyrenean-Cantabrian orogenic belt





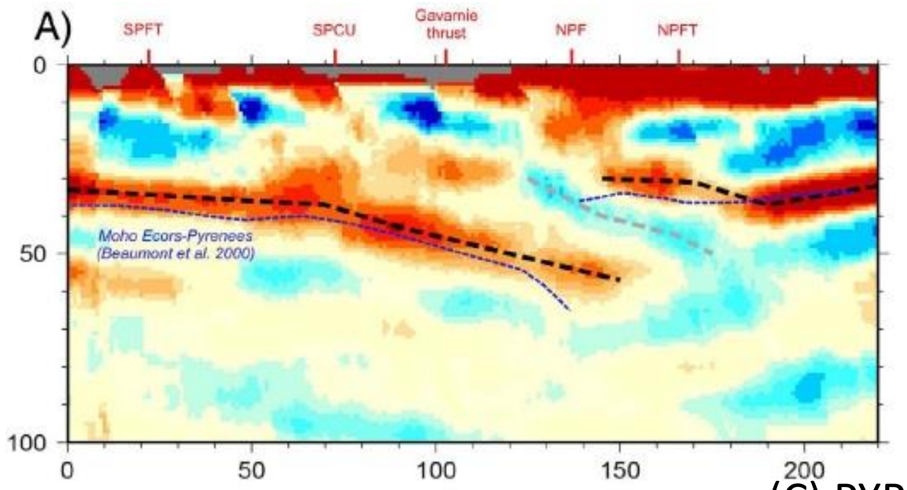


Teixell et al., 2018, Tectonophysics, 724-725

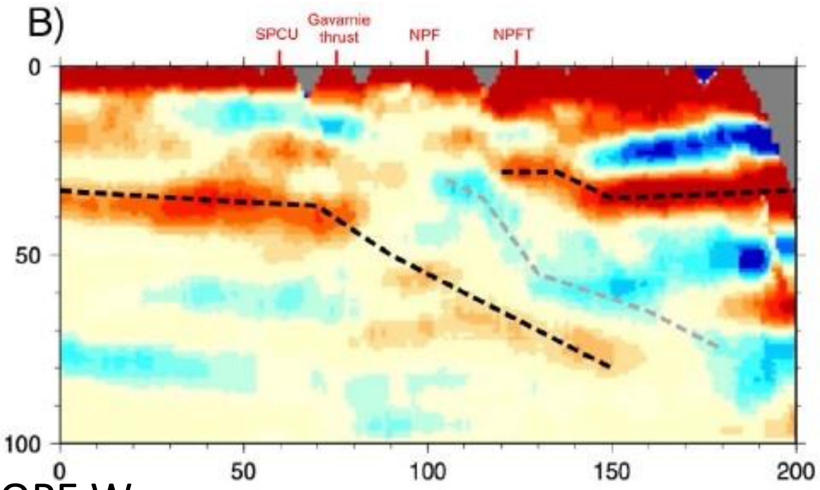
- As in the case of the central Pyrenees, the crust of the northernmost Cantabrian Mountains and margin is interpreted to indent into the Iberian plate and to force the northward subduction of its lower part

Pyrenean-Cantabrian orogenic belt

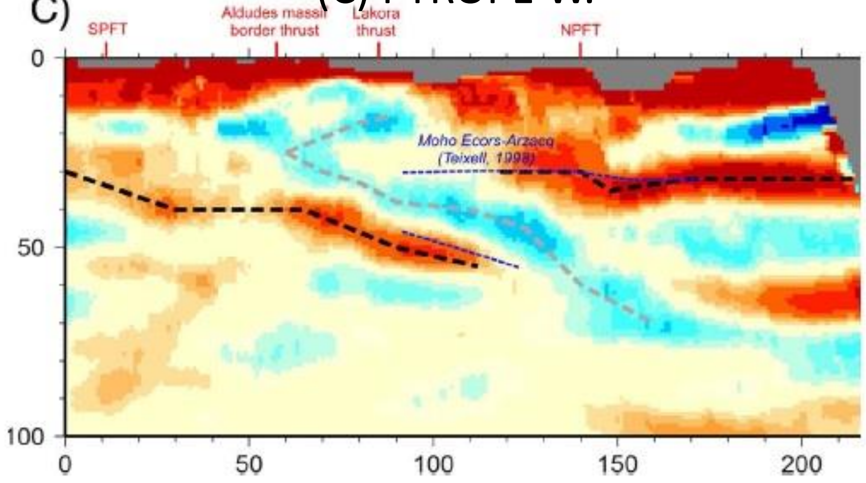
(A) PYROPE E



(B) PYROPE Centre



(C) PYROPE W.



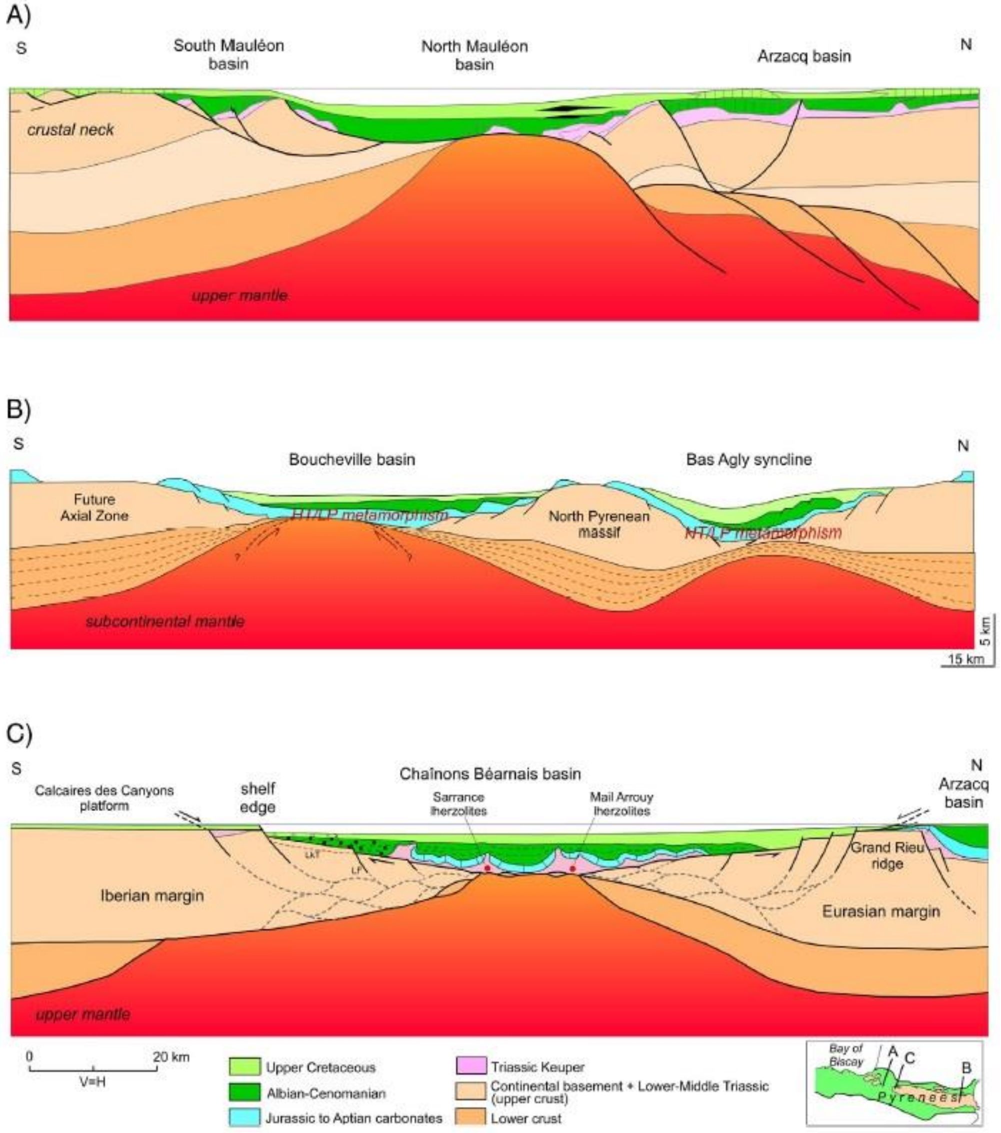
Teixell et al., 2018, Tectonophysics, 724-725

Black and gray dashed lines indicate the positive and negative polarity interfaces attributed to the Moho and the major thrust/subduction zones (intracrustal variscan crust), respectively.

- The thickness of the slab varies between 14 and 17 km in average in the eastern and central profiles, and is somewhat thinner (ca. 12 km) in the western profile.
- The angle of subduction in the eastern and central profiles ranges between 22 and 30° whereas in the western profile it is<20°.

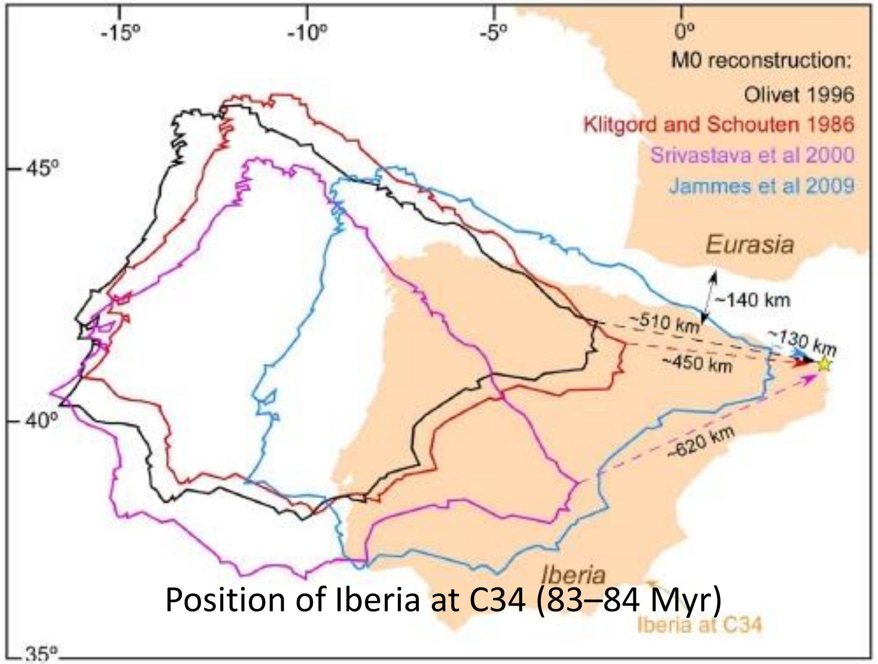
Pyrenean-Cantabrian orogenic belt

Models for the Cretaceous continental margins of the Pyrenees, before the onset of plate convergence around 84 Myr.



Teixell et al., 2018, Tectonophysics, 724-725

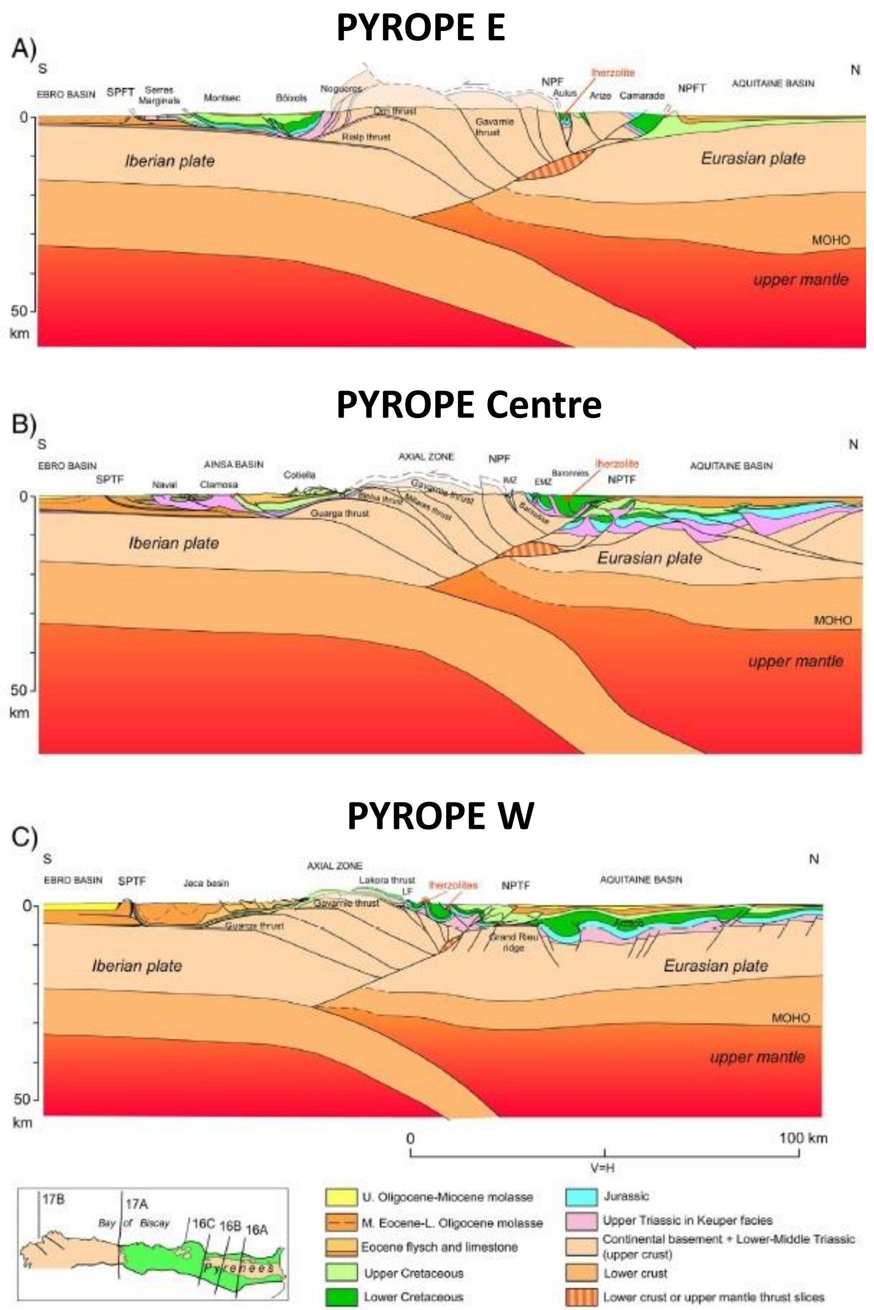
Reconstruction of the Iberian plate with respect to Eurasia (Aptian, 121–125 Myr)



Position of Iberia at C34 (83–84 Myr)

- The present-day architecture of the Pyrenean-Cantabrian orogenic belt is strongly influenced by the precursor Mesozoic rift system, which accumulated several thousands of meters of dominantly marine sediments in Cretaceous times and lead to exhumed mantle between the two plates.
- Cretaceous extension led to a highly segmented continental rift in the present Pyrenees and eastern Bay of Biscay, that passed laterally into (short-lived) oceanic spreading in the western Bay of Biscay.
- Only the southern margin of the Bay of Biscay was subjected to (moderate) compressional deformation, so the Bay of Biscay oceanic Basin is preserved. The rest of the Mesozoic basin system was intensely inverted during the Pyrenean orogeny, and information has to be retrieved from map and section restoration.

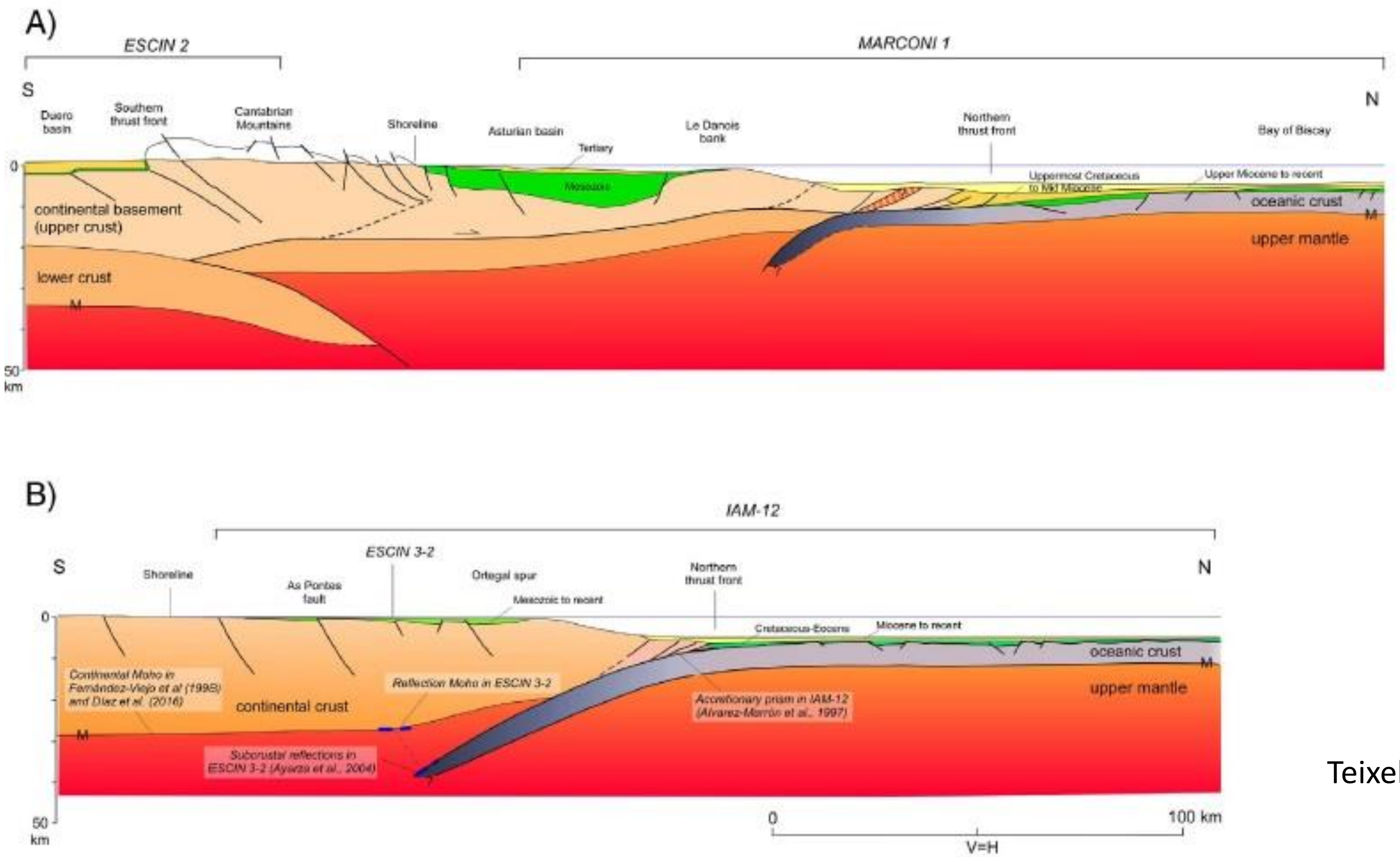
Pyrenean-Cantabrian orogenic belt



- After initial closure of the exhumed mantle domain in the Late Cretaceous, the Pyrenees raised from the collision of the Iberian and Eurasian margins 10–30 Myr after the initiation of convergence, without previous events of oceanic subduction.
- Much of the structure of the Pyrenees is that of a shortened rift, with inversion of marginal extensional faults or low-angle detachments.
- Basement stacking of the Iberian plate resulted in the central high-relief belt of the Axial Zone.
- Seismic data image a north directed subduction of a slab of Iberian lower crust to a depth of 60–80 km, which accounts for only a fraction of the total orogenic convergence (> 100 km).

Pyrenean-Cantabrian orogenic belt

Cantabrian Mountains, North Iberian margin, and Bay of Biscay

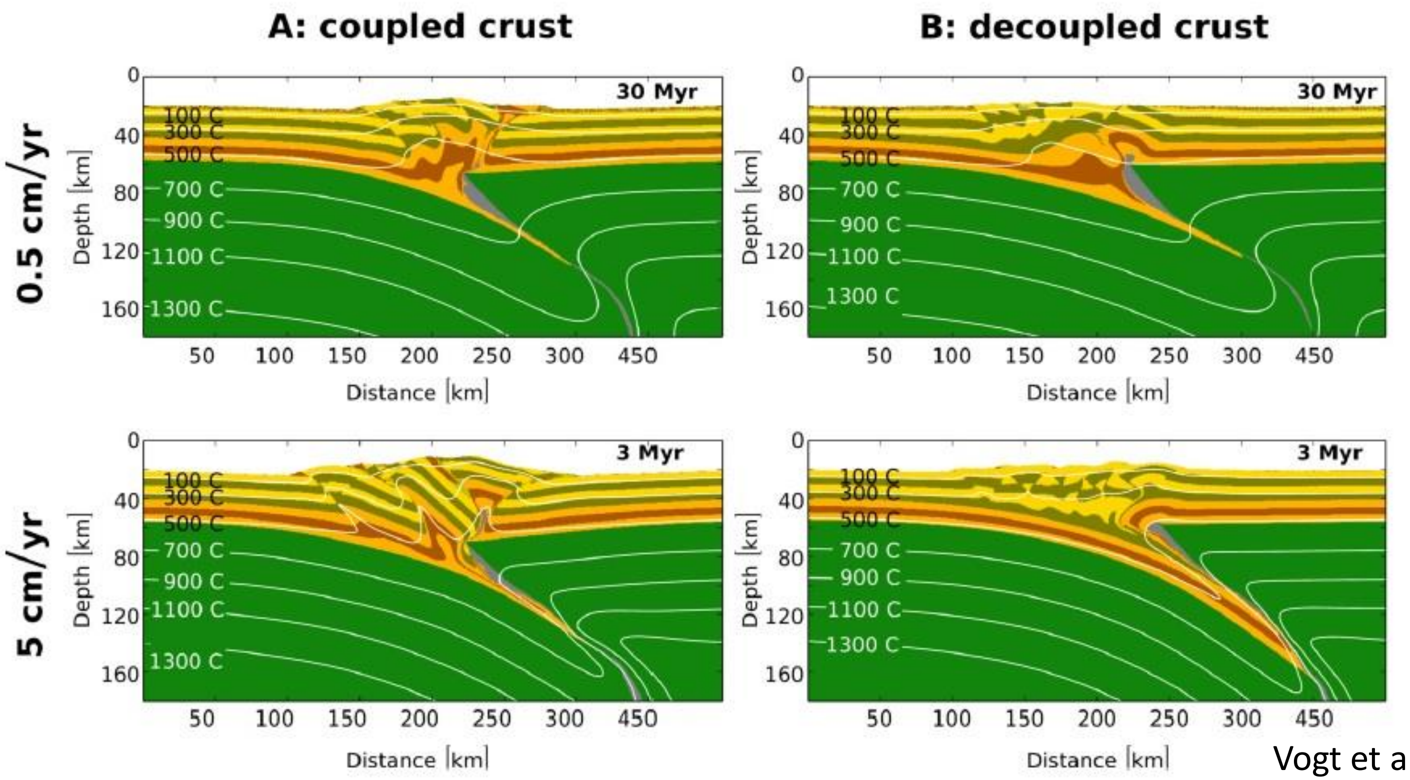


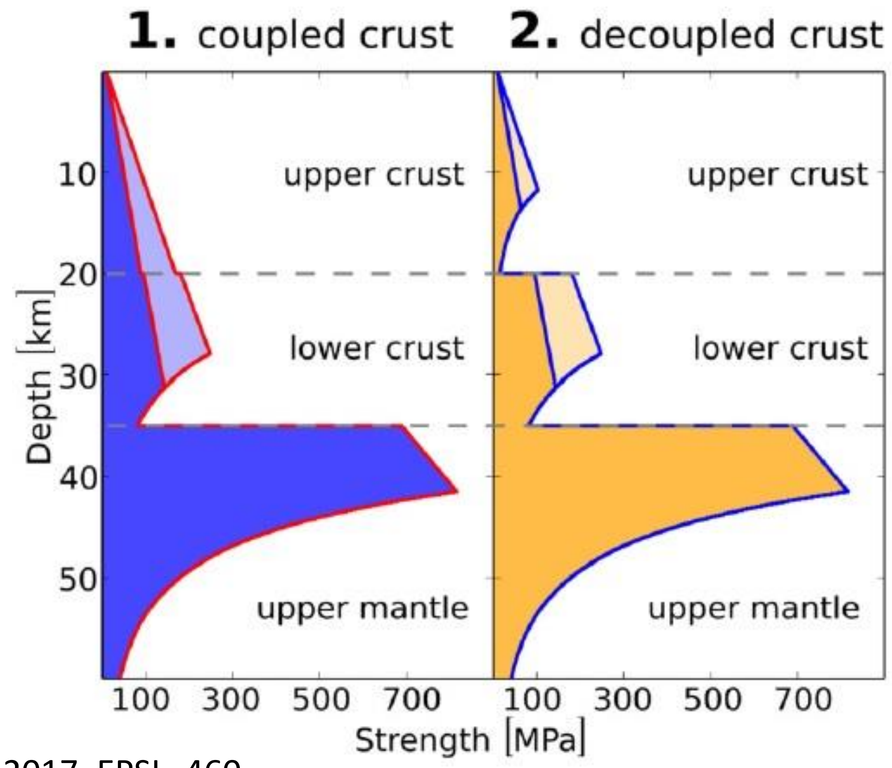
Teixell et al., 2018, Tectonophysics, 724-725

- The central Cantabrian Mountains are also underlain by a north-plunging slab of Iberian lower crust, as it implies deep-seated shortening of the proximal continental margin of the Bay of Biscay ocean without a clear collision driver.
- The deep structure of the Cantabrian Mountains is thus similar to the Pyrenees, although the continental subduction probably did not reach depths beyond 45 km.
- A limited subduction of the Bay of Biscay under the North Iberian margin is here supported, at least west of 4.5°W.

Continental Subduction

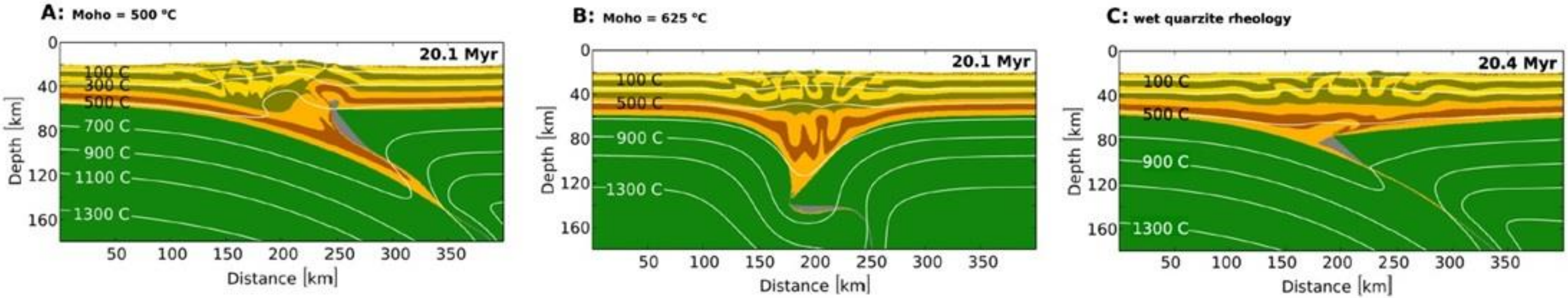
- **Deformation in continent–continent collision zones may result in double vergent orogens** (thrust displacements are directed outwards from the core of the orogeny on both flanks) **or continental subduction**, depending on the rheological structure of the continental lithosphere.
- Strong rheological coupling of upper and lower continental crust results in the formation of a décollement at Moho depth that separates the downgoing mantle lithosphere from the accreting continental crust double vergent orogens.
  Example of these collisional orogens: European Central Alps.
- Low rheological coupling of upper and lower continental crust forms a décollement at mid-crustal levels, which allows only the upper continental crust is accreted frontally, while the lower crust sinks into the mantle (continental subduction).
  Example of these collisional orogens: Carpathians, Dinarides, Apennines.
- High convergence in case of coupled crust (**case A**) increases the overall width of the orogeny, inhibits exhumation of lower crust in the back of the orogen. High convergence in case of decoupled crust (**case B**) promotes subduction of lower continental crust and the resulting orogeny is wide.





Vogt et al., 2017, EPSL, 460

Continental Subduction



Convergence rate: 1 cm/yr

Vogt et al., 2017, EPSL, 460

- Strength reduction with increasing temperature (**case B**) causes continental thickening rather than subduction and the orogeny is narrow, symmetric and deeply rooted (Precambrian conditions).
- Strength reduction due to a weak rheology of the entire crust (**case C**) causes crustal thickening (less pronounced than in the previous case because of the higher mantle strength), while lower crust resists subduction and spreads along the former suture zone.

Surface processes influence the structural architecture of orogens:

- Sedimentation increases the thickness of the brittle upper crust and results in wider orogens, while erosion favours localization of deformation (reduces the thickness of the brittle upper crust and enhances rock uplift/exhumation), forming narrower orogens.

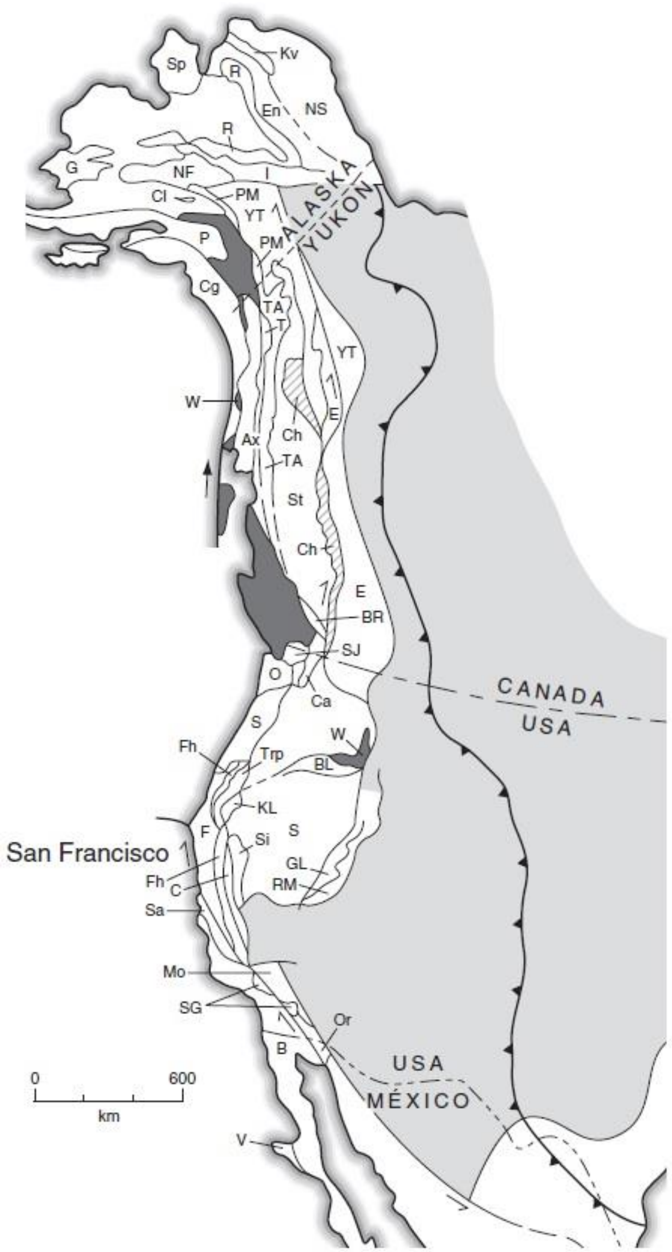
Accretionary Orogens

The *terranes* composing the accretionary orogens may range in size from a few hundreds to thousands of square km and are distinguished on the base of:

1 Provenance, stratigraphy, and sedimentary history;
2 Petrogenetic affinity and the history of magmatism and metamorphism;
3 Nature, history, and style of deformation;
4 Paleontology and paleoenvironments;
5 Paleopole position and paleodeclination.

On the base of their composition the terranes can be distiguished:
1 Turbidite terranes characterized by thick piles of land-derived sediment that are transported offshore by density currents and deposited in deep marine environment.
2 Tectonic and sedimentary *mélange* terranes consisting of a heterogeneous assembly of altered basalt and serpentinite, limestone, shale, and metamorphic rock fragments in a fine grained, highly deformed, and cleaved mudstone matrix.
3 Magmatic terranes, which may be predominantly mafic (basalts, ophiolites generated by seafloor spreading, LIP formation, arc volcanism, ocean islands) or felsic (calc-alkaline plutonic rock and dispersed fragments of old continental crust) according to the environment in which they form.
4 Nonturbiditic clastic, carbonate, or evaporite sedimentary terranes.
5 Composite terranes, which consist of a collage of terranes of any variety that amalgamated prior to accretion onto a continent.

Accretionary Orogens
Cordillera of Western North America



Principal terranes

Alaska and Western Canada

NS	North Slope
Kv	Kagvik
En	Endicott
R	Ruby
Sp	Seaward Peninsula
I	Innoko
NF	Nixon Fork
PM	Pingston and McKinley
YT	Yukon – Tanana
CI	Chuiitna
P	Peninsular
W	Wrangellia
Cg	Chugach and Prince William
TA	Tracy Arm
T	Taku
Ax	Alexander
G	Goodnews
Ch	Cache Creek
St	Stikine
BR	Bridge River
E	Eastern assemblages

Washington, Oregon and California

Ca	Northern Cascades
SJ	San Juan
O	Olympic
S	Siletzia
BL	Blue Mountains
Trp	Western Triassic and Paleozoic of Klamath Mountains
KL	Klamath Mountains
Fh	Foothills Belt
F	Franciscan and Great Valley
C	Calaveras
Si	Northern Sierra
SG	San Gabriel
Mo	Mohave
Sa	Salinia
Or	Orocopia

Nevada

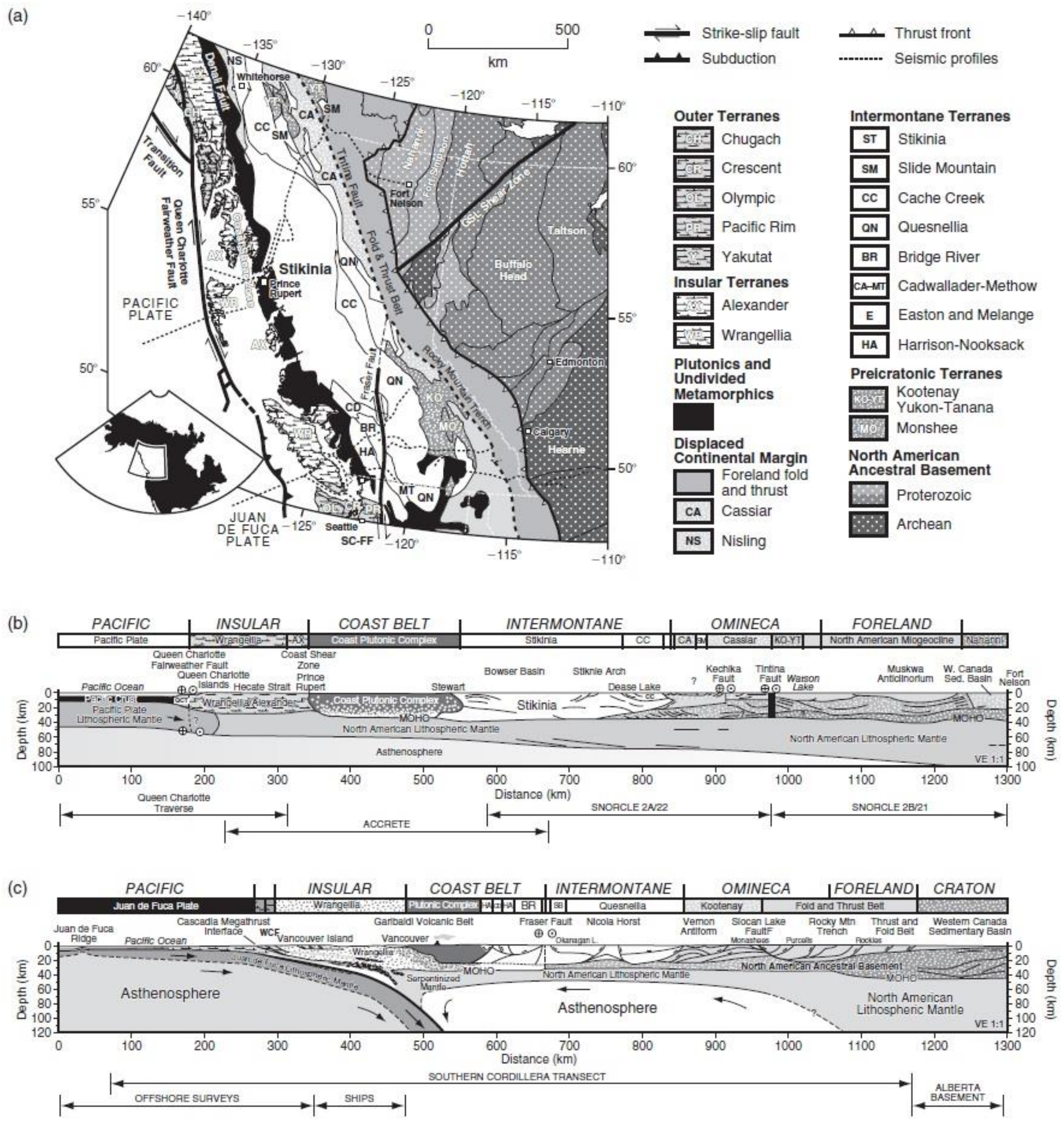
S	Sonomia
RM	Roberts Mountains
GL	Golconda

Mexico

B	Baja
V	Vizcaino

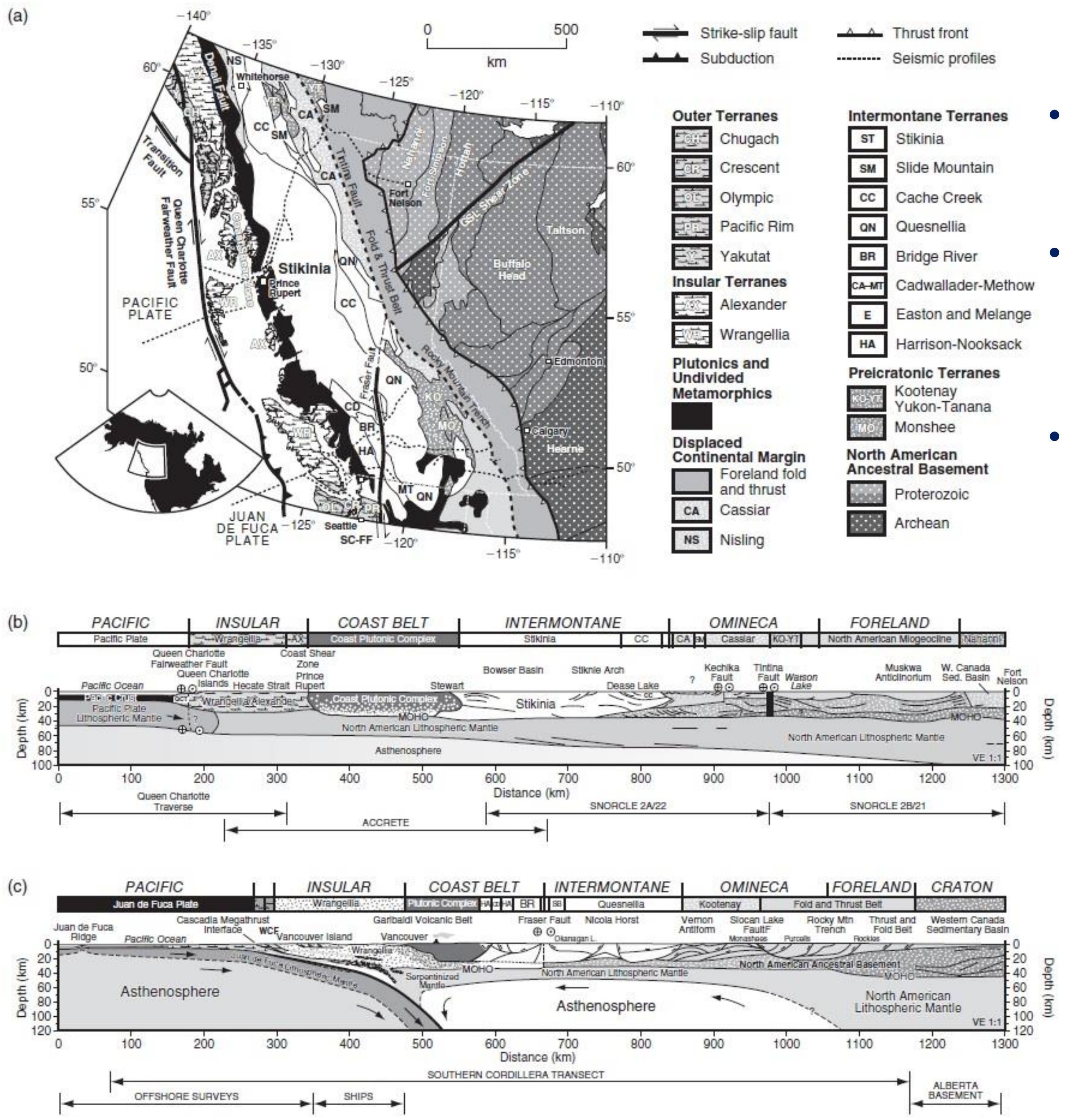
- The distribution of terranes composing the Cordillera of Western NA forms a zone ~500 km wide (30% of the continent).
- Most of the terranes in the Cordillera accreted onto the margin of ancestral North America during Mesozoic times and some of them experienced lateral translations along strike-slip faults.
- Following the amalgamation of the Canadian Shield during the Proterozoic, a number of rifting events between 1.74 Gyr and the Middle Devonian created thick passive margin sequences that were deposited on top of Proterozoic crust of the North American craton.
- During the Middle Jurassic, a composite Terrane (the Intermontane Superterrane), began to accrete onto the continental margin. The collision shortened the passive margin sequences and translated them eastward forming a foreland fold and thrust belt (Eastern Cordillera).
- West of the foreland, the Omineca belt consists of highly deformed and metamorphosed rocks of Middle Jurassic age, representing the suture zone created by the Intermontane–North American collision.

Accretionary Orogens
Cordillera of Western North America



- During the Late Cretaceous, another composite Terrane (the Insular Superterrane), consisted mostly of two island arc terranes accreted to the western NA margin.
- The arrival of the Insular Superterrane deformed the interior of the North American continent and formed a major part of the Coast belt. Prior to and during the amalgamation, subduction beneath the margin formed the Coast Plutonic Complex by magma addition.
- Most of the accreted terranes crustal thickness almost uniform across the entire Cordillera, ranging between 33 and 36 km and a heterogeneous seismic velocity.
- The Moho remains mostly flat regardless of the age of crustal accretion or of the last major tectonic deformation. Lateral changes in crustal thickness tend to be gradual, with abrupt variations occurring at major terrane boundaries.
- Lithospheric thickness is unusually thin and gradually thickens to the east beneath the Precambrian shield.

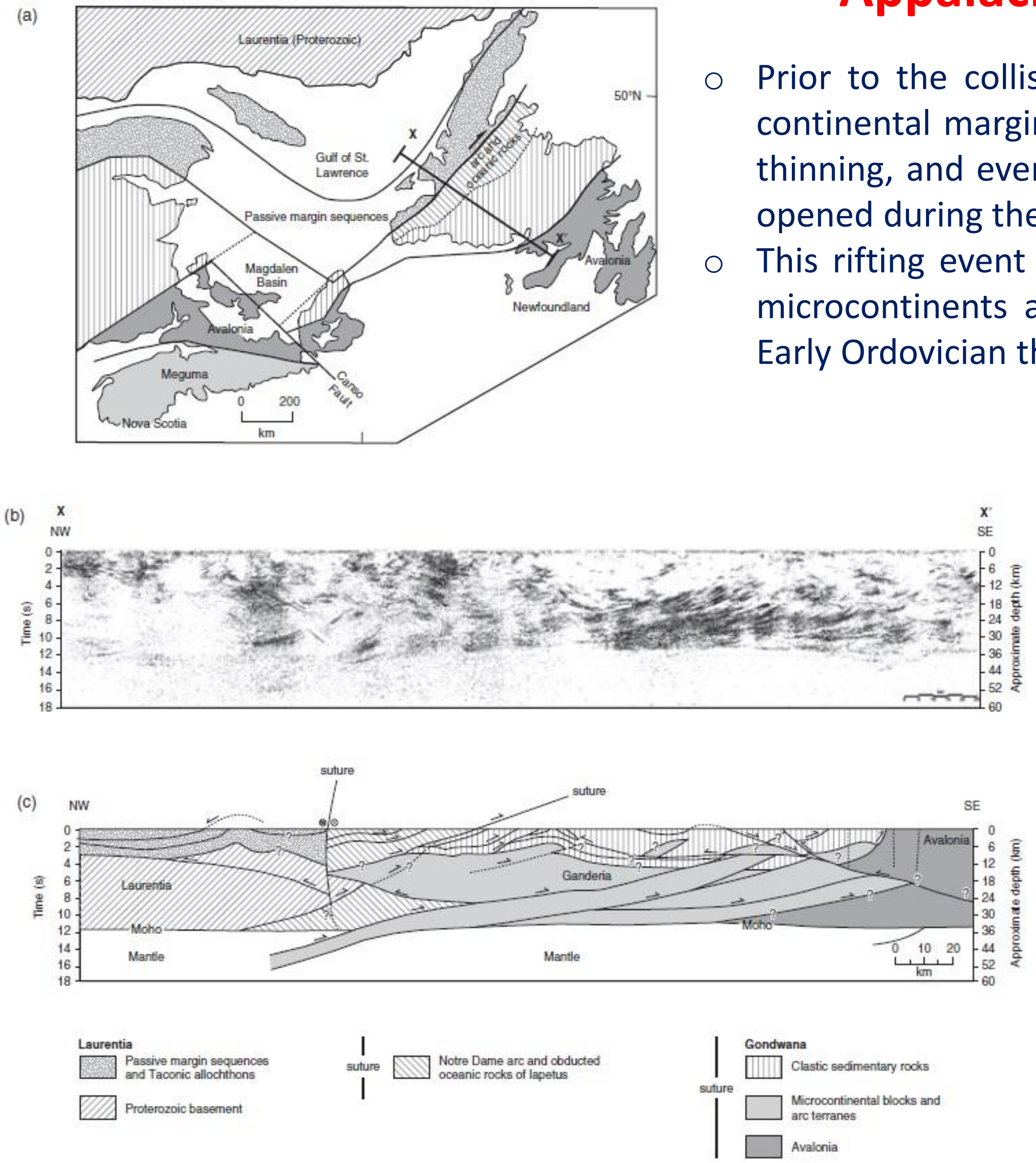
Accretionary Orogens
Cordillera of Western North America



- Subduction occurs beneath Southern Cordillera and thus this southern part of the margin shows shortening and crustal thickening in the forearc region and an active volcanic arc within the Coast belt.
- The mantle lithosphere shows evidence of hydrothermal alteration (serpentinization) in the upper mantle wedge beneath the arc and substantial thinning toward the interior of the continent, reflecting processes associated to subduction.
- Strike-slip displacements also accommodated some relative motion between the accreted terranes and North America in the Canadian Cordillera (e.g., Tintina fault, a major lithospheric-scale structure).

Accretionary Orogens
Appalachian Orogen

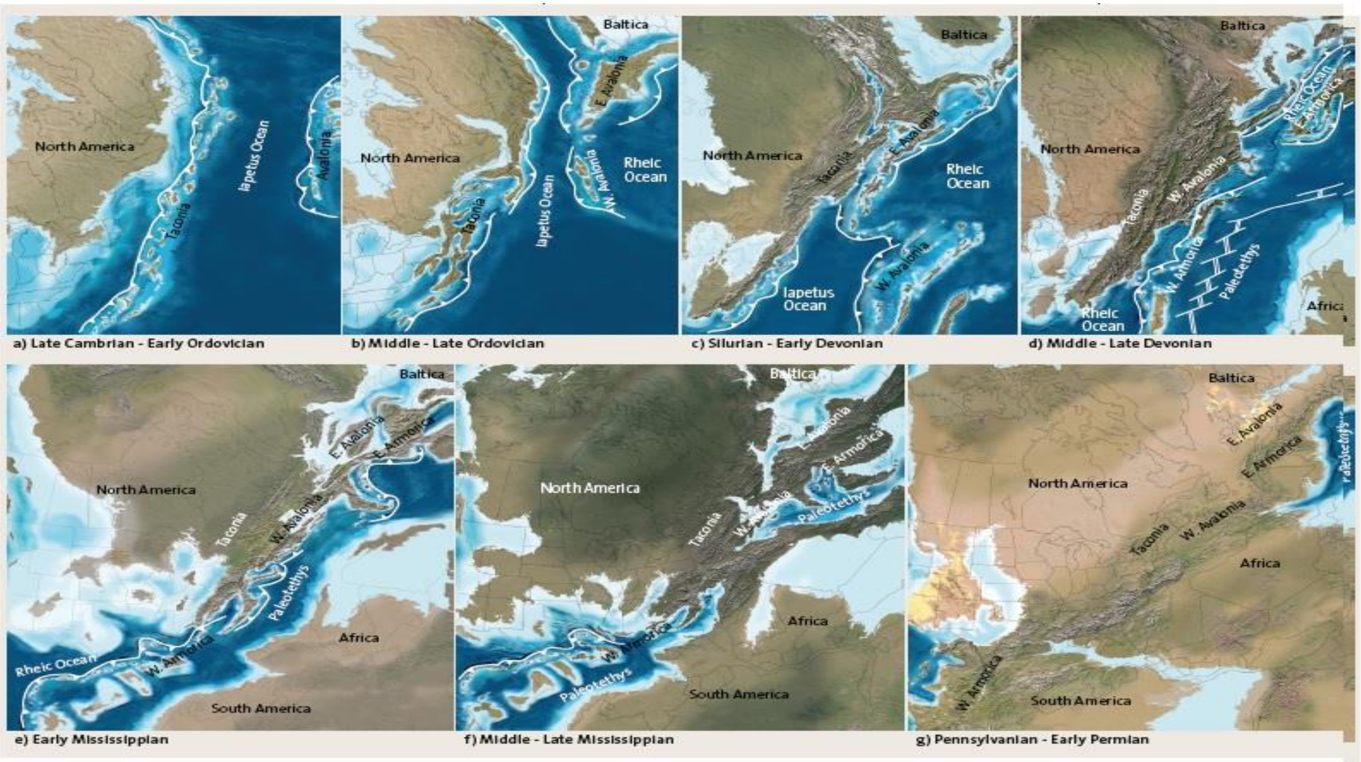
Newfoundland (eastern Canada)



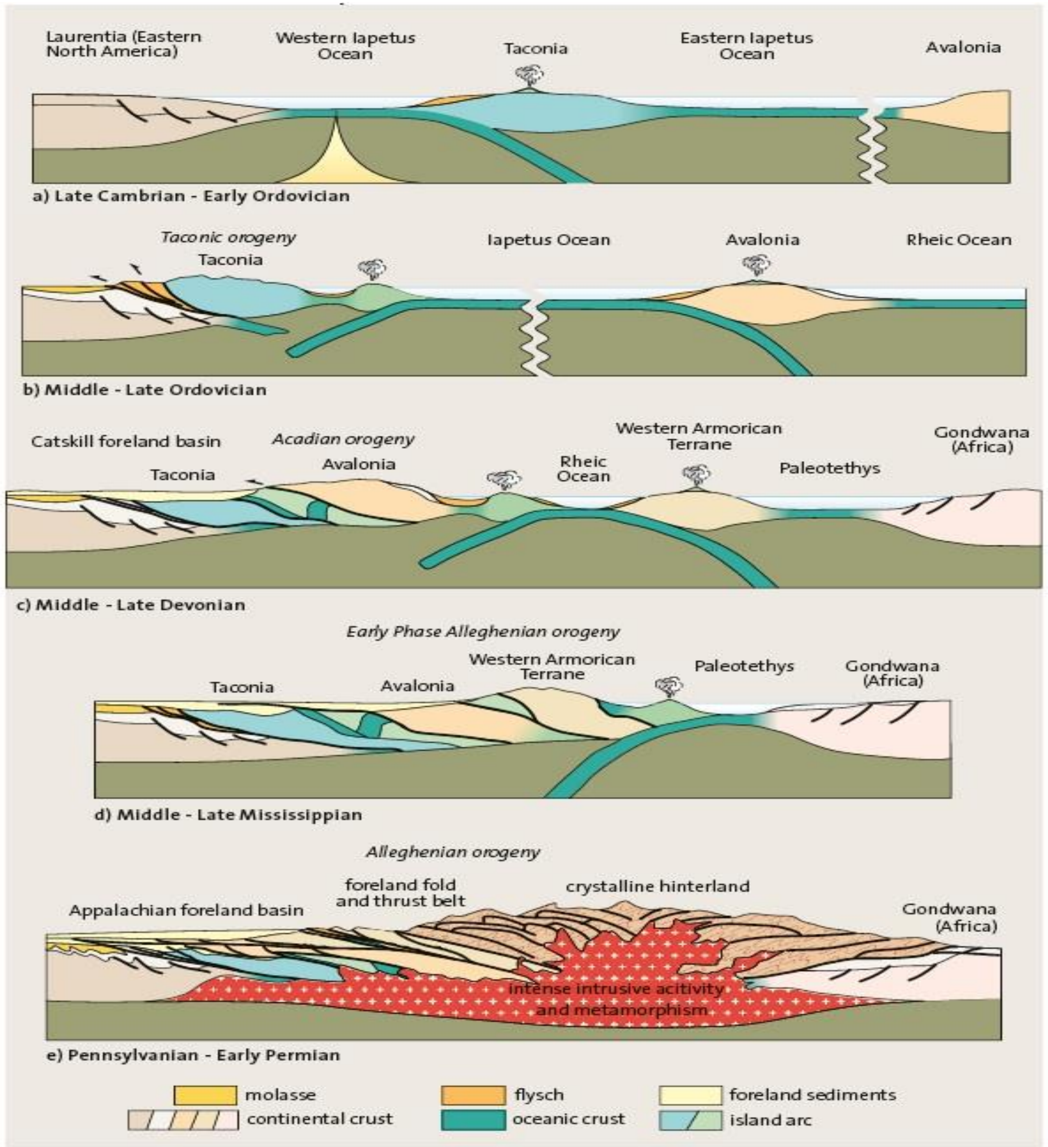
- Prior to the collision, thick sequences of sedimentary rock were deposited on a passive continental margin located outboard of the craton. These sequences record the stretching, thinning, and eventual rupture of Proterozoic continental lithosphere as the Iapetus Ocean opened during the Late Proterozoic and Early Cambrian.
- This rifting event was followed by a series of terrane collisions and accretionary cycles of microcontinents and composite terranes rifted from northwestern Gondwana during the Early Ordovician that formed the Paleozoic orogenies of the Appalachian Mountains.

- From seismic reflection data clearly mark the location of an old Ordovician–Devonian subduction zone. Above and to the east of the paleosubduction zone there are series of dipping thrust faults and tectonic wedges composed of interlayered slices of the amalgamated terranes.

Accretionary Orogens
Appalachian Orogen



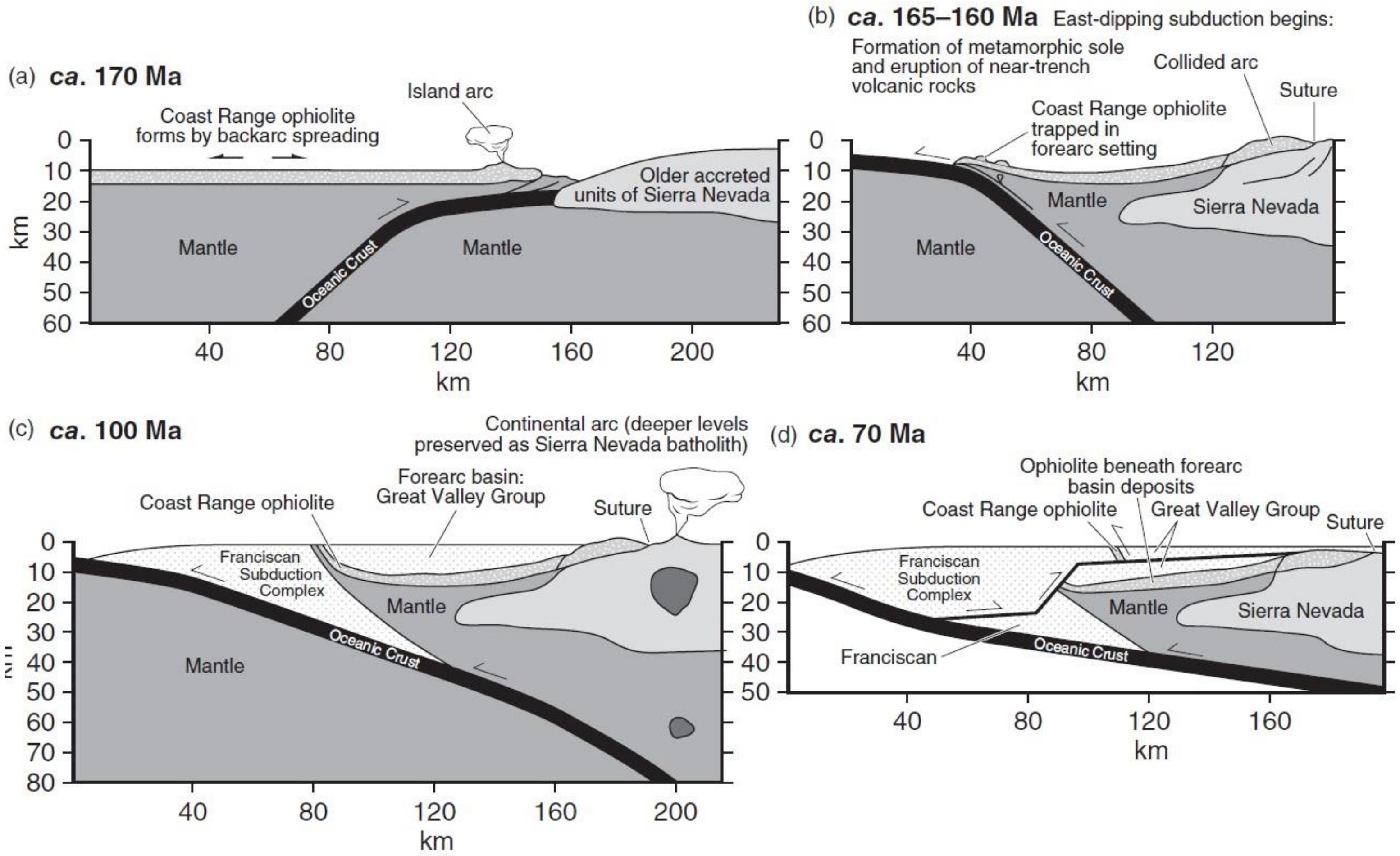
(**a**) Iapetus Ocean prior to Taconic orogeny; (**b**) Taconic orogeny; (**c**) Avalonia collides with Laurentia during Acadian orogeny and closure of Iapetus Ocean as Hun Terrane (Western Armorican Terrane) rifts from Gondwana; (**d**) accretion of Hun Terrane; late phase of Acadian (north) to early phase of Alleghenian (south) orogenies; (**e**) Alleghenian orogeny marks final collision of Gondwana and Laurentia to form Appalachian Mountains and Pangea.



Mechanisms of terranes accretion

In addition to the collision and accretion of exotic terranes, significant continental growth may occur by the obduction of ophiolites (the Coast Range ophiolite of western North America), **magma addition, sedimentation, and the formation and destruction of backarc, intraarc, and forearc basins** (e.g., the Middle Paleozoic Lachlan orogen of southeastern Australia).

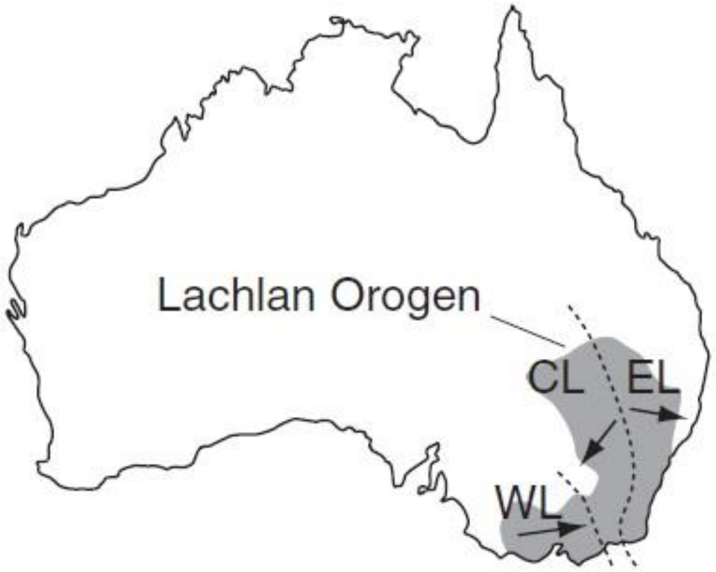
Possible evolution of the Coast Range ophiolite in a backarc setting offshore of California and its subsequent emplacement in a forearc setting

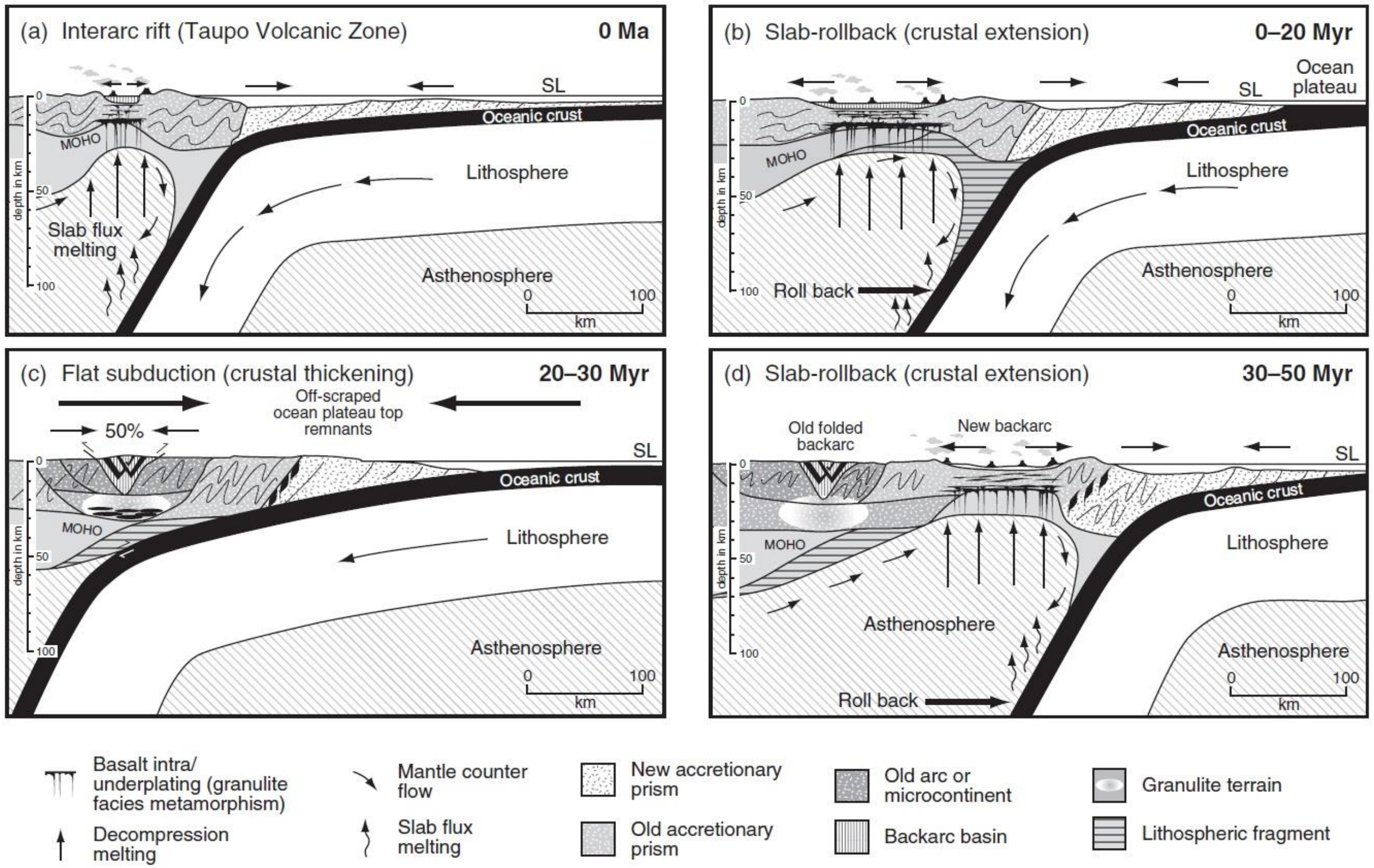


(a) Coast Range ophiolite forms behind a Mesozoic island arc. (b,c) Island arc collides with the continent and a new east-dipping subduction zone initiates, capturing the ophiolite in forearc. (d) Ophiolite obduction occurs in a forearc setting when the crustal layers become detached and uplifted as a result of compression.

Mechanisms of terranes accretion

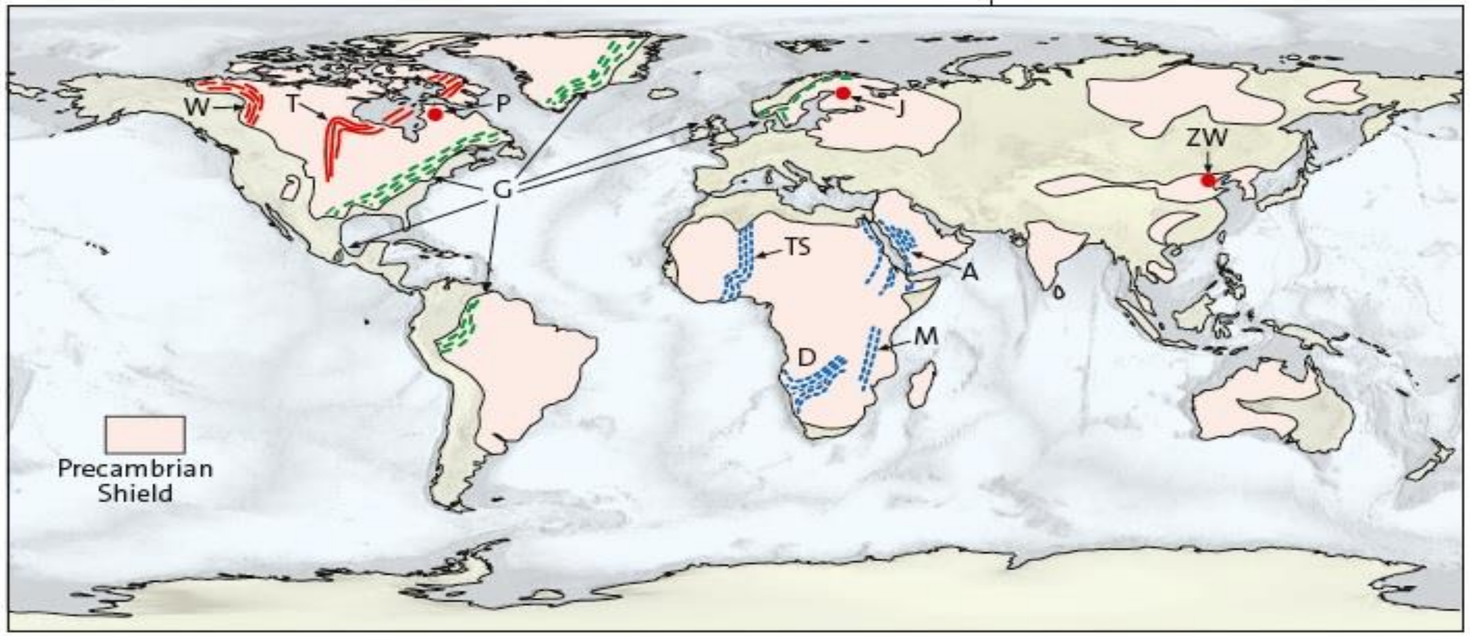
- The Lachlan orogen, formed in absence of main collisional events, during cycles of extension and contraction from Late Ordovician through early Carboniferous times, which lead to the formation and closure of autochthonous backarc basins, is characterized by granitoid rocks, volcanic sequences, and extensive low-grade quartz-rich turbidites.
- This style of shortening did not lead to the development of a well-defined foreland basin nor a foreland fold and thrust belt, but it was controlled by the thick (10 km) succession of turbidites and locally high geothermal gradients, indicating that the Lachlan orogen was dominated by magmatism and recycling of continental detritus during cycles of extension and contraction.





(a) *a zone of intra-arc extension evolves in response to the roll back of a subducting slab.* ***(b)*** *Backarc basin and remnant arc form.* ***(c)*** *Subduction zone flattens and the upper plate of the orogen is thrown into compression. The contraction closes the backarc basins and may lead to the accretion of the arc and forearc onto the continental margin.* ***(d)*** *Extension is re-established and a new arc–backarc system forms.*

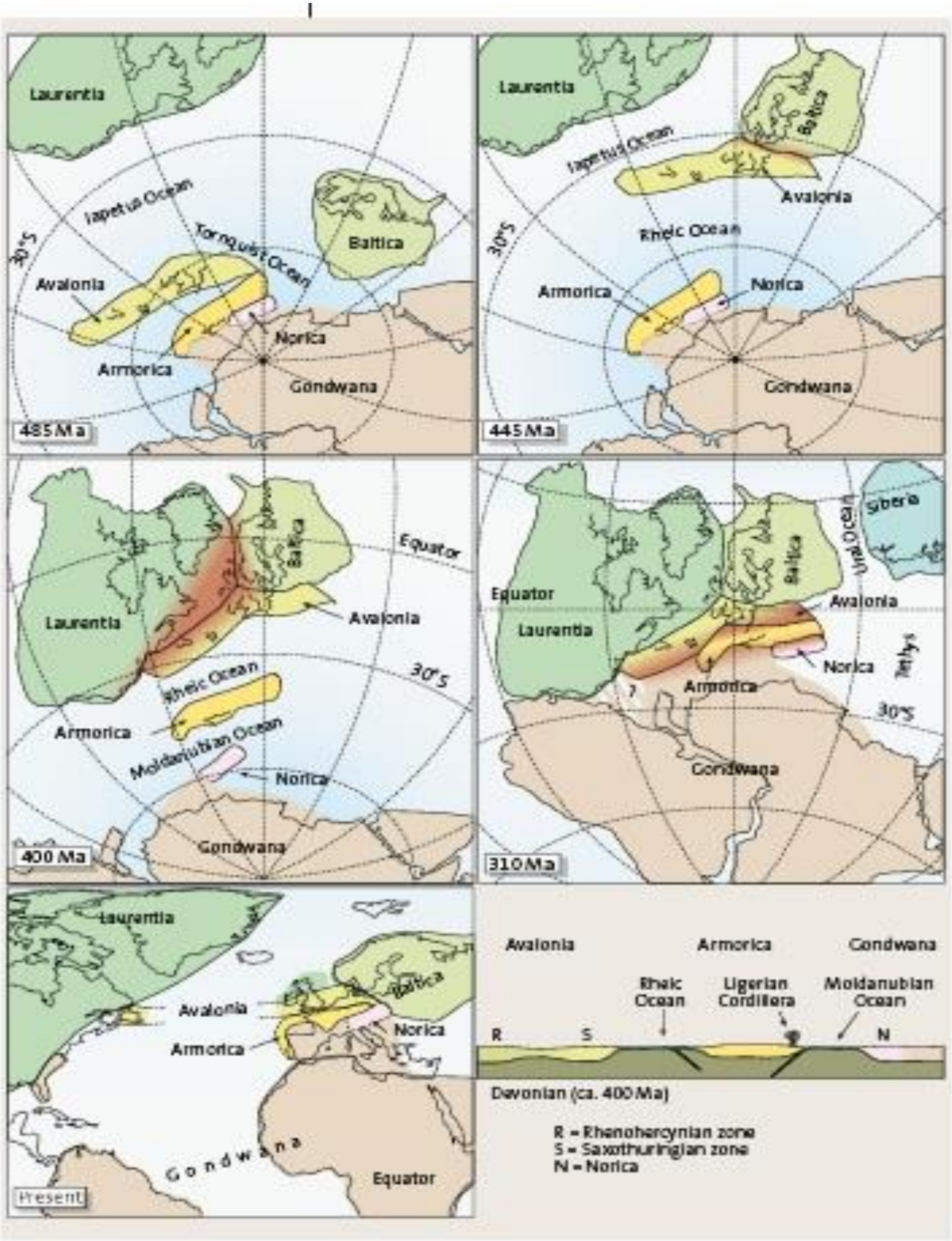
Old Orogens



ZW: Zunhua-Wutaishan ophiolite. J: Jormua ophiolite. **P**: Purtuniq ophiolite. **T**: Trans Hudson orogen. **W**: Wopmay orogen. *Middle Proterozoic*: **G**: Grenville orogen. *Panafrican orogenic belts (Late Proterozoic)*: **A**: Arabian- Nubian Shield. **D**: Damara-Katanga orogen. **M**: Mozambique belt. TS: Trans-Sahara belt.

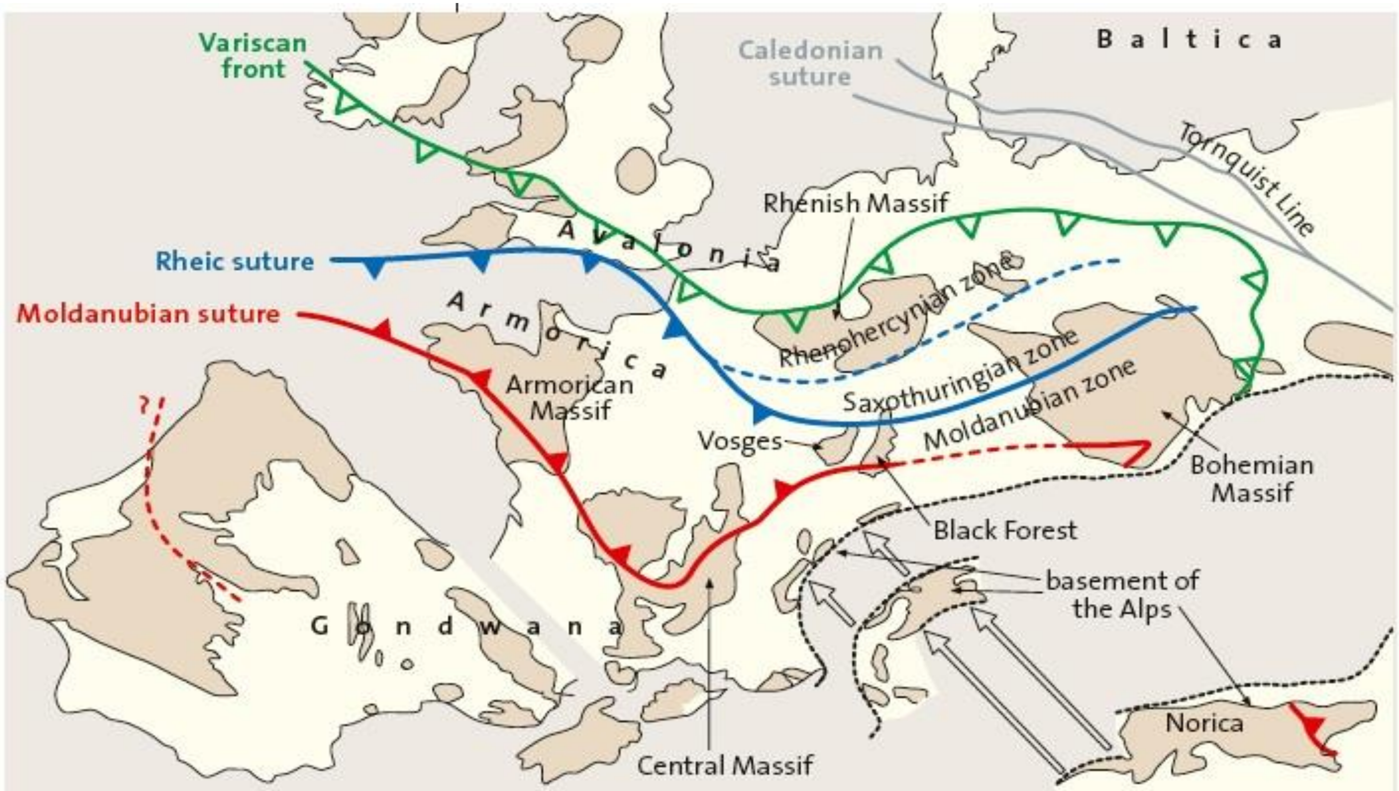
- The **Wopmay orogen** developed between ca. 2100 and 1800 Myr, from a collision between two Archean cratons after a complete Wilson cycle.
- The **Grenville orogen**, generated a long mountain belt that stretched from southern Scandinavia through a strip in Scotland, eastern Greenland, and large parts of eastern North America to South America, after a Wilson cycle that initiated with continent break-up around 1300 Myr. The Grenville orogeny and related orogens assembled in the supercontinent Rodinia, which by 750 Myr started to disrupt.
- The term **Panafrican orogeny** originated because the event welded together, over a period of 200 to 250 Myr, a number of continental blocks and island arc systems, representing the different parts of present-day Africa, as well as other regions (Gondwana continent formation in the late Precambrian).
- **In Europe the Panafrican orogeny is expressed in the** *Cadomian mountain belt*. The rocks were originally positioned at the northern margin of Africa (Gondwana) and formed between 700 and 550 Myr. Cadomian rocks are found in the Armorican Massif (Bretagne and Normandie), in the Bohemian Massif, on the Iberian Peninsula, and in basement complexes of the Alps.

Caledonides



- The Caledonian orogen describes a Wilson cycle that began ~600 Myr in the Late Proterozoic and culminated in the Silurian and Devonian ~400 Myr: During the early Paleozoic, Avalonia rifted from Gondwana and collided with Baltica in the Late Ordovician, and Avalonia/Baltica with Laurentia at the Silurian-Devonian boundary, thus terminating the Caledonian cycle.
- Caledonian orogen is found on both sides of the Atlantic Ocean: in East Greenland, along the western coast of Scandinavia, on the British isles, and along the eastern coast of North America. It continues southward into the Appalachians, where it became overprinted by the collision between Laurentia and Gondwana in Late Paleozoic times.

European Variscides



- The Variscan orogeny took place from the Devonian through the Carboniferous and marks the direct continuation of the Caledonian orogeny. It formed an orogenic belt 1000 km wide, the southernmost part of which was later overprinted by the Alpine orogeny.
- **Variscan belt of Europe is a 3000 km long Paleozoic intra-continental orogeny, formed by compressive deformation of a broad strip of continental crust (not by continent collision)**, extending from the North Sea to Iberia.
- Wide-spread anatectic granites indicate intensive melting of thickened continental crust at the late stages of the Variscan orogeny. Crustal extension with possible rifting and delamination of the lower crust produced the modern crustal structure with a flat Moho at c. 30 km depth.
- Evolution of the **Variscan Orogen, forms part of the Hercynian mega-suture** along which Laurussia and Gondwana were welded together and involved the stepwise accretion of a number of Gondwana-derived terranes to the southern margin of Laurussia and ultimately the collision of Africa with Europe.