Thermo-mechanical modeling of subduction of continental lithosphere

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Abstract
We consider two-dimensional thermo-chemical mantle convection models to investigate the deformation of the continental lithosphere that follows the oceanic lithosphere into the subduction zone. The models account for the compositional buoyancy forces by considering lithospheric plates with distinct crustal layers. Continental convergence results in crust-mantle detachment in the subducting plate and crustal thickening in both subducting and overriding plates. The depth of detachment ranges from 90 to 200 km, depending on the strength of the lithosphere and the density of the continental crust. As the crust thickens, the convergence velocity in the collision zone decreases and the locus of subduction gradually shifts toward the interior of the subducting plate. In models with greater viscosity, the subducting mantle lithosphere maintains its integrity and does not break up. In models with weaker rheology, the positive buoyancy of the thickened crust can overcome the strength of the subducting lithosphere, and causes the oceanic slab to break off and sink into the mantle. The breakoff occurs over a time interval of 10–20 million years, which is roughly the time needed for the crust to reach its maximum thickness. Crustal detachment takes place over a wide range of lithospheric strength, suggesting that crustal buoyancy has an important role in the dynamics of continental subduction.

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1. Introduction
The evolution of continental lithosphere in convergence zones has been the subject of extensive research in the past three decades. Owing to their different density structures, the deformation of the continental and oceanic lithosphere differ from one another in many ways. The oceanic lithosphere actively participates in mantle convection and is completely recycled into the mantle. The subduction of continental lithosphere takes on a much more complex pattern; while the lower part of the denser mantle lithosphere may continue to subduct to great depths, the upper mantle lithosphere and the crust often detach and accumulate on top of subduction zones, giving rise to zones of thickened crust and elevated topography. The thickened crust and mantle lithosphere exhibit a complex pattern of behavior that characterizes the evolution of orogenic belts. Despite the difficulty of subduction, there is considerable geological and experimental evidence suggesting that at least part of the continental...
crust subducts to some depth in order to account for the observed shortening in many mountain belts. Molnar and Gray (1979) argued that the negative buoyancy of the denser mantle part of the subducting continent and the pulling force of subducted oceanic plate contribute to the subduction of as much as several hundred kilometers of continental crust. The analogue modeling of Chemenda et al. (1995) has also demonstrated that the continental crust can subduct to a depth of 250 km, before breakup in the structure of the subducting slab can take place. According to Froidevaux and Ricard (1987), orogenic belts exhibit three distinct phases of deformation; crustal shortening under the driving forces of convergence, resistance of the thickened crust to further shortening and areal expansion of the thickened region, and the collapse of the elevated crust as a result of subsequent decrease in the magnitude of the tectonic forces. During the last two phases of this deformation history, the mantle part of the lithosphere plays an important role in the dynamics of continental deformation. Many authors have emphasized the role of mantle lithosphere in the development of tectonic regimes inside the thickened crust and in the growth and removal of the lithospheric roots in orogenic belts (e.g. Bird, 1978; Davies and von Blanckenburg, 1995; Houseman and Molnar, 1997 and Schott and Schmeling, 1998). In many active mountain belts there exists a rapid transition from compressional shortening in the initial stages of orogeny, to extensional thinning during the later stages that is often accompanied by topographic uplift and magmatic activity (e.g. Bird, 1978 and Molnar et al., 1993). These observations have led to the suggestion that rapid thermal instability of the thickened mantle lithosphere and replacement by hot asthenospheric material are responsible for the late stage evolution in collision zones. Several different scenarios have been proposed to explain the mechanism of such mechanical instabilities. They include the delamination model of Bird (1978) that describes the removal of the lithospheric root by the peeling off of the mantle lithosphere from the crustal layer, the convective thinning models, first proposed by Houseman et al. (1981), which are based on the Rayleigh–Taylor thermal instability of the lithospheric root, and slab breakoff concept by Davies and von Blanckenburg (1995) that explains the separation of the subducted oceanic plate from the continental plate as a result of the opposing buoyancies of the two plates.

The continental deformation in subduction zones has been studied by various approaches that cover a wide range of geodynamic problems and different modeling techniques, varying from simple buoyancy and rheological arguments (e.g. Cloos, 1995; van den Beukel, 1992 and Davies and von Blanckenburg, 1995) to thermo-chemical mantle convection models (e.g. Lenardic and Kaula, 1996 and Schott et al., 2000). Most of the numerical studies have either focused on the crustal-scale aspects of the deformation (e.g. Willett et al., 1993 and Beaumont and Quinlan, 1994), or investigated the deformation of the entire lithosphere without considering the interaction of the lithosphere with the convective flow field in the underlyling mantle (e.g. England and McKenzie, 1982 and Bird, 1989). Furthermore, most have employed kinematic boundary conditions which fail to simulate the time-dependence of deformation near the surface.

Continental subduction and/or collision is a complex dynamic process that should be viewed as the surface manifestation of a thermo-chemical mantle convection involving lithospheric thermal boundary layers as well as chemical boundary layers in the form of crustal layers (Lenardic and Kaula, 1996). The driving force of convection comes from the negative buoyancy of the denser mantle lithosphere, whereas the positive buoyancy of the lighter crustal layer acts as the opposing force. With this “self-consistent” arrangement, there is no need for prescribed kinematic boundary conditions, because the interior dynamics of convection fully determine the velocity of the surface plates.

In this paper we examine a set of thermo-chemical convection models to study large-scale features of the deformation of a continental lithosphere that follows an oceanic lithosphere into a subduction zone. The aim is to understand the role of buoyancy in the subduction of continental lithosphere, the dynamics of crustal thickening in the subduction zone, and the fate of the subducted part of the lithosphere and the thickened lithospheric root. We try to keep the number of approximations made to the geodynamic properties of the convection as few as possible. The lithospheric plates are defined by their thermal structure and their distinct crustal layers. We use temperature-, depth-, and stress-dependent viscosity in order to achieve realistic lithospheric strength and plate velocities. We also calculate models with kinematic surface boundary conditions in order to demonstrate the effect of
prescribed plate velocity on the dynamics of deformation.

2. Governing equations and model set-up

The convection in a chemically heterogeneous viscous fluid is described by conservation equations and the chemical advective transport equation. We adopt a two-dimensional Cartesian Coordinate system and the Boussinesq approximation. The non-dimensional form of the governing equations are:

Conservation of mass
\[ \nabla \cdot \mathbf{v} = 0 \]  

Conservation of momentum
\[ \nabla \cdot \mathbf{\tau} - \nabla P = (Ra T - Rb C) \mathbf{\hat{z}} \]  

Conservation of energy
\[ \frac{\partial T}{\partial t} + \mathbf{v} \cdot \nabla T = \nabla^2 T \]  

and advection of chemical composition
\[ \frac{\partial C}{\partial t} + \mathbf{v} \cdot \nabla C = 0 \]  

where \( \mathbf{v} \) is the velocity vector, \( \mathbf{\tau} \) the deviatoric stress tensor, \( P \) the dynamic pressure, \( T \) the temperature, \( C \) the composition function, \( t \) the time, and \( \mathbf{\hat{z}} \) the unit vector in the direction of gravity. \( Ra \) and \( Rb \) are the thermal and compositional Rayleigh numbers, respectively:

\[ Ra = \frac{\alpha \rho g d^3 \Delta T}{\kappa \eta}, \quad Rb = \frac{\Delta \rho g d^3}{\kappa \eta} \]  

where \( \alpha \) is the thermal expansion coefficient, \( g \) the gravitational acceleration, \( \rho \) the reference density, \( \Delta \rho \) the density difference between the two chemical components, \( d \) the depth of the model, \( \Delta T \) the temperature change from top to bottom, \( \eta \) the reference viscosity and \( \kappa \) the thermal diffusivity. Table 1 lists the values of model parameters. We have neglected the chemical diffusion term in Eq. (4), since the rate of transport of chemical species by diffusion is far smaller than that by the convective flow. The continental and oceanic crustal layers are each assigned a composition function \( C \) that varies from 1 for that layer to 0 for mantle composition.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( d )</td>
<td>Depth (km)</td>
<td>1400</td>
</tr>
<tr>
<td>( \Delta T )</td>
<td>Temperature difference (K)</td>
<td>1700</td>
</tr>
<tr>
<td>( \rho )</td>
<td>Reference mantle density (kg/m(^3))</td>
<td>3300</td>
</tr>
<tr>
<td>( \rho_o )</td>
<td>Oceanic crust density (kg/m(^3))</td>
<td>2900</td>
</tr>
<tr>
<td>( \rho_c )</td>
<td>Continental crust density (kg/m(^3))</td>
<td>2750</td>
</tr>
<tr>
<td>( \eta )</td>
<td>Reference viscosity (Pa\ s)</td>
<td>( 10^{21} )</td>
</tr>
<tr>
<td>( \kappa )</td>
<td>Thermal diffusivity (m(^2)/s)</td>
<td>( 10^{-6} )</td>
</tr>
<tr>
<td>( \alpha )</td>
<td>Thermal expansion coefficient (/K)</td>
<td>( 3 \times 10^{-5} )</td>
</tr>
<tr>
<td>( Ra )</td>
<td>Thermal Rayleigh number</td>
<td>( 5 \times 10^6 )</td>
</tr>
<tr>
<td>( Rb )</td>
<td>Compositional Rayleigh number for oceanic crust</td>
<td>( 1.1 \times 10^7 )</td>
</tr>
<tr>
<td>( Rb_c )</td>
<td>Compositional Rayleigh number for continental crust</td>
<td>( 1.5 \times 10^7 )</td>
</tr>
<tr>
<td>( n )</td>
<td>Power-law exponent of viscosity</td>
<td>3</td>
</tr>
<tr>
<td>( \eta_t )</td>
<td>Transition stress (MPa)</td>
<td>2</td>
</tr>
</tbody>
</table>

2.1. Initial and boundary conditions

We assume a model box with dimensions of 5600 km \( \times \) 1400 km, extending to the middle of the mantle (Fig. 1), and consisting of oceanic and continental plates with their respective thermal structure and crustal layers. The restriction on the dimensions of the model is mostly due to computational limitations. A model extending to the base of the mantle was examined to assess the effect of the excess weight of the deep subducted slab on the near-surface dynamics. It was found that the effect is not significant because of a larger opposing viscous drag of the high-viscosity lower mantle. An overriding and a subducting plate are defined by their geotherms, initially determined on the basis of half-space cooling thermal gradients. The overriding plate on the left is assumed to be continental, with a 35 km thick crust on top. The subducting plate on the right consists of an oceanic and a continental segment, and a small spreading oceanic segment at the far right. The initial thicknesses of the oceanic and continental crusts on the subducting plate are 7 and 35 km, respectively. The spreading oceanic segment is the only part that has laterally varying initial temperature profile. For the mature segment of the oceanic lithosphere a nominal thermal age of 80 million years is
Fig. 1. Initial model set-up. The temperature of the lithospheric plates is defined by error function cooling profiles. The oceanic ridge on the right and the subduction fault zone are defined by their lower viscosity values. The thermal anomaly beneath the fault zone initiates the flow field in the models.

considered. We use an initial geotherm equivalent to a thermal age of 150 million years for the overriding continent. The initial geotherm of the subducting continent varies from 80 to 150 million years in different models. An initial cold thermal anomaly is placed underneath the contact zone of the two plates to ensure that subduction initiates at that location.

The boundary conditions are described in terms of the dependent variables and their gradients. At the surface and bottom of the model the temperature is fixed at constant values of $T_c$ and $T_h$, respectively, and the boundaries are treated as free-slip. In a set of models (called kinematic models here) a prescribed horizontal velocity of 5 cm per year is imposed on the surface of the subducting plate. The vertical side boundaries are treated as free-slip and non-penetrative with zero heat flux across them. To avoid complications associated with the initial transient stages, a chemically homogeneous model is first run for a couple of convective overturns to achieve a near-steady initial flow field.

2.2. Viscosity and density structures

Numerical investigations of convection processes (e.g. King et al., 1992; van den Berg et al., 1993; Zhong and Gurnis, 1995 and Moresi and Solomatov, 1998) have demonstrated that the deformation of the lithospheric plates and the morphology of the subducting slabs and trench zones are strongly controlled by rheology. Any self-consistent dynamic model of convection requires a “plate-generation” mechanism to achieve a realistic lithospheric behavior. Lithospheric plates move with a more or less uniform velocity and most of the deformation is concentrated near the edges of the plates. This deformation involves brittle failure and faulting at plate margins. Moresi and Solomatov (1998) have demonstrated the role of brittle deformation in the tectonic styles of convection, and Zhong and Gurnis (1995) have shown that the inclusion of mobile faults in subduction models has a strong effect on the concentration of strain at plate margins. Furthermore, the stress dependence of viscosity produces weak zones in the regions of high stresses such as subduction zones, thus localizing the deformation to narrow zones near plate boundaries. Following Han and Gurnis (1999), we use a composite rheology with temperature-, depth-, composition-, and stress-dependent viscosity. The upper 400 km of the model has non-Newtonian rheology, and Newtonian rheology is assumed below 400 km depth. This implies that deformation in the upper 400 km is dominated by dislocation creep, whereas, in the lower part diffusion creep prevails (Zhong and Gurnis, 1995). The temperature-dependent Newtonian viscosity is (Han and Gurnis, 1999).
nominal model are chosen to achieve \( 10^4 \) variation of stationary with respect to the interior of the model. For assuming stronger overriding plate is to keep it els (e.g. Forte and Mitrovica, 2001). The main reason broadly consistent with estimates from seismic mod-

\[ \eta = \eta_0 \exp \left( \frac{C_1}{1 + C_2} - \frac{C_1}{1 + \epsilon/\tau} \right) \quad (6) \]

where \( \eta_0 \) is a pre-exponential constant viscosity, and \( C_1 \) and \( C_2 \) are constants that determine the degree of temperature dependence of the viscosity. The effective non-Newtonian viscosity is:

\[ \eta_{eff} = \frac{\eta}{1 + (2q_1/x_0)^{1/n}} \quad (7) \]

where \( q_1 \) is calculated from Eq. (6), \( \eta_{eff} \) is a transition stress that determines the level of stress at which Newtonian and non-Newtonian viscosities have equal contributions to the effective viscosity, \( \epsilon \) is the second invariant of the strain rate tensor, and \( n \) is the power law exponent. The model is divided into several subdo-
mains, each governed by a different set of viscosity parameters (Table 2). The viscosity parameters in the model are listed in Table 2.

<table>
<thead>
<tr>
<th>Region</th>
<th>( \eta_0 )</th>
<th>( C_1 )</th>
<th>( C_2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oceanic lithosphere</td>
<td>1.177</td>
<td>1.177</td>
<td>1.177</td>
</tr>
<tr>
<td>Continental lithosphere</td>
<td>1.177</td>
<td>1.177</td>
<td>1.177</td>
</tr>
<tr>
<td>410–670 upper mantle</td>
<td>3.0595</td>
<td>0.75647</td>
<td>0.75647</td>
</tr>
<tr>
<td>Lower mantle</td>
<td>3.0595</td>
<td>0.75647</td>
<td>0.75647</td>
</tr>
</tbody>
</table>

This choice is also justified by the fact that continental lithosphere has less volatile content than oceanic litho-

dure its strength with respect to the oceanic plate. For the duration of oceanic subduction a narrow zone of weak viscosity, 2–3 orders of magnitude lower than that of the normal lithosphere, is placed at the contact zone of the two plates to approximate the role of subduction fault zone. This weak zone is removed after the two continents collide. The zone has a width comparable to the thickness of the lithosphere, and extends to the base of the thermal lithosphere. A sec-

ded weak zone is placed at the spreading center in the far right of the model to facilitate decoupling of the oceanic plate from the right boundary. The two weak zones are kept fixed in their places. In addi-
tion to the weak zone, the low-viscosity crustal layers act as tectonic lubricants and facilitate the subduction process.

2.3. Numerical procedure

We have adopted an iterative control volume finite difference method (Patankar, 1980) to solve Eqs. (1)–(3). A second-order central difference scheme is used for the discretization of the velocity field, and the temperature function is discretized by a hybrid method consisting of central difference and weighted upwind scheme. The energy equation is integrated forward-in-time by the fourth-order Runge–Kutta method. A pressure equation based on the SIM-
PLER scheme of Parakash and Patankar (1985) is used to solve for the pressure field. Eq. (4) is solved by an improved second-moment method (Ghods et al., 2003) which conserves the mass and the first and second moments of material distribution inside a control volume. The method is conservative, explicit forward-in-time, and quasi-Lagrangian in the sense that the material distribution in each control volume is advected along the streamlines over a time step. It minimizes numerical diffusion without producing appreciable numerical oscillations. The governing equations are solved on a grid of \( 201 \times 201 \) nodes with uniform horizontal and vertical spacing of 28 and 7 km, respectively. The horizontal resolution of the grid is increased by a factor of two when solving for the advective transport equation.
3. Kinematic models

We first briefly examine the role of a prescribed velocity boundary condition in order to see if such a boundary condition can correctly predict the observed deformation of the continental crust in convergence zones. Only temperature-dependent viscosity is considered here. Fig. 2 shows the temperature and flow fields, and the continental crust of four kinematic models with $10^3$, 100, 10 and 1 times viscosity variations, $\lambda$, as temperature varies from 0 to 1. In all models, the imposed surface velocity has a dominant role in organizing the flow field under the subducting plate, and maintaining the continuity of the downflow at the subduction zone. The continental crust evolves in different ways as the viscosity is varied. In models with greater viscosity variations ($\lambda > 100$), due to greater degree of crust–mantle coupling, the continental crust subducts deep into the mantle similar to an oceanic crust. There is no indication of crustal thickening and plateau formation, or any significant disturbance of the structure of the subduction. As the temperature dependence of viscosity is weakened, the positive buoyancy force of the crust overcomes the viscous force that couples the crust and mantle lithosphere, and the crust separates from the mantle part. The detached crust accumulates over the subduction zone. This behavior is similar to that of the oceanic crust in the 670 km transition zone. Gaherty and Hager (1994) and van Keken et al. (1996) have shown that detachment of the oceanic crust of a sinking slab in the transition depth is only possible if the crust is weakly coupled to the mantle part of the slab. The crust detaches from the slab at depths as shallow as 60 km, and is gradually swept towards the overriding plate on the left. There is no significant crustal thickening in the subducting plate. This is a direct consequence of the fixed boundary velocity. The lack of crustal thickening in the subducting plate results in the position of the contact zone of the two plates remaining fixed in time. This feature is inconsistent with observations, since in most collision zones new fault zones are created as a result of formation of crustal overthrust sheets (Chemenda et al., 1995). Crustal thickening is always accompanied by reduction in convergence velocity near the old contact zone and initiation of a new plate boundary on the upstream side of the subducting plate.

The kinematic models demonstrate that the depth of crustal breakup is dependent on the strength of the lithosphere. In this regard, these results are consistent with other investigations (e.g. van den Beukel, 1992 and Davies and von Blanckenburg, 1995) that predict breakup depths of as low as 30 km. Our results suggest that for a realistic lithospheric strength, breakup is improbable due to the influence of the imposed surface velocity on the near-surface flow field. Furthermore, in the case of crustal detachment, the style of crustal thickening is in contradiction with tectonic observations. Therefore, we conclude that kinematic plate models are not suitable for modeling the continental subduction.

4. Subduction of the continental lithosphere

We now describe the deformation of a continental plate in a self-consistent convection model and discuss the factors controlling this deformation. In the following models, 100 million years of convergence is studied, and the resulting surface velocity is of the order of 5 cm per year.

4.1. Evolution of the continental crust

Fig. 3 shows the evolution of the nominal model. The initial stages correspond to the oceanic subduction and the two continents collide after about 25 million years. The dip and thermal structure of the subduction zone remain unchanged during the oceanic subduction, and the oceanic crust subducts with little resistance. With the arrival of the continental crust at the subduction zone, pronounced changes take place in the dynamics of subduction. The continental crust is initially driven down to a depth of about 170 km (plot at $t = 40$ million years). However, the buoyancy of the crust inhibits further subduction and the continental layer detaches from the rest of the down going lithosphere, forming a block-like structure at the top of the subduction zone. In this and subsequent models there is always a minor entainment of the crustal material of the overriding side into the subduction zone. The entailed material eventually returns towards the surface under the influence of crustal buoyancy and is added, along with the material of the subducting side, to the region of thickened crust. Crustal thickening leads to
Fig. 2. The temperature (gray contours), flow field (dark contours) and continental crust (shaded areas) of four different kinematic models at 50 and 100 million years. For the flow lines, the broken lines denote counter-clockwise flow, and solid lines denote clockwise flow. Only the region $0.5 \leq x \leq 2.5$ is shown.
Fig. 3. The temperature, flow field and continental crust of the nominal model at 10 million years time intervals. Figure details are the same as in Fig. 2. Only the upper half of the model and the region $1 \leq x \leq 2$ is shown.
the downward deflection of the isotherms both in the crust and the mantle lithosphere. Consequently, a local zone of thermal boundary layer thickening develops. Within this zone, the depth to the base of the thermal lithosphere (corresponding to non-dimensional temperature of $T = 0.75$) increases by about a factor of two. The accumulation of the continental crust induces significant changes in the near-surface flow field. The convergence velocity near the subduction zone decreases as crustal thickening propagates towards the upstream side (Fig. 4). This reduction of velocity is a result of the decrease in the integrated negative buoyancy of the subducting lithosphere. By the time crustal accumulation reaches its maximum at $t = 75$ million years, the subduction zone has moved as much as 450 km towards the subducting plate, and the dip of the slab reduces from almost vertical to about 45°.

Towards the later stages of crustal thickening, a strong horizontal pressure gradient develops due to lateral variations in the thickness of the crust. This pressure gradient resists further thickening and forces the thickened crust to migrate towards the overriding plate, as also seen in the dynamic topography of the model (Fig. 5a). The topography is calculated by requiring the normal stress on the surface of the model to be equal to the weight of the topography. During the oceanic subduction, the topography is characterized by a trench zone on the oceanic side with a maximum depth of 8 km, and a small rise and depression on the overriding side. After the collision, the topography increases at the plate boundary and spreads towards the interior of the subducting plate, reaching a maximum value of 16 km at around 50 million years. In the later stages, a moderate topographic relaxation of 4 km takes place as the thickened crust flattens towards the overriding plate. This crustal deformation and topographic build up is similar to that described by other investigators (e.g. England and McKenzie, 1982; Willett et al., 1993 and Wdowinski and Bock, 1994). The crustal deformation shown in Fig. 3 is very similar to the orogenic deformation suggested by Willett et al. (1993); In the initial stages of deformation and as the shearing on the crust–mantle boundary increases, an asymmetric pattern of thickening develops with a low-angle wedge on the “pro” direction (towards the subducting plate), and a high-angle wedge on the “retro” side (towards the overriding plate). In the later stages, the base of the crust on the retro side detaches from the mantle and the deformation propagates towards the overriding plate, resulting in a smaller angle of wedge taper on the retro side. This deformation gives rise to different stress regimes during the course of crustal thickening. Fig. 6 shows the horizontal deviatoric stress, $\tau_{xx}$ (for a 2D incompressible flow the vertical deviatoric stress...
Fig. 5. Topography at different times for various models. (a) The nominal model at 10, 30, 50, 80 and 100 million years, (b) a model with $\lambda = 10^2$, and $n = 3$, (c) a model with $\lambda = 10^3$ and $n = 9$, (d) a model with $\lambda = 10^2$, and $n = 3$ and (e) a model with $\lambda = 10^3$, $n = 3$, and a continental crustal density of $\rho_c = 2900\text{kg/m}^3$. 
Fig. 6. Time evolution of horizontal deviatoric stress of the nominal model. Light gray shade and broken contours denote compressional stress and dark gray and solid contours denote tensional stress. Contour lines from 30 to 120 MPa with 30 MPa intervals are shown.

is equal to the opposite of the horizontal stress). During the oceanic subduction, horizontal compression prevails at the margin of the subducting plate (light shade), while the interior of the plate experiences horizontal tension (dark shade). During the crustal thickening ($t > 30$ million years), the state of deviatoric stress above the downgoing slab reverses; the crustal plateau becomes a region of horizontal extension, while the margins of the two converging plates become zones of horizontal compression. At the same time, the extent of the regions of compression grows as the thickened crust exerts greater stress on the two adjacent plates.

4.2. Evolution of the continental mantle lithosphere

Crustal thickening depresses the isotherms and leads to the formation of a colder and denser lithospheric root. Fig. 7 illustrates the depth to isotherms at the center of the downgoing slab against time. The top and base of the mantle lithosphere roughly correspond to isotherms $T = 0.4$ and $0.75$, respectively. During the oceanic subduction ($t < 30$ million years) the thickness of the thermal boundary layer above the subduction zone remains more or less constant, and a balance exists between the thermal structure of the downgoing slab and that of the subducting plate near the surface. The interval $t = 30-80$ million years corresponds to crustal thickening, during which a progressive downward deflection of the isotherms takes place, and a cold lithospheric root is formed. Fig. 7 suggests that the lithospheric root remains stable and does not founder into the mantle, since founding would lead to a rapid sinking and a subsequent rebound of the isotherms. The stability of the lithospheric root is also inferred from the kinetic energy of the system (Fig. 8) defined as (Marotta et al., 1998):

$$E_k = \frac{1}{2} \sum m_i v_i^2$$  (8)

where $m_i$ is the mass of the $i$th grid element and summation is over all the grid elements. The main acceleration of the convective system occurs during the oceanic subduction when the downgoing slab is still developing. The commencement of the continental subduction is concurrent with a sharp decrease in the total kinetic energy of the system, when a steep reduction occurs in the magnitude of the near-surface flow field. There is no major increase in the kinetic energy afterwards that would indicate a significant instability of the lithospheric root.
4.3. The role of viscosity and density in crustal subduction

Here, we study the effects of lithospheric strength and continental crustal density on the deformation of the continental lithosphere by examining models with different viscosity and density structures.

4.3.1. Effects of viscosity

Fig. 9 shows four snapshots of two models with different lithospheric strength. In the top model, viscosity varies by $10^{3}$ times with temperature, and the strength of the overriding continent is equal to that of the subducting one. The overriding plate of this model is thus 50 times weaker than that of the nominal model. The continental crust deforms similar to that of the nominal model, with the same asymmetric pattern. The crust thickens by as much as 90 km and maximum topography reaches 10 km (Fig. 5b). Because of weaker shear traction at the base of the crust of the overriding plate, the deformation front propagates to a greater distance on to the overriding plate. The propagation of the original crustal suture zone (denoted by the arrows in Fig. 9) towards the overriding plate is also observed in plane-strain models of Willett and Beaumont (1994) and Ellis et al. (1999). The bottom
Fig. 9. Time evolution of two models with different viscosities. The top panels are for a model with $\lambda = 10^3$ and $n = 3$. The arrows indicate the position of the original suture zone. The bottom panels are for a model with $\lambda = 10^4$ and $n = 9$.

model has $10^4$ times viscosity variations with temperature and $n = 9$. Lithospheric weakening near plate margins is more significant as there is a greater concentration of strain in the regions of higher stress. Because of weaker plates, the convection is stronger and faster, and crustal thickening reaches its maximum value of 9 km at an earlier time of about $t = 50$ million years. Like in the top model, the lower viscosity of the lithosphere facilitates the crust–mantle decoupling, and crustal separation takes place at shallower depths. Because of lower shear stresses acting at the base of the crust, smaller gradients of crustal thickness
is supported, and a thinner but wider crustal plateau and smaller topography (Fig. 5c) are developed.

In general, weaker lithospheric plates result in shallower depth of crustal detachment, wider and thinner crustal plateau, and lower topographic relief. For example, a model with 100 times variations of viscosity with temperature (i.e. 100 times less than the nominal model), results in a maximum depth of crustal subduction of 110 km compared to 170 km of the nominal model, and a surface topography that never exceeds 8 km (Fig. 5d). Furthermore, the width of the crustal plateau is around 1400 km, about 25% more than that of the nominal model.

4.3.2. Effects of density

Fig. 10 shows snapshots of a model identical to the nominal model except that the density of the continental crust is increased to 2900 kg/m$^3$, which reduces the density difference with the mantle by 27%. The two models behave in a similar way and the differences are mainly in the maximum subduction depth, amount of crustal thickening and surface topography (Fig. 5c). The depth of subduction in this model is about 210 km, 40 km more than that of the nominal model, and the crustal layer forms a thicker and narrower plateau. Because of smaller difference between the buoyancy of the crust and mantle lithosphere, a lower topography is developed. Topography reaches a maximum of 13 km and relaxes to about 8 km towards the end of simulation. A small part of the continental crust is entrained into the mantle by the subducting slab, and later (plot at $t = 100$ million years) is accumulated underneath the overriding plate. This portion is very small and comprises $<3\%$ of the total volume of the crust of the subducting plate.

4.4. Breakoff of the oceanic slab

An important aspect of deformation of a subducting continent is the breakoff of the subducted oceanic slab from the continental plate. The breakoff is a consequence of the opposing buoyancy forces of the denser oceanic plate and the lighter continental plate. This leads to a net extensional force acting in the vicinity of the ocean-continent transition in the subducting slab (Davies and von Blanckenburg, 1995). We have tested several models with different viscosity functions, lithospheric thermal profiles, and subduction velocities, in order to put constraints on the range of parameters that can result in slab breakoff. We find that under strong lithosphere conditions (such as that in the nominal model), slab breakoff does not occur. The integrity of the mantle part of the slab remains intact as the positive buoyancy of the continental crust grows.
larger during crustal accumulation. This is evident in models with $n = 3$ in Figs. 3 and 9, which show a continuous pattern of downflow underneath the zone of crustal thickening. In the model with $n = 9$ (Fig. 9) the subducted slab disappears after the crustal thickening reaches a maximum. However, this process is very slow and is aided by thermal erosion rather than by the sinking of the slab. Fig. 11 shows a model with distinct slab breakoff. In this model, the subducting and overriding plates are assumed to have initial thermal ages of 40 and 70 million years, respectively, and $10^3$ times variation of lithospheric viscosity is assumed. Furthermore, the convergence velocity is somewhat slower (about 4 cm per year). The subducting lithosphere of this model is thinner and weaker than that of the previous models. The deformation of the crust is similar to that seen before. The continental crust reaches a maximum depth of 140 km and eventually forms a plateau with a thickness of about 120 km. The subduction of the mantle part of the lithosphere, however, takes a different course. Soon after the crustal accumulation begins (after 40 million years), the subducting lithosphere necks in the region underneath the continental crust, and the oceanic slab detaches from the continental plate. The time needed for full detachment of the slab is relatively short and does not exceed 15 million years. The crust continues to thicken and achieves its maximum thickness after slab breakoff is completed. Davies and von Blanckenburg (1995) described the breakoff as the localization of strain in narrow zones of rifting in the ocean–continent transition, and argued that strain rate should be sufficiently high, such that the mantle material that replaces the broken slab does not have time to cool during the process. They estimated a relatively fast time-scale of 5–10 million years for the instability. We notice that in our model, the thinning of the slab occurs over a relatively broad depth interval (~150 km), and the rate of cooling near the region of extension is great enough to prevent the base of the crust to attain sub-lithospheric temperatures.
In this regard, the process of slab breakoff in our model differs from that devised by Davies and von Blanckenburg (1995) which predicts a more localized and rapid mode of detachment.

5. Discussion and conclusions

We have presented a set of models that take into account the interaction of the lithosphere and mantle, and the chemical buoyancy of the continental crust to study the deformation of a subducting continental lithosphere. The major results can be summarized as follows:

5.1. Kinematic boundary conditions

The kinematic models show that the imposed surface velocity and the crust–mantle viscous coupling have the greatest roles in preventing the formation of thick crustal roots at the collision zone. Crustal detachment happens over a wide depth range, depending on the strength of the subducting lithosphere, and crustal deformation near the surface of the model does not result in the development of thermal boundary layer thickening, the breakoff of the subducted slab, and/or cessation of subduction. This differs strongly from the self-consistent models in which the buoyancy of the light continental crust plays an important role in the development of a crustal plateau and the breakoff of the subducted slab.

5.2. Depth of continental subduction

The crust–mantle detachment results under a wide range of viscosity and density values. In the self-consistent models in which the buoyancy of the light continental crust plays an important role in the development of a crustal plateau and the breakoff of the subducted slab, the depth of subduction varies from 200 km for the highest lithospheric strength and crustal density to 90 km for the weakest lithosphere and lowest density. For thicker and colder lithosphere, the continental crust subducts to a greater depth. These results are confirmed by other theoretical investigations. van den Beukel (1992) showed that the depth of crustal subduction is inversely proportional to the surface heat flow of the continental lithosphere. Furthermore, geological evidence in many orogenic belts support such notion as well. For example, in the Himalayas, seismic measurements suggest a greater depth of continental subduction in the western part than in the central part. This crustal structure is complemented by higher surface heat flow of the Indian shield near the central Himalayas than in the western part (Gupta, 1982 and van den Beukel, 1992). The subduction depth of the crust and the thickness of the crustal plateau are also inversely related to the power-law exponent in the non-linear viscosity. In a model with \( n = 9 \) the continental crust thickens about 40% less than that of the nominal model with \( n = 3 \). These results are consistent with those obtained by other investigators (van den Beukel, 1992; Chemenda et al., 1995 and Davies and von Blanckenburg, 1995) who have calculated the depth of continental subduction as a function of the balance between the driving and resisting forces of subduction. It is difficult to make a quantitative comparison between the results of the various models, since the mechanical properties of each are very different. Nevertheless, the depth values obtained by our models are within the range of those determined by others. The analogue model of Chemenda et al. (1995) predicts that the lithosphere can subduct to a depth of 200–300 km before the upper crustal component fails and bounces upward, and thermo-mechanical models of van den Beukel (1992) and Davies and von Blanckenburg (1995) give crustal subduction ranges of 25–80 and 50–120 km, respectively. The latter authors emphasize that under lower plate temperatures or higher convergence velocities, the crust can subduct to greater depths. Geological observations also show a similar range of crustal subduction in many mountain belts. For example, Chopin (1984) and Michard et al. (1993) suggested a maximum depth of about 100 km for the high-pressure metamorphic rocks in the western Alps, and England and Holland (1979) estimated a depth of 60 km or more for the formation of high-pressure eclogite blocks in the eastern central Alps.

A mechanism that has been ignored in our two-dimensional modeling is the along-strike deformation in the collision zones. Several lines of geological and theoretical evidence (e.g. Tapponnier and Molnar, 1976; England and McKenzie, 1982 and Tapponnier et al., 1986) emphasize the importance of strike-slip faulting normal to the direction of crustal shortening in accommodating a significant part of convergence during the later stages of orogenies. The lateral extrusion of the crust accounts for some of the deformation...
of the upper part of the subducting lithosphere. We believe that because of the absence of such deformation in our models, our results overestimate the depth of continental subduction and the amount of crustal thickening.

5.3. Patterns of crustal deformation

Crustal thickening in our models has a pattern similar to that seen in the development of accretionary wedges in oceanic subduction zones (Dahlen, 1984); the newly arrived crust is added to the toe of the crustal mass, resulting in an asymmetric wedge-shaped plateau that has a low-angle taper on the subducting side and a high-angle taper on the overriding side. The asymmetry is a result of uneven advection of material towards the wedge. The thickened crust causes a downward deflection of the isotherms in the subducting lithosphere which is manifested by a reduced surface heat flow. As the crustal plateau grows in thickness, the thickened crust tends to relax in the direction of the overriding plate. Subsequently, the crustal wedge becomes more symmetric. The later phase of crustal relaxation is accompanied by increase in the magnitude of extensional stress in the plateau. This pattern of deformation is in good agreement with other results that predict a similar wedge-shape for the crust (e.g. Willett et al., 1993) and stages of early compression and late extension in continental collision zones (e.g. Houseman and England, 1986; Vilotte et al., 1986 and Marotta et al., 1998). The thickness and areal extent of the crustal plateau is found to be a function of lithospheric viscosity. In models with weaker lithosphere, the crust collapses faster and propagates farther in the overriding plate, since the shear stresses at the base of the crust are weaker. This causes the initial contact zone of the two converging continental crusts to progressively displace toward the overriding continent.

Another feature of the models is the reduced plate velocity and displacement of the surface point of subduction towards the interior of the subducting plate. This process is well observed in the collision zones. For example, the main central Fault (MCF) and the main boundary fault (MBF) in the Himalayas were formed after the underthrusting of India beneath Asia began, and each fault has taken up about 125 km of underthrusting (Lyon-Caen and Molnar, 1983). And in the Zagros mountains, many thrust faults to the southwest of the main Zagros thrust which denotes the old subduction suture zone, are currently accommodating much of the convergence between the Arabian shield and Eurasia (Jackson and McKenzie, 1984).

A secondary feature of the deformation of the continental crust is the small-scale convection that develops in the thickened crust because of the shear stresses produced by the down going plate at the base of the crustal wedge. This flow field can be inferred from the upward deflection of the isotherms inside the wedge. The flow field is very small and generates an upward vertical velocity of about 1 mm per year in the rising limb of the convective cell. Some researchers (England and Holland, 1979 and Pavlis and Bruhn, 1983) have suggested that the exhumation of high-pressure rocks in the mountain belts is a result of a corner flow in the crustal material that develops as a consequence of the ongoing subduction. It is possible to speculate that such small-scale flow field that produces vertical displacements of about 10 km over a time interval of 10 million years, can help bring lower crustal material to near-surface depths.

5.4. Slab breakoff

Our modeling shows that slab breakoff is controlled by the viscosity structure of the lithosphere. Strong slabs manage to preserve their integrity against the growth with time of the integrated crustal buoyancy. Reducing the strength of the slab, either by increasing the stress dependence of viscosity or by increasing the plate temperature, can cause detachment of the subducting slab. Breakoff always occurs well below the base of the thickened continental crust, and by the time it is completed (a time interval of about 10 million years), conductive cooling creates a cold lithospheric mantle root below the crust. In our models the breakoff does not occur at shallow depths and is not fast enough to let the hot asthenospheric material replace the detached material. The base of the crust never experiences subcrustal temperatures, and syn-collisional magmatic activity cannot be inferred from the models.

The lithospheric root that forms as a consequence of crustal thickening appears to be thermally stable. In this regard, the results of the models are consistent with those of Buck and Toksöz (1983) and Lenardic and Kaula (1995, 1996) who predict that under
Newtonian rheology the removal of the lithospheric root beneath orogenic belts is accomplished by slow thermal erosion as opposed to rapid gravitational instability. Houseman and Molnar (1997) have emphasized the importance of non-Newtonian viscosity in the rapid removal of the lithospheric root, and Schott and Schmeling (1998) have shown that delamination of the mantle lithosphere can happen only if both non-Newtonian and brittle rheologies are considered. Also, Lenardic and Kaula (1995) have demonstrated the importance of lateral variations in the lithospheric strength. In our models, non-Newtonian rheology is used to model the lithospheric plates, and it fails to model the removal of the root. This could be due to two factors: (1) as argued by Schott and Schmeling (1998), the weakening effect of brittle upper crust and pore pressure in the lithosphere may have strong influence on lithospheric delamination. Our models lack both of these features. The viscosity functions were chosen in order to produce a realistic plate-like behavior. This restricted to some extent the range of acceptable viscosity and may have resulted in a viscosity structure not suitable for modeling the convective instability of the lithosphere; (2) as shown by Lenardic and Kaula (1995), thermal erosion of the peripheral regions of the lithospheric root can result in a peripheral mantle lithosphere that is comparably thinner than the overlying crustal layer. This will result in a lithospheric root that is significantly weaker in the peripheral regions and causes the instability and foundering of the central part of the root. Our models do not provide any mechanism for such peripheral weakening. The thickness of the peripheral mantle lithosphere is always comparable to that of the peripheral crust (Fig. 3). A crustal plateau with a greater horizontal extent may result in a larger lithospheric root with a greater thickness variations across its length. This would require a longer subducting continental plate.

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