Analogue modelling of continental collision: Influence of plate coupling on mantle lithosphere subduction, crustal deformation and surface topography

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ABSTRACT

The role of the plate boundary in a continental collisional setting is investigated by lithospheric-scale analogue models. Key variables in this study are the degree of coupling at the plate interface and along the Moho of the lower plate as well as the geometry of the plate contact. They control the onset of intra plate deformation, orogenic architecture, amount of mantle lithosphere subduction and basin development. Increasing coupling at the plate interface through time leads to intra plate deformation by thickening and gentle folding and influences surface uplift and subsidence above the plate interface. The transition from a mechanically decoupled plate boundary with a significant amount of mantle lithosphere subduction towards stronger plate coupling resulting in intra plate deformation and topography development can be recorded in for instance the Caucasus, the Colombian Cordillera, the Pyrenees and the Alps. Thickening of the lower crust as portrayed for the Western Alps does not demand a strong, frictional-type behavior of the lower crust, but can also be the consequence of ductile processes.

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1. Introduction

Subduction of oceanic lithosphere from its initiation to the accumulation in slab graveyards at the core–mantle boundary is highly investigated over the last decades (e.g. Cloetingh et al., 1982; Toth and Gurnis, 1998; Regenauer-Lieb et al., 2001; Hall et al., 2003; Jarvis and Lowman, 2005; Kito et al., 2008). Their clear present day characteristics in terms of surface topography, volcanism, and earthquakes distribution make oceanic lithosphere subduction accessible for research. On the other hand, less is known about the operating processes in the transition towards continental lithosphere subduction during continental collision.

Ampferer (1906) and Laubscher (1977) recognized that restoration of contractional belts lead often to an unbalanced amount of upper crustal shortening relative to the underlying (basement) layers. Therefore, the existence of decollement systems was required in order to form lower crustal “roots” or to create the so called “floating” orogens. More recently, continental lithosphere subduction has been considered as a continuation of oceanic lithosphere subduction except that buoyant continental crust resists subduction and detaches from the underlying mantle lithosphere (Willett et al., 1993; Ellis, 1996). However, evidence that not all of the continental crust is detached and accreted but at least part of it can get subducted to depths beyond 100 km is provided by exhumed ultra high pressure rocks (e.g. Krabbendam and Dewey, 1998; Faure et al., 2003; Brun and Faccenna, 2008).

2. Modelling of subduction and collision: a concise summary

Several numerical studies have attempted to model continental subduction followed by the exhumation of crustal rock (e.g. Toussaint et al., 2004a; Burov and Yamato, 2008; De Franco et al., 2008a; Faccenda et al., 2008). Burov and Yamato (2008) have shown that high plate velocities (~5–3 cm/yr) and low Moho temperatures (<550 °C) favor subduction of the entire lithosphere, while intermediate Moho temperatures (550–650 °C) in combination with a lower crustal strength lead to crust–mantle decoupling. Other parameters...
affecting the evolution of continental collision zones are convergence rate, lithosphere rheology, buoyancy and inter plate pressure (e.g. Sobouti and Arkani-Hamed, 2002; Toussaint et al., 2004b). In addition, De Franco et al. (2008a,b) pointed out that the most relevant parameter during the initial stage of continental collision is the geometry and (de)coupling along the plate contact. In that sense the plate contact is in an early stage decisive whether the lithosphere will entirely subduct, delaminate, or will not subduct at all (De Franco et al., 2008a). To obtain subduction Tagawa et al. (2007) suggested that weakening of the plate boundary is even more important than the rheology of the lithosphere.

The amount of plate coupling depends on the geometry of the plate contact, the amount of water-rich sediments, the presence of a slab pull force caused by still attached oceanic lithosphere, or shear heating; and hence on the plate velocity (e.g. Toussaint et al., 2004a; Facenda et al., 2009). Plate coupling is spatially and temporally variable and can increase after the consumption of water-rich sediments, by slowing down of convergence, arrival of buoyant material, or a change in plate contact geometry. An increase in coupling changes the style of lithosphere deformation by the transmission of far field stresses, which can result in buckling and regional uplift of both plates and the orogenic wedge (Ziegler et al., 2002; Willingshofer and Sokoutis, 2009).

The role of the plate boundary and its development during continental collision has been studied in both numerical and physical modelling studies (e.g. Sokoutis et al., 2005; De Franco et al., 2008a,b; Willingshofer and Sokoutis, 2009). In most models the plate contact was represented by a predefined weak zone dipping 45° with respect to direction of shortening (Hassani and Jongmans, 1997; Chemenda et al., 2001; Regard et al., 2003; Willingshofer et al., 2005; Tagawa et al., 2007; De Franco et al., 2008a).

In this study we investigate the influence of (a) plate- and (b) lower plate crust–mantle coupling on the mode of deformation of the different rheological layers. Special emphasis has been put on monitoring surface topography development in order to be able to link deep to shallow processes.

3. Experimental design

3.1. Analogue materials and model set-up

The presented models consist of three different layers overlying a viscous fluid mixture of polytungstate and glycerol representing the asthenosphere. From bottom to top these layers are composed of: strong silicon putty, weak silicon putty, and feldspar sand; they are used as analogues for the ductile mantle lithosphere, the ductile lower crust, and the brittle upper crust, respectively (Fig. 1). The viscous layers are mixtures of polydimethyl-siloxane polymer (PDMS) with barium sulphate showing slightly non-Newtonian behavior (Table 1). With the low shortening rate used (0.5 cm/h) the silicon mixtures deform in a ductile manner, while dry feldspar sand is a Mohr–Coulomb-type material. The density structure of the model allows for subduction as the mantle lithosphere (1.5 g/cm³) is denser than the asthenosphere (1.46 g/cm³).

The models were performed in a transparent tank with one moving wall, which induces convergence of the plates (Fig. 1). The horizontal dimensions for all models were 40 × 36 cm, and the thickness of the layers varied between the models (Table 2). However, thickness variations between the experiments were not the main parameter to investigate, but was a result of different length scales. In order to simulate subduction of the lithosphere we implemented a weak zone of Rhodorsil-type silicone, which separated the mantle lithospheres of the converging plates. This weak zone represents a weak plate interface or subduction channel (Fig. 1). Additionally, in experiments 4 and 5 the weak zone continued along the Moho of the lower plate for 6 cm to allow full crust–mantle decoupling. During the experiments top and side view pictures were taken every 30 min. The development of the topography was accurately monitored by a 3-D laser with DEM outputs.

3.2. Scaling

The physical parameters in the models need to be accurately balanced with respect to nature in order to derive meaningful interpretations (e.g. Weijermars and Schmeling, 1986; Davy and Cobbold, 1991; Brun, 2002). Dynamic and geometric scaling between model and nature can be achieved when respecting the stress-scale factor:

\[ \sigma^* = \rho^* g^* L^* \]  

where \( \sigma \) refers to stress, \( \rho \) to density, \( g \) to gravitational acceleration and \( L \) to the length scale. The asterisk refers to the ratio between model and nature. Since in our study \( g^* \) equals 1 and the densities of the used silicon putties (\(-1500 \, \text{kg} \, \text{m}^{-3}\)) and rocks in nature (2300–3000 \, \text{kg} \, \text{m}^{-3}) are in the same order of magnitude Eq. (1) can be simplified to: \( \sigma^* = L^* \). Considering the physical properties of the used materials and assumed values for their equivalents in nature (Table 2) we obtain a stress ratio of \( 3.3 \times 10^{-7} \), which implies a geometric scaling of 1 cm in the model to 30 km in nature.

To fulfill dynamical similarities of the ductile layers the viscous forces depending mainly on viscosity and strain rate need to be balanced between nature and the model. This balancing of the involved parameters can be done by calculating the Ramberg number \( (R_m) \) (Weijermars and Schmeling, 1986), which represents the ratio of gravitational to viscous forces:

\[ R_m = \frac{\rho V h^2}{\eta V} \]  

where \( h \), \( \eta \), and \( V \) are the ductile layer thickness, the viscosity, and the compression rate, respectively. By adopting natural strain rates, viscosities, and thicknesses, we calculated according to \( R_m \) the model viscosities and obtain a shortening rate of 0.5 cm/h (Table 2).

3.3. Modelling assumptions and simplifications

The existence of a thermal gradient determining largely the strength of the lithosphere is not incorporated within the models and hence the rheological behavior of our layers is solely influenced by material composition and strain rate. For this reason the strength of the viscous layers remains constant at varying depths, while in nature the strength of the ductile layers decreases exponentially (Ranalli, 1995, 1997). Consequently, temperature driven time-space variability of rheology during shortening cannot be accounted for, and we assume an initial homogenous plate rheology which did not suffer pre-extensional thinning or earlier thickening affecting the bulk strength of the lithosphere. As such, only the mantle lithosphere with a significant amount of strength is considered, which is slightly thinner compared to its natural equivalent (e.g. Willingshofer et al., 2005).

The models are deformed by a constant shortening rate, which does not respond to stress variations related to e.g. mountain build-up, thickening of the viscous layers, or the presence/absence of a slab pull force. In this study stresses are transmitted horizontally stemming from the advancing wall and convection related drag at the base of the lithosphere is neglected.

We assume that the use of one moving wall instead of two has no significant effect on the deformation pattern, and whether the upper plate or the lower plate borders the moving wall should not influence the stress and strain patterns.

Furthermore, natural recovery processes such as erosion and sedimentation are not taken into account.
3.3.1. Plate contacts in nature and models

To restore the initial plate boundary within an orogen is a difficult task requiring detailed geological mapping and high resolution (tele)seismic profiling. From a global perspective we can robustly distinguish between orogenic cycles following oceanic lithosphere subduction, and orogenic episodes without oceanic lithosphere involved. As an example, the formation of the Alps and also the Himalaya was preceded by subduction of the Tethys Ocean, while the Pyrenees and the Caucasus are mostly interpreted as inverted continental basins (Roure, 2008 and references therein). Both scenarios imply that the plate boundary is already formed during an earlier extensional phase, and that these locations of weakness are subsequently undergoing reactivation. During convergence the plate boundary can become further weakened by multiple processes, such as shear heating, water release and partial melting due to dehydration and/or metamorphism, and heat production by radiogenic sediments (e.g. Pawley, 1994; Tagawa et al., 2007; Faccenda et al., 2008).

Fig. 1. Cartoon showing the experimental set-up within a plexi-glass tank containing a three-layer lithosphere overlying a low viscosity asthenosphere. The different set-ups with respect to the geometry, thickness and extent of the weak zone in the lower crust refer to the models central zone. Black arrow indicates the direction of shortening.

Table 1
Physical properties relevant for scaling of the experimental materials.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Analogue material</th>
<th>Density $\rho$ (kg/m$^3$)</th>
<th>Coefficient of friction $\mu$</th>
<th>Viscosity $\eta$ (Pa s)</th>
<th>Power $n$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper crust</td>
<td>Feldspar sand</td>
<td>1300</td>
<td>0.7</td>
<td>35 Pa (cohesion)</td>
<td></td>
</tr>
<tr>
<td>Lower crust</td>
<td>Silicon mix 1 (PDMS)</td>
<td>1390</td>
<td></td>
<td>4.8 $\times$ 10$^4$</td>
<td></td>
</tr>
<tr>
<td>Mantle lithosphere</td>
<td>Silicon mix 2 (PDMS)</td>
<td>1500</td>
<td>1.2 $\times$ 10$^3$</td>
<td>1.9</td>
<td></td>
</tr>
<tr>
<td>Asthenosphere</td>
<td>Sodium polytungstate solution</td>
<td>1450</td>
<td>1.2</td>
<td>1.7</td>
<td></td>
</tr>
<tr>
<td>Weak zone</td>
<td>Silicon mix 3 (Rhodorsil gum)</td>
<td>1100</td>
<td>1.7 $\times$ 10$^4$</td>
<td>1.1</td>
<td></td>
</tr>
</tbody>
</table>
Together with the scaled shortening velocities these values are used to balance the Ramberg number \( R_m \) between the model and the natural analogue.

addition, the shear zone is maintained by far field stresses generated by the attachment of an earlier subducted oceanic slab (e.g. Toussaint et al., 2004a).

Far field stresses within the model generated by the moving wall are mostly accommodated within the much thicker ductile layers. Therefore, we limited our predefined weak plate interface to the mantle lithosphere and to keep the influence on the evolving crustal architecture as low as possible. The implementation of a weak lower crust close to the plate boundary within the lower plate in experiments 4 and 5 (Fig. 1) simulates the effect of strong crust–mantle decoupling. Such conditions can be explained by a relative high thermal gradient (25 °C/km) resulting in Moho temperatures in the order of 750 °C and hence deformation by flow of the lower crust (Bird, 1979; Toussaint et al., 2004b). Lateral variations in the thermal gradient might be related to earlier phases of rifting (Pyrenean case) or mountain building like in the Alps where Late Cretaceous stacking of Austroalpine units resulted in regional metamorphism and resetting of its thermal tectonic age. Since we aimed for one-sided continental lithosphere subduction, (sensu Gerya et al. (2008)), we limited the existence of a weak lower crust to the subducting plate.

The thickness of the weak plate contact is another important variable, which has been addressed by different authors (e.g. Gerya and Stockhert, 2002; De Franco et al., 2008b). Numerical models by De Franco et al. (2008a,b) distinguish between “fault-type” and “channel-type” plate contacts with thicknesses up to 6 km. Low velocity layers appearing on seismic sections at the plate interfaces support channel thicknesses of a few kilometers (e.g. Tsuru et al., 2002; Abers, 2005). Also the presence of ultra high pressure rocks in several plate bounding regions is in favor of a channel along which crustal blocks can be exhumed after subduction. However, depending on the availability of lubricants and with the assumption that sediments play an important role, the plate contact zone can be very narrow, along which stress can be built up as indicated by interplate seismicity in the upper crust (0–50 km) (Tichelaar and Ruff, 1993). In numerical models, specific values of friction coefficients can be assigned to the predefined weak plate contact (e.g. Hassan and Jongmans, 1997; Tagawa et al., 2007), which we consider in our study as the equivalent of a certain thickness and viscosity of the weak zone (Table 1). The thickness of the weak zone (0.5 cm in Exp. 3 and 5) is exaggerated with respect to nature (15 cm) to compensate for the fact that no weak material is added to the system, e.g. by incoming sediment. However, continental subduction will cease at a certain moment in nature as well. This can be caused by a shortage of lubricants, a slowing down of plate convergence resulting in less shear heating, the arrival of strong coupled buoyant crust, or by removing the slab pull force by slab break-off (e.g. von Blanckenburg and Davies, 1995; Ranalli, 2000; Burov and Yamato, 2008; Faccenda et al., 2008).

### 4. Modelling results

Within this section the results of the experiments will be briefly discussed and the reader is referred to Figs. 2–6 showing the final top view, a cross-section and the topographic evolution of each experiment. The following subsections are categorized according to the main changing parameters: the orientation and the extent of the predefined weak zone. Thickness variations are integrated within the subsections, and a comparison between all the experiments will be given at the end of this section. Finally, interpretations of the results together with an integration of previously published work are presented in the discussion.

#### 4.1. Experiment with a vertical plate boundary

**4.1.1. Experiment 1**

The first pop-up structure appeared in the upper crust above the weak zone at 4% bulk shortening (% BS) and remained active until the end of deformation, i.e. 12% BS (Fig. 2). Progressively several laterally discontinuous pop-up structures developed on both plates propagating mainly away from the weak zone. As it appears from the cross-sections very gentle folding of the competent upper mantle at wavelength of 8 cm (~240 km in nature) occurred. Synclines of the mantle folds coincide often with pop-up structures in the upper crust. Folding was accompanied by vertical thickening of up to 30% of the viscous layers. In contrast to the other experiments no displacement occurred along the vertical weak zone boundary. The surface scans show the development of local topography related to the upper crustal pop-up’s together with a continuous mean surface uplift with ongoing shortening as response to crustal thickening (Fig. 2c).

**4.2. Experiments with an inclined plate boundary**

**4.2.1. Experiment 2**

A large fore thrust (1) developed after 2% BS above the weak plate interface, which remained active during the entire experiment (Fig. 3a). With ongoing shortening little displacement occurred along small fore- and back thrusts on the lower plate directly in front of the plate contact. Contemporarily, small and curved structures forming a discontinuous pop-up structure evolved on the interior of the upper plate. After 13% BS continuous migration of the major thrust zone caused the underthrusting of some small thrusts on the lower plate, and new thrusts were formed in front of the plate contact. Other compressive structures started to deform the interior of the foreland plate, but remained small until the end of the experiment (19% BS) (Fig. 3a).

Although no initial vertical decoupling zone existed between the lower crusts of both plates, a significant amount of vertical displacement occurred along thrust 1 resulting in 1.3 cm (~40 km in nature) offset of the Moho between the plates (Fig. 3b). As a consequence of strong horizontal coupling between the crustal layers the upper crust of the lower plate was partly overridden by the upper pate. The mantle of the upper plate was folded on a wavelength of ~4.4 cm (~130 km in nature)
a

b

C

Topographic evolution

Bulk shortening: Start 5% 10% 14%
Topographic evolution

Upper plate
Ramp anticline
Lower plate

First Pop-up
Basin

Bulk shortening: start 5% 10% 17.5%
and a high amplitude ramp anticline formed above the plate boundary. Additionally, the viscous layers were thickened in the range of 20–30% albeit with some lateral variations. On the upper plate thrusts 5 and 6 are probably related to a synformal deflection of the Moho. The laser scans monitor topography related to the early-stage deformation in the overriding plate and the formation of a ramp anticline near the suture zone (Fig. 3c). Throughout the models run this structure remained the highest topographic domain with a steep gradient towards the partly overridden adjacent foreland basin. Progressive deepening of the basin resulted in the formation of a slightly elevated flexural fore-bulge on the foreland plate (Fig. 3c). During the final stage of shortening the interior of both plates was uplifted with the formation of upper crustal thrusts and the accentuation of fold amplitudes.

4.2.2. Experiment 3

Localized deformation commenced after 2.5% BS along thrust 1, which developed above the weak zone interface perpendicular to the direction of shortening (Fig. 4a). Ongoing shortening resulted in a steep topographic gradient between the upper and lower plates, and after 5% BS deformation occurred along a back-thrust (2) located on the upper plate as well. From this moment on mainly small-scale structures started to develop within the interior of the upper plate, of which some finally merged to form an anastomosing pop-up structure (thrusts 3 and 4). At the same time, short-lived, foreland propagating, upper crustal structures formed before finally getting rotated and/or overridden by the overriding plate (e.g. thrusts 5 and 6).

Within the cross-sections, and similar to experiment 2, thrust 1 can be extrapolated throughout the entire lithosphere defining the plate boundary with a Moho offset of at least 2 cm (~60 km in nature) (Fig. 4b). The upper- and lower crust remained coupled within the lower plate and was overridden by upper plate derived mantle in the suture zone. The development of thrusts 3 to 6 was strongly governed by folding of the underlying ductile layers. The pop-up (thrusts 3 and 4) on the upper plate is rooted into a synform of the upper mantle, which was folded at a wavelength of 8 cm (~240 km in nature). Thrusts 5 and 6 originated from the inflection points of the fold and belong to a set of small-scale accommodation structures.

The digital elevation models document the initiation of topography directly along the plate contact with positive relief on the overriding plate and a topographic low on the lower plate. Ongoing shortening resulted in topography on the plate interiors with a fault governed relief with steep gradients overlying a plateau on the upper plate, but the models highest point is located along the hinge of the folded foreland plate (Fig. 4c). Moreover, it can be observed from the topographic profiles that deepening of the foreland basin and growth of the ramp anticline diminished already after 10% BS (Fig. 4c) and topographic growth was essentially confined to the lower plate until the end of the experiment.

4.3. Experiments with an inclined plate boundary and a weak lower crust

4.3.1. Experiment 4

A direct response to shortening was the formation of a pop-up structure above the weak suture zone. With further shortening the pop-up became part of the upper plate and accommodated most deformation along its fore thrust (1) (Figs. 5a and b). After 5% BS a second pop-up structure evolved above the weak zone/lower crust transition on the lower plate. Subsequently, three generations of fore thrusts related to the second pop-up were active throughout the model run. Finally, after 19% BS the interior of the lower plate started to deform by local brittle deformation together with long-wavelength folding of the viscous layers.

The cross-sections reveal that the first crustal pop-up is located on the upper plate and no coherent lithospheric fault, such as in the previous experiments, can be identified (Fig. 5b). As a result of horizontal decoupling the upper crust of the lower plate remained at relative shallow depths. On the lower plate the lower crust together with the weak zone material has thickened up to 30% meanwhile the mantle lithosphere delaminated from the crust. The second upper crustal pop-up borders this zone of lower crustal thickening and developed further onto the lower plate as a series of fore thrusts (4–6) (Fig. 5). Both pop-up structures are cored by up-thrusted lower crustal material leading to local wedging and crustal thickening. Remarkably, lower crust thickening above the pre-defined suture zone is only minor suggesting that stresses were transferred into the lower plate and localized deformation along the weak- to strong lower crust transition. Displacement along the plate contact was 2.5 cm (~75 km in nature) measured from the subducted slab, and evolved in the smearing out of the weak zone material.

From the analysis of topographic data it can be observed that a narrow basin developed on the foreland side of the first pop-up, but with the formation of the second pop-up on the lower plate this basin became incorporated into the orogen (Fig. 5c). After ~10% BS the second pop-up was elevated, meanwhile the frontal part of the upper plate and the first pop-up subsided. From this moment onward high topography remained restricted to the second pop-up and to a large-scale fold hinge on the upper plate. After 15% BS a narrow foreland basin developed in front of the second pop-up, together with modest topography on the interior of the lower plate.

4.3.2. Experiment 5

As in experiment 4, the first response to shortening was the formation of a pop-up structure above the weak suture zone, and was followed after 5% BS by the development of a second pop-up structure on the lower plate (Fig. 6a). Both pop-ups remained active throughout duration of the model. From the cross-sections it can be observed that ongoing deformation lead to the coalescence of both pop-ups into a single symmetrical pop-up structure, which was rooted and cored by thickened lower plate derived weak zone material (Fig. 6b). In addition, shortening was largely accommodated by lower plate crustal deformation at the suture zone and subduction of mantle lithosphere and some weak lower crust. Subduction of the mantle occurred where the upper crust was initially underlain by a weak lower crust, whereas in a subsequent stage the subduction of stronger lower crust resulted in the burial of highly coupled upper crust down to 55 km. The final subducted slab length was 7.5 cm, which is 75% of the total amount of shortening and corresponds to 225 km in nature.

The topographic profile (Fig. 6c) shows that between 5% and 10% BS a symmetric basin formed between two pop-ups, which got subsequently narrowed and uplifted together with both pop-ups. At the same time adjacent asymmetric foreland basins deepened on both plates. After 15% BS the orogen underwent a phase of rapid topographic growth, coinciding with deepening of the foreland-type basin and folding of the interior of the lower plate. Finally, between 20% and 25% BS the entire orogen subsided and shifted towards the subducting plate. The foreland and retroforeland-type basins narrowed and deepened, respectively. Besides lithosphere-scale folding resulting in uplift of the plate interiors no intra plate crustal deformation occurred.

5. Interpretation and comparison between the experimental results

In this section the most important similarities and differences between the experiments will be briefly discussed. This comparison is primarily qualitative because small variations in crustal thicknesses

*Fig. 3. Final top view and cross-sections of experiment 2 after 19% of bulk shortening (a, b). For layer colors see caption of Fig. 2. c) Topographic evolution profiles taken from the center of the model parallel to shortening direction (arrow) after different amounts of bulk shortening. Vertical exaggeration by factor 3.*
(Table 2) and the amount of shortening make a quantitative comparison difficult.

In general, all experiments with the exception of experiment 1 illustrate two distinctive phases: 1) localized deformation and topography development limited to the plate contact, 2) regional deformation affecting the entire model space leading to uplift of the plate interior and the accentuation or impediment of pre-existing fold and basin geometries. Analyzing the results, it became clear that the transition between both phases is strongly influenced by temporal changes in the degree of plate coupling and the presence of a weak lower crust in the lower plate.

5.1. Plate boundary deformation and orogenic wedge formation

In all experiments deformation initiates along the plate contact by the formation of a pop-up structure in the upper crust. For the experiments with an inclined plate boundary this upper crustal pop-up can be compared to an orogenic wedge rooted in a singularity point located at the brittle–ductile transition (Willett et al., 1993; Beaumont et al., 1996). With ongoing deformation all models start to behave differently. Flow of the weak lower crust in experiments 4 and 5 widens the orogen by the formation of second pop-ups initiating at the transition towards stronger lower crust (e.g. Beaumont et al., 2000). The large amount of shortening and lower crustal flow in experiment 5 favors the coalescence of both pop-ups into a single, symmetrical structure, which can be compared to a bivergent orogenic wedge on the scale of the crust, while the structure of the mantle lithosphere is strongly asymmetric due to subduction of the lower plate. In the absence of a very weak lower crust (experiments 2 and 3) no second pop-up developed on the lower plate. Instead, subsidence of the lower plate occurred, probably due to the weight of the overriding plate.

5.2. Deformation of the plate interior

5.2.1. Influence of plate boundary on crust–mantle decoupling

The topographic profiles of the intra plate regions reveal significant differences among the experiments (Figs. 2c, 3c, 4c, 5c and 6c), which can be interpreted as reflecting variations in plate- and crust–mantle coupling. A thicker weak zone (experiment 3) and hence a higher degree of decoupling results in less intra plate deformation, but in uplift of the entire upper plate in an early stage. In experiment 2 (Fig. 3c) a relative thin weak zone, which is analogue for a relative high friction plate boundary, leads to more pronounced folding of the upper plate and higher topography of the antiform above the plate boundary. However, since deformation in the beginning is strongly localized along the plate interface the weak zone material is displaced rapidly (especially in the low friction plate boundary experiment (3)). This reduces the efficiency of decoupling and hence locking of the fault zone occurs. The increasing coupling along the plate boundary through time is portrayed by a gradual switch from localized deformation along the plate boundary to more distributed deformation affecting also the upper and lower plate through folding.

The results of the above described experiments differ notably form those where crust–mantle decoupling was incorporated for parts of the lower plate (experiments 4 and 5). In experiment 4 both the plate interface and the lower crust were represented by a thin weak zone (see Fig. 5). This time no ramp anticline forms on the upper plate, but subsidence leads to the formation of a retroforeland-type basin (Fig. 5c). Subsidence continued until 15% BS, and from that point onwards the basin underwent uplift indicating an increased coupling.

The combination of a thick weak plate interface and a thick weak lower crustal zone in experiment 5 resulted in a high degree of decoupling and finally subduction of the mantle lithosphere. As a result, intra-plate deformation was only minor during most of the models run (Fig. 6c). However, after 20% BS both plates started to fold causing rapid uplift of the hinge zones and deepening of the basins, probably indicating an increase in resistive forces like buoyancy or plate coupling. Additionally, in contrast to experiment 4 the weak zone material was consumed entirely leading to a high degree of stress and strain transfer through the orogen towards the interior of the subducting plate.

Notably, late-stage intra plate deformation observed in experiments 2 and 3 disclose different deformation styles between upper and lower plates. The upper plate is characterized by crustal pop-ups rooted in low amplitude, long-wavelength synclines comparable to experiment 1 and previous studies (Davy and Cobbold, 1991; Sokoutis et al., 2005), while the foreland plate is folded into a single high amplitude anticlinal structure (Figs. 3 and 4). Additionally, accurate observations on the timing of onset folding in experiment 5 reveals that lower plate folding started slightly before folding of the upper plate’s interior (Fig. 6c). Therefore, lower plate folding might be interpreted as being related to flexure of the lithosphere during mantle lithosphere subduction. With ongoing shortening the plates coupling increases and deformation spreads throughout the entire model accentuating this pre-existing flexural fore-bulge.

5.3. Basin formation

The timing of basin subsidence, widening, and subsequent uplift is of importance for identifying the transition from a decoupled towards a coupled system (Figs. 2c, 3c, 4c, 5c and 6c). Together with the monitored adjacent topography constraints can be made on the timing of the deeper processes, such as crust–mantle decoupling. A distinction can be made according to the amount of basins present in the different experiments. In experiments with a strong lower crust (exp. 2 and 3) a single basin develops on the lower plate in front of the plate contact (Figs. 3c and 4c). The basins have asymmetric foreland-type geometries with the deepest point located near the plate contact and shallow gradually towards the foreland. Remarkably, the timing of basin deepening and widening differs between experiments 2 and 3. In experiment 2 both deepening and widening of the basin was gradually lasting until the end of shortening (Fig. 3c). On the contrary, deepening was rapid in experiment 3 and the lowest topographic point was reached already at 10% BS (Fig. 4c). With ongoing shortening the basin widened, but subsequent narrowing commenced after 15% BS. In order to interpret this difference we need to underpin the processes underlying basin formation. Although the basin geometry is similar to a foreland basin it remains uncertain whether the operating mechanisms are the same. A foreland basin typically forms by flexure due to loading forces applied by adjacent topography or subsurface loads (Karner and Watts, 1983; Ziegler et al., 2002; Naylor and Sinclair, 2008). However, the plate contact developed into a lithosphere-cutting thrust zone along which drag might have caused subsidence, as suggested by the final geometry of the lower plate (Figs. 3c and 4c). For this reason we argue that basin evolution and the degree of plate (de)coupling are intimately related. Consequently, the gradual subsidence and widening of the basin in experiment 2 indicates a slow but constant movement along the plate boundary whereas early high subsidence rates in experiment 3 support a high degree of plate decoupling preceding the moment of increased coupling resulting in stagnation of basin subsidence and the onset of its narrowing.

Fig. 4. Final top view and cross-sections of experiment 3 after 19% of bulk shortening (a, b). Legends as in Fig. 2. c) Topographic evolution profiles taken from the center of the model parallel to shortening direction (arrow) after different amounts of bulk shortening. Vertical exaggeration by factor 3.
Including a weak lower crustal zone (exps. 4 and 5) gives rise to foreland-type basins on both sides of the orogen. In the early phase of experiment 4 a basin evolved on the foreland side of the first pop-up, but eventually became a retroforeland basin with ongoing shortening (Fig. 5). A similar basin overlying the upper plate can be observed in experiment 5 (Fig. 6). This basin is probably not only the result of a topographic load, but also to a subsurface load, namely a slab pull force, which became important as the slab length increased. We therefore suggest that relative subsidence of the basins and the model orogen during the last increment of shortening is due to slab pull force (compare topographic profiles of 20% and 25% BS in Fig. 6c). On the other hand, the retroforeland basin in experiment 4 is uplifted after 15% BS due to increased plate coupling and the lack of a significant vertical slab pull force. During this final stage another foreland basin started to develop on the lower plate in front of the second pop-up.

Within experiment 5 even a third, intramontane-type basin developed between both pop-ups, which showed early deepening and narrowing, and by the time both pop-ups started to behave as a single pop-up the basin was uplifted as well.

6. Discussion

6.1. The role of plate interface rheology on continental collision

The implementation of a weak interface separating the upper and lower plate with varying thickness, length and angle resulted in different styles of continental collision in terms of orogenic structure and topography.

Although vertical rheological boundaries in all our experiments were locations for deformation, a vertical weak zone has only minor influence on the deformation pattern, and the lithosphere responds by buckling without any displacement along the plate boundary. This result is consistent with the analogue modelling studies of Martinod and Davy (1994), Sokoutis et al. (2005) and Willingshofer et al. (2005). In contrast, experiments with an inclined plate interface (exps. 2–5) favor simple shear deformation and hence subduction of the lower plate. However, the presence of a weak plate interface, which is restricted to the mantle lithosphere, does not evolve in incipient continental subduction (exps. 2–3). Instead, this results in the development of a lithosphere cutting fault of which the lower part is localized along the predefined plate contact, while its upper equivalent separates the crustal segments. The strong coupling between the mantle lithosphere and the buoyant crust of the lower plate resulted in lithosphere-scale underthrusting rather than subduction of the continental mantle lithosphere.

To obtain subduction of the mantle lithosphere and formation of a collisional mountain belt a combination of a weak plate interface with sufficient thickness and weak lower crust in the lower plate is needed. Only in such a scenario the required decoupling between the buoyant crust and denser upper mantle can occur. The demand for a weak plate interface to obtain subduction has also been inferred from numerical (e.g. De Franco et al., 2008a; Gerya et al., 2008; Facenda et al., 2009) as well as analogue modelling studies (e.g. Chemenda et al., 1996, 2001).

6.2. Temporal changes of plate coupling

Thinning of the weak zone plate boundary and the absence of softening mechanisms or addition of new weak material result in strengthening of the plate contact. Together with the increasing length of the brittle segment of the plate contact where high stresses can accumulate (Hassani and Jongmans, 1997; Lamb, 2006), the degree of coupling among plates can increase through time.

Displacement along the plate interface results in thinning and/or consumption of the weak zone material and increased plate coupling facilitating stress transmission. For example, in experiment 4, high friction along the plate contact caused the upper plate to be dragged down together with the lower plate. This can be interpreted as an early phase on the initiation of double sided subduction as proposed by Gerya et al. (2008). However, when both plates became too strongly coupled with ongoing shortening subduction ceased, and regional uplift including the basin began. This result is consistent with the findings of Willingshofer and Sokoutis (2009) who argue that strong coupling between an orogenic wedge and its foreland can lead to uplift of the latter. Deducing vertical motions as a function of the degree of plate coupling is limited to cases where no significant slab pull force is present. Experiment 5 (Fig. 6) shows that the effect of the slab pull force may dominate over those related to the changes in plate coupling leading to net subsidence of the foreland basins. Additionally, in experiments (1 to 4) plate coupling results in uniform thickening of the ductile layers and uplift of the mean surface, while contemporaneous folding redistributes local topography.

Other processes, which can influence plate coupling at depth are erosion and sedimentation (Burov and Toussaint, 2007). Transportation of material from the upper plate towards the subducting plate will flex down the foreland plate and, hence favors subduction. However, in our experiments continental subduction involves only the mantle lithosphere, and flexure due to surface loading plays probably a minor role relative to mantle delamination.

6.3. The effect of lower crustal rheology on crust–mantle (de)coupling

The importance of the lower crustal physical properties in terms of density and strength on the deformation pattern of the entire lithosphere has long been recognized (e.g. Ranalli and Murphy, 1987; Burov and Watts, 2006). It is especially this part of the lithosphere which is subjected to strength variation depending on composition, crustal thickness, and geothermal gradient (e.g. Windley and Tarney, 1986). In orogenic systems the role of the lower crust can be diverse depending on among others the thermal history of the accreted blocks. Considering the style of deformation there are geological examples where lower crustal material is incorporated into the orogenic wedge and subsequently brought to the earth's surface by nappes or by the updoming of gneiss domes in the internal parts (e.g. Laubscher, 1988; Yin, 2004). Alternatively, the upper and lower crust are detached leading to the formation of a lower crustal root or even subduction (e.g. Laubscher, 1977). In this case no lower crustal material crops out at the surface. However, the behavior of the lower crust is often debated and the intermediate case might be a better interpretation in which the crust is brought to depth and metamorphosed into eclogites of which detached slivers are returned to shallower levels, while the bulk of the lower crust is subducted (e.g. Schmid and Kissling, 2000).

An important parameter determining the fate of the lower crust in an orogenic setting is the presence of a weak middle crust promoting delamination at the Conrad discontinuity. Numerical models reveal that this will especially occur when Moho temperatures are low (<550 °C), resulting in subduction of both the lower crust and the mantle lithosphere (Regard et al., 2003; Toussaint et al., 2004a; Burov and Yamato, 2008). On the other hand, higher Moho temperatures (550–650 °C) will support crustal separation from the mantle lithosphere, and hence subduction of only the mantle lithosphere.

In experiments presented here the lithosphere consists of three layers with a decisive role for the lower crustal rheology in terms of crust–mantle decoupling. Only in experiments with (some) weak lower crust (exp 4 and 5, Figs. 5 and 6) crust–mantle decoupling occurred. Experiment 4 shows...
that a thin weak zone at the base of the lower crust is sufficient to achieve crust–mantle decoupling. As the mantle lithosphere detaches thickening of the lower crust occurs inducing uplift and deformation of the brittle crust above (Figs. 5b and 7b).

If we compare the difference in the premature slab geometry of the mantle lithosphere between experiments 3 and 4 we notice a concave and convex shape, respectively. This difference can be ascribed to the combination of a buoyant crust and the degree of coupling along the Moho, which prevent mantle delamination in experiment 3 and due to the weight of the overriding plate the lithosphere can deform only by concave bending. While a slight decoupling zone along the Moho is sufficient to bend the lower plate in a convex shape.

In experiment 5 the weak zone along the Moho comprises the entire lower crust and strongly deforms by ductile flow and thickening. As a result the upper crust is uplifted and is maximally shortened above this region leading finally to the coalescence of the two pop-ups (Fig. 7d).

6.4. Comparison between experiments and natural analogues

6.4.1. Pyrenees

The Pyrenees are a suitable candidate for a comparison with the obtained modelling results because the orogen is highly investigated, the amount of shortening is within the same range (100–160 km), no accretion or exhumation of additional crustal pieces occurred, and collision was without strong obliquity. The onset of contraction in the early Paleocene between the European plate in the north and the Iberian plate in the south resulted in a typical asymmetrical bivergent crustal wedge with a relative long southern flank (Fig. 7a and c). At deeper crustal levels the ECORS seismic transect monitored a fan shaped crustal geometry partly within a thickened southward overthrust Iberian crust (Choukroune, 1989; Roure et al., 1989). Also the progressive down flexing of the Iberian Moho towards the internal zone followed by a ~15 km vertical Moho offset along the north dipping plate contact below the Pyrenees internal parts underline its asymmetrical character. For a more detailed review on the lithospheric structure and deformation history of the Pyrenees the reader is referred to e.g. Choukroune (1989), Roure et al. (1989), Muñoz (1992), and Souriau et al. (2008).

Experiments 4 and 5 show the development of a crustal orogenic wedge albeit with more symmetry of the latter. This difference in crustal structure between the models is mainly due to a larger amount of lower crustal decoupling and bulk shortening in experiment 5, causing two pop-ups to agglomerate into a symmetrical wedge with progressive mantle lithosphere subduction. In experiment 4 crustal deformation initiated also along rheological boundaries, but due to a
A thinner weak zone resulted in a modest amount of mantle lithosphere subduction, and thus in preservation of crustal asymmetry. Interpretations of the ECORS seismic transect emphasize the asymmetry at lower crustal levels, which is also evident in both experiments (Fig. 7c–d). The amount of vertical Moho offset along the plate contact is only modest in experiment 4, and wedging of the lower plate's mantle lithosphere caused lower crustal thickening below the orogen, as also interpreted by Roure et al. (1989) (Fig. 7a and b). On the other hand, in experiment 5 the lower crust partly subducted together with the lower mantle lithosphere similar to the lithosphere-scale interpretation of Muñoz (1992) (Fig. 7c). However, recent teleseismic tomography through the Pyrenees is incapable to trace the remnants of subducted lower crust (Souriau et al., 2008).

Balancing of the ECORS profiles indicates a pre-collisional asymmetry and a thicker Iberian crust with respect to Europe prior to contraction (Roure et al., 1989). Numerical modelling of the orogenic phase by Beaumont et al. (2000) advocate the importance of this inherited asymmetry on the final structure of the Pyrenees. Of special importance is the crustal extension during Cretaceous times resulting in mantle uplift and strengthening of the European plate boundary with respect to Iberia. In addition, north dipping listric faults, formed during the Variscan orogeny, flatten out at lower crustal levels favor crustal decoupling at the onset of contraction, while steep south dipping reactivated normal faults on the European side enhance an asymmetric development of the orogeny (Beaumont et al., 2000). Therefore, implementing a weak lower crustal zone simulating a

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**Fig. 8.** a) Tectonic map of the Caucasus from Ruppel and McNutt (1990). The dashed line indicates the location of the crustal section. b) Interpreted crustal section from Philip et al. (1989). Dark grey refers to lower crust, which is hardly subducted but mainly under thrusted resulting in an asymmetric orogen. c) Simple drawing of a cross-section from experiment 2 in which the timing and geometry of the structures are comparable with the eastern Caucasus. Notice the large distant of the “Dagestan” pop-up from the suture zone relative to the geological section. See text for more discussion.
horizontal decoupling in our experiments seems a realistic assumption. Moreover, our models suggest that the condition of decoupling at the lower crust is sufficient to explain the large-scale asymmetry in the Pyrenees and no call on a multitude of inherited complexities is needed. The transition from localized deformation along and in the vicinity of the plate boundary to an intra plate deformation as observed in the experiments (switch from decoupled to coupled mode), is in agreement with the recorded shift from mountain building in the Pyrenees towards thinning of the lithosphere in the Neogene (Cloetingh et al., 2002).

6.4.2. Examples from the Caucasus, Colombian Cordillera, Sierra Pampeanas, and the Alps

Other inverted rift basins to which the results of this study are relevant include the Colombian Cordillera and the Caucasus (Philip et al., 1989; Dengo and Covey, 1993; Saintot and Angelier, 2002; Roure, 2008). As in the Pyrenees no oceanic lithosphere was involved in mountain building since no high grade metamorphic terrains, nor ophiolites or volcanic arcs make part of the recent orogenic structure and the amount of shortening is comparable to experiments 2 to 4. Also the amount of different accreted blocks is limited and there are no deep earthquakes due to the lack of a Benioff zone.

During the Caucasus orogeny Cenozoic inversion of an Early to Mid-Jurassic rift occurred resulting in the formation of the Transcaucasus in the south and the somewhat younger and higher Greater Caucasus to the north (e.g. Philip et al., 1989; Dengo and Covey, 1993; Saintot and Angelier, 2002) (Fig. 8). A well-developed fold- and thrust belt has formed on the northern side of the Greater Caucasus with Dagestan in an apparent “back-arc” setting. The relative timing of deformation within the different regions of experiment 2 is comparable with the eastern part of the Caucasus. Namely, the first pop-up structures appearing on the lower plate are progressively subside by the load of the overriding plate and can be seen as part of the Transcaucasus which are now partially overlain by the Kura basin (Fig. 8b). The highest topography is formed on the overriding plate comparable to the Betcha anticline of the Greater Caucasus, which show high Neogene uplift rates and is thrust onto the foredeep. The even more recently uplifted Dagestan region is located on the upper plate and has been interpreted as an inverted Jurassic basin (Roure, 2008). This inversion accommodated most of the intraplate stresses during what can be considered as the coupled phase. A similar expression of intraplate deformation can be seen in experiment 2 albeit farther away from the plate boundary as is the case in Caucasus and relates most likely to lithospheric buckling, since pre-existing crustal faults were not included (Fig. 8c).

Even more dramatic examples of mountain belts interpreted as being related to far field stresses are the Eastern Cordillera of Colombia and the Sierras Pampeanas in Argentina. In both orogens crustal shortening was induced by the ongoing oceanic subduction below the Andes much farther towards the west (e.g. Dengo and Covey, 1993; Costa et al., 2001). The Eastern Cordillera is interpreted as a Late Jurassic–Early Cretaceous inverted rift basin, which is according to Dengo and Covey (1993) connected by a mid crustal detachment zone to the Andean subduction zone in the west and allows for the transmission of shortening from the convergent plate margin. The Sierras Pampeanas is located even more than 600 km away from the Chilean trench, but is due to the flat lying orientation of the subducted Nazca plate linked to the plate convergence within this latitude (e.g. Stauder, 1973). The orogen comprises uplifted basement blocks thrust on Pleistocene sediments along faults which produce ongoing seismic activity (Costa et al., 2001). However, examining the results of experiments 2 and 3, where intra plate deformation of the upper crust evolves during the coupling stage, an alternative explanation for both orogens can be an increase of plate coupling. Therefore, neither pre-existing crustal structures, nor changes in subduction angle are directly needed to develop mountain belts away from the zone of convergence.

Lower crustal wedges, as imaged in reflection seismic profiles, are prominent features in the Western and central Alps (e.g. Schmid and Kissling, 2000), but are missing in the Eastern Alps (TRANSALP Working Group, 2002) (Fig. 7e). The processes by which these wedges form are still unclear. The results of experiment 4 suggest that one possible mechanism for producing lower crustal wedges invokes flow of lower crust (Fig. 7b). This thickening of the lower crust also enhances uplift of the overlying orogenic wedge with a younging trend towards external regions. The formation of the more external second pop-up in experiment 4 is conditioned at the end of the weak lower crust (Fig. 7b). In natural systems, the transition from a weaker lower crust in internal parts of orogens, where temperatures are higher to a less weak lower crust in the external regions presumably occurs gradually. Inspired by our modelling results, we infer that wedging of the European lower crust in the Western Alps can be explained by shortening of a ductile lower crust, which deforms and thickens dominantly by flow processes. The aerial extent of lower crustal thickening is governed by the lateral strength variation within the lower crust, which might be the underlying cause for deformation and uplift of the external massifs. This view is supported by Ar/Ar biotite cooling ages from the Western Alps documenting uplift of a relative wide area with slightly younger towards the north, the direction of lower crustal wedging (Schlunegger and Willett, 1999; Schmid and Kissling, 2000).

7. Conclusions

Our modelling study emphasizes the importance of the plate boundary on the structural and topographic evolution of collision zones. The geometry of the plate contact together with the abundance and distribution of lubricants along the plate contact controls the initial response of surface deformation and the amount of mantle lithosphere subduction. A vertical plate contact with respect to the shortening direction will result in buckling of the lithosphere and the formation of fold related pop-ups in the upper crust. An inclined boundary leads to underthrusting of the lower plate, but does not necessarily result in mantle lithosphere subduction. Favorable conditions for subduction of the continental mantle lithosphere and continental collision include strong crust–mantle decoupling and the presence of a weak plate interface. Consequently, the distribution and amount of weak lower crust influences the architecture of the orogenic wedge in terms of mantle delamination, but also affects wedge geometry and topography as the lower crust thickens by flow. Whether lower crust subducts beneath the orogen depends mainly on the thickness of the weak plate interface and the amount of weak lower crust.

The transition from a decoupled to a coupled collisional stage by thinning or consumption of the weak zone is recorded by a change from localized to distributed deformation. The coupled stage comprises concomitant folding and thickening of both the upper and lower plates, and therefore late-stage uplift or subsidence of orogens, foreland basins, and flexural fore-bulges.

A change from decoupling to coupling during the collisional stage as it occurs in our experiments matches well with observations from orogens around the world, like for instance the Pyrenees, Caucasus, Colombian Cordillera, and the Sierra de Pampeanas in Argentina. In addition, wedge-shaped geometries of the lower crust as inferred for the Western Alps may be produced even when the lower crust deforms by viscous processes.

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