Control of crustal strength, tectonic inheritance and extrusion/stretching/indentation shortening rates on crustal deformation and basin reactivation: insights from laboratory models

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Abstract. Geological settings characterized by the simultaneous action of multiple coeval tectonic regimes provide a unique opportunity to understand complex interactions among different geodynamic processes. From an experimental point of view, these contexts remain comparatively less studied than areas with more simple patterns of deformation resulting from primary plate-boundary interactions. Here, we carried out analog experiments involving simultaneous shortening and orthogonal extension under different rheological conditions, and including the effect of crustal inheritance. We performed brittle experiments and brittle-ductile experiments to simulate cases of “strong” and “weak” crusts, respectively. We present two types of experiments: i) one-stage experiments with either shortening-only or synchronous orthogonal shortening and stretching, and ii) two-stage experiments with a first phase-stage of stretching and a second phase-stage with either shortening-only or synchronous orthogonal shortening and stretching. In our models, deformation is accommodated by a combination of normal, thrust, and strike-slip faults with structures location depending on boundary conditions and crustal inheritance. For brittle models, we show that the three types of structures can develop at the same time for intermediate ratios of extrusion over indentation/stretching (extension) over shortening rates (1.4< Vₑ/Vₛ<2). For lower ratios, deformation is accommodated by in-sequence shortening-orthogonal thrust faults, and stretching-orthogonal normal faults at the edges of the model (when Vₑ>0). For larger ratios and for the same amount of stretching, deformation is accommodated by normal faults at edges and in the center of the model as well as by conjugate strike-slip faults at the edges of the model. For brittle-ductile models, we observe either always observe strike-slip faults that cross-cut the entire model. They are associated with shortening-orthogonal thrust faults associated with conjugate strike slip faults (for models with low Vₑ/Vₛ and no initial extensional phase) or stretching-orthogonal normal faults associated with conjugate strike slip faults (for models with high Vₑ/Vₛ and initial extensional phase). Whatever the crustal strength, the past deformation history, and the extrusion/stretching/indentation shortening ratio, both normal and thrust faults remain with similar orientations, i.e. stretching-orthogonal and shortening-orthogonal, respectively. Instead, strike-slip faults exhibit variable orientations with respect to the indentation direction, which may be indicative of the strength of the crust and/or of the extrusion/indentation ratio shortening direction that vary between ∼0° and ∼65°. Strike-slip faults parallel to the shortening direction develop in previously extended portions of models with a brittle-ductile crust, while strike-slip faults with a...
high angle form at the boundaries of the brittle model, their orientation being to some extent influenced by pre-existing or newly forming graben in the center of the model. We also show that extensional structures formed during a first stage of deformation are never inverted under orthogonal shortening but can be reactivated as normal or strike-slip faults depending on $V_e/V_s$. The models replicate some deformation patterns documented in nature. Independently of the crustal rheology or the presence of crustal weaknesses, conjugate strike-slip faults develop along with variable Our experiments reproduce V-shaped conjugate strike-slip systems and normal faulting during compression indentation, reminiscent of tectonic escape processes along the Himalayas-Alpine chains similar to structures observed in the Tibetan Plateau, eastern Alps, western Anatolia, and the Central Asia orogen. Models with two-stage deformation show variable extensional to strike-slip reactivation of former extensional basins during basin-parallel shortening, which resemble synorogenic foreland transtensional reactivations documented in the Baikal and Golfo de San Jorge basins.

1 Introduction

Primary interactions Forces generated at plate boundaries and the derived stress fields field result in tectonic regimes characteristics of different geodynamic settings (e.g., Anderson, 1905). For instance, contractional regimes are more common in convergent margins in non-collisional (e.g., Central Andes, Barnes and Ehlers (2009)) and collisional orogens (e.g., Tibetan orogenic Plateau, Royden et al. (2008)) associated with fold-and-thrust belt development and crustal thickening. Extensional regimes are characteristic of divergent margins associated with mid-ocean ridges, intraplate regions in continental rifts (e.g., East African rift system, Chorowicz (2005); Basin and Range Province, Dickinson (2002)), and retreating subduction settings in intra-arc and backarc areas (e.g., Western Pacific marginal basins, Hilde et al. (1977)). Strike-slip regimes can be found in convergent settings where transcurrent faults run along magmatic arcs in contexts of oblique subduction (e.g., Liquiñe-Ofqui fault zone, Cembrano et al. (1996); the Great Sumatra fault, Berglar et al. (2010); Median Tectonic Line in Japan, Takagi (1986)), in hinterland regions of collisional orogens and related areas of tectonic escape (Tapponnier et al., 1982, 2001), in intraplate transcurrent regions (Molnar and Dayem, 2010), and at plate-boundary transform zones (e.g., Woodcock, 1986). In all settings, upper-plate weaknesses exert a major control in nucleation, reactivation, and orientation of structures (e.g., Sutherland et al., 2000; Tapponnier et al., 2001; Chorowicz, 2005; Pfiffner, 2017). Noteworthy, the three tectonic regimes can take place variably in all geodynamic settings, and at any one time during deformation, contrasting regimes may be active at different places (e.g., Harland and Bayly, 1958; Woodcock, 1986; Zoback, 1992). Cases where multiple tectonic regimes acted closely in space and time have long been recognized. (Fig. 1). The coexistence of thrust, strike-slip, and normal faulting has been documented in thick orogenic regions reaching crustal thicknesses above $\sim$60 km (e.g., Molnar and Tapponnier, 1978; Giambiagi et al., 2016), and in oblique convergent settings associated with strain partitioning (Chemenda et al., 2000). (Chemenda et al., 2000; Krézsek et al., 2013). It is also observed in areas of indentation tectonics and lateral escape (Tapponnier et al., 1982, 2001; Scharf et al., 2013), and synorogenic foreland rifting/transtension settings, where extension-transtension takes place in close spatiotemporal relation with plate-margin shortening (Sengör, 1976; Dézes et al., 2004; Gianni et al., 2015).
Comparatively, boundary replicate case North and associated aim to have Chain shortening, to::: an analog in have reproduced an African with devoted acted producing efforts simultaneously east of recent, strike-slip, studies:\n\n\n| 75 | 70 | 85 | 90 |
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- 75 taken into account, while it may be of importance in controlling the type and location of structures accommodating deformation triggered by the collisional far-field effects of India (Dhifaoui et al., 2021) have the close interrelation between the indentation of Arabia, lateral escape of Anatolia, and backarc extension in the Aegean region (e.g., Martinod et al., 2000; Sternai et al., 2014) (e.g., Martinod et al., 2000; Sternai et al., 2014; Philippon et al., 2014) and lateral escape and extension in the eastern Alps resulting from the indentation of the Adriatic plate (Ratschbacher et al., 1991).

Similarly and/or Pacific subduction (e.g., Tapponnier et al., 1982; Davy and Cobbold, 1988; Fournier et al., 2004) or the formation of the V-shaped south China oceanic basin (Le Pourhiet et al., 2018; Jourdon et al., 2020). In addition, brittle-ductile analog and 3-D numerical experiments have been applied to understand complex regional deformation resulting from the close interrelation between the indentation of Arabia, lateral escape of Anatolia, and backarc extension in the Aegean region (e.g., Martinod et al., 2000; Sternai et al., 2014) (e.g., Martinod et al., 2000; Sternai et al., 2014; Philippon et al., 2014) and lateral escape and extension in the eastern Alps resulting from the indentation of the Adriatic plate (Ratschbacher et al., 1991). More recently, Dhifaoui et al. (2021) analyzed through (Ratschbacher et al., 1991; van Gelder et al., 2017). Models imposing coeval orthogonal shortening and stretching with a brittle crust (Corti et al., 2006) or a brittle-ductile analog models a case study of simultaneous compression and lateral extension at the eastern boundary of the North African Alpine Chain crust (Dhifaoui et al., 2021) have also been carried out to understand the strain distribution along the Maghrebides-Apennines accretionary prism and the Sicily Channel rift, where coexisting graben, strike-slip, and thrusts faults formed a complex structural pattern in the general context of Nubia-Eurasia plates convergence. While these studies provide some elements for understanding the coexistence of different tectonic regimes and associated structures, there is a lack for a systematic investigation of the role of the relative ratio between shortening and stretching rates, as horizontal extrusion may not always only result from orthogonal indentation, but may also be controlled by far-field forces leading to non-plane strain deformation. In addition, the role of the strength of the crust, owing to its composition or to inherited structures, is not always taken into account, while it may be of importance in controlling the type and location of structures accommodating deformation (e.g., Munteanu et al., 2013; Zwaan et al., 2021) (e.g., Cruden et al., 2006; Munteanu et al., 2013; Zwaan et al., 2021).

In this study, we carry out a series of brittle and brittle-ductile analog experiments to gain insights into the role played by simultaneous shortening and orthogonal extension under different boundary and rheological conditions on the crustal tectonic
regime. We also analyze the role of crustal inheritance on fault reactivation and potential basin inversion during coeval shortening and lateral extension, which has not been explored so far. Although we do not intend to reproduce any specific natural case, we find some similarities between our experiments and deformation patterns in several natural cases, which provide additional insights into the rheological conditions and kinematics associated with their formation.

2 Laboratory models

2.1 Materials

We perform both brittle (one-layer) experiments (models BI) and brittle-ductile (two-layers) experiments (models CE) to reproduce a brittle upper crust and a brittle-ductile crust simulating cases of “strong” and “weak” crusts, respectively (Fig. 2). For both types of experiments, the 4-cm thick single or double layers rest on top of a foam layer (8-cm thickness). To simulate the brittle part of the crust, whose behavior is of Mohr–Coulomb type (Byerlee, 1978), we use the Fontainebleau quartz sand (NE34, Sibelco, France, D50 = 210 µm) (Klinkmüller et al., 2016). This material has a peak friction of 0.74 and an immediate reactivation friction (after 10 s) of 0.64 (Rudolf et al., 2022). Healing of this material increases the reactivation friction to 0.68 after 2.6 hours (maximum duration of our experiments) (Table 1). The cohesion (C) of Fontainebleau quartz sand is around 60-70 Pa (e.g., Klinkmüller et al., 2016; Schreuers et al., 2016). We sieve the dry sand from a height of ~15-20 cm, ensuring that its density is 1400 kg/m³ (Table 1). Dry sand exhibits frictional plastic behavior, and the geometry of structures that form when deformed does not depend on the applied strain rate.

The strength of the sand layer is calculated following Schellart (2000):

\[ \sigma_1 - \sigma_3 = (K - 1) \rho gz(1 - \lambda) + S \] (for compression) \hspace{1cm} (1)

\[ \sigma_1 - \sigma_3 = \frac{(K - 1)}{K} \rho gz(1 - \lambda) + \frac{S}{K} \] (for extension) \hspace{1cm} (2)

with \( S = \frac{2C \sin(90+\phi)}{1+\cos(90+\phi)} \), and \( K = \frac{1-\cos(90+\phi)}{1+\cos(90+\phi)} \), where \( \phi \) is the angle of internal friction (Fig. 2).

To simulate the ductile part of the crust, we use PDMS silicone whose density is 965 kg/m³ and viscosity \( \mu \) is around 3.5 \( \times 10^4 \) Pa.s (Table 1). This material has a Newtonian rheology (n=1) for strain rates lower than \( 10^{-2} \) s\(^{-1} \) (e.g., Rudolf et al., 2016; Guillaume et al., 2021), which is the case in our study where imposed strain rates are in the range 1.2-4.5\( \times 10^{-5} \) s\(^{-1} \). The strength of the layer of silicone putty (\( \sigma_1 - \sigma_3 \)) varies with imposed strain rates as:

\[ \sigma_1 - \sigma_3 = \eta \dot{\varepsilon} \] \hspace{1cm} (3)

where \( \eta \) is the viscosity and \( \dot{\varepsilon} \) the strain rate. For the values of applied strain rates, differential stress is on the order of 0.4-1.6 Pa. Initial strength envelopes for both types of models and different parts of the models are shown in Fig. 1.
The polyurethane foam RG35 used at the base of the model has a Poisson coefficient of 0.12 and allows producing a linearly varying velocity field (no velocity discontinuities) at the base of the deforming pile by compressing it to obtain shortening or letting it decompress to obtain stretching. However, there is a limitation on the amount of applied stretching/shortening. After 20% of shortening, the foam starts buckling and we therefore limit the amount of applied stretching/shortening under this threshold.

2.2 Scaling

We follow the scaling procedure shown e.g., in Zwaan et al. (2019) and based on Hubbert (1937); Ramberg (1981); Weijermars and Schmeling (1986). The scaling parameters are given in Table 2. Stress ratios between the laboratory and nature $\sigma^*$ are calculated as follows:

$$\sigma^* = \rho^* g^* L^*$$

(4)

where $\rho^*$ represents the density ratio, $g^*$ the gravity ratio, and $L^*$ the length ratio. Considering that we simulate the upper 15 km of the crust with our 4-cm thick pile of material, it gives $\sigma^* = 1.33 \times 10^{-6}$, i.e. that 1 Pa in the lab corresponds to 0.75 MPa in nature.

One may notice that for two-layered models, the density ratio between the brittle and ductile parts of the crust is high (1.45), leading to the two-layered models, the density ratio between the brittle and ductile parts of the crust is high (1.45), leading to increasing the density of the viscous layer would have resulted in a strong increase of its viscosity, which would have required applying speeds too low for the capacities of the engines used. We acknowledge that it leads to buoyancy forces that may trigger gravitational instability of the silicone layer and possible diapirism amplification of the folding of the brittle-ductile interface during shortening/stretching. However, given the relatively high viscosity of the silicone layer and the limited amount of deformation and short duration of these models (between 1.6 and 2.6 h), such a process may remain limited during the experimental time frame.

The strain rate ratio $\dot{\epsilon}^*$ is obtained as the ratio between the stress ratio $\sigma^*$ and the viscosity ratio $\eta^*$:

$$\dot{\epsilon}^* = \sigma^*/\eta^*$$

(5)

We assume a natural viscosity for the crust of $10^{21}$ Pa.s, within the range of proposed values under varying tectonic contexts ($\eta = 10^{19}$-$10^{23}$ Pa.s; (e.g., Buck, 1991; Brun, 1999; Bürgman and Dresen, 2008)). It gives $\dot{\epsilon}^* = 3.84 \times 10^{10}$, i.e. imposed strain rates correspond to strain rates of $0.32$-$1.18 \times 10^{-15}$ s$^{-1}$ in nature.

The time ratio $t^*$ can be obtained from:

$$\dot{t}^* = 1/t^*$$

(6)

It implies that 1h in the lab corresponds to 4.34 Ma in nature. Given that the duration of our models is at maximum 2.6 h, we simulate geological processes lasting for $\sim$11 Ma at maximum.

The velocity ratio $v^*$ is obtained from:

$$\dot{v}^* = v^*/L^*$$

(7)
The imposed values for extension/shortening rates between 0 and ~60 mm/h in the lab corresponds to 0-5.2 mm/yr in nature, typical values for continental rifting (e.g., Saria et al. , 2014), subduction-related (e.g., Andes, Oncken et al. (2006); Alps, Sternai et al. (2019); Zagros, Tatar et al. (2002)) or intra-continental collision orogens (e.g., Pyrenees, Moutheau et al. (2014); Tian Shan, Saint-Carlier et al. (2016)).

The dynamic similarity between our experiments and the natural case for the viscous regime is verified by computing the ratio between lithostatic pressure and viscous strength (Ramberg number $R_m$):

$$R_m = \frac{(\rho g L^2)}{(\eta v)}$$

which gives a value of ~37 considering the maximum deformation velocity for both the model and the natural case (Table 2). For the brittle regime, the dimensionless friction coefficient is similar (0.6-0.7) in the laboratory and in nature. The ratio between gravitational stress and cohesive strength $R_s = \rho g L / C$ is achieved also similar with values ranging between 7.85 and 9.16 considering a cohesion in nature of 45-52 MPa (Jaeger and Cook, 1976; Raleigh and Paterson, 1965; Twiss and Moore, 1992; Handin, 1969).

2.3 Procedure and analysis

The 60 x 60 x 8 cm layer of foam is initially compressed in one direction and is maintained in this state for the rest of the preparation phase. For the two-layers models, we place a pre-cut silicone plate with dimensions of 50 cm x 50 cm x 2 cm. Seeding (Fig. 2). Sieving of the sand is performed by depositing it from a distance of ~15-20 cm and leveling it above the PDMS layer and the sand is leveled with a rigid plate until the desired thickness is achieved. The brittle part of the crust is made of white quartz sand that is randomly sprinkled on top with black colored sand in order to allow particle detection for digital image correlation. Before deformation, the models cover an area of ca. 46 cm x 46 cm that would represent an area of 172 km x 172 km in nature.

We also include “seeds” in our models to help localize deformation. They may represent weak zones inherited from previous phases of deformation. This weaker zone in our experiments is also compatible with the requirement of a finite width low-viscosity zone underneath the fault zones possibly caused by grain size reduction, shear-heating, and localized presence of fluids (e.g., Le Pourhiet et al., 2014). These seeds are linear pieces of silicone that are placed on top of the silicone layer, or directly on top of the foam for the brittle models (Fig. 42). They are placed at the center of the models, orthogonal to the extension direction, have a rectangular shape and extend on the entire length of the model. The dimensions of the seed are 46 cm x 1.5 cm x 1.5 cm for the brittle-ductile models and 46 cm x 1.5 cm x 1 cm for the brittle models (Fig. 42). The strength of the crust is decreased at these locations owing to the reduced thickness of the overlying sand layer (Fig. 42), which in turn may help deformation to localize.

The layer(s) are then deformed by applying a constant velocity boundary condition at the edges of the model through pistons controlled activated by step motors. Stretching (extrusion), which allows us to precisely control the stretching rate to shortening rate ratio. Stretching is applied on both edges of the models by letting the foam decompress while shortening (indentation) is...
only applied at one side of the models, the other side having a no-motion boundary condition (Fig. 1). We arbitrarily consider the non-moving wall as the north in our experiments. We performed two types of experiments: i) one-stage, one-stage experiments with either shortening-only or synchronous orthogonal shortening and stretching, and ii) two-stage, two-stage experiments with a first phase of 5% stretching and a second phase with either shortening-only or synchronous orthogonal shortening and stretching, in order to study the possible reactivation/inversion of structures formed during the first stage (Fig. 2). Applied velocities for stretching vary between 0 and 75 mm/h and for shortening between 0 and 43 mm/h (Fig. 3).

We do not include surface processes in the models (erosion, deposition), especially in between the two phases of deformation, meaning that the created grabens remain unfilled when the second phase of deformation starts. While we acknowledge that redistribution of mass associated with surface processes may impact stress distribution and further deformation (e.g., Mugnier et al., 1997; Pinto et al., 2010), we wanted to ensure similar conditions between models that are difficult to achieve when manually intervening during the course of the experiment.

Experiments are recorded from the top by a DSLR camera (Nikon D3300) taking pictures every 2 minutes. Pictures are then automatically analyzed using an image cross-correlation technique, Particle Image velocimetry (PIV), using the PIVlab software (Thielicke and Sonntag, 2021). We pre-process the images with a CLAHE filter with a window size of 64 px to enhance contrast in the pictures and allow particle detection. PIV analyses are made with the FFT window deformation PIV algorithm in 3 passes with interrogation areas of 128, 64, and 32 pixels, and with a step of 50%. PIV results are then calibrated using spatial scales set on top of the models. We obtain velocity maps with a spatial resolution of 16 pixels, corresponding to ~4 mm (~1.5 km in nature).

Velocity fields obtained from the PIV analyses are then processed with the strain map, StrainMap algorithm (Broere et al., 2021) that allows tracking the cumulative deformation field and as such to map the distribution of deformation over time. In particular, this algorithm is based on the description of shape changes in terms of Hencky strains. It allows to discriminate co-existing strike-slip faults, thrust faults and normal faults, and their evolution over time, and as such efficiently complement inherently subjective visual inspection. Videos and strain analysis of the twelve experiments are available in Guillaume et al. (2022).

3 Results

3.1 Brittle-only models: role of inheritance under varying stress fields

3.1.1 Inheritance as a crustal heterogeneity (seed)

We investigate the role of a crustal heterogeneity (basal seed orthogonal to the stretching direction) on deformation location distribution under different kinematic boundary conditions – for a brittle crust (Fig. 2A). For this, we performed a series of five single-stage models in which the ratio of stretching velocity (extrusion rate) over shortening velocity (indentation rate) $V_e/V_s$ is varied between 0 (no stretching) and 2.8 (stretching dominated) – (Fig. 3). We compute the principal stretches $\lambda_{max}$
and $\lambda_{\text{min}}$ and the corresponding strain type for the five models after 4% of shortening (SM1) and 10% of stretching (Fig. 3) for models BI05 to BI09 and 12% of shortening for the model BI10 with zero stretching (Figs. 4 and SM2).

For the model with shortening-only (BI10), deformation is first concentrated along an E-W striking thrust fault located $\sim$10 cm from the moving piston (fault 1 in Figs. 4A and SM1A). After 4% of shortening, the thrust has a linear shape, orthogonal to the shortening direction (Fig. SM1A). Increase in the amount of shortening leads to the activation of three successive thrusts in a prograde sequence (faults 2, 3 and 4 in Fig. 4A).

For the model with $V_e/V_s = 0.9$ (BI09), deformation is also mainly accommodated by E-W thrust faults developing in a prograde sequence (Figs. SM1B and 4B). However, unlike the previous model, the shape of the thrust front is not linear but rather slightly convex toward the north. In this model, deformation is also accommodated by N-S extensional faults located close to the edges of the model (Figs. SM1B and 4B). There are no traces of significant extension above the central seed.

For the model with $V_e/V_s = 1.4$ (BI05), after 4% of shortening (5.5% of stretching), deformation is partitioned between shortening that is accommodated along an E-W thrust fault, diffuse extension in the retro wedge, and strike-slip faults at the corners of the model (Fig. SM1C). After 10% of stretching (7.7% of shortening), the pattern of deformation has evolved with: i) stretching in the center of the model accommodated by N-S conjugate normal faults forming a 5.2-cm large graben structure, ii) shortening accommodated by a second thrust fault, and iii) conjugate strike-slip structures that make the connection between the frontal thrust and the central graben, but also that deform the southern wedge (Fig. 4C). In the distal part of the retro wedge, other strike-slip faults are visible with a sense of shear that is compatible with previous strike-slip faults but with orientations that largely differ (N66 and N114 in the northern sector vs. N25-N40 and N145-N150 in the south central sector) (Fig. 4C).

For the model with $V_e/V_s = 1.9$ (BI06), the N-S striking central graben is already formed after 4% of shortening (7.6% of stretching) and deformation is also localized along conjugate strike-slip faults (Fig. SM1D). Instead, shortening is not accommodated by discrete E-W thrust faults but rather corresponds to a zone of diffuse deformation. After 10% of stretching (5.1% of shortening), the normal faults remain active and strike-slip faults propagate toward the central graben (Fig. 4D). Like in the model BI05, a transition zone with strike-slip faults develops in between the wedge and the central graben.

For the model with $V_e/V_s = 2.8$ (BI07), the evolution of deformation is almost similar to the previous experiment (Figs. SM1E and 4E). The main difference is visible after 10% of extension: there are no strike-slip faults in the transition zone between the mildly shortened area (shaded area in Fig. 4E) and the central graben.

### 3.1.2 Inheritance as a former extensional phase

In the following models, the set-up is identical to previous models except that they undergo an initial phase of E-W extension with up to 5-5.3% of stretching, prior to a second phase of N-S shortening (and possible coeval E-W stretching for models BI08 and BI11). We test three kinematic boundary conditions for the second phase of deformation: $V_e/V_s = 0$ (BI01), $V_e/V_s = 1.4$ (BI08), $V_e/V_s = 2$ (BI11) (Fig. 4F to 4H) that should be compared with models BI10, BI05, and BI06, respectively.
For model BI01 with $V_e/V_s = 0$, the first phase stage of stretching results in the development of N-S conjugate normal faults and an associated 5.3-cm large graben (Fig. 4F). In the second phase stage of deformation with N-S shortening only, a first thrust develops. After 4% of shortening, deformation is accommodated along a single thrust (Fig. 4F). In between 4 and 12% of shortening, deformation is accommodated along new thrusts in a prograde sequence (Figs. 4E and 5A). However, these thrusts do not cut through the entire width of the models like in model BI10 but rather branch on previous thrusts (Fig. 4E). Interestingly, the intersections between the new and former thrusts are close, but not exactly coincide, to the limits of the downlifted central area central graben.

For model BI08 with $V_e/V_s = 1.4$, the first phase stage of extension also results in the formation of a central N-S striking graben, similar to model BI01. However, in the second phase stage of deformation, deformation is accommodated by the coeval activity of N-S normal faults, an E-W thrust fault and NW-SE and NE-SW conjugate strike-slip faults that develop in the wedge, at the transition between the frontal thrust and the central graben and at the northern boundary of the model (Fig. 4G). The pattern of deformation after 4% of shortening differs from model BI05 in that after the same amount of shortening, with similar boundary conditions but no initial stage of stretching (Fig. 4C). In model BI05, localization of extensional structures and strike-slip faults was not achieved—is not yet achieved (Fig. SM1C). In addition, the shape of the frontal thrust is convex toward the north in the model with an initial stage of stretching while it is linear in the single stage model (Figs. SM1C and 4G). It is only after 10% of extension stretching in the single stage model (BI05) that the overall pattern of deformation appears almost similar. However, second order differences remain: i) the shape of the frontal thrust is— with the activity of strike-slip and normal faults, and the convex toward the north in the model with an initial phase of stretching, ii) shape of the newly formed thrust (fault 2 on Fig. 4C). However, second-order differences remain: the strike-slip faults that develop in the wedge have orientations that slightly differ: N55 (sinistral) and N120 (dextral) in model BI08 instead of N40 and N140 in model BI05 (Figs. 4C and 4G).

For model BI11 with $V_e/V_s = 2$, stretching dominates in the model and is accommodated through N-S normal faults, whose activity pursues during the second phase stage of deformation (Figs. 4 and 5). Evidence 4H and 5C. Evidence of significant shortening localization are— is— lacking, but the experiment stopped after only 2.2% of shortening. The imposed kinematic boundary conditions also resulted in the development of conjugate strike-slip faults at the southern and northern boundaries of the model (Fig. 4H).

3.2 Brittle-ductile models: role of inheritance under varying stress fields

3.2.1 Inheritance as a crustal heterogeneity (seed)

The model CE16 has been performed to test how the deformation distribution evolves as a function of the strength of the crust, by including a ductile layer in the model — (Fig. 2B). It is comparable with the brittle-only model BI09, which shares similar boundary conditions ($V_e/V_s = 0.9$). After 4% of shortening (and 2.8% of stretching), deformation is accommodated by a combination of diffuse shortening along a 4-cm large E-W band (Fig. 66A) and conjugate strike-slip faults at the corners of the models. Unlike model BI09, there are no N-S normal faults at this stage. After an amount of— 10% of stretching (and
14.2% of shortening), the area of diffuse shortening has evolved into a localized thrust fault. In addition, N-S shortening is also accommodated along another thrust with northward dipping. This northward dipping back-thrust, forming an uplifted wedge (Fig. 6A).

3.2.2 Inheritance as a former extensional phase

The model CE17 is similar to model CE16 except that we impose an initial phase of 4.7% E-W stretching. This first phase of stretching results in the coeval localisation of strain along three pairs of N-S trending conjugate normal faults: one in the western part of the model, one in the eastern part of the model and the last one in the central part, eastern and central parts of the model, the latter being located just above the seed (Fig. 6B). During the second phase of deformation with coeval shortening and stretching ($V_e/V_s = 0.9$), some of the pre-existing normal faults are reactivated as normal faults, other as shortening-parallel strike-slip faults, while additional conjugate strike-slip faults develop and branch at the corner of the models (Figs. 7A and 7B). Increase in the amount of shortening does not lead to significant localisation of deformation on reverse faults but rather leads to the development of new strike-slip faults (Fig. 6B). The brittle-only model whose boundary conditions are the closest to model CE17 is model BI08 (Fig. 4G). Despite a larger $V_e/V_s$ ratio in model BI08 that would favor stretching over shortening, the brittle-only model exhibits a clear E-W thrust fault that is not observed in the brittle-ductile model.

Models CE18 and CE20 share a similar set-up with CE17 but explore different boundary conditions during the second phase of deformation with $V_e/V_s = 0.7$ (Fig. 6C) and $V_e/V_s = 1.1$ (Fig. 6D), respectively. We describe deformation after 5% of stretching during the second phase of deformation (Fig. 9, Figs. 6C and 6D). For the model dominated by shortening (CE18), the two main differences with model CE17 are i) the largest extent of zones characterized by diffuse shortening (Fig. 9) that is maintained during the entire second phase of deformation (Fig. 8B) and ii) the limited sections of pre-existing normal faults that are reactivated as normal faults, which are only found in a restricted area over the central seed (Fig. 6C). In addition, in the eastern part of the model one of the N-S normal faults is reactivated as a dextral strike-slip fault (Fig. 6C). For the model dominated by stretching (CE20), the pattern of deformation is clearly different: large portions of the pre-existing western, eastern and central normal faults are reactivated as normal faults during the second phase of deformation (Fig. 6D). Only in the central portions of the western and central faults deformation is accommodated along strike-slip faults that are aligned with the initial orientation of normal faults (point b2 in Fig. 8C).

4 Discussion

4.1 Tectonic regime distribution under non plane-strain conditions

Our models confirm that under plane-strain conditions, i.e. no strain in one of the horizontal directions (models BI01, BI10 and first stage of models BI08, BI11, CE17, CE18, and CE20), the crust accommodates deformation through structures orthogonal
to the direction of transport, i.e. either normal faults when the maximum principal stress \(\sigma_1\) is vertical or thrust faults when the minimum principal stress \(\sigma_3\) is vertical. However, when boundary conditions satisfy non plane-strain conditions (i.e. strain occurs along the three principal axis), deformation is accommodated along a combination of normal faults, thrusts faults and strike-slip faults, whose location and presence/absence appear to depend on the applied boundary conditions and pre-existing heterogeneities and possibly the amount of accumulated deformation.

In model B105 for instance (brittle model with \(V_e/V_s = 1.4\)) the three modes of deformation are active at the same time with combined-exemplifies the possible coeval activity of N-S conjugate normal faults, W-E-W thrust faults, and NE-SW and NW-SE conjugate strike-slip faults (Fig. 10). Interestingly, \(10\%\) of E-W stretching and \(7.7\%\) of N-S shortening, in the center of the model, the change of deformation mode is progressive from extensional structures to the north toward type of structures evolves from compressional structures to the south. An analysis of the stress state suggests that different tectonic regimes are found within the model: extensional tectonic regime where \(\sigma_1\) is vertical, wrench (strike slip) regime where \(\sigma_2\) is vertical and compression regime where \(\sigma_\perp\) is vertical (Anderson, 1905) (close from the piston) to extensional structures to the north. The transition from one regime to another therefore results from the permutation of one of the horizontal principal stress axes with the vertical one, controlled by the relative magnitude between the principal stresses as described by \(b = (\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3)\). A change from extensional to wrench regime is favored for high \(b\) values while a change from wrench regime to compression regime is favored for low \(b\) values. The Mohr-Coulomb analysis for a frictional material like sand also implies that the differential stress necessary for material failure is maximum for thrust faults (Fig. 4). This could explain why thrust faults are only found close to the southern moving piston where the maximum horizontal stress is applied (zone 31). In this area, \(\sigma_1\) is horizontal and N-S oriented, while the minimum stress \(\sigma_3\) is vertical. Going further north from the piston (zone 2), \(\sigma_1\) is still N-S oriented but its magnitude decreases, which requires implies a lower \(\sigma_3\) to be lower when the material fails. As a consequence, the at failure and a vertical intermediate principal stress \(\sigma_2\) becomes vertical, favoring the development of strike-slip faults. Further north (zone 43), the normal faults that were created during the first phase stage of extension are reactivated as normal faults, thus indicating that the principal stress that is oriented in the \(\sigma_1\) is vertical. Progressive increase in the southern wedge thickness accompanying N-S direction is no more \(\sigma_1\) but \(\sigma_2\), and that in turn \(\sigma_1\) is vertical—shortening in the absence of erosion would imply an increase of the vertical stress. As a consequence, thrust faults would propagate toward the north (as evidenced between 4.2 and 7.7% of shortening for model B105; Figs. SM1C and 4C) and could possibly reach places where strike-slip faults and normal faults were previously active. Therefore, the redistribution of stress resulting from deformation could drive temporal variations of the tectonic regime at a specific location.

For brittle models, the coexistence of the three types of structures is only possible for intermediate (after 10% of E-W stretching) starts for \(V_e/V_s\) ratios \(\geq 1.4 < V_e/V_s < 2\) (Fig. 4). For lower \(V_e/V_s\), \(\sigma_1\) is horizontal and N-S oriented and maximum close to the piston applying shortening, favouring the development of thrust sequences in the southern part of the models. Close to the western and eastern pistons applying stretching, in the case where \(V_e/V_s \neq 0\), the vertical stress becomes the maximum stress and N-S normal faults develop. For values of \(V_e/V_s \geq 2\), the W-E-W horizontal stress is low enough in the entire model to maintain as the minimum stress \(\sigma_3\), which in turn implies that even if the N-S horizontal stress is the maximum stress \(\sigma_1\), the crust cannot fail along E-W thrust faults, but fail along strike-slip faults. However, we cannot
preclude that thrust faults could also develop at later stages for these models with high $V_e/V_s$ ratios, as experimental limitations prevent us from imposing large amounts of N-S shortening. In model BI07 for instance, it is at maximum 3.6%, which may be insufficient to locate deformation along an E-W thrust fault.

For brittle-ductile models, over the range of tested $V_e/V_s$, we do not observe the coexistence of the three types of structures within the same model (Fig. 6). We only observe E-W thrust faults associated with conjugate strike-slip faults for models with low $V_e/V_s$ and no initial E-W extensional phase. This implies that the conditions where both horizontal stress conditions required to develop N-S normal faults (E-W stress is low enough to become horizontal stress is $\sigma_3$ and the vertical stress high enough to become vertical stress is $\sigma_1$ are never satisfied) are never encountered in the model. The first condition is met as testified by the presence of strike-slip faults. Instead, the second condition is not met, which could partly be explained by the limited thickness of the brittle portion of the crust, which is half that of the brittle-only models. For models with an initial E-W extensional phase, we observe at the final stage of deformation a combination of N-S normal faults and conjugate strike-slip faults, and no clear W-E-F-W thrust faults, even when $V_e/V_s$ is low (model CE18). For the latter, shortening is active in some areas of the model (Fig. 9) but not 6C and point 3 in Fig. 7B), but it does not localize along discrete structures. For weak crusts, crustal thinning associated with an initial phase of extension therefore appears to control the future development of thrusts during subsequent phase of combined shortening/stretching. However, the amount of applied deformation remains limited with only 5% of horizontal E-W stretching (and $\sim$10% of horizontal N-S shortening) and we cannot preclude that further shortening would eventually result in thrust development.

4.2 Compatibility of the structures with principal stresses orientation

In the brittle-only models, a majority of the structures are consistent with the Coulomb fracture criterion considering either a N-S oriented horizontal $\sigma_1$ or a W-E-F-W oriented horizontal $\sigma_3$. In particular, thrust faults are W-E-F-W oriented, normal faults N-S oriented and strike-slip faults generally oriented at $\sim$30° from a N-S $\sigma_1$. However, in the northern part of the models, strike-slip faults do not obey this criterion and exhibit larger angles with respect to a N-S horizontal $\sigma_1$, with values around 60-70° (e.g., model BI05, Figs. 3 and 14C and 9A). These anomalously oriented strike-slip faults could result from a combination of factors. As they nucleate from the edge of the model, we cannot exclude that they result from some unwanted boundary effects associated with the high friction wall-sand interface. As a result of the applied stretching and boundary effects, the maximum principal stress $\sigma_1$ may have rotated from a N-S direction toward a NE-SW direction in the eastern part of the model and NW-SE direction in the western part of the model, possibly explaining why these strike-slip faults do not lie at 30° with respect to the imposed N-S shortening. However, one can also notice that not all of these strike-slip faults have the same exact orientation (Figs. 4D and 4E), some of them being directed toward the northward termination of the normal faults bounding the central graben. Their geometry therefore appears to orientation could also be controlled by the graben structure that forms in the center of the model above the crustal seed.

Interestingly, in brittle models with an initial phase of extension (e.g., model BI08, Figs. 4 and 14G and 9B), not only the northern strike-slip faults orientation departs from the orientation expected from the Coulomb fracture criterion with a N-S shortening, but also some of the strike-slip faults that develop above the wedge in the southern part of the model.
These While we cannot preclude some boundary effects here too, these faults with a larger than expected angle with respect to \( \sigma_1 \) also connect with the normal faults bounding the central graben formed during the initial phase stage of stretching.

These results highlight the fundamental role of pre-existing structures. In comparison, model BI05, which shares the same stretching/shortening ratio and almost the same amount of total stretching (\( \sim 10\% \)) does not show any anomalous strike-slip faults in the southern part of the model (Fig. 9A). Pre-existing structures may also exert a control on the geometry of subsequent structures even for areas close from where shortening is applied.

For models with a brittle-ductile crust, strike-slip faults are expected to range in between 30\(^\circ\) and 45\(^\circ\) following the Coulomb criterion and the slip-line theory (e.g., Anderson, 1905; Tapponnier and Molnar, 1976). Most of the strike-slip faults that cross-cut the models indeed show strikes compatible with the overall state of stress, i.e. N30-N45 for left-lateral strike-slip faults and N135-N150 for right-lateral strike-slip faults. However, in models with an initial phase stage of stretching (e.g., model CE20, Figs. 9 and 116D and 9C), strike-slip faults also develop with orientations anomalous orientations. In particular, at model corners, strike-slip faults bend with angles that become larger with respect to the N-S shortening direction, possibly indicative of some boundary effects. More interestingly, some strike-slip faults are oriented almost parallel to the shortening direction. This clearly indicates in areas that were previously affected by normal faulting (Figs. 6 and 9C). This clearly indicates that inherited structures, here in the form of previously developed N-S normal faults, can control the subsequent location and geometry of structures.

Overall, our models show that whatever the strength of the crust, its past deformation history and the relative ratio of extrusion rate over indentation stretching rate over shortening rate, both normal faults and thrust faults remain with similar orientations, i.e. N-S and W-E-E-W, respectively. Instead, strike-slip faults exhibit a wider range of possible orientations with respect to the indentation (or extrusion shortening (or stretching) direction. As such they could give us insights into the tectonic context in which they formed. In particular wide angle conjugate, strike-slip faults (with angles of up to \( \sim 65\% \)) with respect to indentation direction) -N-S shortening direction are found for strong crusts. They particularly form far away from the indentation shortening location when the crust has not been previously deformed, but their anomalous orientation may be indicative of a local perturbation of the stress field owing to a possible combination of boundary effect, dominance of stretching in this area, and coeval formation of a graben in the center of the domain. High angle strike-slip faults can also be observed closer from the indentation location to the area where shortening is applied if the crust has been previously extended. Instead, for weaker crusts, strike-slip faults branch to the former location of normal faults that bounded the central graben. For weak crusts, strike-slip faults can even be parallel to the indentation shortening direction in the case they reactivate former extensional structures.

### 4.3 Role of crustal strength in distribution of crustal deformation

Under comparable boundary conditions, the strength of the crust plays a fundamental role in controlling the location of deformation and the types of structures that accommodate deformation. Models without initial stretching phase and similar \( V_t/V_s \) (e.g., models BI09 and CE16, \( V_t/V_s = 0.9 \)) exhibit significantly different patterns of deformation (Fig. 12 Figs. 10A and 10B). A strong crust (thick brittle part) favors the development of in-sequence E-W thrust faults verging to the north to accommodate
indentation while extrusion shortening while stretching is accommodated through N-S normal faults at the western and eastern boundaries of the model (Fig. 4B). Instead, when the entire model is made of a weak crust, indentation shortening results in a doubly-vergent system of conjugate E-W thrust faults and in large-scale conjugate strike-slip faults that also accommodate the extrusion stretching. The lack of normal faults indicate that the horizontal stress remains high enough in the entire model to prevent the vertical stress to become \( \sigma_1 \). This can readily be explained by the decrease by a factor two of the thickness of the brittle crust in the “weak” models. Interestingly, in the strong model (BI09), there is also a variation in the thickness of the brittle part of the crust owing to the presence of the thin layer of silicone corresponding to the viscous seed in the center of the model. However, it does not result in significant lateral variations in deformation style, due to the smaller variation in brittle thickness (decrease by 25%) and/or to the fact that the area of relatively weak crust only represents a small fraction of the entire model and that deformation is therefore controlled by the rest of the model (~1.2% of the crustal volume in model BI09 instead of 51.2% in model CE16). A more systematic change in the width and thickness of the weak part of the crust would be required to isolate the main controlling parameter.

Interestingly, model CE17, which shares the same boundary conditions as model CE16 but with an initial phase of W-E stretching stage of E-W stretching, exhibits a significantly different distribution of deformation during the second stage of deformation. This initial phase of stretching locally modifies the crustal strength by creating zones of thinned ductile and brittle portions of the crust accommodating the extension (Fig. 42). This has fundamental impact on the subsequent deformation pattern. Indeed, these 10C. These areas of even weaker crust concentrate deformation during the second stage through normal faulting or N-S strike-slip faulting that both were absent in model CE16. In turn, there are no thrust faults accommodating the indentation shortening.

Experimental results again highlight the importance of past tectonic history and associated inherited structures and changes of crustal strength on the distribution of deformation and the types of structures that form during subsequent phases of deformation (Fig. 4311). As such, the association of specific types of faults cannot be used as an indicator of the relative rate of indentation and extrusion shortening and stretching, unless previous history of deformation is properly constrained.

4.4 Basin reactivation under orthogonal-normal fault-parallel contraction

Many modeling studies have previously investigated deformation associated with the inversion of extensional half-grabens subject to subsequent contraction (e.g., McClay, 1995; Bonini et al., 2012; Zwaan et al., 2022, and references therein). In general, the applied directions of stretching and successive contraction are parallel (in 2D numerical or analog models) or oblique, but at low angles. Here, inspired by the San Jorge basin in Patagonia where some However, in nature this may not always be the case. In Patagonia for instance, Cretaceous extensional structures were orthogonal to the Cenozoic contraction direction, we formed in the San Jorge basin as a response to N-S stretching. This area has subsequently undergone E-W Cenozoic contraction, parallel to the direction of the normal faults. We therefore more specifically explored the conditions leading to reactivation of former extensional structures during successive orthogonal the reactivation of extensional structures under fault-parallel contraction (and parallel possible fault-orthogonal extension).
From our experimental dataset, there are no clear evidence that normal faults lying parallel to the contraction direction are inverted as contractional structures during the second phase of deformation: stage of deformation (Fig. 11). While these areas are favorable to further contraction owing to their thinned crust and associated low vertical stress, the angular relationship between pre-existing normal faults and the direction of contraction (∼0°) prevents the normal faults to be inverted. Instead, indentation is accommodated along contraction-orthogonal new thrust faults. We further show that if extrusion stretching orthogonal to the contraction direction is inhibited, there is no reactivation of pre-existing normal faults as normal faults or strike-slip faults (model BI01). Instead, allowing orthogonal extrusion stretching always results in the reactivation of normal faults. However, different regimes are found depending on the boundary conditions and crustal strength. In models with low $V_c/V_s$ and a weak crust (model CE18), extensional structures are preferentially reactivated as strike-slip faults, participating in the accommodation of the indentation shortening that dominates. As $V_c/V_s$ increases and becomes closer to 1, pre-existing normal faults are reactivated as either normal faults or strike-slip faults (models CE17 and CE20). Both types of faults therefore develop and are juxtaposed as almost parallel structures without involving changes in boundary conditions. Finally, when extrusion-stretching rate dominates over indentation shortening rate, pre-existing normal faults are reactivated as normal faults (models BI08 and BI11).

4.5 Application to natural cases and comparison with previous models and application to natural cases

In our experiments, we did not intend to reproduce any specific natural case but rather systematically test internal and external parameters. However, the analog models reproduce some of the deformation patterns documented in, the observed deformation patterns resemble those documented in certain areas where deformation is accommodated by different combinations of normal, thrusts, and strike-slip faults. Hence, our experiments can be helpful to shed some light on the kinematics and rheological conditions behind these intricate geological settings.

In general, these complex structural frameworks take place during continental plate convergence, indentation, and subsequent lateral escape of lithospheric blocks during continental collisions associated with continental collision (Tapponnier et al., 1982) (Fig. 14A). From our models, we observe that at least two different deformation regimes coexist when $V_c/V_s$=0.7-0.9 and the three regimes operate simultaneously mostly when $V_c/V_s$ is ≥1.4 (Figs. 3, 4, 5, and 6, 7, 8, and 9). The latter could indicate a minimum value for this ratio for effective lateral escape to occur. In nature, the tectonic escape of crustal blocks takes place along conjugate strike-slip faults, commonly referred to as indent-linked strike-slip faults (Woodcock, 1986) or V-shaped conjugate strike-slip faults (e.g., Yin, 2010, and references therein) that take place along with variable degrees of crustal extension within the escaping crustal blocks. Typical V-shaped conjugate faults have been documented in the Eastern Anatolia region (e.g., Şengör et al., 1985; Dhont et al., 2006; Hisarlı et al., 2016, Fig. 14A) (e.g., Şengör et al., 1985; Dhont et al., 2006; Hisarlı et al., 2016), the eastern Alps (e.g., Ratschbacher et al., 1991; Scharf et al., 2013, Fig. 14B) (e.g., Ratschbacher et al., 1991; Scharf et al., 2013, Fig. 14B), the Tibetan plateau (e.g., Şengör and Kidd, 1979; Tapponnier et al., 1982, Fig. 14C) (e.g., Şengör and Kidd, 1979; Tapponnier et al., 1982, Fig. 14C), and the central Asian intraplate region (e.g., Cunningham, 2005; Yin, 2010, Fig. 14D) (e.g., Cunningham, 2005; Yin, 2010, Fig. 14D). Previous mantle-scale brittle-ductile analog experiments successfully reproduced the formation of V-shaped conjugate faults during continental indentation (e.g., Davy and Cobbold, 1988; Ratschbacher et al., 1991; Martinod et al., 2000; Fournier et al., 2004).
However, a direct comparison with our models is difficult because these previous models were aimed at simulating larger domains (several hundreds km) and the lateral escape of material was not kinematically controlled. Instead, 3D lithospheric-scale numerical models by Le Pourhiet et al. (2014) imposing coeval orthogonal extension and shortening at the edges of the models also show that deformation is accommodated along strike-slip faults with a degree of localization that increases with a decrease of the viscous strength of the lower crust. Results obtained in our models at the crustal scale could therefore be possibly upscaled at the lithospheric scale.

In our models, whether brittle or brittle-ductile, we also observe the development of conjugate strike-slip faults that are accompanied by different degrees of normal faulting during compression, which are reminiscent of those structural systems seen in nature (BI05-08, Figs. 3 and 4C and 4G; CE 16-20; Figs. Fig. 6, 7, and 9). We nevertheless note that experiments including an overall weaker crust, simulated by adding a silicon layer as an analog for ductile lower crust materials, are more prone to have well-developed conjugate strike-slip systems that cross-cut the entire model (CE 16-20; Figs. 6, 7, and 9). Our results are partly similar to those by Dhifaoui et al. (2021) (model M-3), in which crustal-scale brittle-ductile analog models were subjected to simultaneous shortening and orthogonal extension at $V_e/V_s=1$ imposed by one piston in each direction. Despite the slightly different boundary conditions (stretching being allowed only along one boundary), both approaches show crustal escape through a V-shaped strike-slip system. Instead, models by Corti et al. (2006) that used a similar set up than Dhifaoui et al. (2021) with extension applied on one wall during compression but with a brittle crust, do not show V-shaped conjugate strike-slip faults. Interestingly, in these experiments, the authors used a low $V_e/V_s$ value of 0.2. The absence of strike-slip faults is therefore consistent with our minimum value ($V_e/V_s > 0.9$) for effective lateral crustal escape. Strike-slip fault systems active during compression have also been previously addressed by brittle analog models (Duarte et al., 2011; Rosas et al., 2012, 2015). However, in these studies, strike-slip fault strike and activity were constrained by the chosen set up that included confined walls and prescribed fault direction along a mobile preexisting discontinuity. Because of these constraints, the experiments did not develop typical V-shaped strike-slip systems.

In our models, we did not include the effect of compression obliquity, which has been acknowledged as an important factor driving simultaneous conjugate strike-slip systems and orthogonal extension during indentation (Rosenberg et al., 2007). Also, we note that experiments including an overall weaker crust, simulated by adding a silicon layer as an analog for ductile lower crust materials, are more prone to have well-developed conjugate strike-slip systems that cross-cut the entire model (CE 16-20; Fig. 6). This is compatible with previous findings from analog models by Cruden et al. (2006) that explored the role of orogen parallel flow and rheological stratification on vertical and lateral development of structures in hot orogens. These authors simulated a hot orogenic crust by including a weak lower crust with a variable buoyancy and a lithospheric mantle that jointly allowed orogen parallel flow during compression. In these experiments, the development of conjugate strike-slip faults within the orogen appeared to be a common feature in models with buoyant ductile lower crust whereas, in models with a non-buoyant lower crust, conjugate shear zones were generally absent, except at the outer regions of the extruding orogen.

In general, our results are compatible with natural examples analyzed in Figs. 4A, B, C, DFigs. 12A, 12B, 12C, and 12D, where Cenozoic conjugate strike-slip systems formed onboard thick and hot, and hence, weak orogenic crusts of the Anatolian and Tibetan plateaus, Central Asian intraplate orogenic system, and the Alpine orogen. Therefore, a weak crust is an impor-
tant factor contributing to lateral tectonic escape in those settings by allowing the formation of fault-bounded brittle upper crust blocks translated laterally by strike-slip systems and constrictional ductile flow in the lower crust (Ratschbacher et al., 1991; Scharf et al., 2013).

In models that include an initial phase of extension, and under the applied boundary conditions, we observe variable degrees of extensional to strike-slip reactivation of the previous extensional basins during basin-parallel shortening (BI08-11 and CE 17-20, Figs. 4, 7, and 8). These results differ from previous basin inversion analog models by Del Ventisette et al. (2006); Sani et al. (2007); Deng et al. (2020), which also applied a basin-parallel shortening after a first stage of orthogonal extension. In these models, despite the high obliquity-low angle between basin strike and shortening direction, thrust faults formed parallel or at a low angle with the shortening direction, ultimately producing basin inversion. A difference in the boundary conditions applied in our models BI08-11 and CE 17-20, where active orthogonal extension took place during shortening, shortening is accompanied by coeval orthogonal extension, which could explain why basin inversion does not occur in our models. However, this may not be the only controlling factor given that in model BI01, shared similar boundary conditions with previous studies with no extension which does not have stretching applied during shortening and it does not show any. There is no evidence of shortening-parallel thrusting during the second stage of deformation (Fig. 4F). This implies that not only boundary conditions but also internal parameters are crucial. Boundary conditions may not be the only controlling factor in allowing basin inversion during orthogonal shortening. The internal parameters are also crucial. In particular, the presence of a larger seed to localize the initial extension weak zone at the base of the model, a thinner rift basin crust, and inclusion of synrift sediments, surface processes may yield a comparably weaker crustal area leading to an easier localization of deformation during compression. Sokoutis et al. (2000) also showed that folding/thrusting parallel to the shortening direction can be achieved at the transition between crustal blocks with different thicknesses/strengths in model with isostatic compensation, showing that anisotropy and isostasy are two additional key ingredients that may favor basin inversion under parallel shortening. Besides these contrasting results, our models are compatible with natural cases of synorogenic foreland rifting-transtensional reactivations that are complex processes of basin reactivation during regional compression (Gianni et al., 2015). These cases have been well documented in the Baikal region in Central Asia (Mats and Perepelova, 2011; Mats, 2013) and some places in South America such as in Central Patagonia (Gianni et al., 2015) (Figs. 14E and D12D and 12E). In both cases, former extensional basins disposed orthogonal to neighboring plate margins, where subsequently reactivated were subsequently reactivated in transtension by a distal basin-parallel compressional stress field (Figs. 14E and D12DE and 12E).

5 Conclusions

Analog experiments involving simultaneous shortening and orthogonal extension under different boundary and rheological conditions, and including the effect of crustal inheritance on fault reactivation allow us to gain new insights into geological areas recording the simultaneous activity of different deformation regimes. Our experiments corroborate show that crustal deformation takes place through a combination of normal faults, thrust faults, and strike-slip faults when boundary conditions
satisfy non plane-strain conditions. In this case, the type of structures and their location depend on the applied boundary conditions and the inclusion of pre-existing heterogeneities.

For brittle models, the coexistence of the three types of structures is possible for intermediate ratios of extrusion rate over indentation rate \(1.4 < \frac{V_e}{V_s} < 2\). For lower \(V_e/V_s\), the larger principal stress \(\sigma_1\) remains horizontal and parallel to the shortening direction, except at the edges of the model where it becomes vertical producing shortening-parallel normal faults. For values of \(V_e/V_s \geq 2\), shortening-parallel horizontal stress is not high enough to become the maximum stress axis, thus inhibiting and for low amounts of shortening, failure of the crust along shortening-orthogonal thrust faults is inhibited. For brittle-ductile models, we do not observe the coexistence of the three types of structures. We observe either shortening-orthogonal thrust faults associated with conjugate strike-slip faults (model with low \(V_e/V_s\) and no initial extensional phase) or shortening-parallel normal faults associated with conjugate strike-slip faults (model with high \(V_e/V_s\) and initial extensional phase).

Our models also show that whatever the crustal strength, its past deformation history, and \(V_e/V_s\) ratio, both normal and thrust faults remain with similar orientations, i.e. shortening-parallel and shortening-orthogonal, respectively. Instead, strike-slip faults exhibit a wide range of possible orientations with respect to the imposed shortening (or extension) direction. In particular, wide angle conjugate-high angle strike-slip faults with angles of up to \(\sim 65^\circ\) with respect to the shortening direction occur when deforming a strong crust. These faults develop far from the indentation location where shortening is applied when the crust has not been previously deformed or, but can be also observed closer from the indentation shortening location if the crust has been previously extended. Instead, for weaker crusts, strike-slip faults can be parallel to the indentation shortening direction in the case they reactivate former extensional structures. Furthermore, we note that under comparable boundary conditions, the strength of the crust plays a fundamental role in controlling the location of deformation and the types of structures that accommodate deformation, highlighting the importance of inherited structures and changes of crustal strength on the distribution of deformation and the types of structures that form during subsequent phases of deformation.

Finally, our models reproduce some of the deformation patterns documented in natural cases where deformation took place through a complex combination of normal, thrusts, and strike-slip faults. From these experiments, we observe that at least two different deformation regimes coexist when \(V_e/V_s\) is \(\geq 0.9\) and the three regimes operate simultaneously when \(V_e/V_s\) is \(\geq 1.4\), possibly indicating a minimum value in this ratio for effective lateral escape to take place. Independently of the crustal rheology or the presence of crustal weaknesses, we observe the development of conjugate strike-slip faults accompanied by variable normal faulting during compression/indentation shortening. This is reminiscent of those structural systems seen in nature accommodating tectonic escape of crustal blocks in Central Asia, eastern Alps, eastern Anatolia, and the Tibetan plateau. Our results indicate that the conjugate strike-slip systems are favored by an overall weaker crust, which is consistent with observations in the former orogenic areas. In models that include an initial phase of extension, we observe variable degrees of extensional to strike-slip reactivation of the previous extensional basins during basin-parallel shortening that are compatible with Late Mesozoic examples in the San Jorge basin in Patagonia and the Baikal rift.
Data availability. Time-series of top-view pictures and strain analysis in the form of movies for the 12 experiments are available through the GFZ Data Services (Guillaume et al., 2022):

https://dataservices.gfz-potsdam.de/panmetaworks/review/7350e22e83d36b8f48363bc1d4266552e98f7f5e9059bda09471907a1988507d/

Author contributions. B. Guillaume helped in the design of the experiments, participated in running the models, conducted the analysis of the experiments and co-wrote the manuscript. G. Gianni ran some of the experiments and co-wrote the manuscript. J.J.-Kermarrec designed the experiments and K. Bock conducted some of the experiments.

Competing interests. The authors declare no competing interests.

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Figure 1. A) Geological sketch maps showing locations with active or past multiple coeval tectonic regimes (Tapponnier et al., 1982, 2001; Davy and Cobbold, 1988; Martinod et al., 2000; Fournier et al., 2004; Corti et al., 2006; Scharf et al., 2013; Sengör, 1976; Dèzes et al., 2004; Gianni et al., 2015).

B) Close-up map showing the tectonic setting of the Alpine-Mediterranean region. Structures are modified from Faccenna et al. (2014).

C) Block diagram illustrating typical structures formed in settings involving coeval shortening and extension. Abbreviations are SA: South American plate, I: Indian plate, E: Eurasian plate, AF: African plate, AR: Arabian plate, IB: Iberian plate, and A: Adria plate.
Figure 2. Experimental set-up (top) and corresponding strength envelopes (bottom) under extension and compression for the two types of A) models with brittle crust (“strong” crust) and B) models with a brittle-ductile crust (“weak” crust). Solid lines correspond to the peak friction for Qz sand and dashed lines to the reactivation friction.
Figure 3. Applied boundary kinematic conditions for models with brittle crust (circles) and brittle-ductile crust (squares). Models with two stages of two-stage deformation have black filled symbols and subscripts to indicate the stage of deformation. Laboratory and scaled values are given for the shortening (indentation) and stretching (extrusion) rates.
Interpreted pictures, principal stretches $\lambda_{max}$, $\lambda_{min}$ ($\lambda = 1$ represents no length change) and strain type for brittle models model BI01, G) model BI08, and H) model BI11-. Strain type maps show shortening (red), strike-slip (green) and stretching (blue), with intermediate, oblique deformation at intermediate colors. The amount For the two-stage models, faults active only in the first stage of imposed stretching during deformation are in red, faults forming and active only in the second stage are in black, and faults that formed in the first stage is $\sim 5\%$ for and are reactivated in the 2 models second stage are in orange. The initial location of the seed is indicated in purple. The corresponding amount of shortening and along-$x$ stretching during the second stage and/or along-$y$ shortening is indicated at the top. Corresponding maps for principal stretches $\lambda_{max}$, $\lambda_{min}$ are in Fig. SM2.

Interpreted pictures, principal stretches $\lambda_{max}$, $\lambda_{min}$ ($\lambda = 1$ represents no length change) and strain type for brittle models model BI01, G) model BI08, and H) model BI11-. Strain type maps show shortening (red), strike-slip (green) and stretching (blue), with intermediate, oblique deformation at intermediate colors. The amount For the two-stage models, faults active only in the first stage of imposed stretching during deformation are in red, faults forming and active only in the second stage are in black, and faults that formed in the first stage is $\sim 5\%$ for and are reactivated in the 2 models second stage are in orange. The initial location of the seed is indicated in purple. The corresponding amount of shortening and along-$x$ stretching during the second stage and/or along-$y$ shortening is indicated at the top. Corresponding maps for principal stretches $\lambda_{max}$, $\lambda_{min}$ are in Fig. SM2.

Figure 4. Interpreted pictures, principal stretches $\lambda_{max}$, $\lambda_{min}$ ($\lambda = 1$ represents no length change) and strain type for models with a brittle crust after 12% of along-$y$ shortening for A) model BI10, and after 10% of along-$x$-along-$x$ stretching for brittle-models BI10, B) BI09, C)
Figure 5. Temporal evolution of principal stretches and strain type during the second stage of deformation for points \( a_1, b_2, c_3, \) and \( d_4 \) in models A) BI01, B) BI08, and C) BI11 (brittle-only–brittle crust). Upper right-hand panel: strain type (final) and overview of the selected areas. Left column: zoom on the strain type and the selected grid cell (which is outlined in red, neighboring cells outlined in black). Right column: time-evolution of the logarithm of the two principal stretches (Hencky strain; blue and red curves) and associated strain type (cumulative; black curve, right axis).
Figure 6. Interpreted pictures, principal stretches $\lambda_{\text{max}}, \lambda_{\text{min}}$ (\(\lambda = 1\) represents no length change) and strain type for models with a brittle-ductile crust and one-stage deformation: A) model CE16 after 4\% of shortening, 5\% and 10\% of stretching, two-stage deformation: B) model CE17, C) model CE18, and D) model CE20. Strain type maps show shortening (red), strike-slip (green) and stretching (blue), with intermediate, oblique deformation at intermediate colors. For the two-stage models, faults active only in the first stage of deformation are in red, faults forming and active only in the second stage are in black, and faults that formed in the first stage and are reactivated in the second stage are in orange. The initial location of the seed is indicated in purple. The corresponding amount of along-$x$ stretching and/or along-$y$ shortening is indicated at the top. Corresponding maps for principal stretches $\lambda_{\text{max}}, \lambda_{\text{min}}$ are in Fig. SM3.
Figure 7. Interpreted pictures. Temporal evolution of principal stretches $\lambda_{\text{max}}, \lambda_{\text{min}}$ ($\lambda = 1$ represents no length change) and strain type for model CE17 after a first phase during the second stage of E-W 4.7% stretching deformation for points 1, 2, 3 and 4% of shortening in models A) CE17, B) CE18, and C) CE20 (brittle-ductile crust). Upper right-hand panel: strain type (final) and overview of stretching the selected areas. Left column: shortening zoom on the strain type and the selected grid cell (which is outlined in red), strike-slip (green neighboring cells outlined in black) and stretching. Right column: time-evolution of the logarithm of the two principal stretches (Hencky strain; blue and red curves) and associated strain type (cumulative; black curve, with intermediate, oblique deformation at intermediate color; right axis).
Temporal evolution of principal stretches and strain type during the second stage of deformation for points a, b, c and d in models CE17, CE18, and CE20 (brittle-ductile crust). Upper right-hand panel: strain type (final) and overview of the selected areas. Left column: zoom on the strain type and the selected grid cell (which is outlined in red, neighboring cells outlined in black). Right column: time-evolution of the logarithm of the two principal stretches (Hencky strain; blue and red curves) and associated strain type (cumulative; black curve, right axis).

Interpreted pictures, principal stretches $\lambda_{max}$, $\lambda_{min}$ ($\lambda = 1$ represents no length change) and strain type for models CE18 and CE20 after 5% of stretching: shortening (red), strike–slip (green) and stretching (blue), with intermediate, oblique deformation at intermediate colors.

Figure 8. Interpreted distribution of structures for model BI05 after 7.7% of shortening, and corresponding principal stress axis and Mohr-Coulomb analysis for the areas with labels 1, 2 and 3 on the left figure.
Figure 9. Stereoplots showing the ranges of orientations and the relative motion of strike-slip faults located in the northern (SSn) and southern (SSs) parts of models BI05 (left) and BI08 (right), after a total of around 10% of E-W stretching (see Figs. 3 and 4) as well as, and C) in the underformed (SS) and previously deformed (SSg) parts of model CE20 (right) after 5% of stretching (see Fig. 9). The black dotted lines indicate angles of 30° (for the three models) and 45° (for the brittle-ductile model CE20) with respect to a N-S $\sigma_1$. 
Figure 10. N-S cross-sections interpreted from top pictures before deformation (top row), after 4% of shortening (central row) and after 12 to 14% of shortening (bottom row) for models with A) a brittle crust (BI09), B) a brittle-ductile crust (CE16), C) a brittle-ductile crust and two-stages deformation (CE17). The three models share comparable ratios of extrusion-stretching over indentation shortening rates $V_e/V_s = 0.9$. For each model, section AA’ corresponds to the center of the model where the seed is located, and B-B’ to the western quarter of the model. The brittle layer is indicated in yellow and the ductile layer in purple.
Figure 11. Synthesis of map view structures observed as a function of crustal strength, ratio of extrusion/stretching/indentation/shortening rates and tectonic inheritance. For the two-stage deformation, newly formed faults are in blue, strike-slip faults in green, black and normal-reactivated faults are in red/orange.
Table 1. Material properties.

<table>
<thead>
<tr>
<th>Granular material: Fontainbleau Quartz sand</th>
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<tbody>
<tr>
<td>Grain size range</td>
<td>$D_{50} = 210 , \mu m$</td>
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<tr>
<td>Density (specific)</td>
<td>2650 kg/m$^3$</td>
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<tr>
<td>Density (sieved)</td>
<td>1400 kg/m$^3$</td>
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<tr>
<td>Friction coefficient (peak)</td>
<td>0.74</td>
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<tr>
<td>Friction coefficient (reactivation)</td>
<td>0.64*-0.68**</td>
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<tr>
<td>Cohesion</td>
<td>60-70 Pa</td>
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<table>
<thead>
<tr>
<th>Viscous material: PDMS</th>
<th></th>
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</thead>
<tbody>
<tr>
<td>Density</td>
<td>965 kg/m$^3$</td>
</tr>
<tr>
<td>Viscosity</td>
<td>$3.5 \times 10^4$ Pa s</td>
</tr>
<tr>
<td>Rheology</td>
<td>Newtonian (n~1)</td>
</tr>
</tbody>
</table>

* after 10 s; ** after 2.6h
Table 2. Scaling between the model and nature. By convention, ratios (*) are given as laboratory/nature. $R_{m}$ number is given for a velocity of 60 mm/h in the model (equivalent to 5.2 mm/yr in nature). $R_{s}$ number is given for a cohesion of 60-70 Pa in the model and 45-53 MPa in nature.

<table>
<thead>
<tr>
<th>Scaling</th>
<th>Experiment</th>
<th>Nature</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thickness ($L$)</td>
<td>0.04 m</td>
<td>$15 \times 10^3$ m</td>
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<tr>
<td>Density ($\rho$)</td>
<td>1400 kg/m$^3$</td>
<td>2800 kg/m$^3$</td>
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<tr>
<td>Gravitational acceleration ($g$)</td>
<td>9.81 m/s$^2$</td>
<td>9.81 m/s$^2$</td>
</tr>
<tr>
<td>Viscosity ($\eta$)</td>
<td>$3.5 \times 10^4$ Pa s</td>
<td>$10^{21}$ Pa s</td>
</tr>
<tr>
<td>Stress $\sigma^* = \rho^* L^* g^*$</td>
<td>$1.33 \times 10^{-6}$</td>
<td></td>
</tr>
<tr>
<td>Strain rate $\dot{\epsilon}^* = \sigma^<em>/\eta^</em>$</td>
<td>$3.8 \times 10^{10}$</td>
<td></td>
</tr>
<tr>
<td>Time $t^* = 1/\dot{\epsilon}^*$</td>
<td>$2.63 \times 10^{-11}$</td>
<td>$1$ h, 4.35 Ma</td>
</tr>
<tr>
<td>Velocity $v^* = L^<em>/t^</em>$</td>
<td>$1.01 \times 10^5$</td>
<td>$40$ mm/h, $3.46$ mm/yr</td>
</tr>
</tbody>
</table>

$R_{m}$  
37.7  
37.5

$R_{s}$  
7.81-9.16  
7.78-9.16