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Slab detachment during continental collision: Influence of crustal rheology and interaction with lithospheric delamination

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ABSTRACT

Collision between continents can lead to the subduction of continental material. If the crust remains coupled to the downgoing slab, a large buoyancy force is generated. This force slows down convergence and promotes slab detachment. If the crust resists to subduction, it may decouple from the downgoing slab and be subjected to buoyant extrusion.

We employ two-dimensional thermo-mechanical modelling to study the importance of crustal rheology on the evolution of subduction–collision systems. We propose simple quantifications of the mechanical decoupling between lithospheric levels (σ^*) and the potential for buoyant extrusion of the crust (ξ^*). The modelling results indicate that a variable crustal rheological structure results in slab detachment, delamination, or the combination of both mechanisms.

A strong crust provides coupling at the Moho (low σ^*) and remains coherent during subduction (low ξ). It promotes deep subduction of the crust (180 km) and slab detachment. Exhumation occurs in coherent manners via eduction and thrusting. Slab detachment triggers the development of topography (>4.5 km) close to the suture. A contrasting style of collision occurs using a weak crustal rheology. Mechanical decoupling at the Moho (high σ^*) promotes the extrusion of the crust (high ξ), disabling slab detachment. Ongoing shortening leads to buckling of the crust and development of topography on the lower plate. Collisions involving rheologically layered crust allow decoupling at mid-crustal depths. This structure favours both the extrusion of upper crust and the subduction of the lower crust. Such collisions are successively affected by delamination and slab detachment. Topography develops together with the buoyant extrusion of crust onto the foreland and is further amplified by slab detachment.

Our results suggest that the occurrence of both delamination (Apennines) and slab detachment (Himalayas) in orogens may indicate differences in the initial crustal structure of subducting continental plates in these regions.

1. Introduction

The slab detachment (or slab breakoff) model has been widely employed in the interpretation of both geological and geophysical observation. This model involves the detachment of a portion or the integrity of a subducting slab beneath a convergent margin. The concept of slab detachment was born from seismological studies and was first hypothesised in the late sixties (Isacks and Molnar, 1969) in order to explain seismicity patterns in subduction zones. The study of deep seismicity patterns has indicated the existence of seismogenic zones that were associated to gaps within slabs (Chatelain et al., 1993; Chen and Brudzinski, 2001; Kundu and Gahalaut, 2011; Sperner et al., 2001). The slab detachment model has further gained popularity with the development of seismic tomography (Levin et al., 2002; Rogers et al., 2002; Schmandt and Humphreys, 2011; van der Meer et al., 2010; Widiyantoro and van der Hilst, 1996; Wortel and Spakman, 1992; Zor, 2008). Moreover, regional scale seismic tomography has enabled the detection of positive seismic velocity anomalies beneath collision zones, these structures were consequently attributed to detached or detaching slab (Lippitsch et al., 2003; Martin and Wenzel, 2006; Replumaz et al., 2010; Wortel and Spakman, 2000).

Slab detachment is likely to take place once large tensional stresses develop within the down-going slab. Slowdown of the subduction rate is a plausible mechanism that accounts for tensional stress build up as long as the slab pull force is constant (Li and Liao, 2002), such situation may take place following the subduction of an attempted ridge (Andrews and Billen, 2009; Burkett and Billen, 2011) or during the subduction of continental material (Baumann et al., 2009; Duretz et al., 2011; van Hunen and Allen, 2011). The slab detachment model is likely to be a fast geological process (Duretz et al., 2012b) leading to (1) a partial or complete loss of the slab pull force and (2) the inflow of hot asthenosphere at the location of the detachment. The loss of slab pull

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leads to force rebalancing in the orogen which can potentially triggers a wide range of dynamical effects. The slab detachment model was consequently used for the explanation of high pressure and ultra-high pressure rock exhumation (Andersen et al., 1991; Babist et al., 2006; Xu et al., 2010), variations in surface uplift rates (Morley and Back, 2008; Rogers et al., 2002; Wilmsen et al., 2009) and in the sedimentary record (Mugnier and Huyghe, 2006; Sinclair, 1997), orogenic extension (Zeck, 1996), rapid changes in plate motions (Austermann et al., 2011), or reversal of the subduction dip (Regard et al., 2008). The inflow of asthenosphere within a detaching slab is usually considered as an efficient mechanism to advect heat at lithospheric to sub crustal level (van de Zedde and Wortel, 2001) potentially leading to partial melting in the mantle (Altunkaynak and Can Genç, 2004; Davies and von Blanckenburg, 1995; Ferrari, 2004) in association to plutonism and volcanism (Ferrari, 2004; Keskin, 2003; Qin et al., 2008).

A number of analytical, analogue and numerical modelling studies have focussed on slab detachment (Andrews and Billen, 2009; Baumann et al., 2009; Buiter et al., 2002; Burkett and Billen, 2011; Cloetingh et al., 2004; Davies and von Blanckenburg, 1995; Duretz et al., 2011; Gerya et al., 2004; Li and Liao, 2002; Macera et al., 2008; Regard et al., 2008; Schmalholz, 2011; Ton and Wortel, 1997; Toussaint et al., 2004; van de Zedde and Wortel, 2001; van Hunen and Allen, 2011). These studies were designed to evaluate the depth slab detachment (Baumann et al., 2009; Duretz et al., 2011; Gerya et al., 2004; van de Zedde and Wortel, 2001), its duration (Andrews and Billen, 2009; Baumann et al., 2009; Duretz et al., 2012b; Gerya et al., 2004), and its topographic expression (Buiter et al., 2002; Duretz et al., 2011; Gerya et al., 2004), as well as the dynamic consequences of slab pull loss (Duretz et al., 2012a). Recently, three dimensional aspects of slab detachment have been addressed in numerical experiments. These simulations (Burkett and Billen, 2011; van Hunen and Allen, 2011) have highlighted the role of slab rheology and margin obliquity on the style of detachment (tear versus boudinage). The studies of Andrews and Billen (2009) and Schmalholz (2011) have further stressed the importance of a non-Newtonian rheology for slabs to undergo lithospheric-scale boudinage, ultimately leading to detachment.

Slab detachment in a collisional context is likely to take place within the subducting continental margin (Baumann et al., 2009; Gerya et al., 2004; van Hunen and Allen, 2011). It can thus be expected that crustal rheology, which controls the mechanical coupling at the Moho (Le Pourhiet et al., 2006; Nakada, 1994; Toussaint et al., 2004), plays an important role in the occurrence of slab detachment. Moreover, one may expect that changing the rheological structure of the crust may control the occurrence of other orogenic processes such as lithospheric delamination (Bird, 1979). Despite that fact, the role of crustal rheology on the occurrence of slab detachment has not yet been investigated. We therefore use two-dimensional thermo-mechanical modelling to study the influence of continental crust rheology on the occurrence of slab detachment, collision dynamics and topographic evolution of subduction/collision zones.

2. Modelling approach

2.1. Methodology

In order to simulate the dynamics of upper mantle scale processes coupled to topographic development, we employ the thermo-mechanical code I2VIS (Gerya, 2010; Gerya and Yuen, 2003a). This numerical code solves for the two-dimensional steady state momentum equations and heat conservation equation using the finite-difference/marker-in-cell method on a Eulerian grid (Gerya, 2010; Gerya and Yuen, 2003a):

$$\frac{\partial \sigma_{ij}}{\partial x_j} - \frac{\partial p}{\partial x_i} = -\rho g_i \tag{1}$$

$$\frac{\partial v_i}{\partial x_i} = 0 \tag{2}$$

$$\frac{\partial}{\partial x_i}\left(k\frac{\partial T}{\partial x_i}\right) = \rho C_p \frac{DT}{Dt} - H \tag{3}$$

where x_j represents the coordinates, ρ, the material density (kg/m^3), k, the thermal conductivity (W/m/K), C_p, the isobaric heat capacity (J/kg) and H (J/m^3/s), the contribution of internal heat sources (radiogenic, shear, and adiabatic heating).

The mechanical model employs a velocity formulation and the deviatoric stress tensor σ_{ij} relates to the material viscosity η_{eff} and the rate of deformation tensor $\dot{\epsilon}_{ij}$ via:

$$\sigma_{ij} = 2\eta_{\mathrm{eff}}\dot{\epsilon}_{ij} = \eta_{\mathrm{eff}}\left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i}\right) \tag{4}$$

All the lithologies deform according to a visco-plastic rheological model. This model combines the contribution of Newtonian and non-Newtoninan flow rules that control the effective viscosity of each material phase. A more detailed description of this rheological model is given in Section 1.

The model's surface h (air/crust interface) evolves following a gross-scale erosion–sedimentation law (Gerya, 2010; Gerya and Yuen, 2003b):

$$\frac{\partial h}{\partial t} = v_y - v_x \frac{\partial h}{\partial x} - \dot{e} + \dot{s} \tag{5}$$

where v_y and v_x are the uplift and advection velocity (m/s) predicted by the tectonic model, $\dot{e}$ and $\dot{s}$ represent prescribed erosion and sedimentation rates ($\dot{e} = 0.1$ mm/yr if $h > 1$ km and $\dot{s} = 0.1$ mm/yr if $h < -1$ km).

The Lagrangian advection equation is solved by an explicit coordinate update of the markers that carry the material properties through the Eulerian grid.

The density and heat capacity of each lithology are pressure P (Pa) and temperature T (K) dependent and are updated at each timestep. We assume a pyrolitic mantle composition and a basaltic–gabbroic oceanic crust. Their properties are pre-computed using Gibbs free energy minimisation with the chemical model CaO–FeO–MgO–Al_2O_3–SiO_2 (Baumann et al., 2009; Gerya et al., 2004). The density of felsic materials are obtained from the equation of state:

$$\rho = \rho_0\left(1 - \alpha(T - 298.15)\right)\left(1 + \beta\left(P \times 10^{-8} - 10^{-3}\right)\right), \tag{6}$$

where ρ_0 corresponds to the reference density (2700 kg/m^3 for the upper crust and 2800 kg/m^3 for the lower crust), β to the isothermal compressibility (0.5×10^{-3} kbar^{-1}) and α to the thermal expansivity (1.5×10^{-5} K^{-1}). The thermal conductivity k is a function of the temperature (Clauser and Huenges, 1995), the functions used to evaluate k are listed in Table 1. The presented simulations do not take into account the effects of partial melting and hydration/dehydration processes.

2.2. Setup

The model domain consists initially of two continental plates separated by an oceanic basin (Fig. 1). The dimensions of the model box is 4000×1400 km^2 (1361×351 nodes), all the mechanical boundaries are free slip. Variable grid spacing enables to reach a 1 km grid resolution in the central part of the domain where the continental collision takes place. In order to study the interplay between subduction and orogenic evolution, we follow the semi dynamic approach employed in Baumann et al. (2009) and Duretz et al. (2011). The initial conditions of the model are built during a stage of kinematic