Tectonophysics 502 (2011) 257–265

Contents lists available at ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

Deep and near-surface consequences of root removal by asymmetric continental delamination

J.L. Valera [a,*], A.M. Negredo [a], I. Jiménez-Munt [b]

[a] Dept. of Geophysics, Faculty of Physics, Universidad Complutense de Madrid, Av. Complutense s/n.28040 Madrid, Spain
[b] Instituto de Ciencias de la Tierra 'Jaume Almera', CSIC, C/Sole i Sabaris s/n. 08028 Barcelona, Spain

ARTICLE INFO

Article history:
Received 31 July 2009
Received in revised form 3 March 2010
Accepted 1 April 2010
Available online 10 April 2010

Keywords:
Continental delamination
Topography evolution
Thermo-mechanical modeling
Numerical algorithm

ABSTRACT

The occurrence of continental delamination has been proposed for a number of areas characterized by highly variable geodynamic settings. In this study we present results of numerical simulations considering different initial setups, representative for geodynamic scenarios where delamination could potentially develop. To mimic a post-collisional orogenic scenario we have designed an initial state characterized by the presence of an area of orogenic lithosphere, with both crustal and lithospheric roots. In a second setup, we have considered a lithospheric root representative of a remnant slab with a flat overlying crust. We focus on predicted evolution of surface and near-surface observables, namely the crustal structure, surface heat flow and isostatic and dynamic topography evolution. Our results show that a high density orogenic lower crust, likely related to the presence of eclogite, significantly accelerates the sinking of the lithospheric mantle. The pattern of local isostatic elevation is characterized by laterally migrating surface uplift/subsidence. This pattern is shown to be little sensitive to lower crust density variations. In contrast, predicted dynamic topography is more sensitive to these changes, and shows surface subsidence adjacent to the delaminating lithospheric mantle for the model with a high density lower crust, and surface uplift above the slab for a model with a less dense lower crust. The reason for uplift in this second model is that the effect of the positive buoyancy of the thickened crust overwhelms the effect of negative buoyancy of the slowly sinking lithospheric mantle. We infer from our modeling that there is not a specific characteristic pattern of topography changes associated with delamination, but it depends on the interplay between highly variable factors, as slab sinking velocity, asthenospheric upwelling and changes in crustal thickness.

1. Introduction

Removal of continental lithospheric mantle has been inferred from a wide range of observations. The most significant includes anomalously high heat flow, regional uplift, change of stress field toward extension, and the presence of cold slabs in the upper mantle and igneous activity in continental areas far from present subduction zones. The processes responsible for removal of continental lithospheric mantle are still under debate, but most of the related models presented during the last 30 years can be grouped into two categories: those based on viscous convective removal and those based on lithospheric delamination.

The convective removal mechanism is based on the fact that a thickened thermal boundary layer (mantle lithosphere) can develop a Rayleigh–Taylor gravitational instability and drip viscously into the less dense asthenosphere. This mechanism has been extensively investigated in a series of studies using dynamic approaches (e.g. Houseman et al., 1981; England and Houseman, 1989; see Houseman and Molnar, 2001 for a thorough revision) and in a number of studies adopting thermo-mechanical approaches (e.g. Fleitout and Froidevaux, 1982; Buck and Toksöz, 1983; Lenardic and Kaula, 1995; Marotta et al., 1998; Schott and Schmeling, 1998).

The continental delamination mechanism was introduced by Bird (1978, 1979), who proposed that if any process provided an elongated conduit connecting the underlying asthenosphere with the base of the continental crust, the dense lithospheric boundary layer could peel away from the crust and sink. Differently from convective removal, where the lithospheric root deforms internally as it drips, in the case of delamination the mantle part of the lithosphere peels away as a coherent slice, without necessarily undergoing major internal deformation, and is replaced by buoyant asthenosphere. To avoid ambiguity commonly found in the literature, where the term 'delamination' is often used to refer to any process causing removal of lithosphere, it is worth clarifying that in this study we will only use the term 'delamination' when two conditions of Bird's model are fulfilled: 1) the asthenosphere comes into direct contact with the crust and 2) the point of delamination, where the lithosphere peels off the overlying crust, migrates.

* Corresponding author.
E-mail addresses: jlvalera@fis.ucm.es (J.L. Valera), anegredo@fis.ucm.es (A.M. Negredo), ivone@ictja.csic.es (I. Jiménez-Munt).

0040-1951/$ – see front matter © 2010 Elsevier B.V. All rights reserved.
doi:10.1016/j.tecto.2010.04.002

Delamination has been proposed for a number of geological areas. Some examples are Tibet and Colorado plateaus (Bird, 1978 and 1979, respectively), western Mediterranean (Channel and Mareschal, 1989), Variscan belt (Arnold et al., 2001), Alboran sea (e.g. Seber et al., 1996; Calvert et al., 2000; Valera et al., 2008); Sierra Nevada mountains (Ducea and Saleeby, 1998; Zandt et al., 2004; Le Pourhiet et al., 2006), Vrancea region (Knapp et al., 2005) and eastern Anatolia (Göğüş and Pysklywec, 2008a). In spite of this large number of examples, the very few physical models that have been developed (e.g. Schott and Schmeling, 1998; Morency and Doin, 2004); make that basic aspects of the delamination process remain poorly studied.

Very recently, Göğüş and Pysklywec (2008b) presented a comparison between near-surface observables resulting from a model representative of delamination, and from two models of viscous dripping representative of partial and full mantle lithosphere removal. Differences in surface topography and *P*–*T*–*t* paths predicted by Göğüş and Pysklywec (2008b) with delamination and convective removal models reflect major differences in the style of crustal deformation and mantle lithosphere evolution resulting from both processes. These differences were also investigated in the study by Valera et al. (2008), who evaluated quantitatively conceptual models of delamination and convective removal proposed for the evolution of the Alboran Sea and surrounding regions.

In this study we present results of numerical simulations considering different initial setups, representative for different geodynamic scenarios likely prone to develop delamination. We focus on predicted evolution of surface and near-surface observables, namely the crustal structure, topographic response (both isostatic and dynamic) and surface heat flow.

It is worth noting that in some areas (e.g. Alboran sea and Pannonian basin) there is a strong controversial between authors proposing propagating continental delamination and those proposing migration of subduction caused by slab roll-back. In this sense, exploring the consequences of delamination on lithospheric and near-surface scales can also be helpful to discriminate between slab 'roll-back' and mantle lithosphere 'peel-back'.

2. Model description

2.1. Governing equations

The physical process of delamination is governed by the coupled conservation equations of mass, momentum and energy. We have applied several hypotheses to simplify these equations. First, we have assumed two-dimensional flow. Second, we have neglected inertial forces. This hypothesis can only be applied to very viscous flows, so with very high Prandtl number fluids. Third, we have applied the Extended Boussinesq Approximation (EBA). According to the Standard Boussinesq Approximation (Boussinesq, 1903), density variations may be neglected except when they are coupled to the gravitational acceleration in the buoyancy force term. The EBA differs from the Standard Boussinesq Approximation in that the thermal effect of compression is also accounted for (e.g., Tritton, 1988; Schmeling, 1989; Ita and King, 1994). Neglecting the inertial forces in a viscous flow under the EBA implies that the fluid is incompressible, which simplifies the equation of mass conservation.

The final equations are the same as those used by Valera et al. (2008), but we have neglected the shear heating and the effect of the phase transformation from olivine to high-pressure polymorphs. The reader is referred to Valera et al. (2008) for detailed explanations on the mathematical developments leading to equations:

$$\frac{\partial}{\partial x}\rho g = 4\frac{\partial^2}{\partial x \partial z}\left[u\frac{\partial^2}{\partial x \partial z}\Psi\right] + \left(\frac{\partial^2}{\partial z^2}-\frac{\partial^2}{\partial x^2}\right)\left[u\left(\frac{\partial^2}{\partial z^2}-\frac{\partial^2}{\partial x^2}\right)\Psi\right] \quad (1)$$

$$\frac{\partial T}{\partial t} + u_x\frac{\partial T}{\partial x} + u_z\frac{\partial T}{\partial z} = \frac{H}{C_P} + \frac{k}{\rho C_P}\left(\frac{\partial^2 T}{\partial x^2}+\frac{\partial^2 T}{\partial z^2}\right) + u_z\frac{\alpha g}{C_P}T \quad (2)$$

where u_k is the *k*-component of the velocity vector; *x* is the horizontal coordinate; *z* is the vertical coordinate, pointing downward; μ is the viscosity; ρ is the density; *g* is the acceleration of gravity; C_P, the specific heat; *T*, the temperature; *t* is the time; *H*, the radiogenic heat production; *k*, the thermal conductivity; and α, the thermal expansion coefficient. The velocity is related to the stream function Ψ as:

$$u_x = \frac{\partial \Psi}{\partial z}; \quad u_z = -\frac{\partial \Psi}{\partial x}. \quad (3)$$

The heat sources considered here are the terms on the right-hand side of Eq. (2) and correspond to: the radiogenic heat production, the heat conduction and the adiabatic heating. The values of the parameters used are listed in Table 1.

Our modeled domain includes five different materials: upper crust, non-perturbed lower crust, orogenic lower crust, lithospheric mantle and asthenosphere (with asthenosphere we refer to the whole sub-lithospheric upper mantle). In some models we have introduced a sixth material on the top of the model, that will be described later on. For simplicity, density and viscosity have constant values in the upper and lower crust (see Table 1). The boundary between the lithospheric mantle and the asthenosphere is assumed to be a thermal boundary, with no compositional difference. Density and viscosity are assumed to be temperature dependent in the lithospheric mantle and asthenosphere. We have used a Newtonian temperature-dependent (exponential) viscosity law and augmented it with a pressure dependence that crudely simulates an increase in 'deeper mantle' viscosity beneath 450 km (Rüpke et al., 2004):

$$\mu(T,z) = \mu_0\mu(z)\exp\left[b\left(\frac{T_0}{T}-1\right)\right] \quad (4)$$

Table 1
Parameters used in all calculations.

Symbol	Meaning	Value
g	Acceleration of gravity	9.8 m s^{-2}
Q_b	Basal heat flow	0.014 W m^{-2}
b	*b*-parameter of Rüpke Law	15
H_p	Crustal radiogenic heat production	8×10^{-7} W m^{-3}
	Horizontal extent	1376 km
	Vertical extent	680 km
C_P	Specific heat	1.3×10^{3} J K^{-1} kg^{-1}
T_0	Temperature at the base of the lithosphere	1350 °C
k	Thermal conductivity	3.2 W m^{-1} K^{-1}
α	Thermal expansion coefficient	3.7×10^{-5} K^{-1}
	Time step	0.25 Ma
L	Lithospheric thickness	120 km
	Lower bound for the viscosity	10^{17} Pa s
	Upper bound for the viscosity	10^{22} Pa s
w_{max_L}	Maximum amplitude of the Lithospheric Mantle perturbation	121.8 km
$w_{max\_ LC}$	Maximum amplitude of the Lower Crust perturbation	33 km
$w_{max\_ UC}$	Maximum amplitude of the Upper Crust perturbation	15 km
λ	Wavelength of the perturbation	487.2 km
x_{pert}	Horizontal position of the center of the perturbation	688 km
$h_{\_UC}$	Upper Crust thickness	15 km
ρ_{UC}	Upper Crust density	2800 kg m^{-3}
μ_{UC}	Upper Crust viscosity	10^{20} Pa s
$h_{\_LC}$	Lower Crust thickness	22 km
ρ_{npLC}	Lower Crust density (non-perturbed zone)	2900 kg m^{-3}
μ_{npLC}	Lower Crust viscosity (non-perturbed zone)	10^{20} Pa s
ρ_{LC}	Orogenic Lower Crust density (perturbed zone)	Variable
μ_{LC}	Orogenic Lower Crust viscosity (perturbed zone)	10^{20} Pa s