Asymmetric vs. symmetric deep lithospheric architecture of intra-plate continental orogens

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\textbf{A R T I C L E I N F O}

\textbf{A B S T R A C T}

The initiation and subsequent evolution of intra-plate orogens, resulting from continental plate interior deformation due to transmission of stresses over large distances from the active plate boundaries, is controlled by lateral and vertical strength contrasts in the lithosphere. We present lithospheric-scale analogue models combining 1) lateral strength variations in the continental lithosphere, and 2) different vertical rheological stratifications. The experimental continental lithosphere has a four-layer brittle–ductile rheological stratification. Lateral heterogeneity is implemented in all models by increased crustal strength in a central narrow block. The main investigated parameters are strain rate and strength of the lithospheric mantle, both playing an important role in crust–mantle coupling. The experiments show that the presence of a strong crustal domain is effective in localizing deformation along its boundaries. After deformation is localized, the evolution of the orogenic system is governed by the mechanical properties of the lithosphere such that the final geometry of the intra-plate mountain depends on the interplay between crust–mantle coupling and folding versus fracturing of the lithospheric mantle. Underthrusting is the main deformation mode in case of high convergence velocity and/or thick brittle mantle with a final asymmetric architecture of the deep lithosphere. In contrast, lithospheric folding is dominant in case of low convergence velocity and low strength brittle mantle, leading to the development of a symmetric lithospheric root. The presented analogue modelling results provide novel insights for 1) strain localization and 2) the development of the asymmetric architecture of the Pyrenees.

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\section{1. Introduction}

Intra-plate orogens are the result of continental plate interior deformation due to transmission of stresses at large distances from active plate boundaries (Raimondo et al., 2014; Vauchez et al., 1998).

The initiation and subsequent evolution of such mountain belts strongly depend on 1) lateral strength contrasts in the lithosphere allowing for localization of deformation (Vauchez et al., 1998; Ziegler et al., 1998), and 2) the rheological structure of the lithosphere, which governs its overall geometry (Burov, 2011).

The response of the lithosphere to applied stresses is controlled by its strength and thus by composition of crust and mantle, thermal gradient, strain rate and fluids related processes (Burov, 2011).

In general, low strength leads to distributed deformation and symmetry, while high strength results in localization and asymmetry (Davy and Cobbold, 1991; Gueydan et al., 2008).

Previous numerical and analogue modelling studies allowed classifying the behaviour of continental lithosphere, when subject to compressional horizontal stresses, into three main modes of deformation: 1) lithospheric folding, 2) volumetric shortening with distributed or localized thickening, 3) underthrusting, and continental subduction (Cloetingh et al., 1999; Sokoutis et al., 2005; Toussaint et al., 2004).

Lithospheric folding is considered to be a primary response to shortening (Cloetingh et al., 1999). The wavelength of folding, as well as its persistence in time, is a strong function of thermo-mechanical age of the lithosphere (Cloetingh et al., 1999; Martinod and Davy, 1992; Toussaint et al., 2004).

The occurrence of continental subduction or pure shear thickening is largely controlled by lithospheric strength. While the former mechanism dominates for stable old continents, the latter...
is the rule for hot and young lithosphere (Cagnard et al., 2006; Davy and Cobbold, 1991; Toussaint et al., 2004).

Lithospheric folding and pure shear thickening result in symmetric structures, while continental subduction geometries are strongly asymmetric (Davy and Cobbold, 1991; Toussaint et al., 2004). The Pyrenees are a good example of the latter case, being characterized by a pronounced asymmetric architecture both at crustal and lithospheric scale (Roure et al., 1989). Our study aims to provide insights into the parameters that controlled the development of such asymmetric mountain belts.

The presence of lateral heterogeneities in the lithosphere can affect and interfere with the manner shortening is accommodated. The mechanical properties of continental lithosphere are strongly heterogeneous in space, as documented by observations and suggested by mechanical models (Burov, 2011; Cloetingh et al., 2005; Gueydan et al., 2014; Ranalli, 1997). The reaction of lithospheric heterogeneities controlled the evolution of intra-plate orogens in Iberia (Pyrenees, Spanish Central System, Iberian Range), Australia (Petermann and Alice Spring orogens), Central America (Laramide orogen), and Central Asia (Tian Shan) (Beaumont et al., 2000; Cerca et al., 2004; Cloetingh et al., 2005; Delvaux et al., 2013; Fernández-Lozano et al., 2011; Gorczyk et al., 2013; Kennett and Laffaldano, 2013; Raimondo et al., 2014; Sainz and Facenna, 2001). At lithospheric scale the nature of the reacted heterogeneities is diverse and debated. Rift basins are convenient structures for localization of intra-plate deformation due to their lower strength with respect to the surrounding lithospheric domains (Brun and Nalpas, 1996; Buitier et al., 2009; Cloetingh et al., 2008; Jammes and Huismans, 2012). On the other hand, strong microplates can also localize deformation at their margins and thus promote intra-plate orogeny (Keep, 2000).

Previous modelling studies investigated the reaction of lithospheric scale weak zones in compression in order to study rifts inversion (Brun and Nalpas, 1996; Buitier et al., 2009; Cerca et al., 2004) or deformation of weak orogenic wedges (Gerbault and Willingshofer, 2004; Jammes and Huismans, 2012; Willingshofer et al., 2005). A more limited number of studies focused on the role of strong domains in the origin of intra-plate mountain belts (Calignano et al., 2015; Keep, 2000). Among these, Calignano et al. (2015) considered the role of a strong domain embedded within a weak lithosphere.

The aim of this study is to contribute to the understanding of deformation mechanisms of continental lithosphere in compressional intra-plate settings under various rheological conditions. To this purpose, we used lithospheric-scale analogue models where we combined 1) lateral strength variation in the continental lithosphere, and 2) diverse vertical rheological stratification. As a novelty, we focused on the role of a strong domain embedded in an overall strong lithosphere. Other main investigated parameters are strain rate and strength of the lithospheric mantle, both playing an important role in crust–mantle coupling and thus in the final asymmetry/symmetry of the orogenic system.

We discuss the results of the experiments with emphasis on the final geometry of the deep lithospheric structure and the possibility of development of intra-plate continental subduction-type geometries. In particular, the experimental results provide valuable insights on the parameters that controlled strain localization and asymmetry of the Pyrenees.

2. Experimental set-up

The initial geometric and rheological conditions adopted in this study are representative for reactivation of strength heterogeneities within the lithosphere under compression. In particular, the behaviour of a four-layer continental lithosphere under variable strain rates and with different vertical rheological stratification is investigated.

2.1. Initial geometry

Fig. 1 illustrates the set-up and geometrical configuration for the experiments described in the results section. Geometric and kinematic parameters for the experimental series are specified in Table 1.

All experiments consist of three domains with different mechanical properties: a central block where the thickness of the brittle upper crust is increased in order to simulate the presence of a stronger domain is located in between two blocks that share the same lithospheric stratification (Fig. 1). The boundaries between rheologically different domains strike perpendicular to the convergence direction. The experiments represent an area of 840 km × 720 km. The central block, with increased crustal thickness, has a width of 4 cm (80 km in nature).

As discussed above, lateral strength contrasts in the lithosphere can result from rheological and mechanical heterogeneities located at different depth. Variation in composition or thermal structure can affect the crust or the mantle or both layers. Increased crustal strength can result from the presence of 1) a dehydrated residue after melt extraction processes, 2) high-viscosity mafic lower crust or 3) low concentration of radiogenic heat-producing elements (Bürgmann and Dresen, 2008). In the present study we focus on lateral variation of bulk lithospheric strength resulting from rifting and subsequent cooling of the rift basin. If the integrated strength of a lithospheric column is considered, a strong domain can be representative for an old, cooled rift. In fact, rifting induces thinning of the lithosphere, with upward displacement of the Moho and mantle material to shallower levels with the respect to the surrounding regions (McKenzie, 1978). During the post-rift phase, the rift cools and will eventually become stronger than the unstretched continental lithosphere (Buitier et al., 2009; Cloetingh et al., 2008; Leroy et al., 2008). Thus, the creation and thermal relaxation of a continental rift is an important process leading to lateral contrast in the continental lithosphere (Vauchez et al., 1998). We simulate a strong cold/old rift by lateral variation of brittle crustal thickness. In particular, a narrow block characterized by increased crustal strength (deeper brittle/ductile transition), serves as analogue for the presence of a strong rift. This geometry serves to create a localized increase in bulk lithospheric strength, even if it
Experimental setup

Fig. 1. a) Experimental set-up: four-layer laterally heterogeneous continental lithosphere, resting on a high density-low viscosity fluid representing the asthenosphere; shortening is applied through a moving wall; the two red dots indicate the margins of the strong central block; b) representative experimental strength profiles showing lateral variation in lithospheric strength at initial stage of deformation; c) effect of convergence velocity on the strength of the viscous layers; strength of the ductile layers is calculated with a strain rate defined as the ratio of convergence velocity and thickness of the viscous layer; d) zoom-in on the geometry of the experiments along a cross section parallel to the convergence direction; the central block is characterized by a thicker brittle upper crust in order to simulate the increases bulk lithospheric strength of an "old rift" with respect to the surrounding regions. BUC: brittle upper crust; DLC: ductile lower crust; BUM: brittle upper mantle; DUM: ductile upper mantle; A: asthenosphere. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

does not represent the vertical strength partitioning in a cold rift, where the stronger layer would be the lithospheric mantle. With this setup we focus the discussion on the influence of strength variations and eliminate complexities arising from geometric configurations (shallow Moho), which have been addressed by numerical modelling studies (Jammes and Huismans, 2012).

The thickness of the upper crust, lower crust and ductile lithospheric mantle are constant for all experiments, while the thickness of the brittle lithospheric mantle is varied. We tested different convergence velocities in order to investigate different strength of the ductile layers and, as a consequence, the variation in crust mantle coupling (Table 1). For each experiment the convergence velocity remains constant during the whole deformation process.

2.2. Rheology

The reference model lithosphere is characterized by a four-layer rheological stratification, where the strength resides mainly in the brittle upper crust and brittle upper mantle (Fig. 1b). This vertical strength distribution is representative for regions with intermediate to low geothermal gradient (Gueydan et al., 2008). The strength envelopes in Figs. 1b and 1c are representative for the very initial stage of deformation (see Appendix A for strength profiles). From top to bottom the reference model lithosphere is composed of brittle upper crust, ductile lower crust, brittle lithospheric mantle and ductile lithospheric mantle (Fig. 1). The brittle crust and brittle mantle are simulated with dry feldspar sand and dry quartz sand, respectively. These granular materials have different densities and
coefficients of friction (Table 2) and, represent Mohr–Coulomb type brittle behaviour in nature. Different coloured sand layers help the visualization of brittle structures in cross section. The experimental ductile lower crust and ductile lithospheric mantle consist of mixtures of Rhodorsill Gomme CSIR (Rhône Poulenc, France)-type silicon putty, quartz sand as filler, and oleic acid in different proportions to account for differences in density and viscosity. Both ductile materials show viscous quasi-Newtonian behaviour (Table 2). The model lithosphere is resting on a high density and low viscosity fluid mixture of polytungstate and glycerol representing the lower lithospheric mantle and asthenosphere. This guarantees the isostatic compensation. The density of the lithospheric mantle is in all experiments lower than the asthenospheric fluid, so that subduction is impeded.

2.3. Experimental procedure

The models are built inside a transparent Plexiglas box with dimensions 36 cm × 50 cm × 15 cm. Convergence velocity can be adjusted by modulating the frequency of an engine driving an internal moving wall. Experiments are performed in normal gravity field. Boundary effects due to friction along the walls are reduced by placing thin glass plates inside the longer sides of the box. Deformation is monitored during the experiments with top view pictures and laser scanning at regular time intervals. The conversion of scanner data into Digital Elevation Models (DEMs) allows the analysis of topographic evolution during shortening. At the end of an experiment the model is soaked in water, frozen and cut in cross sections parallel to the convergence direction. This allows the visualization of the internal deformation at the latest experimental stage.

2.4. Scaling

The present series of experiments is not designed to reproduce a specific regional setting, but rather to generically investigate the deformation and topography evolution of a lateral heterogeneous lithosphere under compression. We, therefore, scaled the experiments to average convergence rates and ductile layers viscosities in nature.

The experiments are scaled according to the criteria of geometrical, rheological, dynamical and kinematic similarity (Davy and Cobbold, 1991; Ramberg, 1981; Weijermars and Schmeling, 1986).

The experiments are built with a length-scale factor of $5.0 \times 10^{-7}$ so that 1 cm in the model corresponds to 20 km in nature. Dynamic and kinematic similarities have been tested with dimensionless ratios relating gravitational forces to viscous forces. In particular, for ductile layers the Ramberg number ($R_m$) has been calculated as:

$$R_m = \frac{\rho gh_d}{(\sigma_1 - \sigma_3)_{\text{viscous}}}$$

where $\rho$ and $h$ are respectively the density and thickness of the ductile layer, $g$ is the acceleration due to gravity ($g = 9.81 \text{ m/s}^2$), and $(\sigma_1 - \sigma_3)_{\text{viscous}}$ is the strength for ductile materials with power-law type behaviour (see Appendix A). For brittle deformation dynamic similarity can be tested with the ratio between gravitational forces and frictional forces which can be considered an analogue to the Smoluchowski number ($S_m$) defined by Ramberg (1981).

$$S_m = \frac{\rho gh}{c + \mu \rho gh}$$

where $\rho$, $h$, $c$ and $\mu$ are respectively the density, thickness, cohesion and coefficient of friction of the brittle layer.

2.5. Simplifications and limitations

Scaled analogue models are simplified representations of natural processes.

Ductile behaviour in nature is strongly dependent on temperature and consequently varies with depth (Ranalli and Murphy, 1987). In the presented experiments the viscosity of the ductile lower crust and mantle remains constant with depth, i.e. temperature effect is not taken into account. Previous experiments demonstrated that representing the ductile behaviour of the lower crust and lithospheric mantle with uniform viscous materials is an acceptable first-order approximation (Davy and Cobbold, 1991).

The evolution of topography is considered without taking into account surface processes, like erosion, transport and deposition of sediments. These processes lead to stress redistribution and influence the time evolution and localization of deformation (Burov and Toussaint, 2007). The geometry of the system does not consider complex 3D lithospheric structure since the experimental set-up has not been designed to represent a particular natural case study.

Despite these simplifications the presented experiments are considered representative for first order deformation and associated topography in intra-plate compressional settings.

3. Experimental results

In the following section a detailed description of four end-member experiments is provided. Experiments with intermediate values of convergence velocity and thickness of the brittle lithospheric mantle will be presented later for discussion purposes. The experimental results will be illustrated with the help of Digital Elevation Models (DEMs) acquired at progressive stages of the deformation, topographic profiles extracted from the DEMs, and pictures of cross sections representative of the final stage.

In the description of the experiments we refer to “proximal block” to the lithospheric block closest to the moving wall, “central block” to the one containing the strong crustal heterogeneity and “distal block” to the lithospheric domain closest to the fixed back wall (Fig. 1a). A “thrust” is defined by footwall displacement in the same sense of the applied shortening, whereas a “back-thrust” shows opposite vergence.

In all experiments soon after shortening is applied a variable number of closely spaced back-thrusts localize right next to the
moving wall, causing flexure of the whole lithosphere. We consider this as boundary effect due to the geometry of the experiments and for this reason the above mentioned structures are not described in the evolution of the single experiments.

### 3.1. Experiment 1 – high convergence velocity and thin brittle upper mantle

Experiment 1 is characterized by a low-strength brittle upper mantle (5 mm = 10 km) and was deformed at high convergence velocity (5 cm/h). Fig. 2 shows the evolution of the deformation and associated topography with the help of DEMs acquired at subsequent steps. After 5% BS (bulk shortening) the first structure that developed is a thrust localized at the boundary of the central strong block facing the moving wall (thrust 1 in Figs. 2a and 2b). This structure remains active throughout the experiment accommodating a large amount of shortening. A foreland type basin develops in the footwall of thrust 1, in response to the flexure of the lithosphere (Fig. 2c). Basin subsidence is limited to the first 3 cm of applied shortening, when a back-thrust (thrust 2 in Figs. 2a and 2b) nucleates in the upper crust of the proximal block. The foreland basin depocentre becomes uplifted and migrates (Fig. 2c).

Subsequent locking of the main thrust generates strong coupling between the strong central block and the proximal block. Strain is transferred to the proximal block with a series of minor thrusts. The latest stages of deformation are characterized by the reactivation of thrust 1. The upper crust of the central block overthrusts the upper crust of the proximal block. The initial foreland basin is closed between the two thrusts with opposite vergence. A new foreland basin forms on the lower plate (Fig. 2c). The final architecture of the orogenic system is asymmetric. Deformation is strongly localized: a main underthrusting plane with dip direction in the sense of the applied convergence accommodates most of the shortening at the margin of the central block (Fig. 2b). The strong crustal domain becomes partly buried along the major thrust due to drag effects during late stages of deformation. Strong localization is reflected by the location of the highest topography above the activated margin of the strong domain (Fig. 1c). Asymmetric
Experiment 2 - low convergence velocity - thin brittle mantle

a) Modelling results for experiment 2 – low convergence velocity and thin brittle mantle. a) Evolution of deformation and associated topography shown through DEMs at 5%, 10%, 15% and 20% bulk shortening; numbers indicate thrust sequence in time; b) Final geometry and interpreted structures after 20% BS illustrated by a representative cross section (A–A’) and DEM; c) Topography at subsequent deformation stages along the representative section in b. (For interpretation of the colours in this figure, the reader is referred to the web version of this article.)

b) Symmetric deep lithospheric structure

c) Topography along cross section A–A’

Different from experiment 1, the first response of the experimental continental lithosphere to shortening is lithospheric folding. Two high amplitude anticlines can be recognized from the topographic profiles (Fig. 3a). The wavelength of the main buckles is initially 12 cm (240 km in nature). The amplitude of the two anticlinal structures is different with the lower amplitude developed at the location of the strong central block (Fig. 3c). A basin, corresponding to the syncline bounded by the two anticlines, develops in the proximal lithospheric block, next to the strong crustal heterogeneity. The shape of the basin is symmetric and its depocentre axis remains stationary for the first 2 cm of applied shortening. Folding continues until a first thrust fault is formed at the inflection point of the lowest anticline (thrust 1 in Figs. 3a and 3b). The location of the first thrust corresponds to the proximal boundary.

3.2. Experiment 2 – low convergence velocity and thin brittle upper mantle

Experiment 2 has the same geometry and layers thickness as experiment 1, but was deformed at a lower convergence rate (1 cm/h). Decreasing the convergence velocity allows us to decrease the strength of the ductile layers in the experiment and, as a consequence, the amount of crust–mantle coupling.

folding characterizes the lithospheric mantle, without any sign of brittle deformation. Asymmetry is controlled by the activity along thrust 1. The ductile lithospheric mantle is strongly thinned at the tip of the underthrusting slab, while it thickens significantly below the undeformed part of the strong central block.
of the crustal heterogeneity as in experiment 1. The activity along the thrust causes uplift of the central block, while the anticline located over the proximal block stabilizes. The central basin becomes asymmetric and narrower. After 10% BS the activity of thrust 1 stops and is taken over by a new thrust, with opposite vergence (thrust 2 in Figs. 3a and 3b). The switch from thrust 1 to thrust 2 is well expressed in the topography evolution (Fig. 3c). While thrust 1 is active the central block is uplifting and the depocentre of the foreland basin is subsiding. The subsequent activation of the back-thrust is manifested in stable elevation above the strong central block, uplift in the proximal block and uplift of the basin depocentre.

In contrast with experiment 1 the deep lithospheric architecture is characterized by a symmetry mega-fold, localized at the proximal boundary of the strong crustal block (Fig. 3b). Upper crustal material is buried at great depths in correspondence with the mega syncline. Symmetric folding of the upper part of the continental lithosphere (brittle crust, lower crust and brittle mantle) is accompanied by thickening of the ductile lower lithospheric mantle at the high amplitude anticlines and thinning at the syncline. Brittle deformation is limited to the brittle upper crust. In contrast to the deep symmetric geometry, asymmetry in topography is controlled by dominant activity along thrust 2.

3.3. Experiment 3 – low convergence velocity and thick brittle upper mantle

In experiment 3 the thickness of the brittle upper mantle has been increased to 1 cm, equivalent to 20 km in nature. The applied convergence velocity is 1 cm/h.
Soon after the onset of shortening, a pop-up nucleates at the proximal margin of the strong central block (Figs. 4a and 4c). The pop-up is defined by a major thrust (thrust 1 in Figs. 4a and 4b) and an associated back-thrust (thrust 1b in Figs. 4a and 4b). Subsequent shortening results in the migration of deformation in the upper crust of the proximal block, towards the moving wall. By 15% BS three pop-up structures delimiting push-down triangular blocks can be recognized from the DEMs and related topographic profile (Figs. 4a, 4c). At the end of the experiment pop-ups 2 and 3 have merged into a prominent uplifted belt, while the first pop-up shows lower elevation. Uplift of the pop-up structures is accompanied by normal faulting above rising lower crustal material. During the latest stages of deformation a minor thrust (thrust 4 in Figs. 4a and 4b) nucleates at the distal margin of the strong central block that becomes progressively pushed down.

Similar to experiment 1, the cross section in Fig. 4a shows an asymmetric final geometry of the deep lithospheric structure. Brittle failure occurs in the upper lithospheric mantle and it is localized at the proximal margin of the strong block. Deformation is characterized by strong decoupling between crust and mantle (Fig. 4b). Numerous faults bound pop-up and push-down structures in the brittle upper crust. Here deformation propagates progressively from the strong domain towards the moving wall. In contrast, shortening in the upper brittle mantle is accommodated by a single major thrust. The downward movement of the proximal block is accommodated by thinning of the ductile lower lithospheric mantle in the central part of the experiment and associated thickening in correspondence of the distal block.
3.4. Experiment 4 – high convergence velocity and thick brittle mantle

The last end-member experiment is characterized by high strength brittle upper mantle and was deformed at high convergence velocity. These boundary conditions make this experiment representative for a strong continental lithosphere characterized by strong crust–mantle coupling.

The first response of the modelled lithosphere to applied shortening is a broad uplift in the centre of the experiment (Fig. 5a). By 5% BS, deformation is already localized at the proximal boundary of the strong central block. Here a thrust nucleates causing underthrusting of the proximal lithospheric block below the crustal heterogeneity (thrust 1 in Figs. 5a and 5b). A conjugate back-thrust develops to the left of the main thrust bounding a crustal pop-down (thrust 2b in Figs. 5a and 5b). An asymmetric foreland-type basin develops on the footwall of the main thrust. Subsidence of the basin is accompanied by uplift of the upper plate up to 10% BS. Later stages of deformation are characterized by uplift of the proximal block in the form of an anticline, while activity along the main thrust stops, as testified by stable elevation of the central block (Fig. 5a). Deformation subsequently propagates towards the moving wall with thrusts nucleating at the inflection point of the growing anticline (thrusts 2 and 3 in Fig. 5b).

The final geometry (Fig. 5b) is again highly asymmetric. As in the case of experiment 1 a subduction-like geometry developed with the proximal block as lower plate. Deformation is strongly localized and characterized by faulting in both upper brittle crust and upper brittle mantle. Most of the shortening is accommodated along a lithospheric structure localized at the proximal boundary of the crustal heterogeneity. Crustal material of the proximal block is transferred at great depth due to strong crust–mantle coupling.

4. Intra-plate continental deformation

Our results constrain the key parameters that control modes of intra-plate deformation and in particular strain localization and asymmetry/symmetry, as it is summarized in Fig. 6. Intra-plate orogens are found at great distances from active plate boundaries. Their initiation is controlled by the presence of lateral strength variation in the continental lithosphere (Aitken et al., 2013; Raimondo et al., 2014; Ziegler et al., 1998).

Moreover the overall strength of the lithosphere is a key parameter in determining the response to stresses and the final geometry (asymmetry/symmetry) of such mountain belts. A comparison with first order features of the Pyrenees allows us to infer the control exerted by different parameters on the development of such an asymmetric orogen. Furthermore, implications for the development of gravitational instabilities can be drawn for the symmetric case.

4.1. Localization of deformation in continental intra-plate settings

Previous modelling studies focused on initiation of intra-plate deformation by reactivation of mechanically and rheologically weak pre-existing structures (Brun and Naqal, 1996; Buiter et al., 2009; Cerca et al., 2004; Gerbault and Willingshofer, 2004; Jammes and Huismans, 2012; Willingshofer et al., 2005).

The present study demonstrates that the presence of a strong crustal segment (deeper brittle/ductile transition) is a favourable rheological configuration for strain localization in continental compressional settings. In all experiments, irrespective of the initial vertical rheological stratification of the lithosphere, deformation localizes at the margin of the strong block facing the moving wall. In contrast, strain is never localized at the boundary of the strong block facing the back-wall, with exception of experiment 3, where its activation is a late stage expression. We attribute this geometry to the applied kinematic boundary conditions (push from one side). Once deformation has been effectively localized, the subsequent evolution of the system is mainly a function of the vertical layering of the lithosphere. The strong central block remains in all cases undeformed, experiencing only vertical movements (uplift or subsidence), but no internal deformation.

The initiation of the Early Palaeozoic Orogen of South-East China is a good example of intra-plate localization linked to lateral crustal heterogeneities. Here, intra-continental subduction localized at the facies transition between carbonate platform and terrigenous trough deposits. Such sharp facies and thickness contrasts are derived from a previous rift event affecting the Cathaysia block (Faure et al., 2009). The resulting intra-continental orogen is
strongly asymmetric with thick-skinned tectonics and high-grade metamorphism affecting the lower plate and thin-skinned tectonics and low-grade metamorphism on the upper plate. The final geometry of the orogen is comparable with our experiments, where lateral changes in crustal thickness localize deformation and allow the initiation of continental underthrusting, resulting in an asymmetric orogen. The thin-skinned tectonics affecting the northern part of the South-East China Orogen can be associated with pre-existing shallow crustal structure and decollement levels. Note that these kinds of crustal heterogeneities are not taken into account in our experiments.

4.2. Modes of intra-plate continental deformation

Analysis of the experiments enables us to recognize two main modes of deformation: 1) continental lithosphere underthrusting and 2) lithospheric folding. The first one accounts for the development of an asymmetric lithospheric structure and asymmetric intra-continental orogeny, whereas the second one results in a symmetric architecture (Fig. 6). The predominance of one mechanism over the other is controlled by a combination of convergence velocity and strength of the brittle lithospheric mantle.

Convergence velocity controls the strength of the ductile layers. This in turn determines the overall strength of the lithosphere and the degree of crust–mantle coupling. In case of high convergence velocity (>1 cm/h), shortening is always accommodated by continental lithosphere underthrusting, resulting in asymmetry. In case of low convergence velocity (1 cm/h), the mode of deformation is dependent on the thickness of the brittle lithospheric mantle. Underthrusting remains the preferred manner of accommodating shortening when the brittle mantle is thick enough to localize deformation along a large thrust fault. Thus, similar to previous studies on continental plate collision (Davy and Cobbold, 1991; Faccenda et al., 2008; Toussaint et al., 2004), we can conclude that subduction-type geometries in continental lithosphere can develop for high convergence velocities and for high lithospheric strength. In all experiments we observed asymmetric geometries developed, the underthrusting plane always dips towards the fixed back wall. We interpret this result as a consequence of the combination of the central position of the strong domain and the push only from one side that causes the reactivation of the proximal boundary of the central block. Lithospheric folding is favoured when the brittle mantle is thin enough to allow bending instead of fracturing. In this latter scenario brittle faults in the upper crust develop later during deformation and at the inflection points of the high amplitude buckles. After a first short phase of lithospheric buckling, the evolution of the system is characterized by alternating folding and faulting. The coexistence of these two fundamental modes of deformation (folding and faulting) as a response to shortening has been extensively described in previous studies (Cloetingh et al., 1999). Displacement along two main thrusts with opposite vergence results in well-developed symmetric sedimentary basins that become progressively pushed down and buried when the bounding faults meet at the surface. Compressional sedimentary basins with such geometry have been recognized in different orogenic settings like the Fergana Basin in Central Asia and the Magdalena Basin in Colombia (Burov and Molnar, 1998; Cobbold et al., 1993). The development of symmetric subduction-type geometries for medium age lithosphere (150–300 Ma) in case of low convergence velocities, has been previously described by Burov and Cloetingh (2009). Similar results were obtained also out by numerical models by Faccenda et al. (2008) with low convergence resulting in lithospheric buckling and two-sided symmetric collisional zone as a consequence of the strong rheological coupling due to reduced shear heating at the plates interface, and high convergence velocity resulting in enhanced shear heating at the contact between two plates and consequent decoupling and development of asymmetric orogen and one-sided subduction.

The presence of a strong block in our experiment allows for investigating the interference of folding and lateral strength contrast in the lithosphere. The role of crustal heterogeneities in promoting the development of lithospheric folding in compressional settings has been pointed out in previous analogue experiments (Sokoutis et al., 2005). This behaviour is well expressed in experiment 2, where a mega-fold nucleates at the proximal margin of the strong crustal block. From the topography evolution of this experiment we can conclude that the strong block actively controls the amplitude of lithospheric folding. In fact, the anticline developing in correspondence to the increase in crustal thickness has lower amplitude with respect to the next anticline, developed in the weaker lower plate.

4.3. Influence of crust–mantle coupling

Our results show that the mode of deformation strongly depends on the degree of crust–mantle coupling, which is controlled by convergence velocity. The influence of convergence velocity on crust–mantle coupling is well expressed in the differences between experiment 3 and experiment 4 (Fig. 6). Both experiments indicate pronounced asymmetry. However, strong decoupling along the ductile lower crust characterizes experiment 3. Multiple pop-up and pop-down structures in the brittle crust develop in the lower plate, while a single major thrust fault accommodates shortening in the brittle lithospheric mantle. A wide area with distributed deformation that progressively propagated from the central block towards the moving wall, characterizes the proximal lithospheric block. The formation of similar wide and strongly asymmetric orogenic wedges, developed on top of the lower plate and characterized by progressive outward propagation, has been observed in continental subduction experiments by Willingshofer et al. (2013) in case of strong decoupling at the level of the lower crust, and compared to mountain belts like the Carpathians, the Dinarides, and the Caucasus. When the lower crust is stronger and coupled with the brittle mantle (experiment 4) deformation is strongly localized, with a major thrust fault cross-cutting the lithosphere. The number of faults is the upper crust decreases, together with the width of the orogen. In this case upper crustal material of the lower plate can be efficiently transported at great depths. Deeply buried upper crustal fragments are present also in experiments 1 and 5, indicating a high degree of crust–mantle coupling (Fig. 6). Our results are in agreement with the studies on the control exerted by brittle/ductile coupling on localized/distributed deformation (Schueller et al., 2005, 2010; Willingshofer et al., 2013).

4.4. Asymmetry in intra-plate orogens: the Pyrenees example

Large-scale intra-plate deformation affected the Iberian plate from the Early Tertiary, as a result of Africa–Eurasia convergence (Choukroune, 1989; Fernández-Lozano et al., 2011; Sainz and Facenda, 2001). The transmission of compressional stresses from the surrounding active plate boundaries resulted in the reactivation of pre-existing Hercynian and Mesozoic structures (Muñoz, 1992).

The Pyrenees formed between Late Cretaceous and Early Miocene as a result of inversion of pre-existing extensional structures related to Triassic to Early Cretaceous rifting in the northern margin of the Iberian plate (Choukroune, 1989; Muñoz, 1992; Roure et al., 1989). The Central Pyrenean ECORS seismic section displays a clear asymmetric architecture of the deep lithosphere. The orogen is characterized by fan shape geometry with dominant southward vergence (Roure et al., 1989). The Iberian Moho progressively deepens from the Ebro Basin towards the Axial Zone of the Pyrenees, while the European Moho remains at shallow level.

Although our experiments were designed to investigate the parameters relevant to intra-plate orogens in general, a comparison with first order features of the Pyrenees allows us to assess some aspects of the orogen evolution. In particular, the localization of deformation (location of the main underthrusting/subduction plane) and the asymmetric geometry of the mountain belt are similar. In Fig. 7 the interpreted ECORS cross section and the relative restored section from Choukroune (1989) are compared with initial set-up and final geometry of our best fit model.

Mesozoic extensional events produced lateral strength contrasts in the axial part of the Pyrenean crust before the onset of compression in Late Cretaceous time. In particular, the North Pyrenean Fault (NPF) can be considered to be the location of the maximum lateral strength contrast, due to the presence of mantle material at shallow depth to the north of the fault (Fig. 7a). Given a time lag of 40 to 60 Ma between rifting and subsequent inversion (Vergés et al., 1995), we can assume that the lithosphere in correspondence to the rift had enough time to cool and strengthen. For this reason, the strong domain included in the modelled continental lithosphere can be representative for the lateral heterogeneities affecting the Pyrenean lithosphere at the onset of the Pyrenean orogeny. Despite the geometry of the rift, characterized by a thinned crust, the initial experimental setup wants to represent the presence of a strong lithospheric domain in the continental lithosphere at the onset of shortening. The increased thickness of the brittle crust in the central block of the experiment allows us to increase the bulk strength of the lithosphere in this region with respect to the surrounding, thus simulating such lateral strength contrasts. Structural studies and crustal section restoration have demonstrated that the NPF was one of the first inherited heterogeneities to localize deformation at the onset of shortening (Choukroune, 1989). Our experiments show that localization of deformation is controlled by the presence of a strong domain in the lithosphere. Here, for most of the experiments, a main thrust developed, controlling the final asymmetric geometry (Fig. 6). After localization at the margin of the strong heterogeneity, deformation progressively propagates outward affecting the upper crust of the lower plate. This can be considered one of the factors controlling the asymmetry of the Pyrenees, where the southern wedge is more developed that the northern thrust system. Our experiments do not reproduce the de-
Detailed geometry of the extensional system and thus are not able to reproduce the development of north verging thrusts in the Pyrenees, but clearly underpin previous findings, which have pointed out the importance of these inherited crustal heterogeneities for the final geometry of the mountain belt (Beaumont et al., 2000). Small backthrusts are observed in experiment 1, but we prefer not to compare them with the north verging thrusts of the Pyrenees since they are most probably boundary effect due to their vicinity to the box walls.

The analysis of the final geometry of the experiments allows us to exclude the possibility that during the Pyrenean orogeny the Iberian lithosphere was characterized by a weak lower crust. In fact, the experimental results show that low convergence velocity, and thus weak lower crust, results in symmetry in case of a thin brittle mantle or in distributed deformation (wide orogen) in case of a thicker mantle (Fig. 6). An important criterion to infer the parameters that controlled the final geometry of the Pyrenees is the location of relief. In experiment 4 a gentle gradient of topography characterizes the strong domain (Fig. 6). The presence of a narrow orogen with a sharp relief gradient at the location of the strong heterogeneity and associated with asymmetry at depth makes experiment 1 a better candidate to be compared with the Pyrenees. In conclusion, from a detailed comparison of the experimental results and the Pyrenees ECORS cross-section, we propose that the evolution of this orogen has been controlled by: 1) the presence of a strong lithospheric domain, 2) strong ductile layers, i.e. high degree of coupling and, 3) asymmetric folding of the lithospheric mantle.

4.5. Possible implications for the development of gravitational instabilities

Experimental results show that in case of low strength brittle mantle and low convergence velocity, continental lithosphere responds to the applied shortening with high-amplitude, symmetric lithospheric folding. In particular, the final geometry of the intra-plate orogen is characterized by a lithospheric scale pop-down. Upper crustal thickness increases at the location of the pop-down and cold upper crustal material is pushed down at great depths. If such lithospheric swelling continues growing, a gravitational instability (Rayleigh–Taylor) may be triggered (Houseman and Molnar, 1997). As predicted from numerical and analytical models, the descent of cold lithospheric material creates lateral temperature gradients and eventually the heavy lithospheric root is removed by mantle convection. Finally, the compressional regime will be replaced by crustal extension (Houseman and Molnar, 1997).

5. Conclusions

Lithospheric scale analogue models were constructed to investigate the parameters controlling initiation and subsequent evolution of intra-plate continental orogens. The experiments revealed that:

1. The presence of a strong domain effectively localizes deformation in continental compressional settings.
2. The final deep lithospheric architecture of the intra-plate orogen depends on the vertical rheological stratification of the continental lithosphere.
3. Folding is the main deformation mechanism is case of low convergence velocity and low lithospheric mantle strength, leading to a symmetric deep lithospheric structure. Intra-plate orogens characterized by such symmetric lithospheric root may experience later crustal extension triggered by gravitational instability.

4. Continental subduction-type geometry is favoured in case of strong crust–mantle coupling (high convergence velocities) or for cold lithosphere, where the presence of a strong upper lithospheric mantle controls localization of deformation. In this case the final architecture is strongly asymmetric. The development of asymmetric intra-plate mountain belts can be considered the rule, since a symmetric geometry is generated only under restricted conditions (low strength brittle mantle and low convergence velocity).

5. Localization of deformation and asymmetry in the Pyrenees can be linked to the presence of a strong and cold rift, high crust–mantle coupling and asymmetric folding of the lithospheric mantle.

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Appendix. Supplementary material

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