Numerical modelling of spontaneous slab breakoff dynamics during continental collision

CYRILL BAUMANN1*, TARAS V. GERYA1,2 & JAMES A. D. CONNOLLY1

1Department of Geosciences, Swiss Federal Institute of Technology (ETH-Zurich), CH-8092 Zurich, Switzerland
2Geology Department, Moscow State University, 119899 Moscow, Russia

*Corresponding author (e-mail: baumancy@ethz.ch)

Abstract: Slab detachment or breakoff is directly associated with phenomena like morphological orogenesis, occurrence of earthquakes and magmatism. At depth the detachment process is slow and characterized by viscous rheology, whereas closer to the surface the process is relatively fast and plastic. Using a 2D mantle model 1500 km deep and 4000 km wide we investigated, with finite-difference and marker-in-cell numerical techniques, the impact of slab age, convergence rate and phase transitions on the viscous mode of slab detachment. In contrast to previous studies exploring simplified breakoff models in which the blockage responsible for inducing breakoff is kinematically prescribed, we constructed a fully dynamic coupled petrological–thermomechanical model of viscous slab breakoff. In this model, forced subduction of a 700 km-long oceanic plate was followed by collision of two continental plates and spontaneous slab blocking resulting from the buoyancy of the continental crust once it had been subducted to a depth of 100–124 km. Typically, five phases of model development can be distinguished: (a) oceanic slab subduction and bending; (b) continental collision initiation followed by the spontaneous slab blocking, thermal relaxation and unbending – in experiments with old oceanic plates in this phase slab roll-back occurs; (c) slab stretching and necking; (d) slab breakoff and accelerated sinking; and (e) post-breakoff relaxation.

Our experiments confirm a correlation between slab age and the time of spontaneous viscous breakoff as previously identified in simplified breakoff models. The results also demonstrate a non-linear dependence of the duration of the breakoff event on slab age: a positive correlation being characteristic of young (<50 Ma) slabs while for older slabs the correlation is negative. The increasing duration of the breakoff with slab age in young slabs is attributed to the slab thermal thickness, which increases both the slab thermal relaxation time and duration of the necking process. In older slabs this tendency is counteracted by negative slab buoyancy, which generates higher stresses that facilitate slab necking and breakoff. A prediction from our breakoff models is that the olivine – wadsleyite transition plays an important role in localizing viscous slab breakoff at depths of 410–510 km due to the buoyancy effects of the transition.

In addition to multiple geophysical, geological and geochemical investigations acknowledging slab detachment as a possible explanation for observations in various regions (Andrews & Billen 2009 and references therein), analytical models and laboratory and numerical experiments have been undertaken to characterize the breakoff process (e.g. Davies & von Blanckenburg 1995; Larsen et al. 1995; Yoshioka et al. 1995; Yoshioka & Wortel 1995; Wong A Ton & Wortel 1997; Pysklywec et al. 2000; Chemenda et al. 2000; Buiter et al. 2002; Gerya et al. 2004b; Faccenda et al. 2006, 2008; Mishin et al. 2008; Ueda et al. 2008; Zlotnik et al. 2008; Andrews & Billen 2009).

Recent thermomechanical numerical models indicate two modes of detachment (Andrews & Billen 2009): (1) deep viscous detachment that is characteristic of strong slabs (maximum yield strength ≥500 MPa), and is controlled by thermal...
relaxation (heating) of the slab and subsequent thermomechanical necking in dislocation creep regime (Gerya et al. 2004b; Faccenda et al. 2008; Zlotnik et al. 2008); and (2) relatively fast, shallow plastic detachment characteristic of weak slabs (maximum yield strength \( \leq 300 \) MPa) and controlled by plastic necking of the slab (Mishin et al. 2008; Ueda et al. 2008; Andrews & Billen 2009). Andrews & Billen (2009) demonstrated that the time before the onset of viscous detachment increases with slab age, indicating that detachment time is controlled by the thickness and integrated stiffness of the thermally relaxing slabs (Gerya et al. 2004b). In contrast, the plastic detachment time is independent of age and is controlled by the rate of flow of the surrounding mantle (Andrews & Billen 2009).

Two-dimensional dynamic models of viscous detachment have shown that, following the cessation of subduction, necking of the slab and detachment occur after 8–30 Ma at a depth of 150–400 km (Gerya et al. 2004b; Faccenda et al. 2008; Andrews & Billen 2009). In numerical experiments with viscous dissipation included (Gerya et al. 2004b) shear heating in the necking area causes slab breakoff to occur 8% faster than in models that neglect viscous dissipation (Gerya et al. 2004b). Therefore, the slab necking process is not purely mechanical. Zlotnik et al. (2008) modelled the influence of mantle phase transitions on the viscous slab detachment and demonstrated that the olivine–spinel transition facilitates the detachment by increasing the negative buoyancy of the hanging slab.

Thermomechanical models of plastic detachment indicate shallower (Andrews & Billen 2009) and even near-surface (Mishin et al. 2008; Ueda et al. 2008) depths of breakoff. The timing and location of plastic failure is controlled by the dynamics of the loading of the subducting plate by negatively buoyant slab during or shortly after the active phase of convergence. Such models have also shown that, in the case of multiple plates, shallow slab breakoff can affect plate dynamics causing processes such as subduction flip and re-initiation (Mishin et al. 2008).

Thermomechanical numerical studies of breakoff (Gerya et al. 2004b; Zlotnik et al. 2008; Andrews & Billen 2009) explored a simplified static breakoff model with two fixed oceanic plates and a kinematically prescribed interruption of subduction. Thus, no dynamic thermomechanical model of this process accounting for the presence of the continental plates and cessation of subduction (e.g. Chemenda et al. 2000; Faccenda et al. 2008) has been studied. The absence of such an effort leaves a gap in our understanding of dynamics and controls of breakoff. In particular, prescribed slab blocking does not allow for the proper balance between the slab strength, geometry (length, inclination) and negative buoyancy, and may create unrealistic geodynamic situations (e.g. heavy slabs that are too weak to remain coherent during subduction). The simplified models also fail to reproduce the thermal and lithological structure of the collisional orogen that forms in response to continental subduction (e.g. Chemenda et al. 2000; Burov et al. 2001; Faccenda et al. 2008). Therefore, in this paper we investigate the dynamics, geometry and duration of viscous detachment of oceanic slabs after cessation of active oceanic–continental subduction due to collision between two continental plates by means of a high-resolution 2D numerical study. We will account for the effects of phase transformations as well as pressure-, temperature- and strain-rate-dependent rheology of the mantle and the subducted oceanic crust. Compared to previous studies dealing with thermomechanical detachment experiments (e.g. Gerya et al. 2004b; Zlotnik et al. 2008; Andrews & Billen 2009), our petrological–thermomechanical model is fully dynamic and considers the presence of the colliding continental plates as causing subduction blocking and breakoff.

Model set-up and governing equations

Initial configuration

To investigate slab breakoff we developed a 2D, coupled petrological–thermomechanical numerical model of oceanic–continental subduction followed by collision (Fig. 1) using the I2VIS code (Gerya & Yuen 2003a). The spatial co-ordinate frame of the model is 4000 \( \times \) 1500 km. The non-uniform 774 \( \times \) 106 rectangular grid is designed with a resolution varying from 2 \( \times \) 2 km in the studied subduction–collision–breakoff zone to 30 \( \times \) 50 km far from it (Fig. 1b). Lithological structure of the model is represented by a grid of around 3 million randomly distributed active Lagrangian markers used for advecting various material properties and temperature. The density of the marker distribution varies from 1 marker km\(^{-2}\) in the high-resolution zone to 0.25 marker km\(^{-2}\) far from it. The oceanic crust consists of a 3 km-thick basaltic layer underlain by a 5 km-thick gabbroic layer. The continental crust consists of a 35 km-thick granodioritic layer.

The velocity boundary conditions are free slip at all boundaries except the lower boundary of the model domain, which is permeable in the vertical direction (Gerya et al. 2008b). An infinite-iike external free slip condition \((\partial v_z/\partial z = 0, \partial^2 v_z/\partial z^2 = v_z/\Delta_{\text{external}}, where v_z and \Delta_z are, respectively, horizontal and vertical velocity components) along the bottom implies that free slip condition \((\partial v_z/\partial z = 0, v_z = 0)\) is satisfied at an external boundary located at a certain depth \(\Delta_{\text{external}}\).
below the actual lower boundary of the model ($\Delta z_{\text{external}} = 1500$ km for the present study).

Similar to the usual free slip condition, external free slip allows global conservation of mass in the computational domain. The top surface of the crust is calculated dynamically as an internal free surface by using initially a 10–12 km-thick deformable top layer (10 km over the continents and 12 km over the oceanic plate) with low viscosity ($10^{19}$ Pa s) and density ($1$ kg m$^{-3}$) for the atmosphere, above $z = 11$ km sea-water level, $1000$ kg m$^{-3}$ for the hydrosphere, below $z = 11$ km sea-water level, where $z$ is the vertical co-ordinate taken downwards from the top of the model). The validity of the weak layer approach for approximating the free surface has recently been tested and proven (Schmelling et al. 2008) with the use of a large variety of numerical techniques (including our I2VIS code) and comparison with analogue modelling. The interface between this weak layer and the top of the oceanic crust deforms spontaneously and is treated as an internal erosion–sedimentation surface that evolves according to the transport equation solved at each time step (Gerya & Yuen 2003b):

$$\frac{\partial z_{\text{es}}}{\partial t} = v_z - v_x \frac{\partial z_{\text{es}}}{\partial x} - v_k + v_e$$

where $z_{\text{es}}$ is a vertical position of the surface as a function of the horizontal distance $x$; $v_z$ and $v_x$ are the vertical and horizontal components of material velocity vector at the surface; $v_k$ and $v_e$ are, respectively, sedimentation and erosion rates. As demonstrated by Pysklywec (2006) and Gerya et al. (2008b) in the case of subduction and collision variations in erosion–sedimentation rates may significantly affect crustal mass flux and consequently alter the behaviour of the crust–mantle interface. In the present paper we used moderate constant gross-scale erosion and sedimentation rates applicable for continental collision zones (Vance et al. 2003; Gerya et al. 2008b), which correspond to the relation:

$v_k = 0$ mm year$^{-1}$, $v_e = 1$ mm year$^{-1}$ when $z < 9$ km (i.e. for $\geq 2$ km elevation above the sea-water level)
\( v_s = 1 \text{ mm year}^{-1}, \ v_a = 0 \text{ mm year}^{-1} \) when \( z > 11 \text{ km} \) (i.e. below the sea-water level).

The influences of erosion–sedimentation rates on the geometry of collisional orogens in the post-subduction collision model similar to the one explored in our study are discussed by Gerya et al. (2008a).

Subduction is prescribed by the total convergence rate \( R_T = R_R + R_L \), where \( R_R \) and \( R_L \) are locally imposed constant velocities for the right and the left plates, respectively. Moving continental plates have finite width and are detached from the model boundaries (Fig. 2). This convergence condition was applied for 700 km of forced subduction of the oceanic plate. After this time the model was allowed to evolve spontaneously. The subduction area is initiated by a 5–50 km-wide weak zone of hydrated mantle that cuts across the mantle lithosphere from the bottom of the crust down to 190 km depth. Taking into account the critical role of water for subduction initiation (Regenauer-Lieb et al. 2001), this zone is characterized by wet olivine rheology (Ranalli 1995) and a low plastic strength limit of 1 MPa (Table 1). With the advance subduction, this zone is replaced by weak upper oceanic crust, which is also characterized by low plastic strength (Table 1). This device implies that high-pressure fluids are present along the slab interface during subduction (e.g. Sobolev & Babeyko 2005; Gerya et al. 2008a).

The initial geotherm for the oceanic lithosphere is defined by a half-space cooling model (e.g. Turcotte & Schubert 2002; Fowler 2005) for the prescribed lithospheric age (Table 2). The initial thermal structure of the continental lithosphere corresponds to a typical continental geotherm (e.g. Turcotte & Schubert 2002): 0°C at the surface and 1345°C at 140 km depth. A gradual transition from the oceanic to the continental geotherm is prescribed over a 200 km-wide section on either side of the ocean–continent boundary (Fig. 1a). The geotherm for the mantle below the lithosphere is defined by a near-adiabatic temperature gradient of 0.46 K km\(^{-1}\). Thermal boundary conditions are 0°C at the upper boundary of the model; no-flux conditions exist at the vertical boundaries. An infinite-like external constant temperature condition along the lower boundary is implemented using the following limitation: \( \frac{\partial T}{\partial z} = \frac{(T_{\text{external}} - T)}{\Delta T_{\text{external}}} \), where \( T_{\text{external}} = 2140°C \) is the temperature prescribed at the external boundary.

**Petrological model**

The stable mineralogy and physical properties for the various lithologies were computed using Perple_X (Connolly 2005) with free energy minimization as a function of pressure and temperature. For this purpose we adopted the Mie–Gruneisen formulation of Stixrude & Bukowinski (1990), with the parameterization of Stixrude & Lithgow-Bertelloni (2005) augmented for lower-mantle phases as described by Khan et al. (2006). This parameterization limits the chemical model to the CaO–FeO–MgO–Al\(_2\)O\(_3\)–SiO\(_2\). The mantle is assumed to have a pyrolitic composition, for which the thermodynamic parameterization is adequate to reproduce the expected lower-mantle phase relations (Mishin et al. 2008). Application of the thermodynamic model to the basaltic and gabbroic composition of the oceanic crust is more problematic because phase equilibrium experiments (Irifune & Ringwood 1993; Irifune et al. 1994; Hirose & Fei 2002; Ono et al. 2005) suggest the existence of several high-pressure phases that are not included in our parameterization. In addition, the CaO–FeO–MgO–Al\(_2\)O\(_3\)–SiO\(_2\) model excludes volatile oxides, notably K\(_2\)O and Na\(_2\)O, that are more significant in the subducted oceanic crust; as a consequence our model is likely to overestimate the basalt–pyrolite density contrast. To calibrate this effect we found that experimentally derived density estimates for K\(_2\)O–Na\(_2\)O–CaO–FeO–MgO–Al\(_2\)O\(_3\)–SiO\(_2\) (Irifune & Ringwood 1993; Ono et al. 2005) were 1.7–2.3% below those calculated here. Accordingly, neutral buoyancy in the Earth’s interior most probably corresponds to conditions at which our basalt–pyrolite density contrast is 1.02 ± 0.03. Our models therefore overestimate the role of density-induced slab-pull forces in subduction.

**Thermomechanical model**

The momentum, continuity and temperature equations for the 2D creeping flow, accounting for both thermal and chemical buoyancy, are solved using the I2VIS code based on conservative finite differences and non-diffusive marker-in-cell techniques (Gerya & Yuen 2003a). Conservation of mass is prescribed by the incompressible continuity equation:

\[
\frac{\partial v_x}{\partial x} + \frac{\partial v_z}{\partial z} = 0.
\]

The 2D Stokes’ equations for creeping flow take the form:

\[
\frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xz}}{\partial z} = \frac{\partial P}{\partial x} \\
\frac{\partial \sigma_{xz}}{\partial x} + \frac{\partial \sigma_{zz}}{\partial z} = \frac{\partial P}{\partial z} - gp(T, P, C, M).
\]
Fig. 2. Reference model development (model cydd in Table 2). (a) Oceanic slab subduction and bending of the diving plate, high strength of the cold deeply subducted mantle lithosphere prevents its unbending and causes strong slab curvature (Gerya et al. 2008b); (b) thermal relaxation and unbending of the slab; (c) slab stretching and necking; (d) and slab breakoff and sinking. White lines are isotherms taken from 100 °C with 200 °C steps.
Table 1. Material properties* used in 2D numerical experiments

<table>
<thead>
<tr>
<th>Material</th>
<th>$H_r$ ($\mu$W m$^{-2}$)</th>
<th>Standard density ($\rho_0$) (kg m$^{-3}$)</th>
<th>Thermal conductivity ($W$ m$^{-1}$ K$^{-1}$)</th>
<th>Rheology</th>
<th>$P$–$T$ conditions of wet solidus</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper continental crust</td>
<td>1.0</td>
<td>2700 (solid) 2400 (molten)</td>
<td>0.64 + $\frac{807}{T + 77}$</td>
<td>Wet quartzite flow law, $c = 1$ MPa, sin($\varphi$) = 0.15</td>
<td>889 + $\frac{17900}{P + 54} + \frac{20200}{(P + 54)^2}$ at $P &lt; 1200$ MPa, 831 + 0.06$P$ at $P &gt; 1200$ MPa</td>
</tr>
<tr>
<td>Lower continental crust</td>
<td>0.25</td>
<td>2800 (solid) 2400 (molten)</td>
<td>0.64 + $\frac{807}{T + 77}$</td>
<td>Plagioclase (An$_{75}$) flow law, $c = 1$ MPa, sin($\varphi$) = 0.15</td>
<td>889 + $\frac{17900}{P + 54} + \frac{20200}{(P + 54)^2}$ at $P &lt; 1200$ MPa, 831 + 0.06$P$ at $P &gt; 1200$ MPa</td>
</tr>
<tr>
<td>Upper oceanic crust (altered basalt)</td>
<td>0.25</td>
<td>3000 (solid) 2900 (molten)</td>
<td>1.18 + $\frac{474}{T + 77}$</td>
<td>Wet quartzite flow law, $c = 1$ MPa, sin($\varphi$) = 0</td>
<td>973 − $\frac{70400}{P + 354} + \frac{7780000}{(P + 354)^2}$ at $P &lt; 1600$ MPa, 935 + 0.0035$P$ + 0.0000062$P^2$ at $P &gt; 1600$ MPa</td>
</tr>
<tr>
<td>Lower oceanic crust (gabbro)</td>
<td>0.25</td>
<td>3000 (solid) 2900 (molten)</td>
<td>1.18 + $\frac{474}{T + 77}$</td>
<td>Plagioclase (An$_{75}$) flow law, $c = 1$ MPa, sin($\varphi$) = 0.6</td>
<td>973 − $\frac{70400}{P + 354} + \frac{7780000}{(P + 354)^2}$ at $P &lt; 1600$ MPa, 935 + 0.0035$P$ + 0.0000062$P^2$ at $P &gt; 1600$ MPa</td>
</tr>
<tr>
<td>Dry mantle (lithospheric and asthenospheric)</td>
<td>0.022</td>
<td>3300</td>
<td>0.73 + $\frac{1293}{T + 77}$</td>
<td>Dry olivine flow law, $c = 1$ MPa, sin($\varphi$) = 0.6</td>
<td>935 + 0.0035$P$ + 0.0000062$P^2$ at $P &gt; 1600$ MPa</td>
</tr>
<tr>
<td>Mantle in the subduction initiation zone</td>
<td>0.022</td>
<td>3200</td>
<td>0.73 + $\frac{1293}{T + 77}$</td>
<td>Wet olivine flow law, $c = 1$ MPa, sin($\varphi$) = 0</td>
<td>−</td>
</tr>
</tbody>
</table>

**References**
- Turcotte & Schubert 2002
- Turcotte & Schubert 2002
- Clauser & Huenges 1995
- Chopra & Paterson 1981; Brace & Kohlstedt 1980; Ranalli 1995
- Schmidt & Poli 1998; Poli & Schmidt 2002

*Other material properties (density, heat capacity, latent heating, thermal expansion) are computed from the petrological model by Gibbs energy minimization; for models without phase transitions and for the continental crust in all models the following properties are used: $C_p = 1000$ J kg$^{-1}$ K$^{-1}$, $\rho = \rho_0(1 - \alpha(T - T_0)) \times [1 + \beta(P - P_0)]$, where $\alpha = 3 \times 10^{-5}$ K$^{-1}$ is thermal expansion, $\beta = 1 \times 10^{-5}$ MPa$^{-1}$ is compressibility, $\rho_0$ is standard density at $T_0 = 298$ K and $P_0 = 0.1$ MPa.
The density \( \rho (T, P, C, M) \) depends explicitly on the temperature \( T \), the pressure \( P \), the rock composition \( C \) and the mineralogy \( M \).

The Lagrangian temperature equation includes latent heat effects of phase transitions in the crust and mantle, and is formulated as (Gerya & Yuen 2003a):

\[
\rho C_p \left( \frac{DT}{Dt} \right) = -\frac{\partial q_x}{\partial x} + H_x + H_T + H_s
\]

where \( D/Dt \) is the substantive time derivative; \( x \) and \( z \) denote the horizontal and vertical co-ordinates, respectively; \( \sigma_{xx}, \sigma_{zz}, \sigma_{xz} \) are the components of the deviatoric stress tensor; \( k_{xx}, k_{zz}, k_{xz} \) are the components of the strain rate tensor; \( P \) is the pressure; \( T \) is the temperature; \( q_x \) and \( q_z \) are the heat fluxes; \( \rho \) is the density; \( k(T, c) \) is the thermal conductivity; \( C_p \) is the effective isobaric heat capacity, that is, incorporating latent heat; \( H \) is rock enthalpy; \( H_s \), \( H_T \) and \( H_s \) denote radioactive heat production, the energetic effect of isothermal (de)compression and shear heating, respectively. To account for physical effects of phase transitions on the dynamics of breakoff (Zlotnik et al. 2008) we used the coupled petrological–thermomechanical numerical modelling approach described in detail by Gerya et al. (2004a, 2006). In this approach all local rock properties, including effective density, isobaric heat capacity, thermal expansion, adiabatic and latent heating, are calculated at every time step based on Gibbs energy minimization.

Viscosity dependent on strain rate, pressure and temperature is defined in terms of deformation invariants (Ranalli 1995) as:

\[
\eta_{\text{creep}} = (\dot{\varepsilon}_n)^{(1-n)/n} F(A_D)^{-1/n} \times \exp\left(\frac{E_a + V_s P}{nRT}\right)
\]

where \( \dot{\varepsilon}_n = 1/2 \varepsilon \dot{\varepsilon} \) is the second invariant of the strain rate tensor, and \( A_D, E_a, V_s \) and \( n \) are experimentally determined flow law parameters (Table 1). \( F \) is a dimensionless coefficient depending on the type of experiments on which the flow law is based. For example:

\[
F = \frac{2(1-n)/n}{3(1+n)/2n} \quad \text{for triaxial compression}
\]

\[
F = 2^{(1-2n)/n} \quad \text{for simple shear}
\]

The ductile rheology is combined with a brittle–plastic rheology to yield an effective viscous–plastic rheology. For this purpose the Mohr–Coulomb yield criterion (e.g. Ranalli 1995) is implemented by limiting creep viscosity, \( \eta_{\text{creep}} \), as follows:

\[
\eta_{\text{creep}} \leq \frac{c + P \sin(\varphi)}{(4\dot{\varepsilon}_n)^{1/2}}
\]

where \( P \) is the complete (non-lithostatic) pressure (i.e. the mean stress), \( c \) is the cohesion (residual strength at \( P = 0 \)) and \( \varphi \) is effective internal friction angle (Table 1). Assuming high pore fluid pressure in hydrated rocks (e.g. Gerya et al. 2008a), the upper oceanic crust (basalts, sediments) was
characterized by $c = 1 \text{ MPa}$, $\sin \varphi = 0$, a choice that effectively decouples the slabs. As the focus of our paper is viscous breakoff we assigned a high plastic strength to the mantle (sin $\varphi = 0.6$; Brace & Kohlstedt 1980). Consequently, the plastic strength of subducted slabs in our experiments is always greater than the critical value (500 MPa) reported by Andrews & Billen (2009), and slab detachment occurs in the purely viscous mode by temperature- and stress-activated dislocation creep.

The effective viscosity, $\eta$, of molten crustal rocks at $P-T$ conditions above their wet solidus (Table 1) was lowered to $10^{19}$ Pa s. $10^{19}$ and $10^{26}$ Pa s are the lower and upper cut values for viscosity of all types of rocks in our numerical experiments.

**Modelling results**

Ten experiments (Table 2) were performed to study the influence of: (1) slab age; (2) plate rate; and (3) phase transitions on the dynamics of subduction. Numerical experiments were performed using the ETH-Zurich Brutus cluster.

**Reference models development**

In our two reference models (models cydc and cydd in Table 2) we distinguish five characteristic phases (Figs 2–4).

(a) Oceanic slab subduction and bending; this stage initiates with prescribed convergence rate but thereafter evolves self-consistently (Fig. 2a).

(b) Continental collision followed by slab blocking (due to the positive buoyancy of continental crust once it has been subducted to 100–125 km depth), thermal relaxation and unbending; this stage consists of intense heat exchange between the subducted slab and the surrounding asthenospheric mantle (Figs 2b, 3a & 4a); this stage is often manifest by slab decoupling from the overriding plate and asthenospheric upwelling due to the slab roll-back under its own weight (Fig. 4b, c). These latter processes are driven by the slab pull and are most pronounced in experiments with old (dense) oceanic plates ($\geq 80$ Ma, cf. Figs 2c & 3b). The bent slab is unbending under its own weight (Fig. 2b), which is also facilitated by the slab roll-back;

(c) Slab stretching and necking (Fig. 2c). Stretching of the slab is directly related to its thermal softening during the previous stage. The former ocean–continent boundary is subducted together with the stretching slab toward depths of more than 200 km; continental crust, however, continuously detaches from the slab and remains at shallower ($\leq 100$ km) levels beneath the orogen (Fig 2c).

(d) Viscous slab breakoff and rapid sinking (Figs 2d, 3b & 4c). This stage is similar to previous models of breakoff with prescribed slab blocking (Gerya et al. 2004b; Andrews & Billen 2009), except that the geometry of the collisional orogen formed at the plate contact evolves naturally with time allowing, in particular, for notable post-breakoff extension (cf. Fig. 4c, d) driven by the positive buoyancy of previously subducted continental crust.

(e) Post-breakoff relaxation; this stage is manifest by thermal and topographical relaxation of the orogen associated with exhumation and melting of previously subducted continental and oceanic crust located below the extending plate interface (cf. pink partially molten crustal rocks in Fig. 4d). Within 10–20 Ma from the breakoff event 1–4 km of surface uplift occurs in a 200–300 km-wide area above the plate contact (cf. Fig. 4c, d), which is also comparable with that observed in previous numerical models (Buiter et al. 2002; Gerya et al. 2004b; Andrews & Billen 2009).

**Influence of slab age**

Experiments on the function of the age of the subducting oceanic plate (Table 3) show that the necking process occurs later in older slabs (Fig. 5). This behaviour is mainly caused by the variations in temperature structure of the subducting oceanic plate (Gerya et al. 2004b) and is characteristic of the viscous mode of breakoff (Andreas & Billen 2009) investigated in our study. In older (i.e. colder) slabs more time is necessary for thermal relaxation to increase temperature and reduce the slab viscosity. Since breakoff process is a direct consequence of the slab necking, the breakoff also correlates with age (Fig. 6).

The correlation between slab age and breakoff duration [i.e. the time elapsed between the beginning of slab necking (Fig. 2b) and separation (Fig. 2d)] is more complex (Fig. 7). Two domains are visible in Figure 7: a domain characteristic of young slabs ($< 50$ Ma) and an area for mature slabs ($> 50$ Ma). For young slabs breakoff duration increases with slab age and is related to the influence of the growing slab thermal thickness and strength, which slow the necking process. This trend inverts in old slabs owing to the increasing slab negative buoyancy, higher stresses and stronger thermo-mechanical feedback, respectively; effects that accelerate necking and breakoff (Gerya et al. 2004b).

**Influence of relative plate rates**

There is no significant effect on the time required for the onset of necking and breakoff when one plate
Fig. 3. Evolution of geometry and temperature of the detaching slab (Model cydc in Table 2). White lines are isotherms taken from 100 °C with 200 °C steps. (a) Slab stretching and beginning of necking process. (b) Slab breakoff and accelerated sinking.
Fig. 4. Development of collision zone geometry during spontaneous slab breakoff process (Model cydc in Table 2). (a) Transition from subduction to collision; (b) beginning of necking process, slab roll-back associated with the opening of the asthenospheric window below the collision zone occurs; (c) time of breakoff; (d) post-breakoff time associated with exhumation and melting of deeply subducted crust. The black colour at the bottom of the subducted crust corresponds to partially molten crustal rocks. White lines are isotherms taken from 100 °C with 200 °C steps.
velocity is adjusted in order to keep the total rate of convergence at 10 cm per year\(^{-1}\) (Table 4). This behaviour is to be expected because thermal structure of the slab is mainly affected by the total rate of convergence and not by relative plate velocities. In contrast, breakoff duration and depth (Table 4) are affected by relative plate rates. In particular, in our numerical experiments the duration of breakoff (Table 4) is shortest (5.3–5.7 Ma) when the velocities of both plates are similar in magnitude. The duration of breakoff is maximized (11.3 Ma) when the magnitude of the velocity of the overriding plate (10 cm year\(^{-1}\)) is much greater than that of the subducting plate (0 cm year\(^{-1}\)). Breakoff depth uniformly decreases with decreasing relative velocity of the subducting oceanic plate (Table 4). These variations in breakoff depth and duration are most probably caused by changing slab angle and curvature, which depend on the relative plate

*Table 3. Influences of slab age*

<table>
<thead>
<tr>
<th>Name of the numerical experiment</th>
<th>Slab age (Ma)</th>
<th>Beginning of the necking process(^1) (Ma)</th>
<th>Breakoff time(^2) (Ma)</th>
<th>Breakoff duration(^3) (Ma)</th>
<th>Breakoff depth(^4) (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>cyda</td>
<td>20</td>
<td>2.9708</td>
<td>6.1320</td>
<td>3.1612</td>
<td>406.780</td>
</tr>
<tr>
<td>cydb</td>
<td>40</td>
<td>12.6545</td>
<td>19.3235</td>
<td>6.6690</td>
<td>455.085</td>
</tr>
<tr>
<td>cydc</td>
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<td>22.1207</td>
<td>28.5153</td>
<td>6.3946</td>
<td>444.915</td>
</tr>
<tr>
<td>cydd</td>
<td>80</td>
<td>33.2673</td>
<td>38.5978</td>
<td>5.3305</td>
<td>508.475</td>
</tr>
<tr>
<td>cyde</td>
<td>100</td>
<td>32.8012</td>
<td>35.4913</td>
<td>2.6901</td>
<td>432.203</td>
</tr>
</tbody>
</table>

\(^1\)With beginning of the necking process meaning an observable necking of the subducted plate (Fig. 2c).
\(^2\)Breakoff time means the moment of complete separation of the detached slab from the rest of the plate (Fig. 2d).
\(^3\)The breakoff duration is the difference between the beginning of observable necking and complete slab separation.
\(^4\)The depth where the separation and breakoff takes place (the deepest point of the separated plate isotherm).
velocities; for example, if the overriding plate is fast, the slab angle is shallower (van Hunen et al. 2000). In the model with the fastest overriding plate velocity (10 cm year$^{-1}$, model cyar in Tables 2 & 4) the slab is flattened at a depth of 660 km (e.g. Mishin et al. 2008), which is not observed in our reference model with a similar slab age (model cydd in Tables 2 & 4; Fig. 2). In contrast, in the model with the fastest subducting plate velocity (10 cm year$^{-1}$, model cyaq in Tables 2 & 4) the subducting slab is steeper and has stronger curvature than in the reference model (Fig. 2a).

Influence of phase transitions

In our experiments a fully coupled petrological–thermomechanical model based on Gibbs free energy minimization is used that accounts for both continuous and discontinuous phase transformations in the mantle and subducted oceanic crust (cf. Mishin et al. 2008 for details). The lowered temperatures in the descending oceanic plate cause it to have a different mineralogy and higher density than the surrounding mantle. Most significant, additional downwards body force on the descending slab is caused by the phase transition of olivine into spinel at a depth of about 410 km, which accelerates viscous detachment dynamics (Zlotnik et al. 2008). Further negative buoyancy effects may come from eclogitization of the subducted oceanic crust.

We ran two additional experiments without phase transitions for various slab ages (models cydh and cydi in Table 2). Results for pairs of models with and without phase transformations are compared in Table 5 and Figure 8. As follows from our comparison, phase transitions notably accelerate the breakoff process and increase the depth of slab breakoff (Fig. 8). These influences are obviously caused by the stronger negative buoyancy of the subducted slab affected by phase transitions. Consequently, in models with phase transitions the characteristic density contrast of the slab with the surrounding asthenospheric mantle is 100–200 kg m$^{-3}$, which is two–three times larger than the density contrast in models without phase transitions (50–70 kg m$^{-3}$).

### Discussion and conclusions

In previous numerical studies (Gerya et al. 2004b; Zlotnik et al. 2008; Andrews & Billen 2009) the cessation of active subduction that is ultimately responsible for breakoff has been kinematically prescribed. The asset of our model is that it simulates both subduction and collision as dynamic processes. Specifically, our experiments model slab detachment induced by slab pull that develops owing to a continental collision after 700 km of oceanic plate subduction. During this latter stage the plate convergence rate evolves realistically (e.g. Chemenda et al. 2000; Faccenda et al. 2008). The chief limitation of our model is that the rheology includes neither elasticity (e.g. Buiter et al. 2002) nor Peierls plasticity (e.g. Kameyama et al. 1999). These deformation mechanisms may influence the transition from viscous to plastic detachment.

### Table 4. Influences of plate rate

<table>
<thead>
<tr>
<th>Name of the numerical experiment</th>
<th>Slab age (Ma)</th>
<th>Plate rates (cm year$^{-1}$)</th>
<th>Beginning of the necking process (Ma)</th>
<th>Breakoff time (Ma)</th>
<th>Breakoff duration (Ma)</th>
<th>Breakoff depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>cydd</td>
<td>80</td>
<td>+05 –05</td>
<td>33.2673</td>
<td>38.5978</td>
<td>5.3305</td>
<td>508.475</td>
</tr>
<tr>
<td>cyaq</td>
<td>80</td>
<td>+10 00</td>
<td>29.0926</td>
<td>35.7112</td>
<td>6.61860</td>
<td>562.500</td>
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<tr>
<td>cyar</td>
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<td>00 –10</td>
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<td>43.2067</td>
<td>11.3376</td>
<td>460.169</td>
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<tr>
<td>cyas</td>
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<td>+03 –07</td>
<td>33.1033</td>
<td>38.4074</td>
<td>5.30410</td>
<td>432.203</td>
</tr>
</tbody>
</table>

### Table 5. Influences of phase transitions

<table>
<thead>
<tr>
<th>Numerical experiment</th>
<th>Beginning of the necking (Ma)</th>
<th>Time of breakoff (Ma)</th>
<th>Breakoff duration (Ma)</th>
<th>Breakoff depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>cydc</td>
<td>22.1207</td>
<td>28.5153</td>
<td>6.3946</td>
<td>444.915</td>
</tr>
<tr>
<td>cydi (no phase transitions)</td>
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<td>34.7640</td>
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</tr>
<tr>
<td>cydh</td>
<td>12.6545</td>
<td>19.3235</td>
<td>06.6690</td>
<td>457.627</td>
</tr>
<tr>
<td>cydh (no phase transitions)</td>
<td>16.5099</td>
<td>36.2608</td>
<td>19.7509</td>
<td>317.790</td>
</tr>
</tbody>
</table>
(Andrews & Billen 2009) and the development of topography (Buiter et al. 2002). Additional limitations are the simplified rock chemistry and that the petrological model does not account for metastable phase assemblages which might occur in the subducted gabbroic crust (e.g. van Hunen et al. 2000) and lithospheric mantle (e.g. Schmelling et al. 1998). Thus, negative buoyancy of slabs in our experiments may be overestimated. In natural subduction settings significant along-strike

Fig. 8. Influence of phase transitions onto the depth of slab breakoff. (a) Model cydc compared with Model cydi. (b) Model cydb compared with Model cydh. The snapshots for different models are taken for different moments in time when breakoff occurs in each model (Table 5). White lines are isotherms taken from 100 °C with 200 °C steps.
variations in the dynamics of breakoff processes are inferred from seismic and topographical data (e.g. Wortel & Spakman 1992, 2000). Such 3D effects cannot be accounted for by the present 2D model.

Five evolutionary phases occur in most of our experiments: (a) oceanic slab subduction and bending; (b) continental collision initiation followed by a cessation of subduction, thermal relaxation and unbending; (c) slab stretching and necking; (d) slab breakoff and sinking; and (e) post-breakoff relaxation.

The development of our dynamic thermomechanical model of continental collision has a number of similarities with previous analogue and numerical collision models that take into account significant buoyancy effects arising from continental crust subduction (e.g. Chemenda et al. 2000; Burov et al. 2001; Facenna et al. 2008). In particular, our results confirm a positive correlation between the oceanic slab age and the time of necking and breakoff found in previous thermomechanical studies of the viscous mode of detachment, exploring models without phase transformations (Gerya et al. 2003; Andrews & Billen 2009). However, previous work did not report the non-linear dependence of breakoff duration on the slab age: a positive correlation being characteristic of relatively young (<50 Ma) slabs, while for older slabs the correlation is negative. We also found that at a constant convergence rate, a relatively high velocity of subducting oceanic plate increases the depth of breakoff, most probably because of the geometric consequences of the subducting plate (van Hunen et al. 2000) that influences the dynamics of phase transformations affecting slab buoyancy.

A prediction from our breakoff models is that the olivine–wadsleyite transition plays a role in localizing viscous slab breakoff at depths of 410–510 km owing to the buoyancy effects of the transition (e.g. Schmelling et al. 1998; Zlotnik et al. 2008). Without phase transitions, the slab is significantly less dense. The breakoff in this case is shallower, occurs later and lasts longer.

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References


