Plume-like upper mantle instabilities drive subduction initiation

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[1]   The control of upper mantle-lithosphere interactions (MLI) on compressional tectonics is not well resolved. This applies to the role of MLI in triggering of subduction initiation or lithosphere mantle (LM) downwellings. We present results of thermo-mechanically thermo-dynamically coupled numerical experiments that are consistent with an array of recent geophysical constraints on lithosphere and upper mantle rheology and structure. We demonstrate that MLI can lead to initiation of continental lithosphere subduction, inducing its spontaneous downthrusting to depths of 300–500 km upon plume impingement of the lithosphere. This downthrusting is pre-conditioned by rheological stratification of visco-elasto-plastic lithosphere and its free surface. The subsequent evolution of the slab is governed by phase changes and its interactions with the surrounding mantle. We demonstrate that the mode of MLI is strongly affected by the lateral heterogeneities and the presence of suture zones in the lithosphere. **Citation:** Burov, E., and S. Cloetingh (2010), Plume-like upper mantle instabilities drive subduction initiation, *Geophys. Res. Lett.*, *37*, L03309, doi:10.1029/2009GL041535.

1.   Introduction

[2]   Subduction initiation is a key element in plate tectonics. Although plate tectonics theory has advanced significantly [*Wessel and Muller*, 2007], not much progress has been made in quantitative understanding of the mechanisms for the initiation of subduction zones, especially in continental domains [*Schmeling et al.*, 2008]. This is largely due to insufficient integration of data on realistic visco-elastic-plastic lithosphere rheology and structure in geodynamic models (Figure 1) [*Schmeling et al.*, 2008]. Concepts for initiation of subduction should also by their very nature include mantle-lithosphere interactions (MLI), another key element for plate tectonics.

[3]   Previous studies [e.g., *Cloetingh et al.*, 1982, 1989; *Stern*, 2004; *Ueda et al.*, 2008] have focused on the oceanic subduction initiation where plate strength and negative buoyancy associated with plate cooling play a major role. In continents, positive mean buoyancy of the lithosphere does not favour subduction, and, in contrast to oceanic lithosphere [*Vlaar and Wortel*, 1976], there is no correlation between subduction mode and plate age. Hence, other factors such as the rheological stratification of continental

lithosphere and MLI might be key in subduction initiation. In particular, plume-induced detachment of dense strong lithosphere mantle from buoyant crust might be a prerequisite for sustainable subduction that requires a strong forcing and destabilization of lithosphere [*Burov and Guillou-Frottier*, 2005; *Burov and Cloetingh*, 2009]. For this reason, MLI and related thermo-gravitational instabilities have recently received much attention [e.g., *Burov and Guillou-Frottier*, 2005]. Yet, controversial interpretations remain [e.g., *Lustrino and Carminati*, 2007], both in terms of modelling concepts and observations (Figure 1).

[4]   Continental lithosphere essentially differs from oceanic lithosphere as a result of its rheological and density stratification [*Burov*, 2009]. As shown by *Cloetingh et al.* [1982], an increase of lithospheric density with aging alone is insufficient to create suitable conditions for oceanic lithosphere subduction initiation, due to its increased strength as a result of cooling. However, in continents numerous examples exist [*Anderson*, 1994] where very old cratonic lithosphere undergoes long lasting subduction (i.e., India-Asia collision). Continental subduction requires either or all, a strong horizontal force (e.g., ridge push), local weakening, and a special mantle entraining mechanism [*Toussaint et al.*, 2004]. Slab pull forces, driving oceanic subduction, are not efficient in continents because considerably slower rates of continental subduction promote continent-ocean slab break-off at initial stages of subduction [*Burov and Yamato*, 2008].

[5]   It was shown that oceanic subduction could be initiated within narrow time interval around 30 Ma after sea floor spreading, when the combination of gravitational instability and relatively low plate strength is optimal [*Cloetingh et al.*, 1982]. Similarly, in case of slow convergence rates (i.e., <1.5 cm/y; e.g., Alpine collision), continental subduction is limited to the first 2–5 Myr of convergence [*Yamato et al.*, 2007]. For passive continental margins, classical subduction initiation also occurred only at young margins. Therefore, ample evidence exists that external factors are required to initiate subduction of oceanic or continental plates [*Cloetingh et al.*, 1989; *Faccenna et al.*, 1999]. Among these factors, the presence of pre-existing weakness zones should facilitate the onset of rupture required to initiate subduction [e.g., *Schmeling et al.*, 2008].

2.   A New Concept for Initiation of Continental Lithosphere Subduction

[6]   We propose that MLI can provide the necessary weakening, destabilisation and entraining of the lithosphere. The mantle part of continental lithosphere may be gravitationally unstable so that it is prone to detach from the positively buoyant crust [*Burov*, 2007]. This has motivated us to investigate whether such a strong destabilizing event as plume impingement or plume-like diapiric mantle instability may result in initialization of subduction-like processes.

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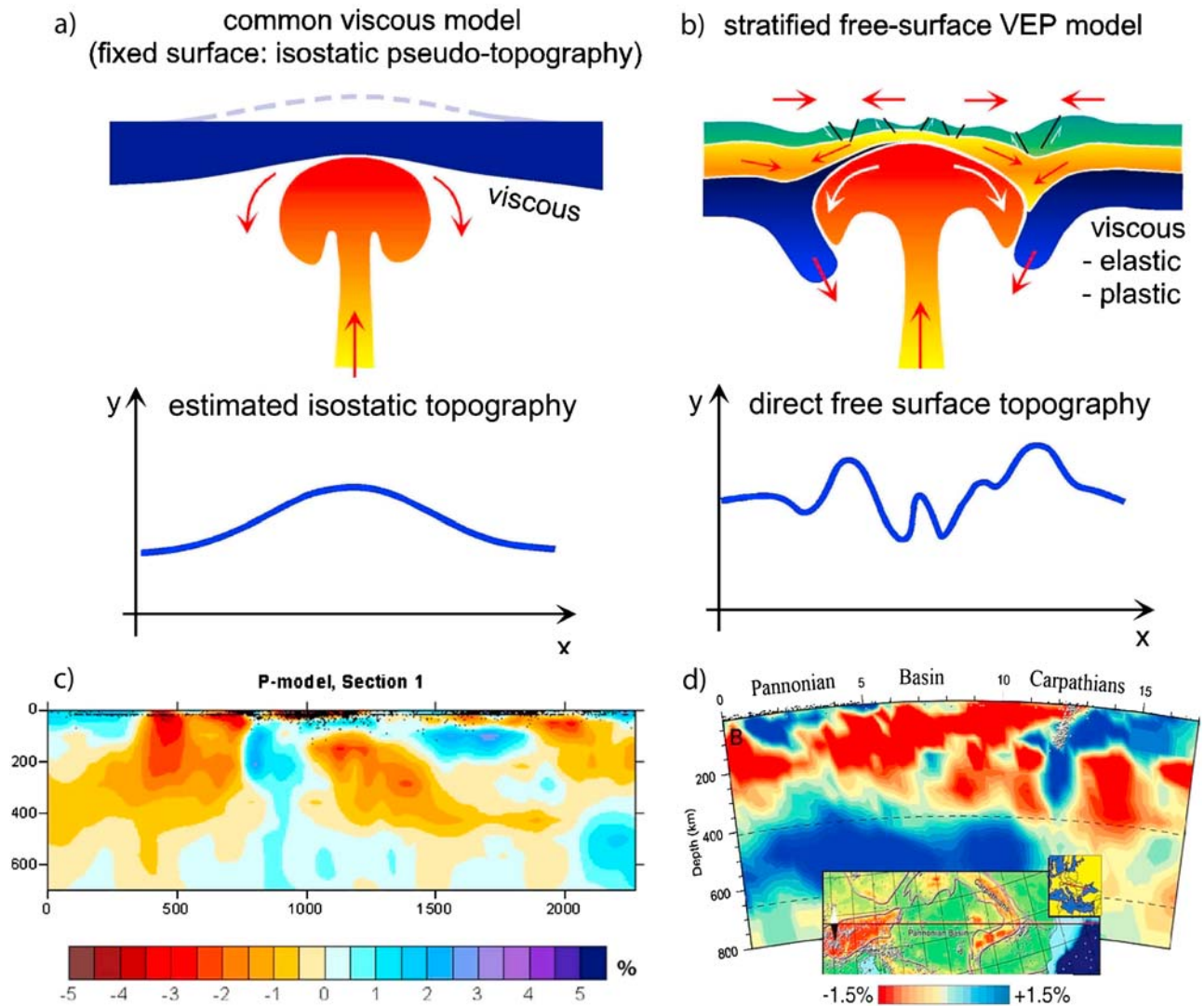
Figure 1. Mechanism for initiation of subduction by plume-lithosphere interaction. (a) Conventional monolayer rigid top viscous model of MLI and predicted long-wavelength dynamic topography. (b) Stratified visco-elasto-plastic model: internal ductile layers (green, upper crust; yellow, lower crust; and blue, mantle) damp plume impact and deform with their own characteristic wavelengths; mantle lithosphere can be downthrusted in subduction-like motion. (c) Local tomographic cross-section through the Massif Central [*Koulakov et al.*, 2009] showing low-velocity upper mantle anomaly below the Massif. (d) Regional tomographic cross-section through the Pannonian basin-Carpathians showing a wide low-velocity upper mantle anomaly below the basin and a slab-like high-velocity anomaly below the range [*Wortel and Spakman*, 2000].

Important in this respect is that continental crust and mantle may mechanically decouple from each other due to the presence of mechanically weak ductile zones (or "channels") between the upper and lower crust and lower crust and mantle [*Burov*, 2009]. Upon its emplacement below rheologically stratified lithosphere, the plume head exerts flexural deformation, basal shear and tension in the overlying plate (Figure 1). This leads to mechanical weakening due to strain-induced inelastic yielding [*Burov*, 2009] and to development of mechanical instabilities. In thermally young lithospheres (<300 Ma old), ductile crustal channels may be thick (10–20 km), with viscosities as low as 10^{20}–10^{21} Pa s [e.g., *Burov*, 2009]. Such easily deformable channels damp vertical undulations (<2 km) of the mantle lithosphere caused by MLI. In response to MLI-driven basal shear, crustal layers deform with their own characteristic short wavelengths (Figure 1) [*Burov and Guillou-Frottier*, 2005] that strongly resemble to observations on differential vertical motions and tectonic deformation in intracontinental domains [*Burov et al.*, 2007]. In case of crust-mantle decoupling, mantle lithosphere may be separated from buoyant crust and downthrusted to considerable depths (>300 km), mimicking subduction. Such downthrusting is favoured at zones of inherited weakness or at mechanical contrasts such as plate boundaries and passive continental margins. These novel aspects of MLI are a direct consequence of the "active" role of lithosphere ignored so far in conventional convection models [*Schubert et al.*, 2001].

[7] These findings are relevant for areas such as the Azores-Gibraltar zone sited at the transition of oceanic lithosphere in the Azores area, affected by plume activity [*Silveira et al.*, 2006] and the Gulf of Cadiz-Goringe Bank underlain by rifted margin lithosphere(N. Zitellini et al., Synclines as prime expression of compressional deformation of the lithosphere: The central Atlantic segment of the Iberia-Africa Plate Boundary, submitted to *Geology*, 2009). The rheological transition between oceanic lithosphere and rifted margin lithosphere might explain the observed change from a narrow plate boundary in the Azores area to a diffuse plate boundary zone in the Goringe Bank area (N. Zitellini et al., submitted manuscript, 2009). In the latter area, tomographic images [*Wortel and Spakman*, 2000; *Gutscher et al.*, 2002] provide evidence for subduction/collision in terms of an eastward dipping slab under the Gibraltar arc system. Evidence for the occurrence of lithospheric folding in the overlying Gulf of Cadiz comes from gravity, topography and deep-seismic reflection profiling (N. Zitellini et al., submitted manuscript, 2009).

3. Mantle Downthrusting and MLI

[8] Based on previous studies [e.g., *Fletcher and Hallet*, 1983; *Burov and Molnar*, 2008], the minimal condition for instable behaviour, and thus mantle detachment, is

$$\Delta\rho g d/2\tau_{\max} < d/d_{\text{fold}} \qquad (1)$$

where $\Delta\rho$ is the density contrast between the lithosphere mantle and the material below, g is the acceleration due to gravity, d is mantle lithosphere thickness, $\tau_{\max}$ is its maximal mechanical strength, and d_{fold} is the characteristic depth decay length of ductile strength. In most cases $d/d_{\text{fold}} \sim 10$ and $\Delta\rho g d/2\tau_{\max} \leq 3$ [e.g., *Burov*, 2007]. Consequently, a small perturbation is sufficient to develop a LM gravitational instability.

[9] The potential role of instabilities during horizontal tectonic deformation can be characterized by Deborah number, *De* [*Burov*, 2007]:

$$De = t_{\min} u_x/L = 13.04 \, \mu \, u_x/(\Delta\rho g d \, L), \qquad (2)$$

where $t_{\min}$ is characteristic growth time of the instability (1–5 Myr), μ is mantle viscosity, L is the size of the area affected by MLI, u_x is the subduction or convergence rate. If $De \leq 1$, which is a typical situation, sustainable LM detachment may occur after 2–5 Myr since the onset of MLI. The sustainment of continental subduction [*Toussaint et al.*, 2004] mainly depends on crustal/mantle rheology and on the convergence rate u_x, and is most likely for median values of u_x (3–6 cm/yr) and intermediately strong rheology (high rates/strong mantle favour folding, slow rates/weak mantle favour instabilities).

4. Numerical Approach

[10] We used our thermo-mechanical code Flamar v12 to assess the response of multilayered visco-elasto-plastic lithosphere to MLI in various tectonic settings. The code is described in the auxiliary material and previous studies [e.g., *Burov and Cloetingh*, 2009]; here we limit ourselves to its essential abilities to handle (1) large strains and visco-

elastic-plastic rheologies, (2) Mohr-Coulomb failure (faulting), (3) pressure (P) - temperature (T) strain-rate dependent ductile creep, (4) mineralogical phase transitions, (5) internal heat sources, and (6) free surface boundary conditions and surface processes.[1] The model setup is the same as that used by *Burov and Cloetingh* [2009]: the initial plume of 200 km in diameter is presented by a 250°C temperature anomaly at 650 km depth. The lithosphere thickness is 150 km for ages < 300 Ma and 250 km for "cratons" (≥300 Ma); 40 km-thick crust consists of 20 km-thick dry granite upper crust and a 20 km-thick dry diabase lower crust. The density is updated dynamically as function of P-T using thermodynamic free-energy minimization [*Connolly*, 2005]. The temperature at the base of the upper mantle is 2000°C, which corresponds to double-layer convection [*Burov and Cloetingh*, 2009]. The initial age-dependent thermal gradient in the lithosphere is computed from half-space cooling model [e.g., *Burov*, 2009] with T=1330°C at the base of the lithosphere. The initial thermal gradient in the sub-lithosphere mantle is adiabatic down to the 550 km depth. The 550 to 650 km depth interval presents the lower boundary layer where the thermal gradient is super-adiabatic. Zero thermal out-flux is the lateral boundary condition. The mechanical boundary conditions are: free upper surface; the lateral borders move at a rate with which the plume material spreads away; hydrostatic bottom. The considered "plumes" are "small" (200 km in diameter) and we only consider the initial stages (first Myr) of their emplacement below the lithosphere.

5. Coupled Thermo-mechanical Thermo-dynamic Numerical Models for MLI

[11] We performed a series of high-resolution (5 km x 5 km elements) experiments to explore the influence of MLI near contrasting plate boundaries on initiation of subduction.

[12] The experiments presented in Figures 2a and 2b display a plume rising below a laterally homogeneous continental 300 Ma old plate, whereas Figure 2c shows the results of a experiment where two old blocks (300 Ma) are embedded in a younger plate (150 Ma). The plume quickly erodes the mantle lithosphere and intrudes in-between the highly positively buoyant crust and denser mantle. In case of contrasting plate boundaries (Figure 2c) this intrusion results in decoupling of the upper crust and lower crust that remains attached to the mantle. The upper crust undergoes intensive brittle faulting (Figure 2b), while the mantle lithosphere is detached and entrained down in a subduction-like motion, sometimes (Figure 2c) involving the lower crust. Subduction is sustained if lateral shortening is applied at the borders of the model.

[13] The experiments of Figure 3 show a plume rising below a boundary between two continental plates of different age, strength and thickness. In this case the induced subduction has a strong polarity: downthrusting is more important on the side of the thicker plate. This process resembles the near-vertical downthrusting of the eastern European craton and the Moessian platform in the bend zone of the Romanian

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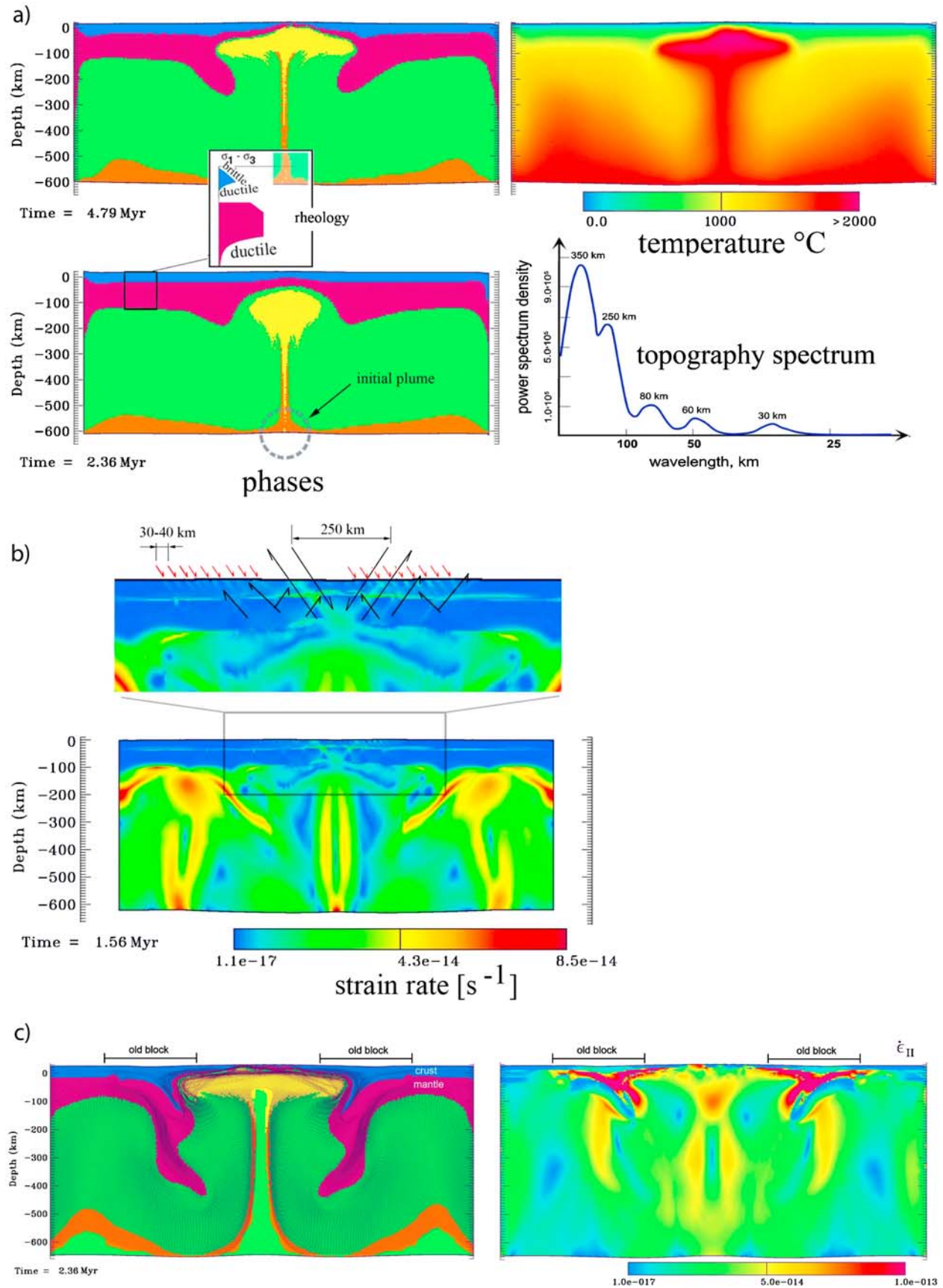[1]Auxiliary materials are available in the HTML. doi:10.1029/2009GL041535.
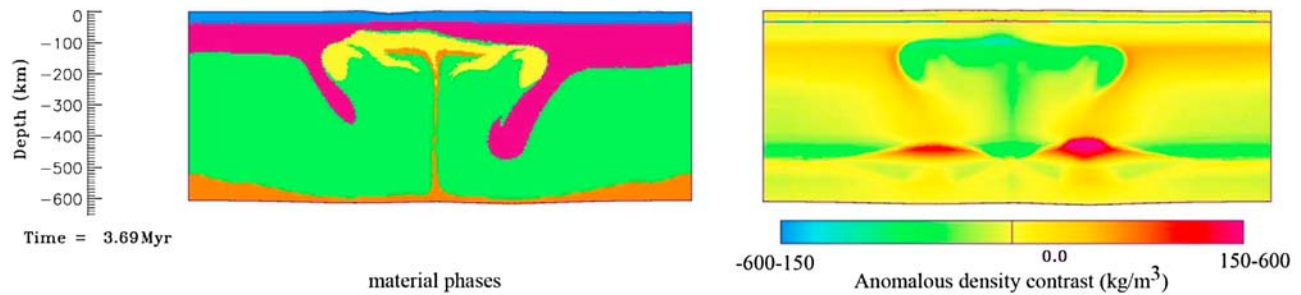
Figure 2

Figure 3. A plume ascending under the boundary (suture) between two continental plates of different age and thickness (150 and 300 Ma old, 150 km and 200 km, respectively). (left) Material phase field. (right) Anomalous density distribution computed from phase changes (deviation from mean petrologic (PERPLEX) density similar to that inferable from the GERM/PREM model). Positive anomalies correspond to excess density compared to the average density for the given depth. Once initialized, mantle lithosphere downthrusting becomes a self-sustaining process.

Carpathians (Figure 1d) [*Cloetingh et al.*, 2004] and the relatively low angle subduction under the Dinarides of the Adriatic plate characterized by a much younger thermo-mechanical age. If the plume rises below the boundary between the plates or just below the thin plate, downthrusting occurs from the thicker plate towards the thinner plate. This pattern is strikingly similar to the configuration of the Neogene Pannonian back-arc basin system considerably weakened by mantle upwelling and the surrounding Carpathian arc [*Wortel and Spakman*, 2000; *Cloetingh et al.*, 2004]. The subduction polarity is inverted when the plume rises below thicker lithosphere, a situation resembling the Gibraltar arc system [*Gutscher et al.*, 2002].

[14] In case of super-continents, it has been shown that mantle instabilities are generated below the middle of the plate [*Anderson*, 1994]. The resulting continental rifting may be followed by crust-mantle decoupling and initiation of subduction at rifted margins. For passive continental margins, gravitational LM instabilities produced by ascent of hot asthenosphere during active rifting, can lead to large-scale cold downwelling [*Burov*, 2007]. The instabilities induced by a plume head are even more important, leading to both mantle and crustal downthrusting. As the lithospheric material goes down, it undergoes phase changes that increase its density (Figure 3) and downthrusting becomes a self-sustaining process.

6. Conclusions

[15] MLI provide an efficient mechanism for initiation of continental subduction in slow convergence settings at intraplate boundaries and passive continental margins. MLI may be an important forcing factor for oceanic subduction too, in particular at passive margin settings.

[16] Observational data on timing and main features of continental subduction [*Matenco et al.*, 2007] are consistent with our numerical experiments that show that slow mantle downthrusting cannot continue for time spans larger than few Myr.

[17] Initiation of subduction by MLI may be most efficient below plate boundaries such as passive margins. The well-known issue of asymmetry of subduction is solved in a natural way, without evoking supplementary mechanisms such as water hydration [*Matenco et al.*, 2007]. The polarity of the mantle downwelling is controlled by the position of the plume head with respect to the plate boundary: the downwelling is directed towards the plume centre. MLI can also explain the processes following the breakup of super-continents: the models predict that when super-continents split, their margins are downthrusted, such as inferred from the geological record [*Condie et al.*, 2000]. Important also is that small (~200 km in diameter) or shortly-acting plumes cannot split supercontinents but can initiate subduction at their margins. "Baby" plumes with a diameter less than 100 km are too small to initiate subduction. An intrinsic coupling between mantle upwellings and continental subduction implies a large degree of heterogeneity of the upper mantle, as predicted by petrologically consistent models for density and elasticity. This provides a new challenge for future seismic tomography studies. The MLI model also provides testable predictions on vertical motions and the interplay between simultaneously occurring extension and compression in continental realms.

Figure 2. Experiments for plume-lithosphere interactions below a single continental plate. (a) Experiment for MLI below a laterally homogeneous strong 300 Ma old continental plate. Color code: purple, mantle lithosphere; blue, upper and lower crust; green, deep mantle; yellow, plume; and orange, bottom marker layer. Predicted topography wavelengths are also shown. Horizontal-to-vertical scale ratio equals 1. (b) Typical surface fault distributions associated with MLI produced in the experiment of Figure 2a. (c) Experiment for MLI below a laterally heterogeneous plate that includes 300 Ma old (strong), 300 km long, 200 km thick blocks embedded in thinner (150 km) and younger (150 Ma) lithosphere. (left) Phase field and passive marker grid that allows to visualize zones of large and small deformation. (right) Strain rate field showing plug-like entrainment of the mantle lithosphere and strong deformation in the crustal parts.

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