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Mountain building, mantle convection, and supercontinents: Holmes (1931) revisited

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A B S T R A C T

Orogeny results from crustal thickening at active margins, and much progress has been made on understanding the associated kinematics. However, the ultimate cause of orogeny is still debated, especially for the case of extreme crustal thickening. Inspired by the seminal work of Holmes (1931), we explore the connections between the style of orogeny and mantle dynamics. We distinguish between two types of orogeny, those that are associated with one-sided, mainly upper mantle subduction, "*slab-pull orogeny*", and those related to more symmetric, whole mantle convection cells, referred to as "*mantle*", or "*slab-suction orogeny*". Only the latter leads to extreme crustal thickening. We propose that mantle orogeny is generated by the penetration of slabs into the lower mantle and the associated change in the length scales of convection. This suggestion is supported by numerical dynamic models which show that upper plate compression is associated with slab penetration into the lower mantle. Slabs can further trigger a buoyant, plume upwelling from the core-mantle boundary which enhances this whole mantle convection cell, and with it upper plate compression. We explore the geological record to test the validity of such a model. For the present-day, compressional backarc regions are commonly associated with slabs that subduct to the deep lower mantle. The temporal evolution of the Nazca and Tethyan slabs with the associated Andean Cordillera and the Tibetan-Himalayan orogenies likewise suggests that extreme crustal thickening below the Bolivia and Tibetan plateau occurred during slab penetration into the lower mantle. This episode of crustal thickening in the Tertiary bears similarity with Pangea assembly events, where the Gondwanide accretionary orogen occurred at the same time of the Variscan-Appalachian and Ural orogeny. We propose that this Late Paleozoic large-scale compression is likewise related to a change from transient slab ponding in the transition zone to lower mantle subduction. If our model is correct, the geological record of orogeny in continental lithosphere can be used to decipher time-dependent mantle convection, and episodic lower mantle subduction may be causally related to the supercontinental cycle.

1. Introduction

Orogeny, i.e. the formation of mountains, represents the most spectacular expression of tectonic activity. Building on the pioneering work of early geologists (e.g., Argand, 1916, 1924), we now have a clear kinematic understanding of the structure and geometry of crustal nappe stack evolution within orogenic belts. Seismological imaging provides constraints on the deep structure with increasing resolution, while geodetic measurements illuminate the current strain-rate distribution within the orogen (e.g., Hafekenscheid et al., 2006; Thatcher, 2009; Zhang et al., 2004; Hatzfeld and Molnar, 2010). Despite those fundamental insights, the actual causes and mechanisms of mountain building are still debated, especially for the case of extreme crustal thickening.

One of the first physical models of orogeny was proposed by Holmes (1931). Building on Dana (1880), and later incorporating Vening Meinesz et al.' (1932) and Hess' (1938) marine observations, Holmes (1931, 1944) proposed that orogeny arises from compressive forces and the inversion of a geosyncline; this leads to pervasive deformation, the formation of a crustal root, associated isostatic uplift, and late-stage magmatism (Fig. 1). Holmes proposed that the compressional forces themselves are generated by convection: "*where currents are flowing horizontally along the un-*

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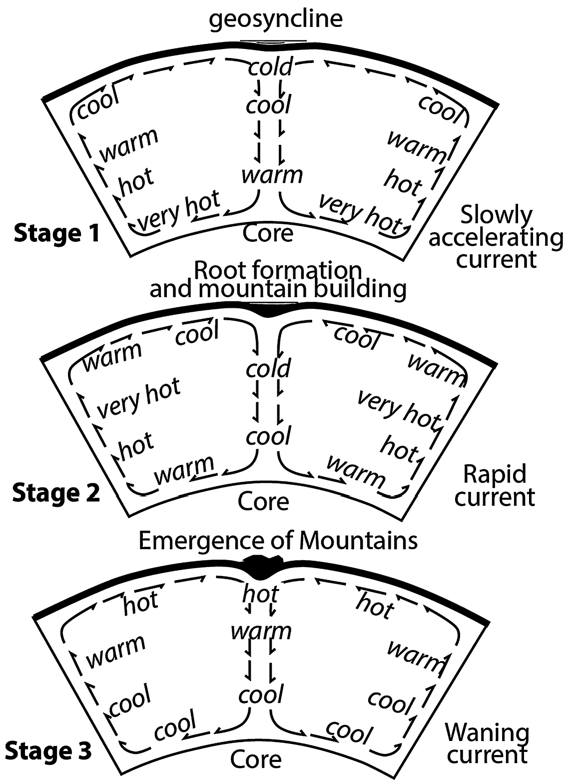

Fig. 1. Successive stages of an orogenic cycle and those of a convection current cycle (redrawn from Holmes, 1944).

der surface of the crust, they exert a powerful drag on the latter, throwing it into tension where they diverge and into compression where they convergence. Thus, we should expect orogenic belt where the two approaching current turn down". This idea was later elegantly tested by Griggs (1939), and Holmes (1944) invoked time-dependent convection with alternating periods of acceleration and deceleration. In line with this concept, Wilson (1961) later stated that "... *circum-Pacific and Alpine-Himalayan mountain systems are due to compression over downward flowing limbs of huge convection cells is straightforward accepted*. . . ."

After plate tectonics was established as a kinematic theory capturing the recycling of oceanic lithosphere and large-scale, horizontal crustal motions in the late '60s, efforts were primarily dedicated to investigating the consequences of continental deformation, with modeling techniques ranging from the thin viscous sheet approach (e.g., England and McKenzie, 1982) to more complex thermo-mechanical models of orogenic wedge formation (e.g., Willett et al., 1993; Burov et al., 2001; Gerya, 2011; Vanderhaeghe, 2012; Jamieson and Beaumont, 2013). The connection between mantle dynamics and orogeny was, however, only further pursued by few (e.g., Wilson and Burke, 1972; Alvarez, 1990, 2010; Doglioni, 1990; Russo and Silver, 1994; Cuffaro and Doglioni, 2018). The relative scarcity of integrated work on the dynamics of orogeny and mantle convection may be partly due to the difficulties inherent in the construction of a fully coupled, mantle-lithosphere model that is able to account for the complex rheology of the lithosphere.

However, the constraints on present-day mantle flow have significantly improved over the last twenty years. Seismic tomography has shown that the lower mantle is dominated by large-scale (∼spherical harmonic degree two) velocity heterogeneities (e.g. Masters et al., 1982; Woodhouse and Dziewonski, 1984; Becker and Boschi, 2002), and similar, long-wavelength temperature anomalies arise in self-consistent, plate-style generating mantle convection computations (van Heck and Tackley, 2008; Foley and Becker, 2009). When seismic tomography is used to infer density anomalies to drive mantle circulation (Busse, 1983; Hager and Clayton, 1989), two upwelling zones above the "large low velocity shear provinces" (on top of Africa and the South Pacific) and two downwelling zones beneath the Cordillera and Tethyan orogenies are predicted (Fig. 2, see also Dziewonski et al., 2010). The mantle flow regime resulting from the numerical computation of Fig. 2 is not too different from Holmes' (1944) vision.

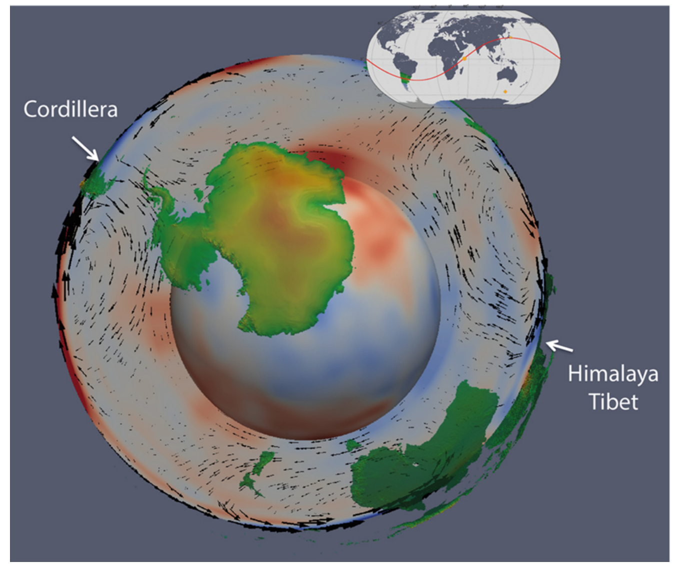

Fig. 2. Equatorial cross section (location shown in the overview map) of a present-day mantle circulation estimate based on seismic tomography and viscosity inversions (cf. Ghosh et al., 2010), modified from Becker and Faccenna (2011, see there for details). The model shows the presence of two large upwelling and downwelling zones forming four large-scale convection cells. The Andes and the Himalaya-Tibet are located on top of, and inferred to be sustained by, deep mantle downwellings (cf. Faccenna et al., 2013).

Here, we explore the role of mantle dynamics in building and supporting orogeny. We start by proposing a classification of orogens based on their deep structures. We then discuss the main distinctive features of orogens and explore how this may apply to the present, as well as the Tertiary and Paleozoic tectonic settings. We conclude by discussing the relationships between orogeny, the Wilson cycle, and supercontinental assembly in the context of mantle convection.

2. Modes of orogeny

Orogeny can be defined as a process producing crustal thickening through penetrative crustal deformation and magmatism during plate convergence (e.g., Dennis, 1967; Cawood et al., 2009). It is commonly classified as *Collisional* and *Cordillera/Subduction* type, based on the continental or oceanic nature of the down-going plate, respectively (Dewey and Bird, 1970; Cawood et al., 2009). The collisional-continental type occurs in continental interiors. The cordillera/subduction type is also characterized as a peripheral, or accretionary orogen, and develops in the oceanic realm surrounding continental land masses (Dewey and Bird, 1970; Wilson and Burke, 1972; Murphy and Nance, 1991; Cawood and Buchan, 2007). Subduction type orogeny can naturally evolve into a continental-collisional type during continental amalgamation after all oceanic lithosphere has been consumed. Arc-collisional orogens may represent a transient phase of accretion between the two end-member cases (Dewey and Bird, 1970). While this classification is useful for identifying the tectonic context, it provides only limited indication of the dynamic processes that drive the deformation or dictate the style of orogeny.

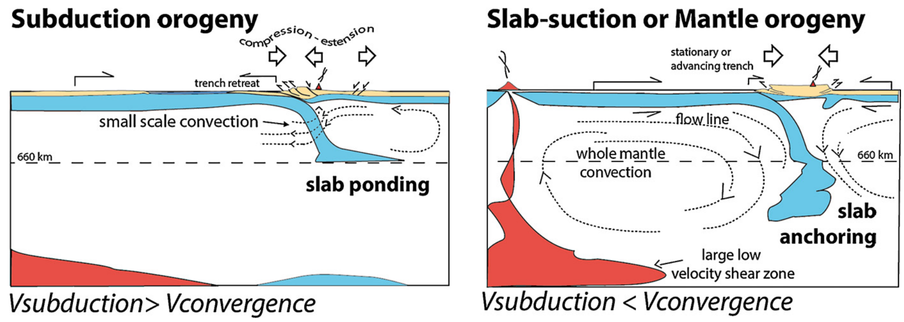

Fig. 3. Conceptual model for two end-member modes of orogeny (modified from Faccenna et al., 2013; Royden and Faccenna, 2018). Left: In a subduction orogeny, subduction is confined to the upper mantle, mantle flow is asymmetric and confined to the upper mantle, and slabs would be favorably in a trench rollback mode, preventing large crustal thickening. Right: slab penetration into the lower mantle promotes trench stabilization and whole-mantle, symmetric mantle flow, which promotes trenchward convergence of both plates, resulting in protracted compression and crustal thickening. Faccenna et al. (2013) used "slab pull" and "slab suction" for subduction and mantle orogeny, respectively, to indicate the one-sided and symmetric nature of force transmission, following Conrad and Lithgow-Bertelloni (2002).

Following Royden (1993), we find it helpful, instead, to distinguish end-member orogenies that are directly related to subduction, also referred to as "*subduction orogeny*" or "*slab-pull orogeny*" (Royden, 1993; Royden and Faccenna, 2018), and the ones that are related to large-scale mantle convection cells with more symmetric mantle flow patterns, also referred to as "*mantle*" or "*slab-suction orogeny*" (Conrad and Lithgow-Bertelloni, 2002; Faccenna et al., 2013) (Fig. 3). This classification has the advantage of being independent of the possibly transient nature of the type of subducting lithosphere or the specific tectonic context.

Orogenic belts show remarkable variability in terms of their shapes, geometry, and evolution. The shape of the orogeny is controlled by several factors such as the history of convergence, block accretions, and the rheology of the lithosphere prior and during deformation. Extracting general rules is thus extremely difficult. Fig. 4 shows inferred cross-sections for several active mountain ranges, distinguishing between the units derived from the downgoing (green) or upper, overriding plate (yellow) during convergence.

There are orogens, such as those of within the Mediterranean region, which are exclusively composed of units derived from the subducting plate (Figs. 4a and b). Sections for the Central Andes and the Tibetan Altiplano (Figs. 4d and e) show similar features, even though those two are often considered end members of the Collisional and Cordillera types. An occasionally useful orogeny classification is the overall shape, in particular if it is symmetric or asymmetric. All orogens are fundamentally asymmetric, but only some of them show a strongly asymmetric wedge structure, tapering toward the subducting plate. This feature is distinctive of the *subduction* or *slab-pull orogen* (Royden, 1993; Jamieson and Beaumont, 2013; Royden and Faccenna, 2018), which is constituted mainly of crustal slices and sediments scraped off from the subducting plate (green material on Fig. 4).

The Mediterranean belts, such as the Apennines, Dinarides, or Hellenides are representative of this style of orogeny (Doglioni, 1992; Royden, 1993; Royden and Faccenna, 2018) (Figs. 3 and 4), as are Paleozoic and Mesozoic belts within the western Pacific (e.g., East Australia and New Zealand: Cawood et al., 2011). In this orogeny type, the depth of the decollément controls the thickness of the thrust nappe. Deformation can vary from thin skinned, as for the case of an accretionary wedge with a shallow decollément level, to thick skinned, as in the case of re-activation of pre-existing extensional feature. The overall shape respects the critical taper criteria (Davies et al., 1983).

For subduction orogeny, crustal thickness does not exceed 50 km. This is because the trench is highly mobile and usually migrates backward, preventing large-scale, protracted shortening (Royden, 1993). Crustal blocks, or allochthonous terranes, may be delaminated and accreted to the upper plate, during active downwelling and accelerating slab rollback (e.g., Brun and Faccenna, 2008). The topographic expression of subduction orogeny is also quite limited, as the active pull of the subducting slab dynamically depresses the overall topography by some hundreds of meters (e.g., Mitrovica et al., 1989; Crameri et al., 2017; Faccenna and Becker, 2020).

The other class of *mantle* or *slab-suction* orogen shows a more symmetrical, double-vergent structure, with thrusting and crustal deformation on both sides of the orogen. The two oppositely vergent belts can be separated by a thickened high plateau. This type of orogen is mainly built up by the shortening of upper plate units, with only minor contributions from the downgoing plate. During convergence, small crustal blocks, or allochthonous terranes, may be accreted to the upper plate producing pulses of shortening. Crustal thickening is protracted for tens of million years and total crustal thickness typically exceeds 50 km. Larger orogenic belts such as the Himalaya-Tibet, Iran or Persian plateau, the Central Andes, or the Nevadaplano during the Laramide orogeny are representative of mantle orogeny (Fig. 4).

Subduction and underthrusting creates asymmetric foredeep basins as a result of plate bending. The basin depth depends on the load of the thrust sheets and sediments, and on the pull of the subducting slab; it is balanced by the flexural elastic and viscous resistance of the subducting plate (Royden, 1993). Foredeep basins of subduction orogeny are commonly quite deep and filled by thick sequences of turbidite marine deposits. Their formation can be explained by the pull of the subducting slab (Royden, 1993). The best examples of foredeep basins are from the Mediterranean, for example from the northern Apennines (e.g. Ori and Friend, 1984; Doglioni, 1992).

In contrast, mantle orogens usually present a pro- and a retro-foredeep basin, e.g., formed on both side of the orogen (cf. Dickinson, 1974). The pro-wedge basin can be quite deep if the subducting plate is oceanic, as for the case of the Andes Cordillera, but the retro-wedge is always shallower and filled by coarse continental deposits (molasse; Royden, 1993). In the case of Himalaya-Tibet, both basins are shallower and filled with continental deposits that were sourced both during the collision and during subsequent orogenic evolution. Many orogenic belts, such as the Alps, show a transition from marine to continental, flysch to molasse, following the progression from oceanic to continental subduction and break off (Sinclair, 1997).

Subduction and mantle orogeny are not just structurally different, but the associated metamorphism is also quite different. Metamorphic gradients for mantle orogeny are commonly moderate to high temperature and related to crustal melting and shear heating. High to ultra-high pressure units are limited, as in the case of Himalaya, to the very initial phase of continental collision (e.g., Leech et al., 2005). Conversely, most of the inner portion sub-

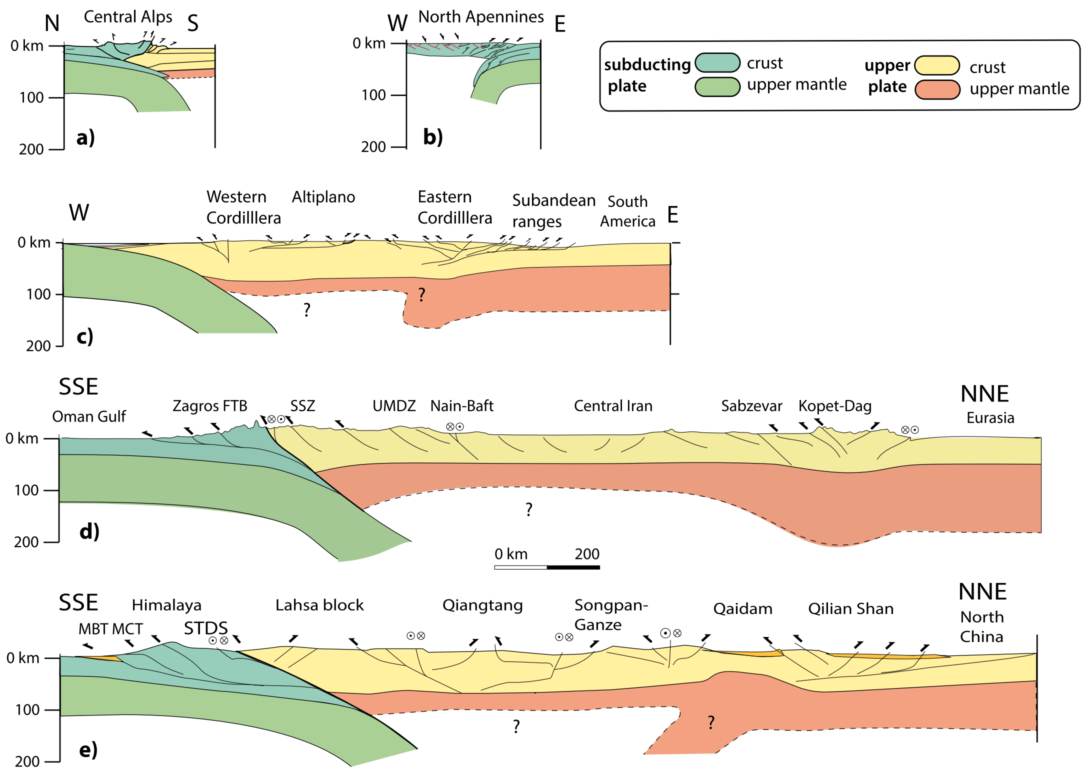

Fig. 4. Cross sections across active and recently active orogens. a) Central Alps (modified from Schmid et al., 1996); b) Northern Apennines (Piana Agostinetti and Faccenna, 2018); c) Andes at latitude 20°S (modified from Oncken et al., 2006); d) Zagros-Iran-Keph-Dagh; e) Himalaya-Tibet cross section (MTB main boundary thrust; MCT: main central thrust; STDS southern Tibetan detachment system; Indus-Tsang-Po suture. Redrawn from Guillot and Replumaz, 2013).

duction orogeny belts are often constituted of high-pressure / low temperature units formed in the frontal region during subduction and high temperature / low pressure in the backarc region during exhumation related to backarc thinning and/or delamination (e.g., Jolivet et al., 2003; Brun and Faccenna, 2008). Those units often exhume within a low angle detachment extensional system. Subduction orogeny, indeed, often produces slab-rollback and extension which provides space for the exhumation of deeper units (Brun and Faccenna, 2008; Malusà et al., 2011).

3. Present-day orogeny: trench kinematics, backarc deformation and deep mantle structure

When we speculate about the origin of the different degrees of symmetry between subduction and mantle orogeny, it is instructive to first view orogeny and backarc deformation in the context of plate and trench motions (Royden, 1993; Heuret and Lallemand, 2005; Heuret et al., 2007; Becker and Faccenna, 2009). Fig. 5 illustrates the possible combinations of trench migration and upper plate motion. Back arc extension occurs if the trench migrates backward at a higher rate than the upper plate does (e.g., Tonga, Sandwich, Mediterranean), or if the upper plate moves away from trench faster than the rate of trench advance (e.g., Marianas, Izu Bonin). In those cases, the subduction rate exceeds the convergence rate. This plate configuration is likely dominantly driven by one-sided slab pull due to the negative buoyancy of the subducting slab (Royden, 1993; Conrad and Lithgow-Bertelloni, 2002). This mode may be favored for the case of older (thicker) oceanic plates, or in the case of continental or transitional lithosphere, where light crustal material can be scraped off and accreted to trench forming orogenic wedge (Brun and Faccenna, 2008). Current examples of such subduction orogens are found in the Mediterranean, Indonesia (e.g., Taiwan and Banda), and Scotia Arc (Royden and Faccenna, 2018).

Backarc compression occurs when the backward trench migration rate is lower than the motion of the upper plate (e.g., Cordillera) or when the trench migrates forward, toward the upper plate (e.g., Adria, Arabia, India). In this case, the subduction rate is lower than the convergence rate. Therefore, one-sided slab pull is not the dominant mechanism for this mode of backarc deformation. Irrespective of the nature of the subducting plate, the upper plate here is shortened and thickened intensively, forming a mantle or slab-suction type orogen, such as the Tethyan belt of the Altiplano.

Following this argument, the style of orogeny then fundamentally depends upon trench migration, which, in turn, is controlled by the possibility of the slab to migrate laterally (e.g., Garfunkel et al., 1986). A range of parameters may influence the dynamics of the trench, including slab buoyancy (Molnar and Atwater, 1978; Royden and Husson, 2006; Funiciello et al., 2008; Stegman et al., 2010), upper plate thickness (e.g., Capitanio et al., 2011; Garel et al., 2014; Holt et al., 2015), width of the slab (Funiciello et al., 2003, 2006; Stegman et al., 2006) and the depth extent of the slab into the lower mantle (Zhong and Gurnis, 1995; Garfunkel et al., 1986; Király et al., 2017). Slab penetration into the more viscous lower mantle produces an anchoring effect that reduces its ability to migrate laterally (e.g., Garfunkel et al., 1986; Zhong and Gurnis, 1995; Faccenna et al., 2017).

Fig. 6 shows the relationship between maximum slab depth for the main subduction zones as inferred from seismic tomography (Li et al., 2008a; Fukao and Obayashi, 2013; van der Meer et al., 2018) and the corresponding, present-day backarc strain field (Heuret and Lallemand, 2005). Most backarc regions associated with slabs deeper than 1000 km are under compression, including Java (Yang et al., 2016). Exceptions are represented by Central America which shows a recently neutral regime, even if it has been associated with strong compression over much of is geological history. Most backarc regions that are under extension

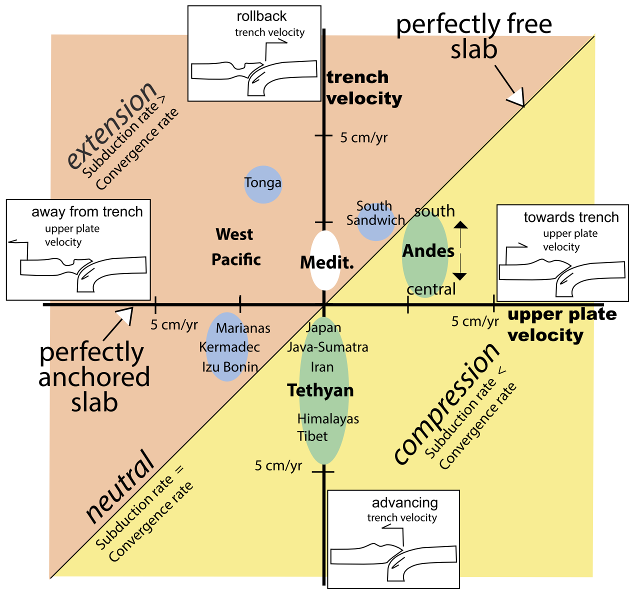

Fig. 5. Relationships between upper plate and trench velocity (in an absolute, lower mantle fixed reference frame) and back-arc deformation. For a laterally free slab, the trench velocity is expected to move according to the upper plate. The trench velocity is zero for an anchored slab, and equal to the upper plate velocity for a free slab (Heuret and Lallemand, 2005).

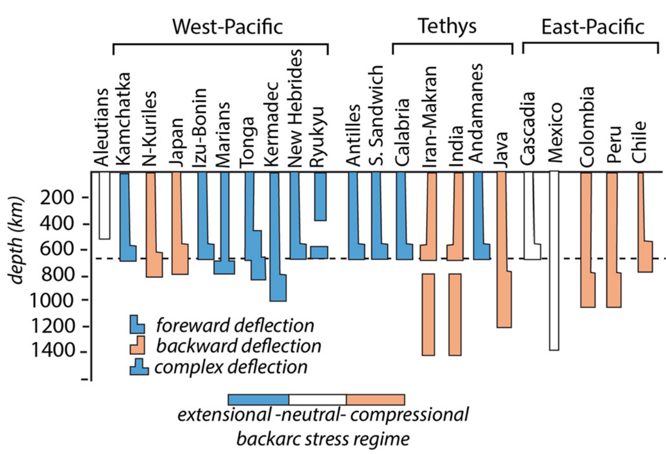

Fig. 6. Depth of the bulk of the subducting slab (from Fukao and Obayashi, 2013; Li et al., 2008a; van der Meer et al., 2018) with color indicating current back-arc strain regime (modified from Heuret and Lallemand, 2005). The lower mantle portion below Kermadec (South Loyalty Basin; van der Meer et al., 2018) between 1200 and 1000 km is not considered to be directly connected to the upper mantle slab and probably belongs to an earlier subduction phase.

or in the neutral regime correspond to slabs that are limited to the upper mantle or transition zone. Exceptions are represented by Japan-Kuriles, where the compressional regime started in the Late Miocene after a large extensional phase, possibly due to the interaction with the Ryukyu subduction zone (Faccenna et al., 2017).

The relationship indicated by Fig. 6 is consistent with the suggestion that slabs that extend into the lower mantle offer greater resistance to lateral migration compared to upper mantle slabs. This is probably due to the larger viscosity of the lower mantle which also controls vertical sinking velocities (e.g., Gurnis and Davies, 1986; Ricard et al., 1993). Upper mantle slabs, conversely, are relatively free to migrate. For this reason, upper mantle slabs may be preferentially associated with an extensional to neutral backarc strain regime. Deeper slabs are more commonly related to backarc compression not only because they are anchored, but also because penetration after ponding at the 660 km discontinuity can trigger large-scale, symmetrically converging convection cells that can drag surface plates toward the downwelling zone (Becker and Faccenna, 2011; Faccenna et al., 2013, 2017; Yamato et al., 2013; Dal Zilio et al., 2018; Yang et al., 2018).

The lateral extent of the subduction zone, the influence of large-scale mantle flow, and the interaction with other slabs likely alter this simple scenario, hiding simple correlations (Heuret and Lallemand, 2005). However, if this model is correct, we should then expect that the entrance of slabs into the lower mantle would result in a surge of compression in the backarc region (Yamato et al., 2013; Faccenna et al., 2013). The geological record of key orogens, where the switch from one strain regime to another can be documented, then consists of a surface record of this process during continental deformation and supercontinental assembly (Hoffman, 2014).

Next, we will present numerical models supporting the idea that slab penetration induces upper plate compression and, further, will explore whether subduction-induced upwellings originating from the core-mantle boundary can enhance this compression. We will then review key tectonic elements supporting the idea that the formation of the Andes Cordillera and the Himalaya-Tibetan orogen are related to episodes of deep mantle subduction.

4. Deep mantle subduction and upper plate compression: a dynamical model

We use idealized, numerical subduction models to explore the relationship between lower mantle penetration and upper plate deformation. We use the *ASPECT* finite element code (version 2.1.0) to construct time-evolving 2-D numerical models (Kronbichler et al., 2012; Heister et al., 2017; Bangerth et al., 2020a, 2020b), with setups analogous to those of our previous work (Faccenna et al., 2017; Holt and Condit, 2021). We extend these previous models, which consider uniform temperature sub-lithospheric mantle, by also incorporating a hot boundary layer at the core-mantle boundary that produces upwelling flow. We find that this affects the large-scale mantle flow field and, in turn, the upper plate stress field (Fig. 7).

Here, we provide a brief overview of the model setup; a more detailed description can be found in the Supporting Information. Our models evolve dynamically in that there are no external forces or velocities applied to the subduction system. We consider a whole mantle domain, with free slip boundaries, and initiate subduction by prescribing an initial proto-slab extending to 250 km depth. Our subduction system consists of three, 60 Ma old thermal plates. The middle plate, which overrides the left-most subducting plate, mimics a continent with reduced density ($\Delta\rho = 100$ kg/m^3) and increased viscosity ($\eta_{cont} = 10\eta_0$) relative to oceanic lithosphere. To the right, the continent transitions to a purely thermal plate at a passive margin. The subducting plate is decoupled from the overriding plate by a thin (10 km) and weak (10^{20} Pa s) crustal layer, which is cut off at 150 km depth.

We consider a composite diffusion creep, dislocation creep, and plastic rheology. Byerlee-type yielding operates within the lithospheric plates (e.g., Enns et al., 2005), dislocation creep operates in the upper mantle to average depths of 200-300 km, and the lower mantle is exclusively diffusion creep. Dislocation and diffusion creep flow law pre-factors are set to give an average upper mantle viscosity of 2.5×10^{20} Pa s (at depth = 330 km). The upper to lower mantle transition, at 660 km depth, is defined by a viscosity jump of a factor of either 20 - 25 (Fig. 7) or 90-95 (Fig. S1). While the smaller value may be in closer agreement with geoid constraints (e.g., Hager, 1984), we test a stronger mantle to examine the effect of a slab (and mantle flow) that is confined to the upper mantle. In all models, the surface temperature is 0 °C, we neglect compressibility and assign the background mantle a tem-

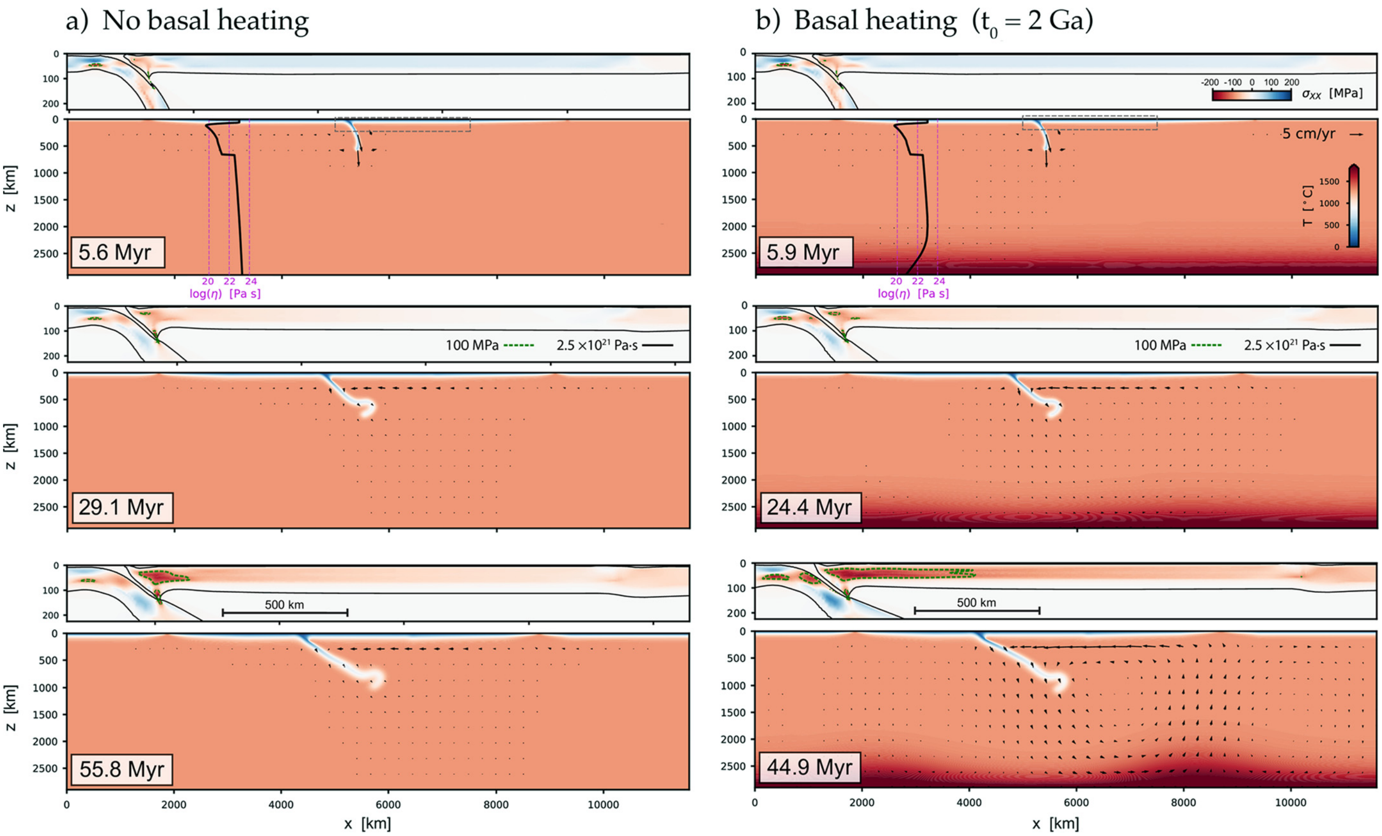

Fig. 7. Comparison of dynamic subduction models with and without a thermal boundary layer at the core-mantle boundary. For each model snapshot, lower panels display the model temperature field overlain by velocity vectors. Upper panels display the horizontal deviatoric stress within the overriding plate, over a length of 2500 km, with negative values (red) indicating compression. A vertical mantle viscosity profile, extracted from near to the trailing edge of the subducting plate, is also displayed for the earliest snapshot.

perature of 1300 °C. For a viscosity jump of 20-25, we examine the role of a basally heated core-mantle boundary (CMB) with initial thermal boundary layer thickness parameterized by half-space cooling ages of 1 Ga (Fig. S1) or 2 Ga (Fig. 7), and a boundary layer temperature contrast of 700 °C (i.e., $T_{CMB} = 2000\,°C$).

With or without lower mantle thermal upwellings, compressive stress in the upper plate becomes larger and spans a greater extent after the slab has penetrated into the more viscous lower mantle (Fig. 7). This occurs because subduction-induced mantle return flow transitions from being confined to the upper mantle to spanning the entire mantle which, in turn, increases the strength of basal tractions that drag the upper plate towards the trench (e.g., Faccenna et al., 2017; Dal Zilio et al., 2018; Yang et al., 2018). This is supported by a ∼50% reduction in maximum upper plate compression in models with a strong lower mantle and hence negligible lower mantle flow (Fig. S1).

The vigor of whole mantle flow, and hence upper plate compression, is enhanced by the presence of a hot thermal boundary layer at the core-mantle boundary (CMB). The effect that lower mantle slabs have on CMB thermal boundary layer morphology and plume formation is relatively well-studied (e.g. Tan et al., 2002; McNamara and Zhong, 2005), as is the effect that large-scale mantle flow has on subduction zones (e.g. Yamato et al., 2013; Faccenna et al., 2017; Baes et al., 2018). However, the two-way coupling between slab dynamics and CMB upwellings, as considered here and relevant to our later discussion of continental cycles, remains relatively unexplored.

As the slab sinks into the lower mantle in our models, it redistributes hot and upwelling CMB material away from the beneath the slab and into two regions beneath the edges of the subducting and overriding plates. The thermal upwelling on the overriding plate side aligns with the subduction-driven return flow thereby increasing the vigor of whole mantle convection and, in turn, the magnitude of upper plate compression. After 40 Myrs of model evolution, for example, upper plate compression reaches a maximum (deviatoric) stress of ≈120 MPa in the model with a constant lower mantle temperature and ≈150 MPa in the model with a 2 Ga basal boundary layer (Fig. S1). More strikingly, the horizontal extent of significant compression ($\sigma_{XX} > 75$ MPa) increases from ≈200 km to ≈600 km with the addition of a basal thermal boundary layer. The thermal upwelling has the additional effect of increasing plate convergence, by >50% at 40 Myr, due to the entrainment of the slab and plates by the more vigorous whole mantle convection cell. We proceed to discuss how this dynamical model may apply to the Tertiary and earlier to the Late Paleozoic orogeny.

5. Tertiary evolution of orogeny and large mantle convection

At present, there are two, thousands of km-long active orogenic belts on Earth. The Tethyan belt extends from the Mediterranean to the Himalaya-Tibet and further east to Java-Sumatra, and at the eastern Pacific margin, the Cordilleran belt stretches along the Americas from Canada to Patagonia (Fig. 8). The Tethyan is commonly considered as a collisional, internal orogeny, while the Cordillera is a peripheral, accretionary orogeny. These orogens were built under a different tectonic context but share similar features. Both have a double-vergent structure with a high plateau in the center (Fig. 4), formed roughly at the same time, and are associated with a lower mantle downwelling (Fig. 2) (Wilson, 1961). In the following sections, we will briefly review their kinematic

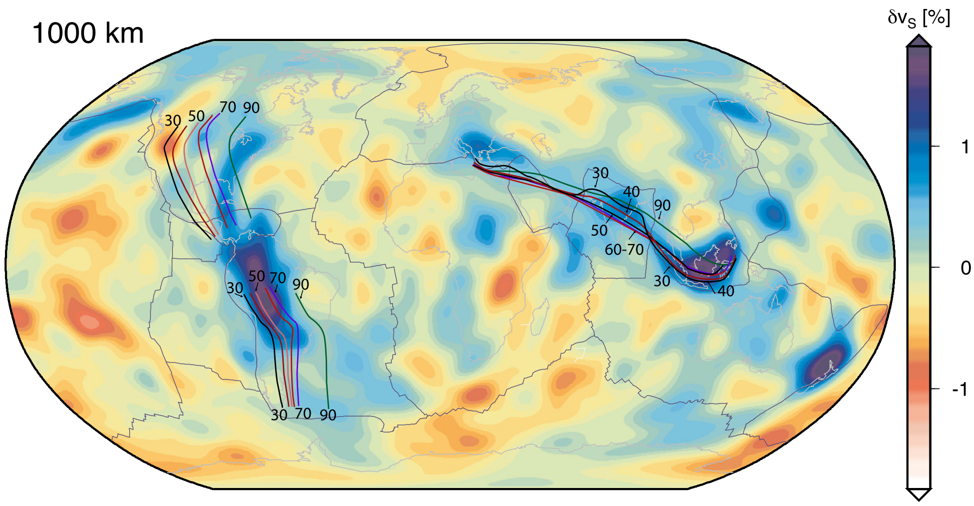

Fig. 8. Global seismic tomography for shear wave speed anomalies at 1000 km depth from the SMEAN composite model (Becker and Boschi, 2002). Lines show the trench position back in time (in Ma; after Faccenna et al., 2013). This diagram provides a first order indication of the timing of slab penetration (cf. Ricard et al., 1993; van der Meer et al., 2018).

history and will then compare their large-scale evolution and the associated deep mantle dynamics.

5.1. Cordillera system

The Cordillera of the Americas extends from North America to Patagonia and Tierra del Fuego. The structure of the belt shows an overall double-vergent and highly asymmetrical structure. The western side is characterized by active subduction of the Nazca plate and the western Cordillera is constituted by the volcanic arc and localized deformation. On the eastern side, a thrust and fold belt of the eastern Cordillera is where most of the shortening is concentrated; the high Altiplano plateau developed between these two belts. The entire orogen is composed of units that are derived from the upper plate (Fig. 4). The deep structure of the belt is constituted of up to ∼70 km thick crust with a reduced lithospheric mantle thickness (e.g., Beck et al., 1996).

The onset of South American orogeny, i.e., the main episode of crustal thickening, started at ∼50 Ma. (e.g., Oncken et al., 2006, 2013; and references in Faccenna et al., 2017). Prior to this, in the Late Cretaceous, South America was under extension, with the formation of the Salta rift (Kley and Monaldi, 2002), and short episodes of compression and accretionary (e.g., Oncken et al., 2006; Horton, 2018a). At that time the westward drift of South America was quite rapid at ∼5 cm/yr. The South American drift velocity then progressively decreased during the formation of the orogen. Reconstruction of the trench migration rate, taking into account time variability in Andean shortening, illustrates that the trench migration rate in the Central Andes also reduced during the last ∼50 Ma, but significantly more rapidly than the South American velocity reduction (Ren et al., 2007; Faccenna et al., 2017). Therefore, the origin of the Cordillera orogeny appears to be related to this decrease in trench migration rate (Faccenna et al., 2017). The Andes Cordilleras, north of 22°S, is located on top of a high velocity zone positioned in the uppermost lower mantle (Fig. 8). In contrast to what is observed below India (see next sub-section), the lower mantle anomaly shows, at least in the Central Andes, a direct connection with the upper mantle present day slab. Inspection of Fig. 8 indicates that the position of the lower mantle anomaly lines up with the location of the trench at ∼50 ± 10 Ma. Hence, slab penetration appears correlated in time with the onset of the main crustal thickening event.

Fig. 9(a-e) shows a tentative reconstruction, at the scale of the mantle, of a section through the Bolivian orocline and across the Altiplano-Puna plateau, where high velocity material is progressively restored according to the amount of reconstructed subduction. A tomographic cross section from Lu et al.'s (2019) model shows the presence of dipping slabs down to ∼1200 km, progressively thickening in the transition zone and in the upper lower mantle. Similar features have been observed in other tomographic models (Ren et al., 2007; Li et al., 2008a; Fukao and Obayashi, 2013; Scire et al., 2015). Another local high velocity anomaly is located deeper, between 1400 and 1600 km, and further to the east.

The retro-deformation in Fig. 9 shows that the slab entered the lower mantle at ∼50 Ma: at that time the trench lines up with the deep high velocity anomaly (Faccenna et al., 2017). The migration of the trench decreased subsequently such that a large fraction of the South American westward motion is accommodated by shortening, as shown by the inversion of extensional features such as the Salta rift. Surface deformation, associated with large-scale thrusting, shifted eastward after 15 Ma (e.g., Oncken et al., 2013). The deeper high velocity anomaly to the east lines up with the shallower one, is restricted to 10-20°S and 40°-50° E, and may correspond to an earlier, ∼100 Ma old, subduction episode. Indeed, subduction beneath the Cordillera was active during the entire Mesozoic (e.g., De Celles et al., 2015; Horton, 2018b). The lack of a stronger seismic velocity anomaly that correlates with this earlier episode of subduction might be due to the rather young seafloor age at the trench at this time, which is harder to resolve with seismic tomography.

Moving northward, the continuation of the lower mantle anomaly can be found below the eastern side of North America. This is typically related to Farallon subduction (Bunge and Grand, 2000; Grand, 2002), but alternative interpretations exist (Sigloch et al., 2008). In North America, the Laramide orogeny is older than the Cordillera, reaching the peak of its development in the Late Cretaceous (De Celles, 2004, and reference therein). The main thickening phase of the Laramide occurred at ∼80-90 Ma, with the formation of a 3-4 km high altiplano, the Nevadaplano, analogous to the Andean plateau (De Celles, 2004). The onset of the Laramide orogeny has been related to slab flattening as suggested by the contemporaneous eastward shift of the volcanic arc (e.g., Coney and Reynolds, 1977; Livaccari et al., 1981; Bird, 1998; Carrapa et al., 2019). Reconstructions also show that an oceanic plateau, the postulated conjugate of Shatsky Rise, may have hit the trench at that time (Liu et al., 2010; Humphreys et al., 2015). The restored

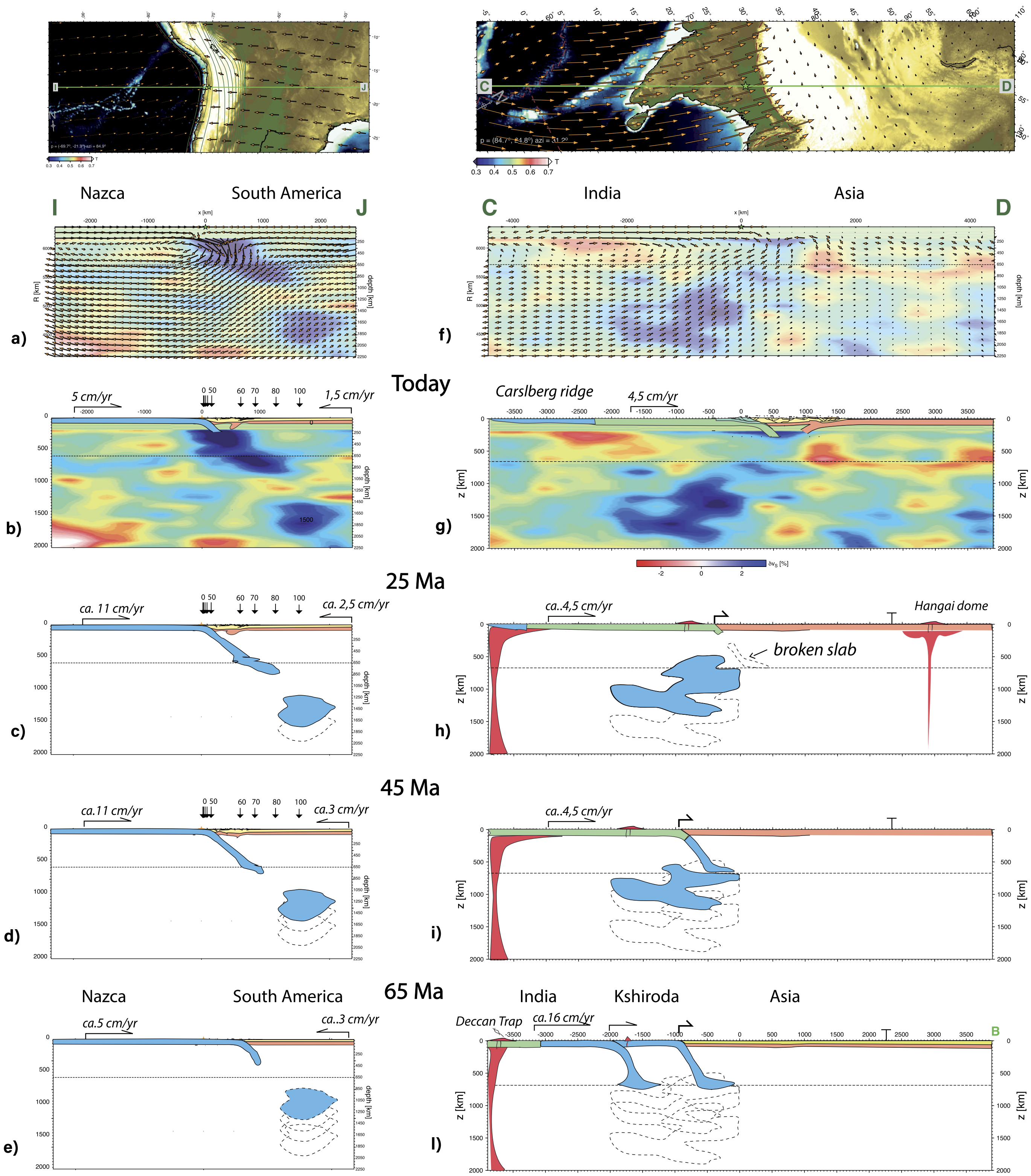

Fig. 9. Reconstruction of the kinematic motion and tentative correlation with mantle anomaly from the TX2019 model (version without imposed upper mantle slabs; Lu et al., 2019) along two cross sections crossing the Bolivian and Tibetan Plateau. a) and f): global mantle circulation model based on mantle tomography; b) and g): mantle tomography model (Lu et al., 2019); c)-e): three stage evolution of the Nazca slab (modified from Faccenna et al., 2017); h)-l): three stage evolution of the Tethyan slab.

location of the Farallon trench that lines up with the location of the 1000 km depth high-velocity anomaly is older than that of South America, at ~70 to 90 Ma. This is well illustrated by previous reconstructions (Bunge and Grand, 2000; Ren et al., 2007; Husson et al., 2012; Faccenna et al., 2013). Therefore, even for the case of North America, it is reasonable to consider that the main crustal shortening episode here occurred during deep mantle anchoring.

Indirect evidence on the role of subduction on surface deformation can be derived from the Late Cretaceous subsidence within western North America. Superimposed on the flexural basin system, there is a clear, long wavelength signal of basin subsidence, extending for several hundreds of kms eastward of the orogenic front. This is apparent in the 85-80 Ma interval. This large-scale episode has been dynamically related to subduction (e.g., Mitrovica et al., 1989; Liu et al., 2008). A clear correlation exists between the high velocity anomaly at 1200-1500 km depth and the late Cretaceous isopach once restored to its absolute position between 95 and 70 Ma (Spasojevic et al., 2009). This suggests that subsidence can also be related to the deep slab evolution in the lower mantle and that the slab penetration probably occurred around this time.

We speculate that slab shallowing and then flattening may be facilitated by, and may actually have only occurred if, the slab tip is anchored into the lower mantle, at which stage the trench is forced to migrate backward if the upper plate is advancing towards the trench (e.g., Espurt et al., 2008). Under this condition, it is possible that a perturbation of the slab buoyancy by an oceanic plateau could result in a flat slab episode. More detailed reconstruction and modeling is required to test this hypothesis. If this model is correct, we can suggest that orogeny and slab penetration progressed southward from the Farallon plate subduction to the Central Andes.

5.2. Tethyan belt

The Himalaya-Tibet orogeny extends from India to the Altai (Fig. 4). The Indus-TsangPo suture separates the Himalayan belt, a stacked of nappe scraped off from the Indian continent, from Tibet-Asia terranes (Fig. 4). Tibet is constituted of six terranes progressively accreted to Asia before collision (e.g., Yin and Harrison, 2000; Kapp and De Celles, 2019 and references therein). Most agree that the onset of the India-Asia collision started around ~50 Ma (e.g., Hatzfeld and Molnar, 2010; Molnar and Tapponnier, 1975; Guillot et al., 2003; Avouac, 2003), even if other models suggest a later onset. The crust beneath Tibet is 70-80 km thick (e.g., Mitra et al., 2005; Schulte-Pelkum et al., 2005), and has been thickened during the last ~50 Ma over a distance of ~1000 km. Therefore, during collision, India advanced inside Asia by about the same distance, at an average rate of ~2 cm/yr. In other words, half of the convergence rate is absorbed by the advancing motion of the trench, reducing the subduction/underthrusting velocity to ~2-2.5 cm/yr. Around ~1000 km of Greater India continental lithosphere underthrusted below Tibet (Capitanio et al., 2010; Replumaz et al., 2010; van Hinsbergen et al., 2011a).

The Himalaya-Tibet orogen system, and the rest of the Alpine-Himalayan orogeny, is positioned on top of a long and well-resolved seismic high velocity anomaly (e.g., Van der Voo et al., 1999; Li et al., 2008b; Fig. 8). At ~1100-1500 km depth, it represents a rather continuous structure, from Java to the East (Replumaz et al., 2004) to the Mediterranean to the West (Faccenna et al., 2003). This velocity anomaly has been interpreted as the thermal signature of the Tethyan slab descending into the lower mantle (Van der Voo et al., 1999; Replumaz et al., 2004; Li et al., 2008b; Yang et al., 2016). Below India, the lower mantle anomaly is positioned to the south of the present-day collisional zone. At shallower depth, the high velocity anomaly is distorted and discontinuous and it has been imaged beneath Tibet down to a depth of 200-300 km (Li et al., 2008b).

Several studies have attempted to integrate plate kinematics, mantle tomography and the development of tectonic structures (e.g., Van der Voo et al., 1999; Hafkenscheid et al., 2006; Replumaz et al., 2004). Fig. 9 illustrates one possible evolutionary scenario along a NNE-SSW cross-section. Between 65 and 45 Ma, two subduction zones were active: the southern one was intraoceanic (Trans-Tethyan subduction zone), and the northern one bounding Asia formed an Andean type margin (e.g., Aitchison et al., 2000; Jagoutz et al., 2015). The two subduction zones were active from the Late Cretaceous to the Eocene, and were separated by the old, oceanic Kshiroda plate. The size of this plate is inferred to have decreased over time and at around ~ 65 Ma it was perhaps ~1000 km in length. At that time, India was located 2500 km south of the northern margin and was rapidly advancing northward at a speed exceeding ~16 cm/yr, while the Deccan lava was flooding northwestern India (Allegre et al., 1999).

The rapid motion may be related to the action of the double subduction system (Jagoutz et al., 2015; Holt et al., 2017), the push of the Deccan hot spot (van Hinsbergen et al., 2011b), or a combination of the two (Pusok and Stegman, 2020). The reconstruction of Fig. 9 shows that the Tethyan lower mantle anomaly is positioned between the former sites of the two subduction zones. Between ~65 and 45 Ma the double subduction system disappeared, due to the consumption of ~1200 km and ~1000 km of seafloor at the rear and frontal subduction zones, respectively. The high velocity anomaly in the lower mantle today is probably composed of both slabs (Van der Voo et al., 1999). Assuming vertical motion in the lower mantle, we can deduce that the penetration of the slab in the lower mantle occurred around ~60 ± 10 Ma. This timing of penetration of the Indian slab inside the lower mantle corresponds to a reasonable lower mantle sinking rate of few cm/yr (Ricard et al., 1993; Steinberger et al., 2012; van der Meer et al., 2018) and is in good agreement with the reconstructed position of India (Replumaz et al., 2004).

After collision of greater India with Asia, India's plate velocity rapidly decreased to ~4–5 cm/yr (Copley et al., 2010; Zhang et al., 2004). Convergence, underthrusting, and trench advance likely continued up to the present-day (e.g., Avouac, 2003; Zhang et al., 2004). More than ~1000 km of Greater India continental lithosphere underthrusted beneath Asia from ~45 Ma onward. The trace in the mantle of this material can be related to the shallower high velocity anomaly positioned below the Himalaya, which broke-off in the Miocene (Guillot and Replumaz, 2013). On the northern margin, the Asian continent underthrusts to the south below Tibet, probably by ~200-250 km since the Eocene (Replumaz et al., 2010). The northward motion and collision of India has been accompanied by two, plume-like, presumably hot, upwelling flows. The rear one, already mentioned, appears related to the Deccan trap outpouring around ~65 Ma, and is presently active beneath the Carlsberg ridge. The frontal one appears related to the Hangay dome, which formed during multiple episodes of flood volcanism from ~28 Ma onward.

At present, India is moving northward at a sustained and significant rate, causing widespread deformation across Tibet and into Central Asia to the north. Mantle flow computations indicate that India's present motion may be sustained by mantle drag via a "conveyor belt", i.e., a convection cell with the upwelling limb centered on the Carlberg ridge and the downwelling limb centered on the deep subduction (Alvarez, 2010; Becker and Faccenna, 2011).

In summary, irrespective of variable tectonic settings and subducting plate natures, the Cordillera and the Tethyan, Tibetan-Himalayan system share the following common features:

- they formed at around the same time in the Tertiary;
- they are both characterized by extreme crustal thickening and the formation of a double-vergent orogenic system flanking a high internal plateau;
- the onset and progression of crustal thickening is accompanied at depth by slab penetration into the lower mantle;
- they are both located over the downwelling limbs of a large-scale, whole mantle convection cell.

We can deduce that the evolution of those two orogenic systems is directly related to the establishment of whole mantle

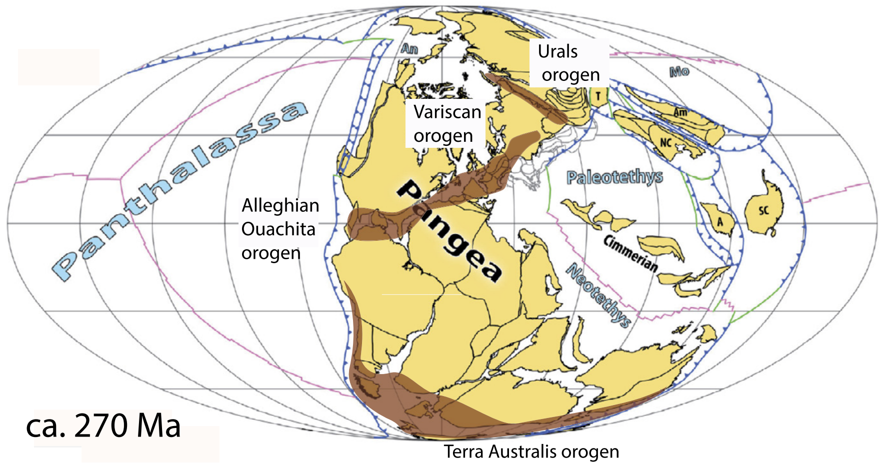

Fig. 10. Continental position during Pangea assembly (modified from Domeir and Torsvik, 2014). Orogenic belt in brown as from Cawood and Buchan (2007).

convection cells. Such a large-scale mantle convection system has previously been discussed in the context of the Indo-Atlantic box convecting system (Davaille et al., 2005), where the motion of Africa-India and South America are related to one another (Silver et al., 1998; Conrad and Lithgow-Bertelloni, 2007). It is then interesting to explore if similar correlations between different orogenic systems also occurred further back in the past.

6. Back in time: the Pangea orogeny

Pangea is the most recent and thus best documented supercontinent. It formed by the amalgamation of Gondwana, Laurentia/Baltica, and the Siberia–Asia continental masses. Several orogens formed during this assembly (e.g., Stampfli et al., 2013; Domeier and Torsvik, 2014) (Fig. 10). The Alleghian-Ouachita-Variscan orogen formed by the collision of Gondwana with Laurentia and Europe, and the Urals formed by the suture of East Europe with Siberia-Asia. Those orogens are collisional, internal orogens related to introversion-type continental amalgamation. The initial collision of cratonic blocks is dated back to ∼360 Ma. The main crustal thickening episode and thermal events occurred between 320 −280 Ma, followed by final suturing/cooling and uplift at 260 Ma (Cawood and Buchan, 2007). Overall, the orogens lasted for ∼100 Ma. Contemporaneously at the periphery of the supercontinent, all along the southern and western Pangea margin (Cawood and Buchan, 2007), the Gondwanide orogen formed lasting from 300 to 230 Ma.

Therefore, during the Pangea orogeny, the internal collisional orogen occurred at roughly the same time, or slightly predates, the accretionary orogens located along the margins. Similar connections have been documented for the Gondwana assembly, from 580 to 510 Ma (Cawood and Buchan, 2007) and possibly even further back in time (Cawood et al., 2016). The correspondence of the active phase of both internal and external orogens has been related to a global kinematic plate re-adjustment. It has been suggested that this re-adjustment can be related to the terrane accretion or, more probably, to an increase in the mechanical coupling within the collisional interior that was transferred to the external margin (Cawood and Buchan, 2007).

Yet, the correspondence observed during Pangea between the formation of an internal collisional orogeny and an external Pacific belt holds also in the Tertiary. Even though the Tertiary deformation is not (yet) related to super-continental assembly, Tertiary plate kinematics resemble those active during Pangea assembly (Collins, 2003). The model proposed here, where significantly thickened and large-scale orogenic belts are only formed over whole mantle downwellings associated with lower mantle subduction, may thus also apply for Pangea.

The Mesozoic is mostly associated with continental dispersal, without large-scale orogenic episodes, with the exception of the Jurassic Cimmerian block accretion in the Tethys. The first large-scale crustal thickening episode may be related to the Laramide in the Late Cretaceous. The lack of large-scale episodes of continental thickening in the Mesozoic is consistent with the idea that during this time slabs were not anchored in the lower mantle, which facilitated slab rollback and continental dispersal (Bercovici and Long, 2014). For example, the opening of the Atlantic and the westward motion of the Americas appears to have been facilitated by the retrograde motion of the Cordillera subducting slab (Husson et al., 2012).

In the Mesozoic, the style and planform of convection cells was likely different to what we infer today (Zhong et al., 2007). The characteristics of this evolving convective pattern can be investigated by tracking the geographic locations of net plate convergence and divergence, under the assumption that they correspond to deep mantle downwelling and upwelling, respectively (Conrad et al., 2013). This analysis suggests that the divergent component did not have change significantly during the last 250 Ma and may have remained positioned over a large-scale upwelling beneath Africa and Pacific. Conversely, convergence shifted from an equatorial focus during the Tertiary, to a more polar position in the Mesozoic, and then back to an equatorial position at the end of the Paleozoic during Pangea assembly. This implies that degree two convection with two mantle downwellings, as inferred during the last 50 Ma, is not a strictly stable feature but changed between ∼100 and 200 Ma at least for what concern the depth extent of the downwelling component (e.g., Zhang et al., 2010).

In summary, the history of orogeny and the connection with deep mantle convection during the Tertiary suggests that only lower mantle subduction produces orogeny and sustains extensive crustal thickening. Currently, this is related to global mantle circulation within the Indo-Atlantic "box" (Davaille et al., 2005), through which the Cordillera and Tethyan orogeny are linked (Silver et al., 1998), including in terms of their timing at ∼50-60 Ma. A similar connection between convection and orogeny may hold in the Phanerozoic during the assembly of Pangea. Therefore, large scale orogeny may be viewed as the surface expression of vigorous whole mantle convection episodes during earth history (Alvarez, 2010; Faccenna et al., 2013; Hoffman, 2014). If this model is correct, we should then expect that phases of lower mantle subduction (avalanching?) produce pulses of large-scale compression, favoring the inversion of continental margins and eventually continental assembly (cf. Condie, 1998).

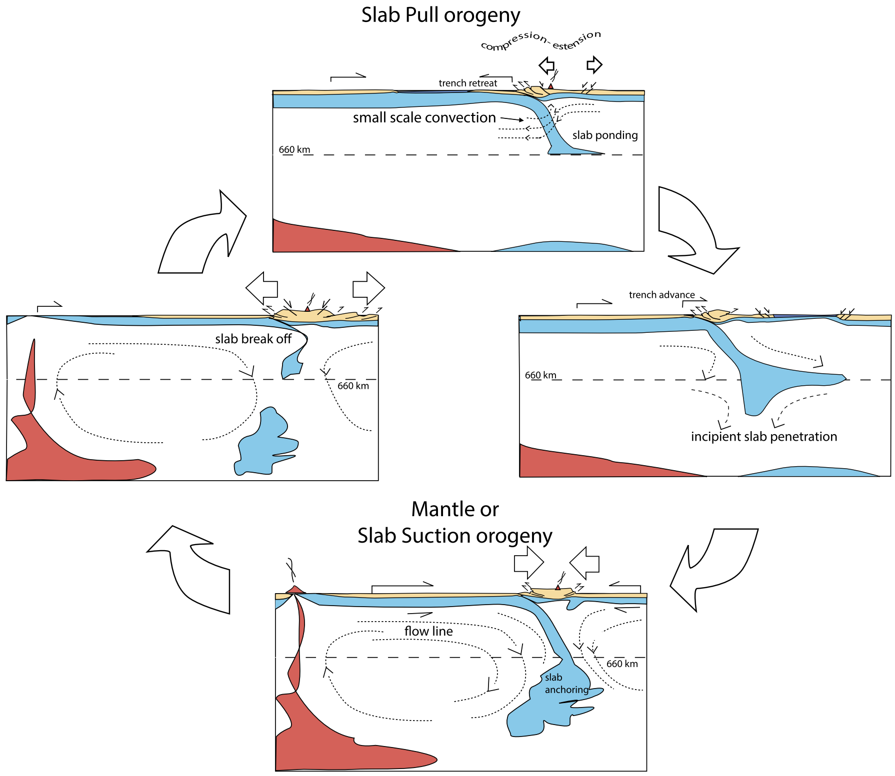

Fig. 11. Conceptual model for an orogenic cycle. The duration of the whole cycle could last for 10 to 100 Myr. Slab penetration induces a change from slab-pull to mantle orogeny type, producing large scale compression in the upper plate. Slab break-off and total uncoupling from upper to lower mantle slab could generate a general collapse of the large orogen, returning back a new stage of subduction orogeny.

7. Orogeny, the Wilson cycle, and super-continent assembly

There is general agreement on relating the formation of wedge-type orogeny to subduction (e.g., Royden, 1993; Jamieson and Beaumont, 2013). The origin of the large-scale, double vergent orogens and the related formation of high plateaus is, however, more enigmatic (e.g., Vanderhaeghe, 2012). Previous models suggested that protracted compression could favor the transition from wedge type to large scale Himalayan-Tibet by migration of the "*S*-point"; the deep contact point between the descending and upper plate (Jamieson and Beaumont, 2013). However, this model does not seem to apply to the case of the Cordillera. In our alternative model, double-vergent orogeny represents the surface expression of mantle processes, related to slab anchoring at depth and to the onset of a whole mantle convection cell, dragging the subducting and upper plate against one another for a protracted period of time.

Subduction and mantle orogeny are here presented as end-members. However, since they are related to slab dynamics, which can be highly time-dependent, it is of course possible to transition from one to the other type and stage of orogen growth as part of an evolutionary cycle (Fig. 11). During the initial phase of subduction, slabs accumulate in the deep transition zone, i.e., stagnating at depths between ∼700 and ∼1000 km. Under this "ponded" slab configuration, subduction zones and their trenches predominantly retreat, favoring the formation of an orogenic wedge, and an asymmetric pile of crustal slices tapering toward the subduction zone with limited (<50 km) crustal thickening.

Several models have shown that the penetration of slabs into the lower mantle is only temporarily inhibited by the action of the viscosity increase and negative Clapeyron slope phase change around the 660 km seismic discontinuity (e.g., Zhong and Gurnis, 1995; Christensen, 1996; Billen, 2008; Yanagisawa et al., 2010; Garel et al., 2014; Agrusta et al., 2017). After a few tens of Myrs, i.e., once enough subducting slab accumulates within the transition zone, the slab starts to penetrate, albeit at lower rate, into the lower mantle. As a consequence, the slab is anchored by the more viscous lower mantle, which inhibits the lateral migration of the trench and activates a vigorous mantle return flow that drags the two plates against one another (Yamato et al., 2013; Faccenna et al., 2013; Dal Zilio et al., 2018).

This is the case for Nazca subduction beneath South America, where the slab penetrates into the lower mantle after a long slab rollback episode, which provides the impetus for the onset of Andean orogeny (Faccenna et al., 2017). The activity of the mantle orogeny may last for several tens of million years. Mantle flow associated with the large-scale convection cell may produce plumes on the upwelling side of the convection cell (Husson et al., 2012; Fig. 7). This system acts as a conveyor belt, sustaining a protracted mantle drag over sustained periods of time (Faccenna et al., 2013; Rowley et al., 2016) as for India and Arabia at present (Becker and Faccenna, 2011). The action of downwelling and upwelling limbs of the convection cells favors a double-sided orogen.

Once the deep portion of the mantle slab is disconnected from the shallow one, as for example after slab break-off, the anchoring effect slowly vanishes along with the mantle drag (Fig. 11). Without the continuous drag from the downwelling mantle, the overthickened continental crust will collapse due to its gravitational potential energy, and the mantle orogen will progressively evolve into a subduction orogen. This is the case for the formation of the Basin and Range after the Laramide orogeny, or the case of the Aegean subduction system, after the Rhodope orogeny was followed by orogenic collapse and a fast phase of slab rollback (Jolivet and Faccenna, 2000). The duration of this entire orogenic cycle may be of the order of 100 Myr.

The contemporaneous activity of accretionary type orogens and collisional type orogens, as observed today, holds also during the formation of Pangea and Gondwana (Cawood and Buchan, 2007).

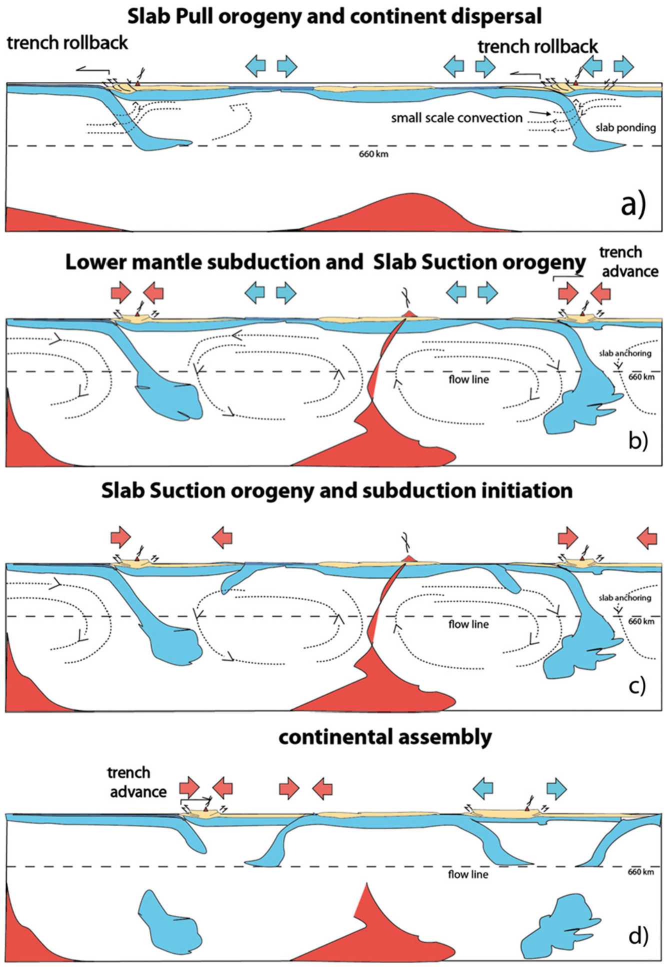

Fig. 12. Conceptual model for an orogenic Wilson (1966) cycle, as is linked to mantle convection and ultimately leads to supercontinent assembly. The model shows the possible linkage between accretionary-Cordillera type and internal-Collisional type orogenies. The penetration of the slab into the lower mantle (stage b) induces compression and consequent subduction initiation at passive margins (stage c), leading to the closure of the Atlantic and Indian ocean (stage d).

Thus, it is possible to include the concept of an orogenic cycle into the large, global-scale supercontinental cycle, and so link the time-dependent recycling of oceanic plates to the continental plate record.

Several models have been proposed to address the dynamics of the supercontinental cycle. The dispersal of continents could be related to the heating and upwelling of continent-insulated mantle (see, e.g., Gurnis, 1988), it could be driven by slab avalanching that in turn excites upwelling from the lowermost mantle (Zhong et al., 2007; Li and Zhong, 2009; O'Neill et al., 2015), or emerge spontaneously by slab rollback (Bercovici and Long, 2014).

Our preferred scenario starts after continental assembly, here explored for the Indo-Atlantic box (Fig. 12). The dispersal of the continent occurred under upper mantle restricted convection that favored slab rollback (stage a; e.g.. Bercovici and Long, 2014; Cawood et al., 2016). After a phase of stagnation of subducted material in the transition zone, the slab can penetrate and thus experiences anchoring in the lower mantle (Fig. 12b). This produces a decrease in the trench retreat rate and triggers large-scale whole mantle convection (Zhong et al., 2007; Li and Zhong, 2009; Husson et al., 2012; Faccenna et al., 2013, 2017; Yang et al., 2018), which then drags the plates towards the downwelling zone (Faccenna et al., 2017). This induces compression not only over the downwelling zone, forming a thick mantle orogen, but also over the entire plate system (Husson et al., 2012; Yamato et al., 2013; Faccenna et al., 2017; Yang et al., 2018). Under these conditions, a passive Atlantic-type margin can be slowly inverted and converted into an active margin (Faccenna et al., 1999; Yamato et al., 2013; Baes at al., 2020), which could ultimately lead to the closure of the Atlantic (Wilson, 1966) if the trench on the Pacific side advances toward the upper plate (Fig. 12c, d). This scenario likely played a role in the assembly of Pangea. Whether Pangea formed on top of the African plume upwelling, and if this is dynamically plausible, is debated (Torsvik et al., 2008; Li and Zhong, 2009; Conrad et al., 2013). Irrespective of that, deep mantle subduction would naturally excite return flow upwellings and buoyant plumes (e.g., Tan et al., 2002; Fig. 7), perhaps from the large, low shear wave velocity province margins. Those plumes might then locally weaken the upper plate and favor the initial breakup (e.g., Condie, 1998; Li and Zhong, 2009). Subsequently, rollback of the slab subducting beneath the Americas would again trigger continental dispersal during a new phase of restricted upper mantle convection (Fig. 12a). Recent modelling effort support this idea, showing that mantle drag is acting efficiently during collision and increasing during period of supercontinental assembly (Coltice et al., 2019).

The suggested relationships between subduction and the supercontinental cycle can be best illustrated along an equatorial global cross section, going back to the Pangea stage and imagined forward to the next assembly (Fig. 13). The contemporaneous formation of an accretionary, Gondwanide orogen and the Variscan-Appalachian and Ural collisional orogens in the Pangea stage suggests that the plate system was under compression (Fig. 13h), and likely organized with a convection pattern similar to that of the present-day (Conrad et al., 2013), which is linked to deep mantle subduction. Afterward, during the entire Mesozoic, slab rollback provided the free lateral boundary condition for continental dispersal to occur under the action of continuous upwelling beneath the continental interior (Fig. 13g). In this phase, slabs were then disconnected from their deeper portions and mainly restricted to the upper mantle. As a consequence, no large orogens were formed. Then from 50-60 Ma onward, a new phase of lower mantle slab penetration occurred, establishing a degree-two style of convection, and creating the necessary conditions for a new compressional and orogenic phase (Fig. 13e-f). If this model is correct, subduction into the lower mantle could be considered as intermittent from the Paleozoic onward. Episodes of lower mantle subduction occurred mainly in the Paleozoic and Tertiary, while in the Mesozoic slabs would have been mainly confined to the upper mantle (Machetel and Weber, 1991; Tackley et al., 1993; O'Neill et al., 2015).

As to what will happen in the future, we can speculate about two possible scenarios. If the Caribbean and South Sandwich subduction zones can propagate laterally and spread over the Atlantic seaboard, the system may evolve to a new supercontinental phase leading to the formation of Pangea Ultima by closure of the internal ocean, i.e. introversion type assembly (Fig. 13d, c; Nance et al., 1988; Scotese, 2000; Murphy and Nance, 1991). This implies that the Cordillera subduction zone will advance over large distances or, more probably, will vanish as it converts from an active to a passive margin. The alternative scenario is extroversion, caused by the closure of the external Pacific Ocean, forming the Amasia supercontinent (Hoffman, 1992) (Fig. 13a, b). Extroversion is favored by slab rollback as it implies the fast consumption of a large oceanic domain. Rodinia and Gondwana (Pannotia) probably belong to this style of continental assembly (Hartnady, 1991). From this perspective, the style of continental assembly is determined by the style of subduction, where deep lower subduction favors introversion (Condie, 1998) and upper mantle, ponded slab subduction

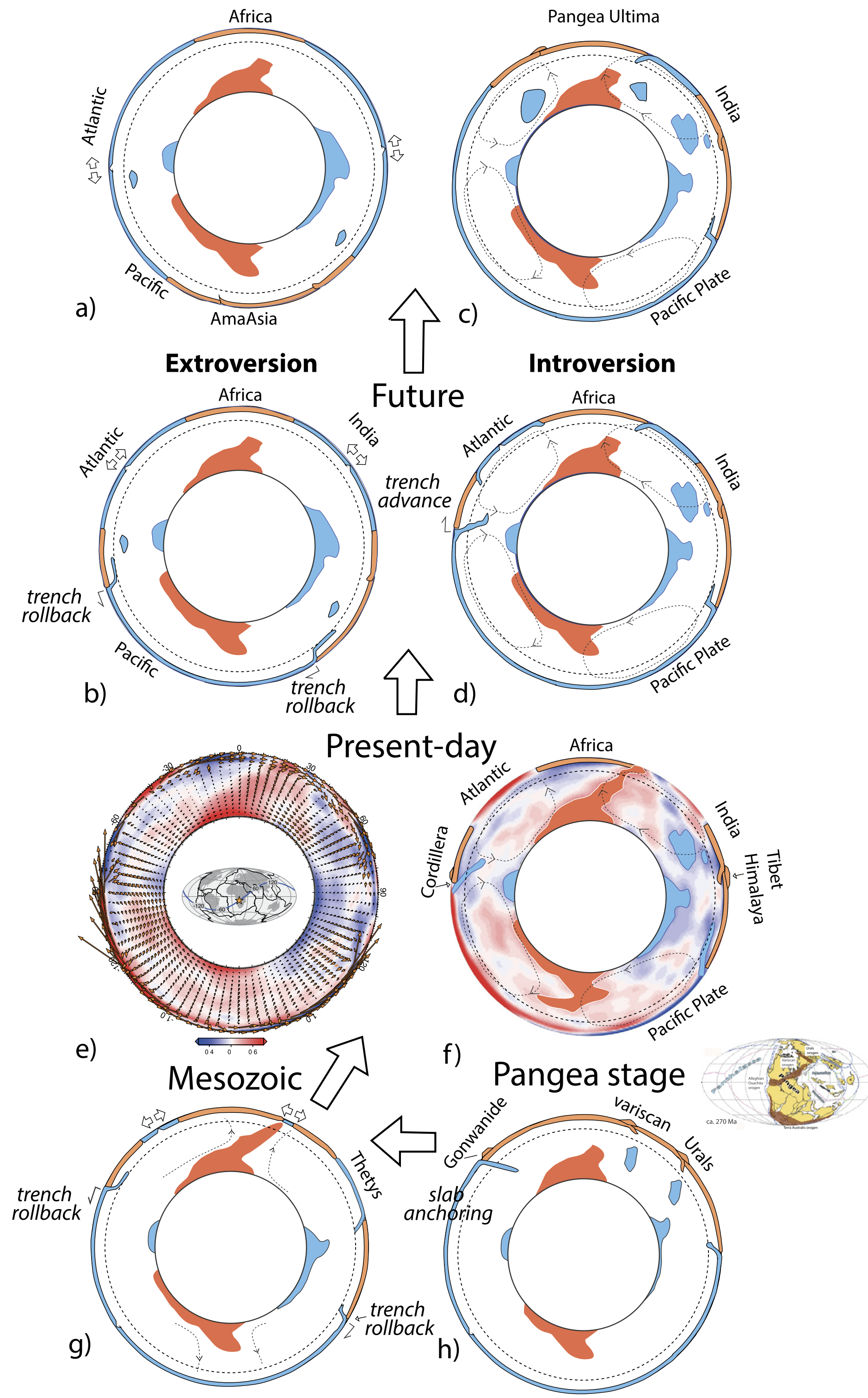

Fig. 13. Transition from Pangea to next future supercontinent, Pangea Ultima (introversion) or to AmaAsia (extroversion) from a mantle convection perspective. Note that Pangea sections have been rotated to a more northeastern direction to better cross-cut the different orogenic systems (see inset). The reconstruction assumes that Africa did not substantially translate longitudinally during the last 250 Ma (Torsvik et al., 2008). e) Convection model as in Faccenna et al. (2013).

and trench rollback favors the extroversion mode (Silver and Behn, 2008). The time-dependent penetration and anchoring of slabs in the lower mantle depends upon a range of parameters, such as slab age and velocity, mantle viscosity structure, and the 3-D extent of the slab. Further exploration of such dynamics should be viewed within an integrated tectonic-convective framework.

8. Conclusions

We find it useful to distinguish between two orogeny end-members: those of *subduction* or *slab-pull* type, related to predominantly one-sided subduction confined to the upper mantle, and those of *mantle* or *slab-suction* type, related to whole mantle scale convection. The associated structure of orogeny is distinct: in the first case, the shape of the belt is highly asymmetric, with tapering toward the subduction zone, crustal thickening that does not exceed ∼50 km, and an orogenic belt mainly composed of lower plate units. Mantle orogeny, conversely, shows a more symmetric shape and large crustal thickening within the upper plate.

We test this idea using idealized numerical models exploring the relationships between mantle dynamics and orogeny. Our results imply that extreme crustal thickening and significant upper plate compression likely form when the slab enters and anchors into the lower mantle, which induces a large-scale convection cell that drags the two plates against each other for a protracted period of time. This effect can be enhanced by thermal upwellings from the lower mantle that are near to, and triggered by, the lower mantle slab.

We explore the applicability of this model by considering the present-day distribution of compressional backarc regions, and Tertiary and Late Paleozoic orogeny episodes, and reach the following conclusions:

i) The relationships between maximum slab depth for representative subduction zones as inferred from seismic tomography (Li et al., 2008a; Fukao and Obayashi, 2013; van der Meer et al., 2018) show that most of the backarc regions where slabs reached depths larger than ∼1000 km are under compression.
ii) The central Andes Cordillera and the Tibetan-Himalayan system formed at around the same time in the Tertiary. Based on kinematic reconstructions, we show that the onset and progression of crustal thickening in both areas is accompanied at depth by slab penetration into the lower mantle, and both regions are located over the downwelling limbs of a large-scale, whole mantle convection cell. Therefore, irrespective of the variable tectonic settings and subducting plate styles, we infer that extreme crustal thickening is driven by mantle convection.
iii) Prior to the Tertiary, large-scale orogeny occurred during Pangea assembly when the Gondwanide accretionary orogeny and the Alleghian-Ouachita-Variscan collisional orogeny formed. We propose that this large-scale episode of orogeny is also induced by large scale mantle flow; lower mantle subduction leads to continental assembly.

If our model is correct, orogeny and extreme crustal thickening episodes can be used to decipher time-dependent mantle convection. This also implies that that lower mantle subduction may have occurred episodically throughout much of the Phanerozoic, inducing the large-scale compressional stresses that eventually lead to supercontinental cycling.

CRediT authorship contribution statement

Claudio Faccenna: Conceptualization, Modeling analysis, Writing. Thorsten W. Becker: Conceptualization, Numerical modeling, Writing. Adam Holt: Conceptualization, Numerical modeling, Writing. Jean Pierre Brun: Conceptualization.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary material

Supplementary material related to this article can be found online at https://doi.org/10.1016/j.epsl.2021.116905.

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