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Crustal eclogitization and lithosphere delamination in orogens

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ABSTRACT

Many orogens, including the Central Andes and Himalayas, are characterized by wide areas that have undergone upper crustal shortening and surface uplift. The behaviour of the deep lithosphere is poorly constrained, and in some mountain belts, lower crust and mantle lithosphere appear to have been removed through delamination during orogen development. Thermal–mechanical numerical models demonstrate that as crust thickens during shortening, the lowermost crust may undergo metamorphic eclogitization, which increases its density. Even a small density increase (7% or more) causes shortening to localize above the eclogitic crustal root, promoting the development of thick lithosphere in this area which is then prone to gravitational removal. Complete removal of orogen mantle lithosphere occurs if the eclogitized lower crust is weak enough to allow full detachment of negatively buoyant mantle lithosphere; this can occur even if the lower crust is less dense than the mantle. The onset of delamination may be determined by the hydration state of the lower crust, as the presence of water promotes eclogitization and significantly reduces rock strength. Two distinct styles of delamination are observed: (1) retreating delamination in which weak mantle lithosphere rolls back and peels away from the crust, producing a contemporaneous migration of crustal thickening, surface uplift and magmatism, and (2) stationary delamination in which strong lithosphere separates from the weak lower crust and slides into the deep mantle at a stationary detachment point, followed by widespread crustal deformation and magmatism.

1. Introduction

During orogenesis, shortening produces thick (>50 km) continental crust and corresponding high elevations that characterize orogens such as the Central Andes and Himalayas. The behaviour of the lower crust and mantle lithosphere remains poorly constrained. Shortening of the upper crust should be accompanied by thickening of the deeper lithosphere. However for many orogens, geophysical and geological observations indicate that mantle lithosphere is anomalously thin or absent.

Seismic studies of the Central Andes show low velocities and high attenuation in the shallow mantle of the eastern Altiplano Plateau (Beck and Zandt, 2002) and throughout the Puna Plateau (Schurr et al., 2006). These data imply high temperatures and the absence of a lithospheric lid for much of the orogen. Higher mantle velocities are observed in some areas, such as the central Altiplano, indicating regions of locally thicker lithosphere (Beck and Zandt, 2002). Seismic data also show that the crust is 50–70 km thick, with velocities that are consistent with a felsic composition, suggesting that the mafic lower crust has been removed (Beck and Zandt, 2002). Seismic data reveal present-day lithosphere structure but does not constrain the timing of removal. There is mounting evidence to indicate that lithosphere loss occurred during Cenozoic orogen shortening, including observations of rapid surface uplift, voluminous magmatism, and a change in crustal stress from compression to extension (e.g., Kay and Kay, 1993; Garzione et al., 2008; Kay and Coira, 2009). Similar observations have been used to infer lithosphere removal in other mountain belts, such as the Tibetan Plateau (e.g., England and Houseman, 1989; Jimenez-Munt et al., 2008), East Anatolian Plateau (e.g., Keskin, 2003; Sengor et al., 2003), Eastern Carpathians (e.g., Knapp et al., 2005; Fillerup et al., 2010), Sierra–Nevada region (e.g., Ducea and Saleeby, 1998), and Varsican orogen (e.g., Gutierrez-Alonso et al., 2011), suggesting that this is an important process in orogen evolution.

Mantle lithosphere, being cooler and therefore denser than the underlying material, is gravitationally unstable and susceptible to removal. One mechanism is convective removal via Rayleigh–Taylor-type (RT) instability ("drip") (Houseman et al., 1981), possibly initiated by shortening-induced perturbation of the lithosphere. However, most studies conclude that removal is generally limited to the lowermost lithosphere, owing to the strong temperature-dependence of typical mantle rheologies (e.g., Buck and Toksoz, 1983; Conrad and Molnar, 1999). An alternate

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removal mechanism is delamination, which involves peeling of mantle lithosphere along the Moho (Bird, 1979), and by definition, removes the entire mantle lithosphere. Delamination requires a weak zone in the deep crust (Meissner and Mooney, 1998), which may be related to high Moho temperatures (e.g., Morency and Doin, 2004) or rheological weakness associated with hydration (e.g., Schott and Schmeling, 1998).

In many modelling studies, delamination is induced using a prescribed weak layer (e.g., Schott and Schmeling, 1998, Gogus and Pysklywec, 2008a, 2008b; Valera et al., 2011; Bajolet et al., 2012). These studies provide insights into the consequences of lithosphere removal on surface topography, crustal deformation, and thermal structure but do not address the factors leading to delamination. Numerical models of Morency and Doin (2004) demonstrate that delamination can develop dynamically through localized lithosphere thinning associated with mantle convection. Alternatively, models of continental collision show that delamination may be induced by retreat of high density subducting lithosphere (Beaumont et al., 2006; Gray and Pysklywec, 2012), where decoupling is enhanced if the plate boundary is rheologically weak (Faccenda et al., 2009; Ueda et al., 2012).

In this study, we consider the role of lower crustal eclogitization in triggering delamination. Prograde metamorphism of the deep crust as it shortens and thickens may produce an eclogitized crustal root. The density of the root depends on its composition, extent of eclogitization, and local pressure and temperature conditions. For a mafic composition, the eclogitized root can be denser than the underlying mantle and thus it is prone to foundering (e.g., Kay and Kay, 1993). Numerical models confirm that high density eclogite decreases RT instability times (Jull and Kelemen, 2001), affects orogen evolution if the crust is decoupled from the mantle (Doin and Henry, 2001) and can lead to delamination (e.g., Sobolev and Babeyko, 2005; Le Pourhiet et al., 2006).

Here, we use thermal–mechanical numerical models to assess the effect of eclogitization of the lower crust during orogen shortening. We systematically examine variations in eclogite density and strength of the crust and mantle lithosphere to understand the conditions required for delamination. In contrast to some previous studies, models do not include an imposed lithosphere thickness perturbation or prescribed eclogite layer. Instead, eclogitization and gravitational instability develop self-consistently during model evolution. Our study focuses on metamorphic eclogitization of thickened crust as the orogen shortens. Eclogitized crust may also be produced through igneous processes, whereby magmatic differentiation leaves a mafic residue in the deep crust (e.g., Kay and Kay, 1993; Jull and Kelemen, 2001; Kay and Coira, 2009). In this case, there may be no relationship between orogen shortening and eclogite formation.

2. Numerical modelling approach

2.1. Methods and model geometry

Two-dimensional numerical models are used to study the behaviour of continental lithosphere during shortening. The coupled thermal–mechanical evolution of the lithosphere–upper mantle system is modelled using the SOPALE code, which has been widely applied to studies of lithosphere dynamics (e.g., Pysklywec and Beaumont, 2004; Beaumont et al., 2006; Gogus and Pysklywec, 2008a; 2008b; Warren et al., 2008; Currie and Beaumont, 2011; Gray and Pysklywec, 2012). Arbitrary Eulerian–Lagrangian finite element techniques are used to solve the equations of mass conservation, force balance and energy balance, under the assumptions of plain strain, incompressibility and zero Reynolds number. Mechanical and thermal calculations are performed on an Eulerian mesh that stretches vertically to conform to the top model boundary, which is a free surface. Material properties are tracked on a Lagrangian mesh, with additional Lagrangian tracer particles which are advected with the model velocity field. Full details of the numerical method and model parameterization are given by Fullsack (1995) and Beaumont et al. (2006).

The model domain has a width of 1200 km and extends from the Earth's surface to 660 km depth (Fig. 1a). The 100 km thick continental plate consists of 25 km upper-mid crust, 15 km lower crust, and 60 km mantle lithosphere. The crust is divided into an 850 km wide weak region (orogen) with strong blocks on either side. This geometry replicates the orogenic vise model, in which shortening occurs within a zone of inherited weakness (Ellis et al., 1998). For example, shortening in the Central Andes is confined to accreted and rifted terranes that are bounded by strong forearc on the west and Brazilian craton on the east (e.g., Oncken et al. 2006). Similarly, the wide area of Himalayan deformation has been related to pre-existing weakness associated with high temperatures (Molnar and Tapponnier, 1981). The Eulerian mesh has 120 elements horizontally (10 km width) and 108 elements vertically, with 56 elements in the upper 140 km (2.5 km height) and 52 elements below (10 km height). Models with a higher resolution mesh give comparable results (see Supplementary material).

Shortening is produced by introducing strong continental lithosphere through the right-hand model boundary at a rate of 1 cm/yr, comparable to the Cenozoic shortening rate of the Central Andes (e.g., Oncken et al., 2006). Shortening is inferred to be related to plate boundary forces originating to the right of the model domain (c.f., Doin and Henry, 2001; Beaumont et al., 2006). Within the domain, materials evolve dynamically in response to boundary conditions and internal buoyancy forces.

2.2. Material properties and model boundary conditions

All model materials have a viscous-plastic rheology. Frictional-plastic deformation follows a Drucker–Prager yield criterion:

$$(J_2')^{1/2} = c_0 \cos \phi_{eff} + P \sin \phi_{eff} \tag{1}$$

where J'_2 is the second invariant of the deviatoric stress tensor σ'_{ij} ($2J'_2 = \sigma'_{ij}\sigma'_{ij}$), c_0 is cohesion, ϕ_{eff} is the effective internal angle of friction, and P is the pressure. Frictional-plastic deformation is modelled using an effective viscosity that places the state of stress on yield (Fullsack, 1995; Willett, 1999). Materials undergo frictional-plastic strain softening where ϕ_{eff} decreases from 15° to 2° over accumulated strain (I'_2) of 0.5–1.5 (e.g., Warren et al., 2008 and references therein).

At stresses below frictional-plastic yield, materials deform viscously in power law creep:

$$\eta_{eff} = f(B^*)(\dot{I}_2')^{(1-n)/2n} \exp\left(\frac{Q+PV^*}{nRT_K}\right) \tag{2}$$

where η_{eff} is effective viscosity, f is a scaling factor (see below), $\dot{I}_2'$ is the second invariant of the strain rate tensor $\dot{\varepsilon}_{ij}$ ($2\dot{I}_2' = \dot{\varepsilon}_{ij}\dot{\varepsilon}_{ij}$), R is the universal gas constant, and T_K is the absolute temperature. The rheological parameters, B^*, n, Q, and V^*, are the pre-exponential factor, the stress exponent, the activation energy and the activation volume, respectively.

Rheological parameters are given in Table 1. The laboratory-derived viscous flow laws of wet quartzite (WQ, Gleason and Tullis, 1995), dry Maryland diabase (DMD, Mackwell et al., 1998), and wet olivine (WO, Karato and Wu, 1993) are used for the upper-mid crust, lower crust and mantle, respectively. Following previous studies (e.g., Beaumont et al., 2006), a scaling factor (f) is