Dynamics of continental collision: influence of the plate contact

Roberta De Franco, Rob Govers and Rinus Wortel

Faculty of Geosciences, Utrecht University, Utrecht, the Netherlands. E-mail: defranco@geo.uu.nl

Accepted 2008 May 19. Received 2008 May 19; in original form 2007 October 11

SUMMARY

Observations show that continental collision may evolve in different ways, resulting in a wide range of tectonic responses. In search of the controlling conditions and parameters, we start from the results of our previous work, which demonstrated that the properties of the plate contact are important for the overall dynamics of convergent plate margins. Two fundamental types of subduction plate contact can be distinguished: one based on a fault and the other based on a weak subduction channel. In this study, we investigate how the plate contact affects the initial stage of continental collision. We use a finite element method to solve the heat and the time-dependent momentum equations for elastic, (power-law) viscous and plastic rheologies. For the same rheological properties and driving forces, varying the nature of the plate contact leads to three types of responses. The presence of a subduction channel promotes coherent and, when the boundary conditions allow it, plate-like subduction of the continental margin. In models with a subduction fault, coherent subduction of the incoming continental lithosphere occurs when the colliding passive margin has a gentle slope. The approaching continental sliver starts to subduct and the subduction is characterized by a non-plate like behaviour—slower subduction velocity than in channel models and strong slab deformation. If the continental margin is steep and the strength of the incoming continental crust is high, fault models result in locking of the trench, eventually leading to slab break-off. If the crustal strength is relatively low, shear delamination of part of the crust is expected. In the channel model, this type of delamination never occurs. The tectonic settings used in our experiments (prescribed plate velocity of the subducting plate versus fixed subducting plate corresponding to a landlocked basin setting) do not significantly influence the nature of the model response. We conclude that initial stages of continental collision are strongly affected by whether the subduction contact is a fault or a channel. Neither the slab pull magnitude nor the tectonic setting is very important to the overall geodynamics at this stage. The plate contact type, along with the slope of the incoming passive margin and the rheology of the continent, controls whether the incoming crust (1) subducts entirely; (2) separates partially or entirely from the lithospheric mantle or (3) blocks the trench, likely leading to slab break-off.

Key words: Numerical solutions; Subduction zone processes; Continental margins: convergent.

1 INTRODUCTION

Continental collision, following oceanic subduction, plays a key role in connecting plate motion and orogenic events. In general, due to its thick and light crust, continental lithosphere is positively buoyant and will resist subduction, whereas cold and dense subducted oceanic lithosphere generates a large downwards force (slab pull). Consequently, continental crust is subducted with difficulty, if at all, whereas most oceanic crust is subducted easily at an oceanic trench. In a subduction process, the collision between a continental fragment or terrane and a continental overriding plate may result in different modes: (1) the incoming continental material arrives at the trench and is subducted; (2) all or part of the continental buoyant crust is separated from the lithospheric mantle and is accreted to the overriding plate, allowing continuation of plate convergence; (3) the approaching continental margin locks the trench, eventually resulting in slab break-off (slab detachment). All these processes appear to be possible, and which of these scenarios takes place or has taken place in a certain region, depends on several factors: large scale tectonic setting of the two plates involved and more local factors such as the rheological parameters, the geometry of the continental margin and the buoyancy.

Subduction of continental crust, together with the lithospheric mantle, results in a positive buoyancy force opposing the negative buoyancy force of the dense mantle part of the lithosphere. Nevertheless, deep subduction of continental crust is possible if, for example, the crust becomes denser because of phase transformations (Austrheim 1987; Le Pichon et al. 1992; Dewey et al. 1993). In
nature, there is ample evidence that continental crust may have experienced pressure of about 30–35 kbar. The evidence comes from ultra-high pressure metamorphic minerals (e.g. coesite, diamond) revealing subduction of continental crust to mantle depth and subsequent exhumation (e.g. Smith 1984; Liou et al. 2000; Chopin 2001; Yang et al. 2003). Numerical studies supported the idea that the upper crust can reach a burial depth varying between 50 and 450 km if the crust is tightly welded to the subducting mantle by a strong rheology and if slab break-off does not occur (Ranalli et al. 2000; Regard et al. 2003; Toussaint et al. 2004). Crustal subduction is also affected by other parameters—plate velocity, thermomechanical state of the lithosphere, collision zone geometry, shear heating at the plate interface, partial melting in the subduction channel, composition and erosion/sedimentation processes (Gerya & Yuen 2003; Gerya et al. 2004b; Gerya & Stockhert 2006; Burov & Yamato 2007; Gerya & Burg 2007; Faccenda et al. 2007; Massonne et al. 2007).

Another mechanism influencing subduction of the continental crust is the interplate pressure between the overriding and the subducting plate. Chemenda et al. (1996) showed that a low interplate pressure allows the continental crust to subduct to a depth of 200 km without failure. They found that the subducting continental crust reaches a maximum depth that is proportional to the strength of the crust and inversely proportional to the interplate pressure.

The whole crust, or part of it, may separate from the lithospheric mantle and remain at the surface (Cloos 1993; Kerr & Tarney 2005; Vos et al. 2007). This implies that when crust and mantle lithosphere subduct, their speed and direction can be different. This process, known as delamination, was proposed originally by Bird (1978) to explain some of the tectonic features in the Himalayas and it is, indeed, a suitable mechanism to facilitate continuation of subduction. Delamination is only possible if there is a layer of weak material within the continental crust, for example, the lower crust (Ranalli et al. 2000; Toussaint et al. 2004). Chemenda et al. (1996) showed that continental crust delamination occurs in front of the subduction zone when the interplate pressure between overriding and subducting plate is high. For several regimes, progressive separation of the lithospheric mantle from the buoyant continental crust has been proposed, for example, the Colorado Plateau (Bird 1979), the Barbados Ridge (Westbrook et al. 1988).

Another possibility is that the incoming continental material may lock the trench, possibly resulting in break-off of the slab if the slab is not sufficiently strong (Davies & von Blanckenburg 1995; Wong A Tong & Wortel 1997; Wortel & Spakman 2000) or after a period of thermal relaxation (depending on slab age, subduction velocity and interplate heating) for any initial strength of the slab (Boutelier et al. 2004; Gerya et al. 2004a; Andrews & Billen 2007; Faccenda et al. 2007). Alternatively, the continental lithosphere thickens until it drips into the deeper mantle as a result of a Rayleigh–Taylor thermal instability (Housen et al. 1981; Houseman & Molnar 1997; Pysklywec et al. 2000, 2002). Whether and at which depth slab break-off occurs is a function of temperature and convergence velocity (Davies & von Blanckenburg 1995; Wong A Tong & Wortel 1997; Sobouti & Arkani-Hamed 2002; Toussaint et al. 2004). Strong slabs preserve their integrity against the increase of the integrated crustal buoyancy with the amount of subducted crust. The feasibility of slab detachment started to be investigated in the early 1970s. Some studies suggested that the gaps in seismic activity as a function of depth might be explained by the existence of detached slabs (Isacks & Molnar 1971; Barazangi et al. 1973). Subsequently, seismic tomography allowed for the identification of real gaps in slab structure (e.g. Spakman 1990; Blanco & Spakman 1993). Slab break-off has been used to explain the upper-mantle structure and tectonic deformation in many regions, for example, the Mediterranean–Carpathian region, the New Hebrides and Turkey (Chatelain et al. 1992; Wortel & Spakman 1992, 2000; Parlak et al. 2006; Lei & Zhao 2007). Understanding the factors controlling the three different modes of continental collision is important for unravelling the tectonic evolution of convergent plate boundary regimes.

At convergent plate boundaries, the properties of the plate contact are important for the overall dynamics. Two fundamental physical states of the subduction contact can be distinguished: one based on a fault and the other based on a subduction channel. A subduction zone may evolve from one state to another, for example, through a variation in sediment supply or hydration (mainly serpentinitization) of the overriding plate (Gerya et al. 2002; Stoeckhert & Gerya 2005; Sobolev & Babeyko 2005; Babeyko & Sobolev 2007). For normal oceanic lithosphere subduction, De Franco et al. (2007) have illustrated that the type of plate contact (i.e. channel or fault) changes the way in which the stresses are transmitted from one plate to the other and affects the velocity of subduction and the magnitude of the interplate pressure. The fact that these are also taken to be governing factors in the evolution of continental collision (see above) motivates us to assess for the first time the role of the plate contact in this process. Although numerous studies have been done to understand the mechanisms governing the different modes of continental collision (e.g. Pysklywec et al. 2000, 2002; Burov et al. 2001; Toussaint & Burov 2004; Burg & Gerya 2005; Burov & Toussaint 2007), and fault models with various incoming slab geometries, including ridges on the subducting plate, have been studied numerically (e.g. van Hunen et al. 2000, 2002, 2004), a comparison between the influence of the two different types of plate contact is missing.

In this study, we show that critical differences develop during initial stages of the collision. Through numerical modelling we illustrate how, for the same rheological properties and driving forces, a variation of plate contact type results in different modes of continental collision—subduction of the continental material or delamination or break-off. Here we focus on two tectonic settings. In the first scenario, the subducting plate is pushed into the subduction zone by virtue of the plate tectonic setting (e.g. Himalayas). This is in contrast with the second setting of a land-locked basin, where there is no net convergence between the surface plates so that subduction must occur through roll-back (e.g. the Apennine and Hellenic arc systems in the Mediterranean region).

2 NUMERICAL METHOD AND MODEL SETUP

In this study, we do not make predictions for any real continental subduction zone, but we are interested in understanding the physical process involved. For this reason, we choose to analyse a generic subduction zone in which a passive margin arrives at the plate contact after the oceanic lithosphere has been subducted beneath a continental overriding plate. In our models, the subduction zone is represented by a 2-D cross-section (Fig. 1). Although continental collision processes have important 3-D features, the first-order effects of convergence can be appreciated by analysing a characteristic cross-section normal to the trench. With this simplification, we assume that the continent extends infinitely in the out-of-plane direction.

Because the numerical method and the model setup used in this study are very similar to De Franco et al. (2007), we refer...
Figure 1. Model setup. On the left-hand side the continental plate with the continental crust on top, on the right-hand side the oceanic plate with the embedded continental margin. Boundary conditions: rigid bottom; on the left-hand side: in the lithospheric mantle horizontal and vertical displacement are not allowed, in the sublithospheric mantle only horizontal displacement is allowed; on the right-hand side: in the lithospheric mantle horizontal and vertical displacement are not allowed or, alternatively a horizontal velocity of 5 cm yr\(^{-1}\) is applied. In the sublithospheric mantle only horizontal displacement is allowed. On top: free surface boundary conditions.

Table 1. Thermal (Ponko & Peacock 1995), elastic (Dziewonski & Anderson 1981) and creep model parameters. \(n\) is the stress power exponent.

<table>
<thead>
<tr>
<th>Thermal properties</th>
<th>Conductivity (k)</th>
<th>Heat production (H)</th>
<th>Specific heat (C(_p))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Continental crust</td>
<td>2.8 W m(^{-1}) K(^{-1})</td>
<td>6.81 \times 10(^{-7}) W m(^{-3})</td>
<td>1.24 \times 10(^{5}) J kg(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>Lithospheric mantle</td>
<td>3.14 W m(^{-1}) K(^{-1})</td>
<td>0</td>
<td>1.17 \times 10(^{5}) J kg(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>Sublithospheric mantle</td>
<td>3.14 W m(^{-1}) K(^{-1})</td>
<td>0</td>
<td>1.17 \times 10(^{5}) J kg(^{-1}) K(^{-1})</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Elastic properties</th>
<th>Young’s modulus (Y)</th>
<th>Poisson’s modulus (v)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Continental crust</td>
<td>5 \times 10(^{10}) Pa</td>
<td>0.3</td>
</tr>
<tr>
<td>Lithospheric mantle</td>
<td>1 \times 10(^{11}) Pa</td>
<td>0.25</td>
</tr>
<tr>
<td>Sublithospheric mantle</td>
<td>1 \times 10(^{11}) Pa</td>
<td>0.25</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Creep properties</th>
<th>Pre-exponent (A)</th>
<th>Activation energy (Q)</th>
<th>Activation volume (V)</th>
<th>(n)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Continental crust</td>
<td>8.3 \times 10(^{-26}) Pa(^{-n}) s(^{-1})</td>
<td>163 kJ/mole</td>
<td>17 \times 10(^{-6}) m(^3) mole(^{-1})</td>
<td>3.1</td>
</tr>
<tr>
<td>Mantle &lt; 350 km</td>
<td>9.18 \times 10(^{-16}) Pa(^{-n}) s(^{-1})</td>
<td>485 kJ/mole</td>
<td>25 \times 10(^{-6}) m(^3) mole(^{-1})</td>
<td>3.25</td>
</tr>
<tr>
<td>Mantle &gt; 350 km</td>
<td>1.54 \times 10(^{-11}) Pa(^{-n}) s(^{-1})</td>
<td>270 kJ/mole</td>
<td>6 \times 10(^{-6}) m(^3) mole(^{-1})</td>
<td>1</td>
</tr>
</tbody>
</table>

to their study for more detailed aspects. The dynamics of the lithosphere and upper mantle is governed by the momentum equation. Using the plane strain approximation, we solve this partial differential equation for instantaneous velocities and stresses. We use the G-TECTON finite element code to solve the momentum equation (http://www.geo.uu.nl/Research/Tectonophysics; Melosh & Raefsky 1983; Govers 1993; Buiter et al. 2001), and the steady state diffusion-advection equation (Govers & Wortel 1993; De Franco et al. 2007). Constitutive laws in the model represent elastic, viscous and plastic deformation. Viscosity is taken to be strongly temperature-dependent, in accordance with rock mechanical experiments in both the power-law and diffusion creep regime (e.g. Kohlstedt & Zimmerman 1996). Density also depends on temperature and is expressed by a linear equation of state in our model. These dependences make it necessary to also solve for model temperature (see De Franco et al. 2007). As the mechanical models of this paper focus on short timescales relative to a thermal diffusion timescale, we ignore the temporal evolution of conduction. We do account for advective heat transport through the Lagrangian motion of the numerical grid.

We adopt two different ages for the surface oceanic lithosphere at the right-hand model boundary: one of 33 Myr and one of 25 Myr. Here, the initial geotherms are defined using a half-space cooling model. Using a spreading rate of the oceanic lithosphere of 4 cm yr\(^{-1}\), the oceanic age at the trench is about 70 and 40 Myr, respectively. Temperatures in the continental overriding plate are based on a representative steady state geotherm with a surface heat flow of 65 mW m\(^{-2}\) (see Table 1 and Ponko & Peacock 1995). In the sublithospheric mantle, we use an adiabatic gradient of 0.3° K km\(^{-1}\) (Ponko & Peacock 1995). The imposed surface temperature is 0°C for both models. The initial temperature field for the first model is displayed in Fig. 2(a). Temperatures at the right- and left-hand side of the model correspond to the two geotherms plotted in Fig. 2(b). These geotherms represent the side boundary conditions used to solve the diffusion-advection equation.

Fig. 1 displays the model domain. It is characterized by a depth of 700 km and a horizontal extension of 2800 km. The subducting oceanic plate with a continental block embedded is on the right-hand side, the overriding plate is on the left-hand side. The incoming continental margin is located at the plate contact. We adopt two inclination angles of 3° and 15° for the incoming margin continental slope. These values are derived from our analysis of the current global distribution of slope inclination angles of passive margins representing the most common and the maximum slope inclination angle, respectively (see Fig. 3). The histogram is obtained from the analysis of a global 5-min latitude/longitude topography data set (ETOPO5). The crustal thickness of the continental fragment is assumed to be 26 km. The continent is in isostatic equilibrium. The shape of the curved portion of slab is defined by an error function (Govers & Wortel 2005); we adopt a radius of curvature \(R\) of 1.6 L (\(L = 80\) km being the thickness of the oceanic lithosphere in our models) and final dip angle \(\theta\) of 45°. The slab initially extends to 350 km depth. In some models, we also use a shorter slab that extends to 175 km depth. The overriding plate is 100 km thick, which represents the lithospheric mechanical thickness defined by the viscosity structure, and has a crustal thickness of 35
Figure 2. (a) Total temperature distribution. (b) Geotherm at the sides of the model domain. Solid curve, temperature boundary condition at the right-hand side of the model; dashed curve, temperature boundary condition at the left-hand side of the model.

Figure 3. Global distribution of inclination angles of continental slopes at passive margins.

The finite element mesh is described by a non-uniform triangular grid, characterized by a maximum cell size of 500 km$^2$ in the asthenosphere and the lower mantle, 100 km$^2$ in the mantle lithosphere and 50 km$^2$ in the trench area.

The rheology of our model is elastic, viscous, or plastic and depends on composition, temperature, pressure and effective stress (Table 1). Model viscosities follow from either steady state dislocation creep or steady state diffusion creep. We choose model parameters following Karato & Wu (1993) for the mantle, where the parameters are chosen to be intermediate between their wet and dry values since they concluded that the rheology of the upper mantle is between that of dry (water-free) and wet (water-saturated) olivine. For continental crust of both the overriding plate and the fragment, we use rheological parameters from Freed & Burgmann (2004) (see Table 1). Dislocation creep is assumed to prevail at a depth shallower than 350 km, whereas diffusion creep dominates within the transition zone (Karato & Wu 1993).

In the region of the mantle lithosphere and asthenosphere above the subducting slab (Fig. 1), the low viscosity mantle wedge (LVW) reduces the dynamic downwarping of the overriding plate in the arc/backarc region (Billen & Gurnis 2001). In our model, we assume a uniform viscosity of $5 \times 10^{19}$ Pa s in the mantle wedge. Following Billen & Gurnis (2001), this value has been chosen 10 times smaller than the minimum asthenosphere viscosity. We apply an isotropic power-law Von Mises criterion to limit deviatoric stresses in accordance with Byerlee’s law (Byerlee 1978) using hydrostatic fluid pressure and horizontal compression. The lower limits of the viscosity are $1 \times 10^{20}$ Pa s for the asthenosphere and the lower mantle, whereas in the mantle lithosphere we did not impose a lower limit. The upper limit is $1 \times 10^{24}$ Pa s and $1 \times 10^{23}$ for the mantle lithosphere and asthenosphere and lower mantle, respectively.

We use two descriptions for the active plate contact—a subduction fault and a subduction channel—with the following characteristics:

1. A subduction channel separates the subducting and the overriding plate. The channel width is assumed to be approximately 6 km, similar to what have been used in previous model studies (Shreve & Cloos 1986; Beaumont et al. 1999) and in agreement with seismic studies that have shown the presence of interplate sedimentary channel-like units of about 1–8 km (Eberhart-Phillips & Martin 1999; Tsuru et al. 2002; Oncken et al. 2003; Abers 2005). Channel viscosity is taken to be Newtonian, with a value of $7 \times 10^{18}$ Pa s (England & Holland 1979; Shreve & Cloos 1986). We name this class of models CM (channel models).

2. A deformable subduction fault is described via updated slippery nodes, in which the fault slip is locally kept parallel to the fault (Buitert al. 2001). Fault friction is negligible. We name this class of models FM (fault models).

We separate model densities into a 1-D reference density profile and the remaining density anomalies (Govers & Wortel 2005). Reference densities are used to initialize hydrostatic pressures, which initially do not contribute to the forcing. The remaining density anomalies (i.e. slab pull forces) are used to drive model deformation and are instantaneously applied at the beginning of the model calculation. Bending stresses would be expected to exist in the initially curved portion of the slab. We do not include these initial bending stresses. However, bending stresses do develop as the slab deforms during the model evolution. The continental fragment is positively buoyant and its density anomaly varies between 100 and 400 kg m$^{-3}$ (Ranalli et al. 2000).

During the numerical experiments, we perform a re-meshing procedure of the deformed grid. The reason is that in our Lagrangian approach, the finite element mesh becomes so distorted that integrals and derivatives needed in the finite element problem cannot be properly approximated. Our re-meshing method consists of updating the connectivity between Lagrangian node positions. Taking particular care to preserve material boundaries and physical boundaries, we adopt a Delaunay triangulation using...
Dynamics of continental collision: influence of the plate contact

3 RESULTS

In the experiments described below, we applied two types of boundary conditions. First, the open subduction setting where we impose a velocity of $5 \text{ cm yr}^{-1}$ at the right-hand side of the subducting plate, and we fix the left-hand side of the overriding plate. Second, we use a land-locked basin setting in which there is no net convergence between the surface plates. We also investigate two different types of geometries for the incoming continental sliver; one with a continental slope inclination angle of 3° and a steeper one with a continental slope inclination angle of 15°.

The results are displayed in Figs 4–7. Figs 4 and 6 show the surface expressions of the models in the open subduction setting and land-locked basin setting, respectively. Figs 4(a) and 6(a) display the horizontal velocity of the free surface for the entire model domain at 0.56 Myr. At this time, model initialization signatures have decayed. The trench is visible as a step change at horizontal coordinate of about 100 km. Figs 4(b) and 6(b) show the evolution with time of the ratio obtained dividing the velocity magnitude of the slab ($V_{\text{slab}}$) by that of the horizontal part of the oceanic plate ($V_{\text{horiz}}$) after the initial spin-up period of the model. The ratio is one if the plate has a plate-like behaviour. A value greater than one indicates that the slab moves faster than the horizontal part of the plate, suggesting that plate thinning occurs somewhere in between the two points where the velocities are taken (red and blue circles in Fig. 4c). When the ratio is less than one, the slab moves slower than the surface plate, resulting in plate thickening.

Figs 5(a1)–(a5) display the total strain rate distribution for the open subduction model at 0.56 Myr, whereas Figs 5(b1)–(b5) show...
3.1 Channel models (CM)

In this setup the plate contact is described by a low viscosity channel that extends to a depth of 100 km. The viscosity of the channel is $\eta = 2 \times 10^{18}$ Pa s.

3.1.1 Channel model with open subduction boundary conditions and $3^\circ$ continental slope inclination angle (CMO$_3$)

The horizontal velocity of the subducting plate varies from $5 \text{ cm yr}^{-1}$ at the right-hand side to $5.6 \text{ cm yr}^{-1}$ for the incoming continental fragment near the trench (Fig. 4a black solid curve). This velocity increment is likely due to the elastic response of the plate to the positive buoyancy of the continental crust. Close to the plate contact, the overriding plate is characterized by a negative velocity, indicating the overriding plate retreats from the trench with a speed of about $5 \text{ mm yr}^{-1}$. In Fig. 4(b), the velocity ratio is represented by the black solid curve and is equal to a constant value close to one, denoting that the velocity along the subducting plate is uniform and it remains uniform as a function of time.

The strain rate distribution is displayed in Fig. 5(a1) at 0.56 Myr. The strain rate in the slab and in the continental crust of the incoming fragment has a small magnitude of about $1 \times 10^{-15} \text{ s}^{-1}$. In the entire model domain, the highest value of the strain rate is in the subduction channel and below the horizontal subducting plate. These two regions represent relative thin layers of rapidly deforming material. In Fig. 5(a1), the deformed geometry of the model shows that the continent has been dragged down with the subducting plate to a depth of about 30 km. In Fig. 5(b1), the average value of the strain rate in the slab shows a nearly stationary behaviour as a function of time. The strain rate is not affected by the subduction of the buoyant continental crust. On the other hand, the crustal strain rate slightly increases with time.
3.1.2 Channel model with open subduction boundary conditions and 15° continental slope inclination angle (CMO$_{15}$)

In this model, the continental slope of the margin is inclined at an angle of 15°. The steeper geometry of the continental fragment results in a similar response of the model as CMO$_3$.

The subducting plate moves with a speed of about 5 cm yr$^{-1}$ and, close to the plate contact, increases to a value of about 5.6 cm yr$^{-1}$ (Fig. 4a black dashed curve). The overriding plate velocity is nearly zero, except close to the plate contact where the velocity is negative; this region of the overriding plate moves towards the left-hand side with a speed of about 4 mm yr$^{-1}$. 
Like in the previous model, the velocity ratio is equal to one. This means that the velocity along the subducting plate is uniform and that this plate-like behaviour is not perturbed in time (Fig. 4b black dashed curve).

Like in CMO3, the strain rate within the subducting plate is small (Fig. 5a2). Strain rate is high in the channel and beneath the oceanic surface plate. The deformed geometry of the model displays that the incoming continental fragment is subducted to a depth of about 30 km. In Fig. 5(b2) the average value of the strain rate shows a nearly stationary behaviour. This curve displays a response very similar to the previous model. The continental crust is characterized by a slightly higher strain rate than in the slab.

We performed an experiment in which the lower crust of the continental margin is characterized by a low strength of about 30 MPa (CMOw15). The results are very similar to those of CMO15; for this reason, we decide not to show them.

### 3.1.3 Channel model with land-locked basin boundary conditions (CML3, CML15)

Here we impose land-locked basin boundary conditions. We adopt two inclination angles, $3^\circ$ and $15^\circ$, of the incoming continental margin.

In Fig. 6(a), the black solid line represents the velocity of CML3 at 0.56 Myr. The speed of the oceanic plate increases linearly towards...
the trench with a maximum value of about 2 cm yr$^{-1}$. The overriding plate moves towards the right-hand side with a maximum speed of 1 cm yr$^{-1}$. The dashed black line shows the velocity of CML15. Like before, the velocity increases towards the plate contact. The velocity magnitude is slightly smaller than in CML1. The overriding plate is characterized by the same behaviour as CML1.

In Fig. 6(b) the black solid curve and the black dashed curve represent the velocity ratio of CML1 and CML15, respectively. They are characterized by a constant value equal to 1.3. Different from CMO, velocities in CML are not plate-like.

The strain rate distribution for these models is very similar to that of CMO3 and CMO15. Therefore, we will subsequently show these results in the Appendix.

### 3.2 Fault Models (FM)

In these models the plate contact is described by a frictionless fault that extends to a depth of 100 km.

#### 3.2.1 Fault model with open subduction boundary conditions and 3° continental slope inclination angle (FMO3)

In Fig. 4(a), the grey dashed curve shows the horizontal velocity at the surface at 0.56 Myr. The convergence velocity is 5 cm yr$^{-1}$ and it decreases towards the plate contact, to increase again in correspondence of the continental fragment. The overriding plate, close to the plate contact moves away from the trench with a maximum value of 1.5 cm yr$^{-1}$. In this region the velocity is higher than in the CM.

The grey dashed curve in Fig. 4(b) shows the velocity ratio as function of time. This velocity ratio increases with time. This indicates that the horizontal part of the oceanic plate moves slower than part of the slab and that the velocity difference between the two regions enlarges with time. Downdip extension thus occurs somewhere within the region between the two points indicated in Fig. 4(c).

Fig. 5(a3) displays the total strain rate distribution at 0.56 Myr. The strain rates in the slab, in the continental crust and in the overriding plate are higher than those in the CM. The slab significantly deforms at a depth of about 150 km, where the strain rate is localized. The deformed geometry of the domain shows that the incoming continental crust is subducted to a depth of about 15 km. The depth reached by the subducting continent is shallower than in the CM. The overriding plate is characterized by high strain rate and marked downwelling. In Fig. 5(b3) the black and the red curves show that the average strain rate in the slab and the crust of the incoming fragment increases with time, reaching higher values than the previous models. Strain rate is high beneath the oceanic surface plate.

#### 3.2.2 Fault model with open subduction boundary conditions and 15° continental slope inclination angle (FMO15)

The black dotted curve in Fig. 4(a) shows that the velocity jump at the trench is strongly reduced with respect to FMO3. The horizontal velocity shows that the speed of the subducting plate decreases from the right-hand side towards the plate contact to less than 4 cm yr$^{-1}$. The overriding plate moves in the same direction as the subducting plate in the region between the plate contact and the horizontally projected tip of the slab end. The rate of retreat of the overriding plate from the trench is faster than in all previous models. The rest of the overriding plate has velocity magnitude equal to zero.

In Fig. 4(b) black dotted line, the velocity ratio increases with time faster than in all the models so far. Such a behaviour shows that the horizontal part of the oceanic plate moves slower than the slab.

The strain rate in the slab and in the overriding plate is higher than in the CM and in FMO3 (Fig. 5a4). From the deformed geometry of the model, it is evident that the incoming continent locks the trench and does not subduct with the oceanic material. The slab deforms more than in FMO3 at a depth of about 150 km. The strain rate in the slab is higher than in FMO3. Fig. 5(b4) shows the average strain rate in more detail—the strain rate in the slab increases with time more than in FMO3 reaching a maximum value of about 1 × 10$^{-14}$ s$^{-1}$. In the crust the strain rate is stationary and lower than in FMO3. Strain rate is high beneath the oceanic surface plate.

#### 3.2.3 Fault model with open subduction boundary conditions, 15° continental slope inclination angle and weak crustal layer (FMOw15)

In this model, the strength of the weak crustal layer of the incoming continental fragment is 30 MPa, about 50 per cent lower than FMO15. The subducting plate moves at a speed of 5 cm yr$^{-1}$ and jumps to 2 cm yr$^{-1}$ in the continental fragment (Fig. 4a solid grey curve). The part of the overriding plate that is closest to the plate contact moves towards the left-hand side, while the rest of the overriding plate advances towards the plate contact. The continental fragment moves with the same velocity and in the same direction as the near-trench part of the overriding plate. The velocity jump along the subducting plate is an expression of shear delamination of the upper crust from the rest of the lithosphere.

The velocity ratio (grey solid curve in Fig. 4b) increases with time similar to FMO3. However, it increases less than in FMO15, indicating that crustal delamination decreases downdip extension.

The strain rate distribution in the slab and in the overriding plate is very similar to the one of FMO1 (Fig. 5a5). The strain rate maximizes in the slab at a depth of about 150 km; at that depth the slab is strongly deformed. A high strain rate region is visible in the crust of the continental margin. This is the expression of the shear delamination process, in which the upper crust separates from the lower one. The lower crust continues to subduct reaching a depth of about 15 km during the modelled period. The detailed representation of the strain rate in Fig. 5(b5) indicates that the average strain rate in the slab increases with time but less than in FMO15. In the crust, the strain rate is one order of magnitude higher than in the slab and higher than in all the other models.

#### 3.2.4 Fault models with land-locked basin boundary conditions (FML3, FML15, FMLw15)

The following experiments are characterized by land-locked basin boundary conditions. We adopt two inclination angles of the incoming continental margin of 3° and 15°, FML3 and FML15 respectively. In FMLw15 (w=weak), we use the same geometry as in FML15, but the strength at the transition between lower and upper crust of the incoming passive margin is lower, and its value is about 30 MPa.

In FML3, the surface velocity of the subducting plate is lower than in either of the channel models (CML1 and CML15); it increases towards the trench with a maximum value of 1 cm yr$^{-1}$ (Fig. 6a grey dashed curve). In the overriding plate, the velocity increases.
towards the trench with a maximum speed of 2 cm yr$^{-1}$. The black dotted curve shows the surface horizontal velocity of FML$_{15}$. The subduction rate is a few mm yr$^{-1}$ and it is smaller than in FML$_{3}$. The overriding plate moves at a maximum speed of about 0.5 cm yr$^{-1}$, which is slower than in all the previous models. The subduction velocity of FMLw$_{15}$ is a bit lower than the velocity of FML$_{1}$ but higher than in FML$_{15}$ (Fig. 4a solid grey curve). The main difference between FML$_{13}$ and FMLw$_{15}$ is that the upper crust of FML$_{15}$ hardly moves, and that FMLw$_{15}$ shows substantial movement towards the right-hand side. There is no relative motion across the trench in FMLw$_{15}$, and the upper crust of the continental fragment moves along with the crust of the overriding plate at a speed of 1 cm yr$^{-1}$.

In Fig. 6(b) the grey dashed curve shows the velocity ratio of FML$_{13}$. The ratio increases with time. This means that the horizontal part of the oceanic plate moves more slowly than the slab. Such a difference increases with time. The black dotted line, represents the velocity ratio of FML$_{15}$, whereas the grey solid curve is the ratio of FMLw$_{15}$. The FML$_{15}$ velocity ratio increases faster than in all other subduction models. The FLMw$_{15}$ velocity ratio is only slightly higher than in FML$_{1}$ ratio. In both cases, the horizontal part of the oceanic plate moves slower than the slab, indicating downdip extension of the subducting plate. The detailed description of the strain rate distribution for these models is displayed in Appendix. We performed some tests with stronger rheology of the plates; since the main characteristics of our results do not change, we decided not to show these results.

### 3.3 Short slab models (CMO$_{15}$, FMO$_{15}$, FMOw$_{15}$)

In the next models, we want to estimate the importance of the slab pull in the evolution of our experiments. To reduce the slab pull, we diminish the length of the slab, that in the following experiments reaches an initial depth of 175 km. In these experiments, we only use the continental margin with slope inclination angle of 15$^\circ$. We impose a velocity of 5 cm yr$^{-1}$ at the right-hand side lithospheric boundary of the models, and we fix the left-hand side of the overriding plate.

The results are summarized in Fig. 4. In Fig. 4(a) the red dashed curve represents the horizontal velocity for CMO$_{15}$ at 0.56 Myr. The only difference with respect to CMO$_{15}$ is that the velocity magnitude is slightly reduced. The velocity ratio of CMO$_{15}$ also is similar to CMO$_{15}$ (Fig. 4b). The velocity of the slab is initially slower than in the horizontal part of the plate (curve value lower than one). With time, it increases until the ratio reaches a constant value equal to one. This suggests that the continental margin is pushed coherently into the subduction zone with the oceanic lithosphere in a steady plate-like fashion.

For the same type of geometry, we further reduce the density anomaly of the slab (we do not show these curves in Fig. 4), and even in this case, the behaviour of the model does not change much. In the extreme case, in which the slab pull due to the negatively buoyant oceanic lithosphere is zero, the subducting plate continues to be driven into the mantle and slowly tends to reach the same tectonic state of the previous CM.

In the FM, reducing the slab length has a stronger impact than in the CM. The blue dotted curve in Fig. 4 shows results of FMO$_{15}$; the horizontal velocity of the plate is lower than in FMO$_{15}$. The horizontal velocity of FMOw$_{15}$ (Fig. 4 blue curve) is higher than in FMO$_{15}$ but slower than in FMOw$_{15}$. Reducing the slab pull even further does not lead to a further decrease of the surface velocity. The velocity ratio does, however, change in this case; a slab pull results in a smaller slab velocity, making it more similar to the velocity of the surface oceanic plate.

### 3.4 Horizontal surface stress

In this section, we summarize the horizontal stress response at the surface of the previous models (Figs 7a and b) after 0.56 Myr. Away from the plate boundary zone, there is an overall tendency for the fault models to be more compressive than the channel models. In CMO$_{3}$ and CMO$_{15}$, the subducting plate and the terrane are in tension, whereas in FMO$_{3}$ and FMOw$_{15}$, subducting plate the stress state is more variable; between $x = 1600$ and 350 km, the surface stress increases from more compressive to somewhat tensile. The stresses are compressive in the ocean plate near the terrane and tensile in the continental fragment itself. In FMO$_{15}$, where the delamination of the upper crust does not develop, the stresses in the overriding plate are strongly compressive. In Fig. 7, the overriding plate is divided in three regions. Region C extends from the left-hand edge of the plate to the vertically projected left-hand side of the LVW, region B extends from the vertically projected left-hand side of the LVW to the vertically projected intersection of the slab with the Moho of the overriding plate; region A extends from there to the plate contact. Most of the channel models are largely in tension in region C in the open subduction setting. Also, tensile stresses dominate in region C of FMO$_{3}$ and FMOw$_{15}$. In the same region, the stress of FMO$_{15}$ is nearly zero, indicating a more compressive tectonic system. Region C in the land-locked basin setting is mostly in tension, irrespective of the plate contact type. In region B, compressional regime prevails for all the models. Fault models in the open subduction setting tend to show more compression than channel models. In region A, horizontal stresses show a positive excursion towards more tensile stresses in the fault models. Channel models in region A do not show such behaviour.

When we reduce the slab pull using the short slab model, the subducting plate experiences less tension than in the long slab models (CMO$_{15}$ and CMO$_{15}$ in Fig. 7a). Horizontal stresses in region C show to be rather insensitive to the slab pull for these models. The stress change in region B of Fig. 7(a) is largely brought by the fact that the slab is shorter in CMO$_{15}$, with the consequence that the upward projection of the slab tip occurs closer to the plate contact. A reduction of the slab pull in the fault models results in a more compressive oceanic subducting plate and region C (FMOw$_{15}$). Stresses at the continental terrane becomes neutral. A simultaneous decrease of slab pull and terrane crustal strength (FMOw$_{15}$) results in a similar stress in the surface oceanic plate as in the FMO$_{15}$. Within region C, these models show approximately the same surface stresses too. Differences between FMO$_{15}$ and FMOw$_{15}$ are substantial only within region B.

In the land-locked basin setting, the horizontal stress along the surface oceanic plate becomes more tensile than in the open subduction setting, for both channel and fault models. The same holds for region C of the overriding plate. The most relevant difference with respect to the open subduction is in region B, where the compressive regime is drastically reduced. Nevertheless, FM and in particular FML$_{15}$ are characterized by the most compressive stress in that area. Table 2 summarizes all the models used in the experiments.

### 4 Model Analysis

Our experiments are divided in two main categories: channel models (CM) and fault models (FM). In both groups, we vary some of the
characteristic parameters: the geometry of the incoming continental sliver, the tectonic setting, the rheology and the length of the sinking oceanic lithosphere.

In a previous study, we established that the nature of plate contact affects the response of the subduction process (De Franco et al. 2007). We found that CM are characterized by faster subduction velocities, less compression in the overriding plate, less dynamic subsidence of the upper plate, more plate-like behaviour and lower interplate pressure than FM. We refer to this paper for a detailed explanation of the physical process that causes the differences between subduction channel and fault. These results are also valid in the present study.

In the CM of the present study, the very nature of the subduction channel, with its weak material, reduces the coupling between the plates and interplate pressure, promoting in this way continental subduction, as already shown by Chemenda et al. (1996). As a consequence, the incoming continental fragment is subducted coherently with the oceanic lithosphere without perturbing the subduction process, at least during the initial stage of the collision. This is expressed by the fact that in the CMO, the subducting plate shows a steady state plate-like behaviour (solid and dashed black curves in Fig. 4). In CML, as a consequence of the boundary conditions, the slab moves faster than the horizontal part of the plate, but in such a way that the two velocities are constantly proportional to each other. This results in the fact that in both settings, the continental sliver is coherently subducted without localization of high strain rate in the slab (Figs 5a1 and a2 and A1a1 and a2).

On the other hand, FM result in a very different response. First of all, FM do not show plate-like behaviour for any kind of boundary conditions, and the horizontal part of the subducting plate moves always slower than part of the sinking slab (solid and dashed grey curves and black dotted curve in Fig. 4b); moreover, these differences increase with time. Such a response implies that the plate deforms with time. In all FM, the strain rate strongly differs from the cases of the corresponding CM-series. The magnitude of the strain rate is higher than in CM and localizes between −100 and −200 km depth. The slab visibly deforms (e.g. Figs 5a3, b3, a4, b4, a5, b5). These features are explained by two factors. First, the high interplate pressure and high normal coupling between lower and upper plate in the FM hampers the subduction of the continent, whereas the gravitational force in the slab pulls the plate down, causing deformation of the slab. Second, the strong flow generated in the wedge below the overriding plate (see De Franco et al. (2007)) contributes to deformation of the slab and promotes localization of strain rate. FMO15 shows the highest strain rate and the most pronounced deformation of the slab. This occurs because the continental fragment, with the steep geometry and the high-strength crust, completely locks the trench, preventing subduction of the incoming continent. Since in FMO15 the strain rate in the slab increases with time (Fig. 5b4 black curve) whereas the strain rate in the crust is nearly constant (Fig. 5b4 red curve), slab break-off will possibly take place.

A general result, true for all the tectonic settings, is that the geometry of the passive continental margin of the terrane does not strongly affect the response of CM. On the other hand, the slope inclination angle of the incoming passive margin plays a very important part in FM. Whereas in FMO3 and FML15, the continent starts to subduct (Figs 4a and 6a dashed grey curve), in FMO15 and FML15, the subduction stops (see Fig.4a and 6a black dotted). The surface oceanic plate surface decreases with time. When the crustal rheology of the continental fragment permits shear delamination (i.e. FMOw15 and FMLw15), the response of the models becomes similar to the one of FMO3 or FML3.

The progressive separation of the upper crust from the lower crust occurs only in the FM when the crustal strength at the contact between lower and upper crust is low with a value of about 30 MPa (e.g. Figs 4a and 6a, solid grey curve, and Figs 5a3 and A1a1 and a2). The upper crust moves with a different velocity and sometimes different direction than the lithospheric mantle. Through shear deformation, the upper crust remains at the surface, whereas the lithospheric mantle continues to subduct. The strong overriding plate and the high interplate pressure hamper the subduction of the upper crust, whereas the weak strength of the lower crust allows the separation of the two units. When delamination occurs, the horizontal state of stress of the overriding plate changes—the far-field plate (Fig. 7a region C) as well as the region immediately close to the plate contact (Fig. 7a region A) become tensile, whereas in FMO15 the state of stress in the overriding plate is neutral or compressive. In the subducting plate, compressive stresses decrease as soon as delamination takes place and subduction again starts to accommodate the convergence. In the land-locked basin setting, the state of stress of FMLw15 and FMLw15 is not much different since the total convergence is limited by the boundary conditions. In CM, delamination of the crust does not take place, for any geometry and strength of the incoming continental material. This is due to the fact that the channel flow facilitates subduction of the continental sliver and does not act as a chisel on the incoming continental material. Moreover, the lower interplate pressure helps making subduction of the continental margin possible.

Table 2. Summary of the models used in this study.

<table>
<thead>
<tr>
<th>Model name</th>
<th>Plate contact type</th>
<th>Boundary conditions</th>
<th>Continental slope angle</th>
</tr>
</thead>
<tbody>
<tr>
<td>CMO15</td>
<td>Channel model</td>
<td>Open subduction</td>
<td>3°</td>
</tr>
<tr>
<td>CMO15</td>
<td>Channel model</td>
<td>Open subduction</td>
<td>15°</td>
</tr>
<tr>
<td>CMLw15</td>
<td>Channel model</td>
<td>Land-locked basin</td>
<td>3°</td>
</tr>
<tr>
<td>CML15</td>
<td>Channel model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Open subduction</td>
<td>3°</td>
</tr>
<tr>
<td>FML15</td>
<td>Fault model</td>
<td>Open subduction</td>
<td>15°, weak crust</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°, weak crust</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°, weak crust</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
<tr>
<td>FMLw15</td>
<td>Fault model</td>
<td>Land-locked basin</td>
<td>15°</td>
</tr>
</tbody>
</table>
To understand the role of the slab pull during the initial stage of continental collision, we reduce the net pull force through the reduction of the slab length and keep the velocity on the right-hand side of the subducting plate equal to 5 cm yr$^{-1}$. In the CM, slab pull does not significantly affect the response of the model. In CMs, after a spin up period, the subduction velocity reaches the same value as in the model with a long slab (Fig. 4a red dashed curve). The velocity along the subducting plate becomes uniform with time (Fig. 4b red dashed curve), meaning that the incoming continental sliver is subducted with the oceanic lithosphere without perturbing the process. However, the horizontal stress at the surface domain becomes more compressive in the subducting plate and in region C of the overriding plate, whereas in region B it is less compressive (see Fig. 7a). The higher compressive stress in region C of the upper plate and in the subducting plate is caused by the reduction of the slab pull that increases the interplate pressure. The lower compressive stress in region B is due to the fact the tip of the slab does not lie below that region anymore, causing downwelling and compression in that part of the overriding plate. The slab velocity is fast, like in the long slab experiments, because the horizontal velocity applied on the right-hand side of the subducting plate is the same as in the long slab experiments and is strongly transmitted along the plate. The low viscosity channel reduces the transmission of such a velocity to the overriding plate. Moreover, the short length of the slab reduces the frictional forces in the mantle that opposes the sinking of the slab.

In FM with short slabs (FMs), the subduction velocity becomes slower than in the other FM, and the overriding and subducting plate are in a more compressive regime. This is due to the fact that the horizontal velocity applied at the right-hand side of the subducting plate is strongly transmitted through the fault to the overriding plate and that the interplate pressure increases (interplate pressure is inversely proportional to the slab pull). In FMOs15, where the delamination of the crust of the continental margin does not develop, the subduction velocity is reduced even more than in the FMO15. As a consequence, even though the net slab pull is less, the slab moves faster than the horizontal part of the plate (Fig. 4b blue dotted curve). The same holds for the case in which delamination develops (FMOsw15). We can conclude that a reduction of the slab pull does not affect the general response of the models.

In our models, we considered different tectonic settings in which the right-hand side of the subducting plate is pushed towards the trench and the left-hand side of the overriding plate is fixed, or in which there is no net convergence between subducting and overriding plate. In this way, we consider some important possible settings present in nature. From our results, we conclude that even though the tectonic setting can change some of the features of the models, the imprint due to the different plate contacts remains the same.

The results of our models are plotted as data points in Fig. 8. The black curves that divide the panels in the different mode domains are qualitatively drawn. In summary, on the basis of our numerical experiments, we support the initial idea that three modes of continental collision can develop:

1. Coherent subduction of the continental sliver. This mode is subdivided in plate-like and non-plate-like continental subduction.
2. Crustal delamination of the upper crust of the incoming continental margin. Subsequent continuation of subduction in a non-plate-like fashion.
3. Locking of the subduction zone and possibly leading to slab break-off.

The three modes are illustrated in Fig. 9. In the upper panel, CM$_{1}$/CM$_{15}$ are characterized by plate-like subduction of the incoming continental margin for any type of geometry and strength of the continental fragment; in FM$_{3}$, the continental material is coherently subducted but the velocity along the slab is not plate-like.

Figure 8. Qualitative–quantitative description of the influence of different parameters on the modes of continental collision. CM are represented by red symbols. FM with 15$^\circ$ slope inclination angle are represented by black symbols. FM with 15$^\circ$ slope inclination angle and a weak lower crust are represented by green symbols. FM with 3$^\circ$ slope inclination angle are represented by blue symbols.
and the slab deforms; in FM the passive margin locks the trench and does not subduct, causing even more pronounced deformation of the slab; in FMw the upper crust is separated from the lower crust in a shear mode and remains at the surface, whereas the lithospheric mantle descends in the asthenosphere in a non-plate-like fashion.

We conclude that the plate contact type is critical in controlling the different modes of continental collision during the initial stage of the process. The plate contact plays a more relevant role than the magnitude of slab pull and than the tectonic setting.

5 A LINK BETWEEN SHORT AND LONG TIMESCALE CONTINENTAL COLLISION MODELS

In previous studies, buoyancy (Ranalli et al. 2000; Sobouti & Arkani-Hamed 2002), rheological properties of the continental and oceanic plate (strength, temperature; Wong A Tong & Wortel 1997; van de Zedde & Wortel 2001; Sobouti & Arkani-Hamed 2002; Li & Liao 2002; Toussaint et al. 2004), subduction velocity (Davies & von Blanckenburg 1995; Li & Liao 2002; Toussaint et al. 2004; Burov & Yamato 2007) and interplate pressure (Chemenda et al. 1996) have been indicated as the main variables controlling the evolution of continental collision. In our study, we took a step in direction of establishing which are the parameters that dominate the behaviour of continental collision during the initial stage of the process, and we do not study the entire evolution of it. There are two main differences between our study and the previous ones: (1) the earlier studies did not take in consideration the nature of the plate contact, and (2) we do not analyse long timescale processes. Facenda et al. (2007) discussed the importance of having a weak rheological coupling between the plates to obtain one-sided asymmetrical subduction zone during continental collision. They changed the coupling between the plates increasing or decreasing the shear heating (that is governed by the convergence velocity) at the plate contact. In our models, however, shear coupling in both channel and fault models is very low or negligible. The fault and channel represent two physical different states of the subduction plate contact and are characterized by different properties (for more details see De Franco et al. 2007).

It is likely that some of the parameters that do not play an important role in our experiments, become relevant during a second stage of the process developing. In the following sections, we make some connections between our findings and the results of other studies.

5.1 Deep burial of continental material and slab break-off

In our CMO, continental lithosphere starts to subduct in a coherent and plate-like fashion, indicating the possibility of deep (> 30 km) subduction of the passive margin. Many numerical studies and geological data suggest that the entire continental lithosphere may be taken to even greater depths.

Ranalli et al. (2000) concluded that subduction stops when the total slab buoyancy, calculated from the summation of positive and negative density anomalies, is zero. The maximum depth reached by the continental material is a function of dip angle, geometry of sliver and convergence velocity. If the imposed velocity at the edge of the subducting plate continues to drive the plate, continental subduction can also carry on when the system is neutrally or slightly positively buoyant, as shown in our short slab experiments. This is in agreement with Regard et al. (2003), who suggested that subduction will terminate when the subducting system reaches a significant positive buoyancy. Facenda et al. (2007) also found that, if the plates are efficiently rheologically decoupled (analogue to our channel models), the slab continues to subduct even without external forcing due to the slab pull and the scraping off of the buoyant upper crust. These findings indicate that a small piece of continental crust can be taken to considerable depth, and that buoyancy is not very important until it reaches a certain threshold value. As a consequence, the amount of positive buoyancy of the continental crust is not a dominant parameter during the initial stage of collision. In our model, the width of the incoming continent is not crucial, whereas on a long timescale the volume of the subducted continental fragment controls the total amount of buoyancy. The depth to which continental crust can be subducted, coherently decreases as the crustal thickness of the subducting continental plate increases (van den Beukel 1992).

On the other hand, collision modes is conditioned by initial convergence velocity and the thermomehalogical state of the lithosphere (Burov & Yamato 2007; Toussaint et al. 2004). Burov & Yamato (2007) showed that continental subduction is possible only by strong and cold mantle lithosphere and fast initial convergence velocity.

Nevertheless, it is plausible that a relevant change in buoyancy due to the subducting crust may cause slab break-off. Eventually slab break-off is likely to occur in collision zones in which a passive continental margin is involved, since the increasing tectonic stresses generated by continued continental subduction makes extensional deformation inevitable and slab break-off probable. To estimate where and when the break-off takes place, Davies & von...
Blanckenburg (1995) compared the integrated strength of the slab with the variation in buoyancy that results in extensional force. When this force overcomes the integrated strength, slab break-off takes place. The extensional force is proportional to the subduction depth of the continental crust and is controlled by its composition. They found that the integrated strength is strongly affected by the convergence velocity. A slow subduction is likely to promote break-off and at shallower depth than for fast subduction; for a velocity of 1 cm yr$^{-1}$ break-off might occur between 50 and 120 km depth. Slow subduction convergence promotes failure of the slab in most of the continental collision models (e.g., Wong A Tong & Wortel 1997; Sobouti & Arkani-Hamed 2002; Toussaint et al. 2004; Burov & Yamato 2007). Gerya et al. (2004a) showed that slab break-off can be initiated in form of slab necking after cessation of active subduction. This supports our finding in which we found necking of the slab in our fault models, where subduction velocity is very small. The main difference between these two models is the timescale in which the phenomena takes place. In Gerya et al. (2004a) the necking starts at least after 6 Myr after a period of thermal weakening due to thermal diffusion. In our models the slab starts necking after 0.5–0.6 Myr and thermal diffusion is not active. The timescale difference may be due to the fact that the forcings in the models are different. Slab necking is only present in our fault models where the strong flow generated in the wedge below the overriding plate (see De Franco et al. 2007) contributes to deformation of the slab.

In summary, we have learned from the previous numerical studies that the parameters governing slab break-off are: strength/temperature of the slab; convergence velocity and buoyancy variation. In the light of our experiments another parameter must be added to this list: the nature of the plate contact.

5.2 Delamination

Another way in which subduction of the continental crust may evolve is through delamination of the entire crust or part of it. In our models, we excite delamination of the upper crust in that part of the model parameter space where the inclination angle of the continental slope is unusually steep, the strength of the lower crust is low and the plate contact is described by a fault. Delamination occurs through the removal of the upper crust in a shearing mode; the upper crust remains at the surface and subsequently is accreted to the upper plate. In nature, this process may take place in essentially two ways: (1) by terrane or block accretion or (2) by thrusting of the delaminated upper crust over the adjacent margin of the upper plate. The latter may be accompanied by intense deformation of the over-thrust units (Willet et al. 1993).

We note, however, that there are other ways in which delamination can take place. Bird (1978) proposed that the previously subducted oceanic lithosphere with its excess density could result in a large tension in the upper slab if it was prevented from sinking at its terminal velocity (Forsyth & Uyeda 1975). This force would tend to vertically separate the subcrustal lithosphere from the already subducted crust. This kind of delamination may generate a so-called nappe stack that consists of a series of slices of continental crust left at the plate contact by the incoming subducting plate.

The last process would require subduction of the continental crust to a certain depth like, for example, in our coherent subduction models (CM and FM3). Since from observations and model studies we know that the upper crust may reach a considerable depth (≥50 km) before being exhumed, we envisage that in models like CM and FM3 delamination may occur at a later stage than what we show in our models. Chemenda et al. (1996) and Toussaint et al. (2004) showed the feasibility of this process. After a period of subduction, the upper crust may separate from the rest of the plate and because of its positive buoyancy, it may return to the surface. This process depends on the rheology of the lower crust, which has to be relatively weak, and of course on the buoyancy. Burov & Yamato (2007) also agreed with the fact that a strong lower crust and a fast subduction velocity are necessary to drag the light upper crust to great depth. Ellis et al. (1999) investigated a small continental terrane entering the subduction zone after a period of sediment subduction. In their model (their experiment 5), the plate contact is described by a subduction channel with an initial thickness of 5 km. The continental terrane is dragged down beneath the upper plate. This finding is in agreement with our results, even though the rheological properties and the forces used in their experiment differ from ours. In their model, they investigate a larger amount of convergence than in our experiments: after 400 km convergence, the weak upper crust of the terrane separates from the lithosphere and creates a large fold nappe. Chemenda et al. (1996) introduced interplate pressure as a pertinent parameter in the process of continental crust subduction. If the interplate pressure is low, the continental crust can be subducted to a depth of about 200 km, before the delamination process starts, and relatively easily exhumed (see also Boutelier et al. 2004; Gerya et al. 2007). When the interplate pressure is high, failure of the crust occurs at shallower depth and exhumation is not so easy and not so fast. This finding has some similarity with our experiments: in our FMOW$_{15}$, where the interplate pressure is high and crustal strength weak, the continental crust fails, whereas the in CM, where the interplate pressure is low the crust is subducted. In FM, subduction of the upper crust is more difficult than in CM or does not take place at all. In conclusion, even though our models differ from Chemenda et al.'s (1996) analogous experiments, the physical process behind them is very similar. Chemenda et al. (1996) vary the interplate pressure through the variation of the slab pull; in our modelling, this is achieved through the variation of the plate contact type.

The previous studies pointed out, which are the main parameters governing the evolution of continental subduction. The range in which these parameters vary is wide; continental collision cannot be explained by a unique combination of parameters. With our models, we show that the initial phase of the process is important and may control what will happen at longer timescales. The plate contact type is one of the variables to take into account. During the initial stage of continental collision, the plate contact is more relevant than slab strength, subduction velocity and buoyancy. In the later stage of the process, these last variables become important.

6 DISCUSSION

6.1 Do subduction channel and fault plate contacts exist?

One of the questions arising from our models is whether it is possible to establish where, in nature, the plate contact is characterized by a fault or a channel. Several seismic observations have shown a low velocity layer at the plate interface, which in general is few km thick (Eberhart-Phillips & Martin 1999; Tsuru et al. 2002; Oncken et al. 2003; Abers 2005). This is interpreted in term of a subducting sediment channel. However, a subduction channel cannot be a proper description of all subduction zones. First, great subduction earthquakes suggest that the subduction interface cannot be weak everywhere (Davies & Brune 1971; Kanamori 1977; Ruff & Kanamori 1983; Tichelaar & Ruff 1993). Most interplate seismicity...
occurs in the depth range of 0–50 km (Tichelaar & Ruff 1993), implying that at least the shallow part of the subduction zone has a finite shear strength. Second, sediment supply varies from site to site, as indicated by accretionary wedge dimensions and erosive margin observations (Clift & Vannucchi 2004). This might suggest that when the amount of sediments is small, the subduction channel becomes so thin that a more appropriate description of the plate contact is by one or several faults. In general, when a continental margin approaches the trench, a high quantity of sediments arrives at the subduction zone, increasing the possibility of a channel-like plate contact. On the other hand, erosion, which produces sediments, is influenced by tectonics, topography and climate. These factors change in time and as a consequence the amount of sediment changes, resulting in the possibility that subduction plate contacts evolve from one type into the other.

We could speculate that some tectonic settings may facilitate the creation of a subduction channel, for example, in a land-locked basin setting in which it is easier to accumulate sediments. On the other hand, an open subduction setting, in which the subducting plate is pushed towards the trench, is likely to be represented by a fault. Unfortunately, we must conclude that for many subduction zones, relevant information on the nature of the plate contact is lacking so far.

### 6.2 Comparison between our models and natural systems

The application of our numerical results to a natural system is restricted by the simplifications we made in our models. However, these idealized convergent systems yield some expressions that are similar to those observed in some continental collision systems. For instance, examples of the type of delamination we produce in our models (block/terrane accretion) may be represented by the Chugach terrane delaminated and accreted to the cratonic North America (e.g. Coney et al. 1980; Clift et al. 2005) or the accretion of the Hanshan and Xilin Hot microcontinents, part of the Turkestan ocean, to the Donggqiyan arc. According to Kusky et al. (2007), this collision resulted in the formation of the North China Craton. An example of delamination and thrust accretion is the Qining–Dabie Shan orogenic belt east of the Tianlu fault. The upper crust of the South China block thrust over the crust of the North China block, followed by continuous subduction of the lower crust and upper mantle of the South China block underneath the North China block (Li 1994). A similar process occurs in the Alps: the European plate subducts, while the Adriatic plate acts as a chisel scraping off the upper from the lower crust (Schmid et al. 1996).

In a general sense, understanding the possible variations in convergent plate boundary evolution, as addressed in this study, helps interpreting observational data from such a setting. For example, in the nappe stacking of the continental basement of the Rhodope, there is ample evidence of ultra-high pressure metamorphism, suggesting that continental upper crust was dragged down to a depth of 70–80 km before being exhumed (Liati et al. 2002). At least 900 km of continental lower crust and upper mantle have been subducted, without resulting in break-off (van Hinsbergen et al. 2005). This region is characterized by extension in the backarke region. The CML seem to be a good setting to generate a Rhodope-type geological structure: the crust can be easily subducted and exhumed because of the low interplate pressure. The low interplate pressure helps both processes and promotes continuation of subduction. Moreover, in the land locked basin setting, the state of stress in the backarke region is compressive, but the compression is strongly reduced compared with the FM. We argue that an increase in channel thickness or a decrease in the viscosity value in the channel may result in extension in the backarc region.

The fact that most of the above cited regions are characterized by the presence of ultra-high pressure rocks—which generally indicates the presence of a subduction channel—may be explained by taking into consideration that the nature of the plate contact may change in time, going from a channel to a fault and vice versa.

Continental collision is highly 3-D: the size of the continent in the out-of-plane direction affects the response of the process. However, we can infer 3-D aspects from our 2-D models. For instance, if in our models we predict subduction of the incoming continental margin or delamination of the upper crust, we can then infer that if the out-of-plane dimension of the incoming continent is small compared with that of the subducting plate, the subduction of the buoyant material or the removal of the upper crust is facilitated.

### 7 Conclusions

From a consistent comparison of channel and fault models, we find:

1. Channel models always promote coherent and steady state subduction of the continental fragment, irrespective of the geometry and strength of the continental crust of the incoming continent. In fault models, coherent subduction of the incoming continental material occurs if the continental rise of the colliding terrane is gentle. In this case, the approaching continental sliver starts to subduct, and subduction is characterized by a non-plate-like behaviour and strong deformation of the slab.
2. Fault models result in locking of the trench and likely in subsequent slab break-off, if the margin of the terrane is steep and if the strength of its lower crust is high.
3. The combination of a strong plate contact (i.e. fault), a steep slope angle and a low crustal strength of the incoming continental margin results in efficient shear delamination of the upper crust. The remaining lower crust and lithospheric mantle of the continental fragment continue to subduct. The subducting plate is characterized by a non-plate-like behaviour and strong deformation of the slab.

In summary, our results indicate that the following types of responses can occur by solely varying the plate contact type:

1. Coherent subduction of the continental sliver. This mode is subdivided in two groups: plate-like and non-plate-like continental subduction.
2. Shear crustal delamination of the upper crust of the incoming continental margin; subsequent continuation of subduction in a non-plate-like fashion.
3. Locking of the subduction zone and possible development of slab break-off.

We conclude that initial stages of continental collision are strongly affected by whether the subduction contact is a fault or a channel. Neither the slab pull magnitude nor tectonic setting (in our case prescribed plate velocity of the subducting plate versus fixed subducting plate) is very important to the overall geodynamics at this stage. The plate contact type, along with the slope of the incoming passive margin and the rheology of the continent, decides whether the incoming crust (partially) overthrusts, subducts entirely or blocks the trench leading to subduction.

### Acknowledgments

Partial support for RDF and computational resources for this work were provided by the Netherlands Research Center for Integrated
Solid Earth Science (ISES). We would like to thank the anonymous reviewer and Taras Gerya for their many critical and constructive comments and corrections. We also thank Anna Maria Marotta and Serge Lallemand for the very useful suggestions that helped to improve the manuscript.

REFERENCES


Li, X.Z., 1994. Collision between the north and south China blocks - a crustal-detachment model for suturing in the region east of the Tanlu fault, Geology, 22(8), 739–742.


© 2008 The Authors, GJI
Journal compilation © 2008 RAS


APPENDIX A: MODELS FOR A LAND LOCKED BASIN SETTING

In this section, we show the strain rate distribution for the experiments in the land-locked basin setting in which there is no net convergence between the surface plates. We again use two different geometries to describe the incoming continental sliver: one with a continental slope inclination angle of 3° and a steeper one with a continental slope inclination angle of 15°. Figs A1(a1)–(a5) display the total strain rate distribution, whereas Figs A1(b1)–(b5) show the evolution of the average strain rate with time in two areas: solid black line corresponds to the solid circle in Figs A1(a1)–(a5), red curve corresponds to the average crustal strain rate (red circle in Figs A1a1–a5).

A1 Channel models (CM)

In this setup the plate contact is described by a low viscosity channel that extends to a depth of 100 km. The viscosity of the channel is \( \eta = 2 \times 10^{18} \) Pa s.

A1.1 Channel model with land-locked basin subduction boundary conditions and 3° continental slope inclination angle (CML3)

The strain rate distribution is displayed in Fig. A1(a1) at 0.64 Myr. The strain rate in the slab and in the continental crust of the incoming fragment has a small magnitude of about 1 \( \times 10^{15} \) s\(^{-1}\). The horizontal part of the subducting plate is characterized by a higher strain rate than in CMO (channel models with open subduction setting). In the entire model domain, the highest value of the strain rate is in the subduction channel and below the horizontal subducting plate. These two regions represent relative thin layers of rapidly deforming material. In Fig. A1(a1), the deformed geometry of the model shows that the continent has been dragged down with the subducting plate to a depth of about 12 km. In Fig. A1(b1), the average value of the strain rate shows a nearly stationary behaviour as a function of time with a value of about 1 \( \times 10^{-15} \) s\(^{-1}\). The same holds for the crustal strain rate. However, in this region the strain rate is higher than in the slab. This is a consequence of the fact that the crust deforms while entering the trench.

A1.2 Channel model with land-locked basin subduction boundary conditions and 15° continental slope inclination angle (CML15)

In this model, the continental slope of the margin is inclined at an angle of 15°. The different geometry of the sliver slightly affects the behaviour of the subduction process and the results are very similar to the previous model.

Like in CML3, the strain rate within the subducting plate is small (Fig. A1a2). Strain rate is high in the channel and beneath the oceanic surface plate. The deformed geometry of the model displays that the incoming continental fragment has been subducted to a depth of about 12 km. In Fig. A1(b2) the average value of the strain rate shows a nearly stationary behaviour. The black curve in Fig. A1(b2) displays a response very similar to the previous model. In the crust, the strain rate magnitude is slightly higher than in the slab. The fact that the average strain rate in the crust is higher than in CML3 suggests that the continental crust experiences more deformation than in the previous model.

We performed an experiment in which the lower crust of the continental margin is characterized by low strength of 30 MPa (CMLw15). Because the results are very similar to those the CML15, we decided not to show them.

A2 Fault Models (FM)

In this model the plate contact is described by a frictionless fault that extends to a depth of 100 km.

A2.1 Fault model with land-locked basin subduction boundary conditions and 3° continental slope inclination angle (FML3)

Fig. A1(a3) displays the total strain rate distribution at 0.64 Myr. The strain rate that develops in the slab and in the overriding plate is higher than that in the CML. The slab significantly deforms at a depth of about 150 km, where the strain rate is localized, but less than in FML3. The deformed geometry of the domain shows that the incoming continental crust is subducted to a depth of about 4 km. The depth reached within the integration time by the subducting continent is shallower than in the CML. The overriding plate is characterized by high strain rate and marked downwelling. In Fig. A1(b3) the black and the red curves show that the average strain rate in the slab and in the crust of the incoming fragment increases with time, reaching higher values than in the previous models. Strain rate is high beneath the oceanic surface plate.

A2.2 Fault model with land-locked basin subduction boundary conditions and 15° continental slope inclination angle (FML15)

The strain rate in the slab and in the overriding plate is higher than in the CML and in FML3 (Fig. A1a4). From the deformed geometry of the model, we note that the incoming continent locks the trench and does not subduct with the oceanic material. The slab deforms more than in FML3 at a depth of about 150 km, but less than in FMO15. The strain rate is higher than in FML3. Fig. A1(b4) shows the average strain rate in more detail: the strain rate in the slab increases with time more than in FML3, reaching a maximum value of about 0.6 \( \times 10^{-14} \) s\(^{-1}\). In the crust the strain rate is stationary and lower than in FML3. Like in FML3, the strain rate is high beneath the oceanic surface plate.

A2.3 Fault model with land-locked basin subduction boundary conditions and 15° continental slope inclination angle and weak crustal layer (FMLw15)

The strain rate distribution that develops in the slab and in the overriding plate is very similar to FML3 (Fig. A1a5). The strain
rate maximizes in the slab at a depth of about 150 km; at that depth the slab is deformed. A high strain rate region is visible in the crust of the continental margin. This is the expression of the shear delamination process in which the upper crust separates from the lower one. The lower crust continues to subduct reaching a depth of about 4 km, during the modelled period. The detailed representation of the strain rate in Fig. A1(b5) indicates that the average strain rate in the slab increases with time but less than in FML15. In the crust, the strain rate is one order of magnitude higher than in the slab and higher than in all other land-locked basin models.
Figure A1. (Continued.)