Numerical modelling of inversion tectonics in fold-and-thrust belts

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Abstract

This work presents numerical experiments of inversion of rift basins and consequent sub-thrust imbrication in tectonic wedges. Half-graben basins initially develop and then covered with a post-rift sequence bearing a décollement-prone horizon (i.e., the upper décollement). A total of twelve models of tectonic inversion have been conducted varying (i) the strength of inherited extensional fault arrays and (ii) applying different fluid pressure ratios (i.e., strength) within syn-rift strata. Combinations of those were simulated using different internal angles of friction for the inherited faults, different strengths for the syn-rift infill and for the upper décollement. Results show that changes in relative strength between inherited faults, syn-rift deposits and the upper crustal décollement leads to important variations in structural styles. Weak faults systematically favour the compressional

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reactivation of inherited extensional faults. Weak syn-rift sediments favour hanging wall by-pass structures instead of fault reactivation and less internal deformation of the syn-rift deposits. Weak upper décollements supports the accretion of basement in a hinterland antiformal stack, decoupling of basement and cover, and forward tectonic transport of rift basins. Strong upper crustal décollements favours basement and cover coupling, can lead to fault reactivation in the absence of weak faults and syn-rift sediments, however combinations of weak faults and strong upper décollement shows fault reactivation, weak syn-rift sediments and strong upper décollement form hanging wall by-pass structures. Modelling results are compared to natural case studies.

Keywords: *Finite-difference; inversion tectonics; thrust wedges; rock strength; pore-fluid pressure.*

1. Introduction

During orogeny, rifted continental margins become buried by foreland sediments and overridden by advancing fold-and-thrust belts. Upon continental collision, transmission of stress into the foreland causes the positive inversion of rift basins in the sub-thrust region and/or ahead of the thin-skinned thrust front (e.g., Jackson, 1980). These processes indicate the co-existence of two crustal detachments soled at different depths within forelands: a shallow level décollement decoupling sedimentary sequences from the underlying basement resulting in thin-skinned deformation, and deeper-rooted shear zones leading to basement involvement (e.g., Lacombe and Mouthereau, 2002; Lacombe et al., 2003; Bellahsen et al., 2014; Camanni et al., 2014; Santolaria et al., 2015; Granado et al., 2016; Izquierdo-Llavall et al., 2018; Muñoz et al., 2018).

In this work, we briefly introduce the main concepts and controls related to basin inversion, and provide a summary of the archetypal structural styles of inverted basins. Then, a series of 2D finite difference numerical simulations with a non-temperature dependent Maxwell-type visco-elastic rheology are presented (Ruh, 2017, Ruh and Vergés, 2018). Our main objective is to investigate the deformation of half-graben basins within tectonic wedges, but still. The presented models have taken into consideration strength variations of inherited, strain-weakened normal faults and deep and shallow décollements within the brittle upper section of the continental crust. Furthermore, locally elevated fluid pressure ratios within the syn-rift sediments have been investigated. In detail, we aim at gaining insights into: i) how the above parameters impact the structural styles of inverted basins in the sub-thrust; ii) the amounts of fault reactivation and the accumulated strain within syn-rift sediments as a function of the strength of inherited faults and syn-rift sediments; iii) the impact of the relative strength between upper and lower décollements. Our modelling results are discussed and compared to several natural case studies.

2. Tectonic inversion of rifted basins

The tectonic inversion of rifted basins (or positive basin inversion) refers to a change in the tectonic regime from extension to compression, so that extensional basins become positive structural features by the compressional reactivation of extensional faults (e.g., Bally, 1984; Williams et al., 1989; Turner and Williams, 2004). In its original definition, there are two fundamental features associated with basin inversion: i) significant reactivation of pre-existing fault systems and, ii) hanging wall uplift associated with this reactivation. Ideally, pre-, syn- and post-rift sedimentary sequences should be identified to define the geological history of any inverted basin. It is necessary to portrait such sedimentary sequences within their present structural elevation, and to compare it by means of structural restoration with their originally corresponding regional elevation (Fig. 1a, b). Upon inversion, if the basement top is brought back to its pre-extensional configuration, then the basin is referred to as totally inverted; on the other hand, when the top pf the basement occurs in net extension but the post-rift beds are in net shortening (Fig. 1b), the basin is considered partly inverted. A diagnostic feature of inversion tectonics is also illustrated by the shift of sedimentary depocenters, from an early extensional phase in which subsidence and sedimentation are localized on the hanging wall of extensional faults, to an inversion phase during which depocenters are shifted away from the uplifting hanging wall (Fig. 1b).

However, when an extensional basin undergoing inversion has been either deeply buried beneath foreland sediments, overridden by thrust sheets, or significant parts of the orogenic building have been eroded, such depocentre shift may not be clearly observed. From a strictly geometrical point of view, however, the existence and degree of inversion can be assessed by using an imaginary point referred to as the null point (or null line in 3d; Fig. 1b). This point separates net contraction above from net extension below, and migrates down the fault plane during progressive reverse reactivation; when inversion is complete, the null point disappears. If shortening continues, the pre-rift and the basement will be on net shortening as well. Our modelling approach provides insights into the structural styles and deformation sequence associated with ‘blind basin inversion scenarios’ (Fig. 2), in the sub-thrust region of fold-and-thrust belts. A brief review on the main controls ruling basin inversion is given in the following to provide a background and rationale for our numerical simulations and to aid on the discussion of the obtained results.

2.1 Main controls on basin inversion

Amongst the first order controls are: i) the lithosphere's integrated strength profile and its variation through time, which is strongly dependent on composition and thermal history (e.g., Ziegler et al., 2002; Butler and Mazzoli, 2006; Lacombe and Bellahsen, 2016); ii) the regional geometry of the rift basin system (e.g., Macedo and Marshak, 1999), including the geometry and mechanical properties of the extensional fault system (e.g., Sibson, 1985; Holdsworth, 2004) and its orientation relative to the direction of tectonic shortening (e.g., Gillcrist et al., 1987); iii) depth of burial, which controls whether the deformation mechanisms are limited to the upper crustal frictional field (i.e., pressure-dependent mechanisms), or controlled by deeper temperature-activated processes (e.g., Rutter, 1986; Scholz, 1988; Holdsworth et al., 2001; Pfiffner, 2016); iv) the ingression of fluids is of particular importance for fault reactivation, as fluids can weaken faults by inducing mineral reactions (e.g., Byerlee, 1978; Holdsworth, 2004; Wibberley, 2005) and elevated pore-fluid pressures reduce the effective stress necessary for reactivating misoriented faults (e.g., Sibson, 1985, 1990); v) rheological contrast between the basin infill and its basement (e.g., Buiter et al., 2009; Bauville and Schmalholz, 2015; Boutoux et al., 2014), and the distribution of weak layers such as salt or overpressured formations (e.g., Davis and Engelder, 1985) in the stratigraphic pile (i.e., pre-rift, Mencos et al., 2015; post-rift, Granado et al., 2016; syn-orogenic, Izquierdo-Llavall et al., 2018). The relative strength of décollements in tectonic wedges is of particular importance as it has strong implications for the timing of basin inversion relative to thrusting, burial and structural styles (e.g., Granado et al., 2018a; Muñoz et al., 2018). vi) erosion and sedimentation are also recognised as fundamental in shaping the evolution of fold-and-thrust belts. In particular, they redistribute the vertical load and modify the taper of active tectonic wedges, controlling sequences of thrusting, thrust-sheet size and deformation localization. The importance of erosion and sedimentation has been thoroughly studied in natural case studies (e.g., Burbank et al., 1992; Muñoz, 2002) and physical analogue modelling (e.g., Storti and McClay, 1995; Mugnier et al., 1997; Malavieille, 2010; Granado et al., 2017, amongst others).

2.2 Archetypal structural styles related with basin inversion

Basin inversion is characteristically expressed by a series of geometrical relationships and structural styles (Fig. 1). These were firstly introduced from field observations in the French Western Alps, as summarized in the original works by Gillcrist et al., (1987), Williams et al., (1989), Butler (1989) and Coward (1996), and physical analogue models (McClay, 1989, 1995; Brun and Nalpas, 1996; Bonini et al., 2012). Amongst the main geometric features are: 'harpoon' anticlines above the reactivated extensional fault and switch in depocenters (Fig. 1b); broad regional or basin-wide arching and uplift; reversal or tilting of fault blocks; buttressing (i.e., internal deformation of the hanging wall layers) developed against upright extensional faults (Fig. 1c-e); back-thrusts (Fig. 1c), hanging wall by-pass thrusts (Fig. 1d); footwall shortcut thrusts (Fig. 1c, e) and pseudo-flower structures (Fig. 1e). In the majority of the cases, back-thrusts, pseudo-flower structures and hanging wall by pass thrusts develop from the reactivation of extensional faults originally formed on the hanging wall to the master extensional fault. A recent work by Granado et al., (2018b) in the Asturian basin of the Northern Iberian Margin shows a comprehensive field example of an inversion structure displaying most of these characteristic structural styles. Other recent works on positive basin inversion have been more provided by Scisciani (2009), Bellahsen et al. (2012, 2014), Boutoux et al., (2014) Mencos et al. (2015), Granado et al., 2016, and Muñoz et al. (2018).

2.3 Rationale for numerical experiments

Numerical modelling is gaining importance in geoscience as applied mathematical codes aid at providing a better understanding of geological processes (e.g., Willet, 1999; Beaumont et al., 2000; Buiter et al., 2009; Ruh et al., 2012, 2014; Fillon et al., 2013; Boutoux et al., 2014; Erdős et al., 2014; Bauville and Schmalholz, 2015; Lafosse et al., 2016). In line with the numerical modelling presented here, the reactivation of inherited fault systems in upper crustal levels is largely controlled by pressure-dependent frictional processes (Rutter, 1986). As mentioned above, the main factors controlling the frictional reactivation of faults and related basin inversion are the frictional properties of faults (i.e., cohesion and coefficient of friction) and their variation through time by fault-weakening/hardening processes, the orientation of these faults in respect to the stress trajectory and the presence of elevated pore-fluid pressures. Following this rationale, we have included weak faults inherited from an extensional phase during which a half-graben basin is formed by reducing the internal angle of friction and the cohesion of the extensional faults. The role imposed by weak syn-rift sediments has been approached by elevating their internal fluid pressure ratios. After that, we have included in the simulations different relative strengths for the upper and the lower décollements. Reducing the least principal stress (σ3) by erosion (i.e., removing lithostatic load) should facilitate the fault reactivation and deformation localization; however, erosion and sedimentation were not included in our experiments. A detailed description of the modelling parameters is given in the following section.

3. Numerical modelling of sub-thrust inversion tectonics

3.1. Governing equations and rheological implementation

Simulations of positive inversion tectonics were conducted with a 2D finite difference numerical code with a fully-staggered Eulerian grid and freely advecting Lagrangian markers storing the rock parameters (Ruh, 2017, Ruh and Vergés, 2018). The model mechanics is based on the equations for conservation of mass (assuming incompressibility)

$$\frac{\partial u_i}{\partial x_i} = 0 \qquad (1)$$

and the conservation of momentum (Stokes equation)

$$\frac{-\partial P}{\partial x_i} + \frac{\partial \sigma_{ij}}{\partial x_j} = \rho g_i, \qquad (2)$$

where P is mean stress, u_i are velocities in x and y-direction, x_i are spatial coordinates (x, y), σ_{ij} is the deviatoric stress tensor, ρ the density, and g_i the gravitational acceleration in x- and y-direction. The model is not solving for temperature and therefore, all applied rheologies are temperature-independent.

The above described governing equations are discretized on a fully-staggered Eulerian grid and solved for two velocity components and pressure with MATLAB's "backslash" direct solver. Rock properties are interpolated on Lagrangian markers freely advecting through the Eulerian grid according to a fourth-order Runge-Kutta derived velocity field.

The implemented rheology follows a Maxwell-type visco-elastic relation between stress and strain rate,

$$\dot{\varepsilon}_{ij} = \frac{1}{2\eta}\sigma_{ij} + \frac{1}{2G}\frac{D\sigma_{ij}}{Dt}, \qquad (3)$$

composed of a viscous and an elastic strain rate. G denotes the shear modulus and η the effective viscosity with a lower and an upper cutoff of 10^{17} and 10^{24} Pa·s, respectively. Elasticity is implemented by modifying the effective viscosity due to the material's stress history and an applied "elastic" time step of Δt = 1000 years (Gerya and Yuen, 2007; Moresi et al., 2003, 2007).

All rock-type markers have an initial viscosity of 10^{24} Pa·s and an elastic shear modulus of 10^{11} Pa. If the visco-elastic differential stresses exceed the pressure-dependent yield stress ($F > 0$), the viscosity η is changed due to plastic/brittle failure:

$$F = \sigma_{II} - \sigma_y, \qquad (4)$$

where σ_{II} denotes the second invariant of the stress tensor

$$\sigma_{II} = \sqrt{\frac{1}{2}\sigma_{ij}^2}, \qquad (5)$$

and σ_y denotes the yield stress formulated by the Drucker-Prager yield criterion

$$\sigma_y = P \cdot (1 - \lambda) \cdot \sin\varphi + C \cdot \cos\varphi. \qquad (6)$$

C is cohesion, φ the friction angle, and λ the fluid pressure ratio of the specific rock. Then, stresses get reduced to be kept within the failure envelope

$$\sigma_{xx}^{new} = \sigma_{xx}\frac{\sigma_y}{\sigma_{II}} \qquad (7)$$

$$\sigma_{xy}^{new} = \sigma_{xy}\frac{\sigma_y}{\sigma_{II}}. \qquad (8)$$

Then, the effective viscosity η is calculated to maintain the yield stress

$$\eta = \frac{\sigma_y}{2\dot{\varepsilon}_{II}}, \quad (9)$$

where $\dot{\varepsilon}_{II}$ denote the second invariant of the strain rate tensor

$$\dot{\varepsilon}_{II} = \sqrt{\frac{1}{2}\dot{\varepsilon}_{ij}^2}. \quad (10)$$

The code solves for the velocity field and pressure on the Eulerian nodal points according the staggered grid. Resulting velocities and pressure values are interpolated onto the Lagrangian markers, where stress changes and plasticity-related viscosity are calculated. The resulting effective viscosity is interpolated back on the Eulerian nodes by weighted-distance averaging and applied to solve the system of equations. Per time step, repeated cycles of the global solution are performed (Picard iterations) until the average velocity change is smaller than 10^{-14} m/s (≈3.2 μm/yr). The applied computational time step has a maximal value of 1000 years and may be reduced to prevent that markers move further than a fourth of an Eulerian cell size during one time step.

3.3. Model setup and boundary conditions

Numerical models of positive basin inversion consist of an extensional phase followed by a compressional one. The Eulerian grid size and its resolution remain unchanged throughout the complete model sequence. The model domain has a size of Lx = 200 km and Ly = 25 km with a nodal resolution of 2001·251, respectively (Fig. 2a). This results in Eulerian grid cells of 100·100 m. Rock information and material parameters are stored on 4.5 million randomly distributed Lagrangian markers. The initial marker distribution, from bottom to top, defines a 300 m thick basal detachment horizon covered by 7.7 km of crustal basement (i.e., brittle fraction of the crust) covered by 17 km of low-viscosity, low-density sticky-air (Crameri et al., 2012) to ensure zero shear stresses along the rock/air interface and to allow for vertical growth during shortening (Fig. 2a). Initial rock properties are listed in Table 1.

The first part of the numerical experiments is defined by an extensional phase of 1 Myr. During extension, the gravitational acceleration is vertical (g_x = 0, g_y = 9.81 m/s^2). The bottom boundary is defined by a velocity singularity at x = 125 km, with a zero velocity condition to the left (x = 0-125 km) and a velocity of v_x = 1 cm/yr to the right (Fig. 2a). The lateral boundaries prescribe no-slip conditions with zero horizontal velocity at the left side and a horizontal velocity of v_x = 1 cm/yr at the right. At the top, a free-slip boundary is prescribed with material coming into the Eulerian domain at a velocity of v_y = 0.125 cm/yr to ensure conservation of volume (Fig. 2a). During extension, deposition of sediments filling up the developing rift is modelled by diffusion of the topography not taking into account any erosion, following the diffusion equation,

$$\frac{\partial h_s}{\partial t} = \kappa \frac{\partial^2 h_s}{\partial x_i^2}, \quad (11)$$

where h_s is the surface level, x_i the spatial coordinates, and the diffusion constant κ = 10^{-4} m/s^2. Sticky-air markers falling below the surface line are then converted to sediment. The frictional strength of developing basement normal faults and related rift basin deposits (φ = 30°, C = 1 MPa; Byerlee, 1978) gets linearly reduced to φ_w = 25° and C_w = 0.1 MPa between accumulated brittle/plastic strain of 0 and 0.5. Fluid pressure in the basement and the rift sediments is λ = 0.4, representing hydrostatic fluid pressure (equation 6). After 1 Myr of extension, all boundary velocities are reduced to zero during 200 kyr to ensure elastic stress relaxation.

After extension and elastic relaxation, the basement and rift are covered by post-rift deposits of 1 km, of which 500 m act as an upper décollement between x = 100-200 km

(Fig. 2b). Above the post-rift cover, deposits thinning towards the left mimic “molasse-type” foreland sediments. During the shortening and tectonic inversion phase, the gravitational acceleration is rotated counter-clock wise at 3° (g_x = 0.5134, g_y = 9.7766 m/s^2), resulting in flat laying foreland infill deposits, and a hinterland tilted basement related to the vergence of subduction in orogeny (Fig. 2b). Boundary conditions are defined by a velocity of v_x = 1 cm/yr along the complete bottom and no-slip lateral boundaries with zero horizontal velocity at the right and v_x = 1 cm/yr at the left. The free-slip top boundary exhibits a velocity of v_y = -0.125 cm/yr to compensate for the incoming material from the left side (Fig. 2b).

During convergence, inherited faults are defined by shear zones that experienced an accumulated brittle/plastic strain of at least 0.5 after rifting. The reference experiment (i.e., model R1) has an inherited fault strength of φ_{if} = 30° and C = 0.1 MPa, a fluid pressure ratio of λ_{sr} = 0.4 within the syn-rift strata representing hydrostatic fluid pressure, a relatively weak upper décollement (λ_{ud} = 0.8) and with no surface erosion taking place. A first series of experiments (models 2-4) has been conducted where inherited normal faults exhibit reduced friction angles (φ_{if} = 25°, 20°, 15°) with a cohesion of 0.1 MPa. A second series (models 5-7) tests the effect of fluid pressure variation in the syn-rift sediments (λ_{sr} = 0.55, 0.7, 0.85). Furthermore (models 8-12), the effects of the upper décollement strength (λ_{ud} = 0.4 *vs.* 0.8) are investigated. For the sake of a better comparison between fault and syn-rift sediment strength, we converted both to an effective coefficient of friction that implies the friction angle and the fluid pressure ratio (Table 2). During tectonic inversion, all experiments exhibit syn-orogenic sedimentation depending on surface diffusion with a constant of $\kappa = 10^{-6}$ m^2/s.

3.4 Limitations of the modelling approach

In terms of geometry and rheology, the presented numerical model setup is very simplified with regard to the much more complicated natural systems. The experiments presented here contain a horizontally layered rock sequence pushed by a perfectly vertical backstop. This is partly given by the numerical technique applied, where the Eulerian grid remains rectangular and undeformed throughout the model sequence. A more sophisticated temporal and geometrical approach would be to investigate basin inversion as a part of a larger, mantle-scale, rifting and subduction system. However, this would drastically lower the numerical resolution and prohibit focusing on nappe-scale deformation and structural styles. Furthermore, the numerical model does not solve for temperature and therefore neglects important rheological behavior of rocks occurring at lower crustal depths in the Earth, such as temperature-dependent power-law viscosity variations (e.g., Shinevar et al., 2015). Here, only the upper part of the crustal basement is simulated, which is expected to behave in a brittle manner. It shears off above the lower, potentially viscous, basement (outside of the model domain) along the imposed weak basal detachment horizon (Fig. 2b). The advantage of this simplified setup is not just its comparability to analogue experiments (e.g., Granado et al., 2017) but also to analytical solutions (e.g., Davis et al., 1983) that apply similar geometrical boundary conditions.

4. Results

In the following, a set of twelve numerical simulations introduced above is presented, investigating the structural evolution of thrust wedges involving rift basins (Table 2). With the exception on one model (model 7, see following sections), all experiments have been run for a total of 9.1 Myr, including 1 Myr of extension followed by 200 kyr of relaxation phase, and 7.9 Myr of shortening. The shortening phase was chosen to last until a deformation front develops well ahead of the thin-skinned thrust front and the rift basins. Effects of inherited fault strength, syn-rift deposit fluid pressure, and upper décollement strength are discussed in terms of extensional fault reactivation and deformation of the syn-rift basin during tectonic inversion. All experiments are presented by plotting the Lagrangian markers based on the color code in figures 2 and 3 indicating finite deformation, and the second invariant of the strain-rate tensor indicating the velocity of deformation at a specific time.

4.1. Temporal evolution of the reference model

The evolution of the reference model is divided into two phases, which are rifting (phase 1) followed by shortening (phase 2), separated by a phase of tectonic quiescence. During rifting, deformation localizes at the velocity singularity point at the base of the model domain, as shown in the model setup (Fig. 2a). A system of conjugate normal faults develops forming a half-graben basin (Fig. 2a). The normal fault to the left hand side remains active during the complete rifting phase, whereas faults to the right hand side become inactive while moving away from the velocity singularity where new faults localize. After 1 Myr of rifting, this process results in a major normal fault to the left and a set of four normal faults towards the right side of the half-graben basin (Figs. 2b, and 3a). These five faults, outlined in figure 2b, are referred to when mentioning and quantifying inherited fault reactivation during shortening. Phase 1 is simulated equally for all remaining models, the structural styles developed and their kinematics being also the same.

After relaxation, shortening is initiated and deformation localizes along the backstop on the right side of the model domain in form of conjugate thrust shear zones within the basement rocks (Fig. 3b). The second invariant of the strain-rate tensor indicates that the upper décollement is activated immediately after the initiation of phase 2, transferring slip to a frontal thrust where the décollement tips out, overriding the rift basin. Furthermore, the foreland deposits overlying the upper décollement are deformed along the rear by the formation of an initial basement pop-up (Fig. 3b). At 3.1 Myr of total model runtime (1.9 Myr of shortening), three major thrust sheets develop in the basement verging towards the toe of the wedge, with the external most one cutting and splitting the rift basin in two, and without any visible inherited fault reactivation (Fig. 3c); this deformation style corresponds to a hanging wall by-pass thrust (see Fig. 1d for comparison). The foreland strata are scraped off along the upper décollement, forming a forelandward directed thin-skinned thrust system. With ongoing shortening, the rift-cutting basement shear zone (i.e., the hanging wall by-pass thrust) propagates into the foreland strata, separating a highly deformed sequence of upper crustal rocks close to the thrust front, from a large cover anticline at the hinterland (Fig. 3d; 5.1 Myr). The thin-skinned thrust front is formed by a large thrust-related anticline, showing a fully overturned, and tectonically thinned frontal limb; this structure is characterized by significant structural relief, supported by the presence of thick syn-orogenic wedge-top basins (Fig. 3d,e,f). Similar results were obtained for all simulations carried out. Provided that our work deals with the tectonic inversion of rift basins, a detailed description of each thin-skinned thrust front simulation is considered out of the scope of this work and will not be included hereafter.

The deformation within the hinterland basement jumps outward from the shallow thin-skinned deformation front, whereas a frontal basement thrust sheet at the wedge toe is developed by slip along the basal décollement. Further shortening of the crustal sequence is mainly expressed by localized deformation caused by protracted slip along the hanging wall by-pass thrust synchronous to basement imbrication at the front (Fig. 3e,f). The second invariant of the strain-rate tensor of the reference model after 5.1 and 7.1 Myr of total runtime illustrates that the hanging wall by-pass thrust is a long-lived fault with a ramp-flat-ramp geometry, active throughout most of phase 2 (Fig. 3e,f).

4.2. Effect of inherited fault strength

In the following, the influence of the frictional strength of inherited rift faults (i.e., φ_{if} = 25°, 20°, 15°) is discussed and compared to the reference model with a rift-fault friction angle of φ_{if} = 30°. Rock composition (Lagrangian markers) and the second invariant of the strain-rate tensor are illustrated after 9.1 Myr of total model runtime (Fig. 4). Contrary to the reference model, all experiments with reduced inherited fault strength show important compressional reactivation along the main extensional fault, and its antithetic faults. Structural inversion is shown as a ramp-flat-ramp fault system that imbricates, and tilts the rift basins towards the foreland, generating significant structural relief and related folding of the cover strata. The conjugate left-dipping extensional faults become inverted to different degrees within all according experiments (Fig. 4b-d), forming back-thrusts, imbricate stacks and pseudo-flower structures (see Fig. 1c-e). Syn-rift deposits however remain largely internally undeformed; only the simulation with φ_{if} = 25° shows a small offset according to an incipient hanging wall by-pass thrust fault (compare figs. 1d and 4b). Basement-involved footwall-shortcut thrusts (Fig. 1c, e) also develop from the main rift fault, and acquire more relevance for the weaker fault simulations (Fig. 4b-d). Overall, reducing the rift fault strength has a profound effect on the shape of the crustal wedge; however, for the three different simulations of weak faults, the overall structural styles are broadly similar: all these experiments exhibit three basement involved thrust sheets at the backstop and develop a newly-formed basement-involved system at the wedge front. When weak faults are present, uplift is always focused at the rift basin margins, a typical feature of basin inversion; this is in marked contrast with the reference model R1 (compare Fig. 4a to 4b-d).

4.3. Effect of syn-rift deposits fluid pressure

The influence of syn-rift strata strength, affected by varying fluid pressure ratios (λ_{sr} = 0.55, 0.7, 0.85), is investigated and compared to the reference experiment (model R1 with λ_{sr} = 0.4; Fig. 5a). Rock composition (Lagrangian markers) and the second invariant of the strain-rate tensor are illustrated after 9.1 Myr for model 5 and 6 (Fig. 5b-c) and after 7.6 Myr for model 7 (Fig. 5d). The experiment with a slightly elevated fluid pressure ratio in the syn-rift basin of λ_{sr} = 0.55 (model 5) shows broadly similar deformation patterns as observed in the reference model. Three thrust sheets verging towards the wedge front form the hinterland part of the wedge, overlain by an open anticline of foreland-type deposits (Fig. 5b). Shortening takes place along the hanging wall by-pass thrust crosscutting the rift basin and by frontal basement-involved imbrication at the wedge toe. A thin-skinned thrust front is also developed in between, with a large thrust-related anticline supported and fully covered by thick syn-orogenic strata. Model 6 (λ_{sr} = 0.7) presents an almost equal deformation pattern including steeply-dipping to overturned syn-rift strata imbricated and transported by the main hanging wall by-pass thrust (Fig. 5c). However, several minor differences can be observed: syn-rift strata in model 6 gets squeezed more intensely in relation to model 5 and the reference model R1 (Fig. 5a-c); weaker syn-rift strata (i.e., increased λ_{sr}) results in a larger offset along the major hanging wall by-pass thrust; the thin-skinned thrust front is represented by a buried break-through fault-propagation fold (Fig. 5b,c). The increased offset is also apparent in model 7 (λ_{sr} = 0.85), even though the snapshot is taken after only 7.6 Myr of model runtime (Fig. 5d). There, intense shortening along the hanging wall by-pass thrust leads to the almost complete overthrust of the syn-rift infill, and buttressing against the rift fault of the half-graben basin to form a pinched-in syncline (Fig. 5c, d). In model 7, increased uplift of the hinterland leads to the gravitational instability of the foreland deposits and erosion of the thin-skinned thrust front: the second invariant of the strain-rate tensor shows intense surface flow along the slope and a set of extensional normal faults above the uplifted inverted rift soled along the upper décollement (Fig. 5d). Basement imbrication ahead of the thin-skinned thrust front is less developed, and constituted by a series of conjugate shears and basement pop-ups.

4.4. Influence of upper décollement strength

The effect of upper décollement strength on the structural evolution during tectonic inversion, and in particular on the reactivation of inherited rift faults and plastic/brittle deformation of syn-rift strata, is visualized and quantified by a set of eight experiments (Fig. 6). Upper décollement strength is varied by imposing a reduced fluid pressure ratio (λ_{ud} = 0.4) in contrast to the reference experiment (model R1; λ_{ud} = 0.8). Here, the reference experiment (model R1), an experiment with reduced fault strength (model 3), an experiment with reduced syn-rift basin sediment strength (model 6) and an experiment with both reduced fault and syn-rift basin sediment strength (model 8) are compared to their counterparts with strong upper décollements (models 9–12; see Table 2).

The strength of the upper décollement affects the deformation of the underlying basement as well as of the overlying foreland deposits, and the degree of structural coupling between both. The reference model R1 develops a hanging wall by-pass thrust folding of the overlying foreland strata across the backstop (Fig. 6a). On the other hand, the identical experiment with a strong upper décollement (model 9; λ_{ud} = 0.4) favors the structural coupling between basement and cover deposits, resulting in a more intense faulting of the latter (Fig. 6e). In this case, the hanging wall by-pass fault is absent while the half-graben is fully inverted as a symmetric pop-up structure by the reactivation of the whole extensional fault array (i.e., rift fault and its antithetic faults), without much internal deformation and no forelandward tilting of the rift basin. Above, the basement thrusts propagate into the cover, breach the surface, control syn-orogenic depocentres and compartmentalize wedge top basins.

The effect of a strong upper décollement for the experiment with weak inherited fault strength (models 3 and 10) is less pronounced than for the reference experiment (Fig. 6b,f). Both experiments show structural inversion along the conjugate rift faults forming fully inverted rift basins tilted to the foreland, independent of their respective upper décollement strength. Mechanical decoupling along the weak upper décollement facilitates the forced folding of the cover (e.g., Sterns, 1978; Tavani and Granado, 2015) by pronounced basement imbrication (Fig. 6b). On the other hand, increased coupling between the basement and the overlaying sediments leads to basement thrusts propagating into upper crustal strata across the strong upper décollement (Fig. 6f).

For experiments with reduced syn-rift strata but strong inherited faults (models 6 and 11), large parts of shortening take place along the weak rift sediments in the form of hanging wall by-pass thrusts (Fig. 6c, g). Upper décollement strength mainly affects the structural evolution of the foreland deposits; a weak upper décollement decouples the foreland sediments from the basement and deformation shows large wavelength folding with minor faults (Fig. 6c). A strong upper décollement on the other hand connects basement faults into the cover and hinders continuous slip along the décollement; this is particularly indicated by comparing the respective second invariant of the strain rate tensor (Fig. 6c, g).

Model 8 with both weak inherited faults and weak rift basin strength, but the same upper décollement strength as the reference model (i.e., φ_{if} = 20°; λ_{sr} = 0.7; λ_{ud} = 0.8) shows the complete inversion of the inherited fault array, including a pseudo-flower structure (see Fig. 1e) in the hanging wall basement (Fig. 6d). The foreland strata are decoupled from the basement and form long-wavelength forced folds controlled by the underlying basement thrust sheets; these forced folds bound thick synclinal wedge top basins. The equivalent experiment with a strong upper décollement (model 12) shows that major basement faults reach up to the surface (Fig. 6h). Instead of long-wavelength forced anticlines above the basement thrust sheets, a harpoon anticline (see Fig. 1b) after the half-graben basin develops, crosscutting syn-orogenic basins and rooting at the basal décollement. Only two large thrust sheets are developed in the hinterland, instead of three as in all other simulations (Fig. 6h).

5. Discussion

Deformation of thrust wedges including an inherited half-graben basin have been numerically simulated by means of finite difference numerical models. In our simulations we aimed at studying the influence of weak inherited faults and elevated fluid pressure ratios in the syn-rift sediment wedges. Given that thrust wedges are commonly constituted by two décollement levels soled at different depths (e.g., Lacombe and Mouthereau, 2002; Lacombe et al., 2003; Bellahsen et al., 2014; Camanni et al., 2014; Santolaria et al., 2015; Granado et al., 2016; Izquierdo-Llavall et al., 2018; Muñoz et al., 2018), we also included in our simulations the relative strength between these two. In terms of rheology, we have opted for a non-temperature dependent Maxwell-type visco-elastic rheology. We are aware that certain zones of the modelled tectonic wedges, such as the accreted basement in the hinterland stacks or the lower reaches of the half-graben basins may actually deform under the transition into the temperature-controlled viscous regime; however, we assume that the used rheologies can still simulate to a large extent the deformation processes in the upper continental brittle crust, where deformation is dominated by pressure-dependent processes and the frictional properties of rocks and fault rocks. A summary of all simulations carried out is provided in figure 7 as an aid to the following discussions. Comparison to natural case studies is provided afterwards.

5.1 Structural styles of tectonic inversion in thrust wedges

The results of the presented numerical simulations have successfully replicated the archetypal structural suite of positive inversion tectonics (compare Figs. 1 and 7) such as: rift extensional faults reactivated in a reverse mode to different degrees, harpoon anticlines, hanging wall by-pass thrusts, hanging wall back-thrusts, pseudo-flower structures, pop-up and footwall-shortcut thrusts (*sensu* Williams et al., 1989 and McClay, 1989, 1995). Harpoon anticlines, hanging wall back-thrusts and pseudo-flower structures and pop-up are systematically associated to the occurrence of weak inherited faults. On the other hand, hanging wall by-pass thrusts develop when weak syn-rift sediments are present. When a relatively weak upper décollement (i.e., λ_{ud} = 0.8) is present, the general structural styles developed on the upper crustal sequence are dominated by foreland-directed thin-skinned thrust sheets; the structural style within the basement is represented by symmetric, long-wavelength anticlines produced by conjugate shear bands that develop during the incipient stages of contractional deformation; upon shortening, these shear bands evolve into forward-verging basement plugs and thrust sheets. With ongoing shortening and forward transport of the basement-involved thrust sheets, forelandward-dipping shear bands are abandoned, and become incorporated into the backlimbs of the thrust sheets. On the other hand, when a relatively stronger upper décollement (i.e., λ_{ud} = 0.4) is present a significant change in the structural styles take place: for relatively stronger upper décollements, the inherited extensional faults are reactivated during the compressional phase when the syn-rift infill stays in hydrostatic conditions and when the faults are weak. When no fault weakening is included (i.e., φ_{if} = 30°), but overpressure in the syn-rift is, no rift fault reactivation takes place.

5.2 Numerical analysis of fault reactivation and syn-rift basin strain

To best visualize the effects of fault weakness, syn-rift strata and upper décollement strength on extensional fault reactivation, the accumulated plastic strain of Lagrangian markers defining inherited faults (Fig. 2b) are plotted against model time (Fig. 8). In these plots, 100% of reactivation denotes an accumulated strain along inherited faults during shortening equal to the accumulated strain gained during rifting. These plots provide a constraints on the amount of inversion along inherited faults based on strain rather than on fault slip, which is a purely geometrical approach (i.e., the null point *sensu* Williams et al., 1989; see Fig. 1a,b). In this sense, the reference model R1 with strong inherited faults reaches a rift fault inversion of ~30% after 9.1 Myr (Fig. 8a). Experiments with inherited faults exhibiting lower friction angles than the surrounding undeformed basement (i.e., models 2-4) accumulate significantly more strain by fault reactivation; hence, fault reactivation increases with decreasing applied friction angle. In addition, these plots show that the lower the friction angle, the earlier the fault reactivation (Fig. 8a).

The effect of varying fluid pressure ratios within rift basins on inherited fault reactivation is also illustrated in figure 8b. Similar to the reference model R1, reactivation of brittle/plastic deformation along extensional faults remains below 100% of inversion. However, an increase of fault reactivation from 30% (model R1) towards ~60% (model 6 and 7) is observed due to decreasing syn-rift strata strength (Fig. 8b). As shown by the rock composition diagrams and the second-invariant of strain rate tensor, this reactivation is mostly focused along the antithetic extensional faults of the half-graben basin (Fig. 5b-d).

In general, experiments with weaker inherited faults show a higher percentage of reactivation along those during shortening (Fig. 8c) in contrast to the reference model; however, the upper décollement strength has a minor effect relative to the inherited fault strength. Fault reactivation of model 9 is significantly increased (~120%) in comparison to its counterpart with a weak upper décollement (~30%; Fig. 8c). Hence, a strong upper décollement favors the reactivation of basement faults, structural coupling between basement and cover represented by surface-breaching deeply rooted basement faults, smaller structural spacing in the cover, and fault compartmentalization of syn-orogenic wedge top basins at the rear of the thrust wedge.

Besides fault reactivation, brittle/plastic deformation within syn-rift strata acquired during inversion can also be quantified (Fig. 9). Deformation of syn-rift sediments is calculated by averaging the accumulated plastic/brittle strain of all according Lagrangian markers during shortening. Results of experiments with different extensional fault strength show that any reduction of the friction angle of inherited faults (φ_{if}) translates into a substantial decrease of brittle/plastic deformation within the syn-rift deposits (Fig. 9a). These results are in marked contrasts with those of the reference model (φ_{if} = 30°), where shortening generates significant strain within the syn-rift units (Fig. 9a) as it is mainly accommodated by hanging wall by-pass thrusting (Figs. 1d and 3). Quantification of the average strain within the syn-rift strata demonstrates that increased fluid pressures within the rift basin sediments result in more intense deformation of those during shortening (Fig. 9b). Accumulated strain within the syn-rift sediments increases in a roughly linear manner over time, with higher pore fluid ratios producing higher strain values. Such temporal evolution of deformation suggests constant slip along the crosscutting hanging wall by-pass thrust. This stands in contrast to experiments with weaker inherited faults (Fig. 4), where deformation during tectonic inversion is mainly accumulated along those faults (Fig. 8a), and much less within the syn-rift strata (Fig. 9a).

Out of the four experiments discussed in this section, the reference model R1 accumulates the most deformation within the syn-rift basin during tectonic inversion (Fig. 9c). The experiment comparable to the reference one, but with a strong upper décollement (model 9), shows very little rift-internal deformation, as observed in the rock composition snapshot (Fig. 6e). Upper décollement strength is not crucial for syn-rift strata deformation for simulations with increased fluid pressures within the rift (λ_{sr} = 0.7), but seems crucial for controlling the style of inversion (i.e., hanging wall by-passing in the reference model *vs*. rift fault reactivation in model 9). Both models 8 and 12 show fast average strain accumulation up to 0.4, and reduced increase of average strain up to ~0.7-0.8 at 9.1 Myr (Fig. 9c).

To summarize, the percentage of fault reactivation is strongly dependent on inherited fault strength; the lower the strength of the faults, the larger the percentage of reactivation (Fig 10a). The increase on the rift basin fluid pressure ratio from hydrostatic conditions (i.e., λ_{sr} = 0.4) towards nearly tectonic pressures (i.e., λ_{sr} = 0.85) on the other hand, hampers the reactivation of inherited faults, and favours hanging wall by-pass thrusting. Ratio contours between effective strength of faults and syn-rift sediments (see Table 2) allow a better comparison of the individual strength parameters (Fig. 10). Accumulation of plastic strain within the syn-rift basin is strongly dependent on apparent fluid pressure ratios (Fig. 10b). In general, decreasing fault strength results in less accumulated deformation of rift sediments during inversion for experiments with inherited faults with friction angles above 22.5º. For lower friction angles, deformation of syn-rift sediments does not show a strong dependency of fault strength (Fig. 10b).

5.3 Natural case studies

In this section, we provide three natural case studies which display similarities with our numerical simulations. The first case study is the Helvetic nappes of Switzerland (Fig. 11a), an excellent example of deformation affecting the basement and the sedimentary cover of the former Mesozoic European continental margin. The second and third case studies are from the Argentinian Andes, from the Malargüe fold-and-thrust belt of (ref. Fig. 11b), and the Salta Rift system (ref. Fig. 11c).

5.3.1 Alpine inversion as shown by the Helvetic nappes (Switzerland).

The European Alps developed from the collision between the Adriatic and European continental lithospheres since about Eocene times, following the closure of the Alpine Tethys Ocean. As a result of the collision, the former rifted European continental margin became inverted to different degrees, subducted and eventually incorporated into the Alpine tectonic wedge (see Schmid et al. 2004; Froitzheim et al. 2008; and Handy et al., 2010 for a throughout review of the European Alps). The inversion of the European passive margin was formerly described by Gillcrist et al. (1987), de Graciansky et al. (1989), and Coward et al. (1991) but more recent works have been carried out, including field (i.e., Bellahsen et al. 2012, 2014; Boutoux et al., 2014) and numerical simulations (i.e., Bauville and Schmalholz, 2015). In the Central Alps of Switzerland, the strongly folded Helvetic nappes of Morcles, Diablerets and Wildhorn are constituted by Jurassic to Cretaceous continental margin sequences deformed into recumbent folds with tectonically thinned and thickened limbs (Fig. 11a) resulting from intense heterogeneous simple shear. These three nappes are bound by thrusts formed after ductile shears within the crystalline gneissic basement (Ramsay et al., 1983). The relative autochthonous basement of the Helvetic nappes is represented by the Mont Blanc Massif, presently exposed to the south of these nappes, whereas the Aiguilles Rouges crystalline massif occurs beneath the strongly folded sequence. In between both massifs, the Chamonix valley constitutes a pinched syncline, remains of a former half-graben basin (i.e., the Mont Blanc half-graben basin *sensu* Boutoux et al., 2014). As a result of Alpine shortening the half-graben infill has been completely squeezed out, imbricated and transported northwards, and both normal and reverse offsets are observed. The Helvetic nappes are a great example of the controls imposed by the rheological layering of basement and cover sequence on the structural style of orogenic belts.

Based on our obtained results, a comparison between structures of the Helvetic nappes and their rifted and subsequently inverted basement are provided. Albeit our numerical simulations do not solve for temperature controlled rheologies (see Bauville and Schmalholz, 2015 for instance), they positively simulate the upper parts of the crustal basement and the overlying sedimentary sequences. Models including high pore pressure ratios above hydrostatic (i.e., $\lambda_{sr} > 0.4$) display broadly similar structural styles (see Fig. 5, Fig. 7e-g and 7j, and 11b) to the Helvetic nappes. The structural styles are dominated by hanging wall by-pass thrusts (after deeply rooted basement shears), the squeezed and the forelandward imbricated half-graben basin, strongly buttressed against the footwall block in a very similar manner to the Chamonix syncline (Fig. 11a). Normal offsets within the lower reaches of the former half-graben basin passing upwards to reverse offsets are also shown in the simulations, along with steeply-dipping to overturned syn-rift sediments. A series of cascading folds in the frontal culmination of the hanging wall by-pass thrust also show similar thinning and thickening patterns, as well as significant bed rotation indicative of simple shearing, comparable to the Helvetic nappes.

5.*3.2 Andean inversion in the Malargüe fold-and-thrust belt (Argentina)*

The Andean orogen is divided into several tectonic provinces including a forearc region, a magmatic arc, a retroarc fold-thrust belt, and foreland basin system (Jordan et al., 1983, Ramos, 1999). For most of the latitudinal extent of the Andes, a narrow western range with a steep forearc flank, generally referred to as the Western Cordillera, marks the active or formerly active magmatic arc; on the opposing eastern slope, a series of ranges commonly referred to as the Eastern Cordillera, and the Sub-andean Zone comprise the east-directed retroarc fold-and-thrust belt (see Horton, 2018 for a recent review).

The Malargüe fold-and-thrust belt is an E-directed basement-involved fold-and-thrust belt that constitutes part of the Argentinian Andes between 34º-36º S (Kozlowski et al., 1993), within the retroarc fold-thrust belt its foreland basin system. Basement involvement in the Malargüe has been interpreted as a result of the inversion of Mesozoic back-arc basins and related faults and the development of new basement-involved thrusts (Giambiagi et al., 2009). According to these authors, the mechanical stratigraphy played as well an important role on the structural styles of this fold-belt, which includes: a Proterozoic to Lower Triassic crystalline basement; Upper Triassic to Lower Jurassic continental rift sequences; Middle Jurassic to Lower Cretaceous platform sequences belonging to a sag phase, which also includes extensive evaporitic sequences; Upper Cretaceous to Paleogene continental and marine strata, and Neogene to Quaternary syn-orogenic continental sediments and volcanic rocks. Middle Miocene to late Pliocene deformation reactivated the Upper Triassic to Lower Jurassic back-arc extensional faults, transferring slip into a shallow thin-skinned system in the overlying Mesozoic to Cenozoic sedimentary section. The basement structure constitutes large thrust-related folds, whereas the upper section is constituted by short wavelength and tightly folded thrust-related folds detached along an upper effective décollement and decoupling horizons. Folding of the upper thin-skinned fold-and-thrust belts indicates that basement faulting was active, transferring slip and deforming the previously developed cover structures.

Simulations using a weak upper décollement with exclusively weak inherited faults or a combination of weak faults and weak syn-rift sediments and (Fig. 7a-g) present geometries and kinematics compared to the Malargüe fold-and-thrust belt. In particular, Model 4 (φ_{if} = 15º; λ_{sr} = 0.4; λ_{ud} = 0.8) including very weak faults and an effective upper décollement shows a fully inverted rift basin, including reversely-reactivated master and antithetic faults. The latter fault array forms a stack of backthrusts contributing to the faulting the basement syn-rift sediment interface and forced folding. (Fig. 11c). In addition, the whole rift basin and its sedimentary cover are imbricated by the deeper décollement ramping upwards along the master extensional fault to form the frontal culmination wall (Dahlstrom, 1970; Butler, 1982) of a basement-involved thrust sheet. This culmination wall deforms the previously detached cover, folding and tilting the upper décollement level and its related imbricates (Fig. 10d), in a very similar fashion to that interpreted for the Malargüe fold-and-thrust belt (Kozlowski et al., 1993; Giambiagi et al., 2009). Comparison of our numerical suggest the presence of two well defined décollement soled and different depths, with the Middle Jurassic evaporites playing a major role as an décollement horizon probably activated early in the shortening history, to be later folded by the basement involved basin inversion. Similar temporal relationships have been recently described in other fold-and-thrust belt systems such as the Alpine-Carpathian Junction in Austria (Granado el al., 2016), and obtained in sandbox analogue models (Granado et al., 2017).

5.3.3 Andean inversion in the Salta Rift (Argentina)

The Cretaceous Salta Rift is located in the Eastern Cordillera of the Argentinean Andes at about 25ºS. It was formed during the Cretaceous in a back-arc setting, including a Cretaceous syn-rift stage followed by Upper Cretaceous-Eocene sag. The rift system became inverted and incorporated in the Andean deformation since the late Eocene. Broadly speaking, tectonic inversion of Cretaceous back-arc extensional basins has largely controlled the structural styles and kinematics of this part of the Andean chain (Grier el al., 1991; Kley et al., 2005; Carrera et al., 2006). However, the orientation and wavelength of several structures show a trend departing from the regional N-S structural grain (i.e., perpendicular to the main Andean shortening). In this sense, Carrera and Muñoz (2013) pointed out that anomalously trending Andean folds and faults do not necessarily result from the inversion of Cretaceous structures, but also from the reactivation of favorably oriented and weak structures in the basement. This is evidenced by the presence of: i) different structural wavelengths, also characterized by partial decoupling between basement and cover, and ii), west-dipping extensional faults located on the footwall of west-directed thrusts instead of the expected east-directed thrusts (Fig. 11e). The dominant structural style is represented by surface-breaching basement faults forming harpoon inversion structures (Fig. 11e). These inverted basins were fundamentally uplifted by fault reactivation, and not incorporated onto thrust flats. Structures with smaller wavelengths in the inverted Salta Rift involving the post-rift and syn-orogenic cover are related with a limited decoupling of basement and cover rocks above the continental post-rift series, and favored by a weak basement behavior (Carrera and Muñoz, 2013; their figure 2).

Simulations using a stronger décollement in addition to weak faults (models 10 and 12; Fig 6 f, h; Fig. 7i, k) show remarkable resemblances with the Salta Rift inversion structures. Of these two, model 12 including weak syn-rift sediments shows an uplifted harpoon anticline with bound by deep seated basement faults after the rift fault and one of its antithetic extensional faults; these faults cut across the upper décollement, and postdate limited slip along the upper décollement, responsible for the development of cover structures with smaller wavelengths (Fig. 11f).

6. Conclusions

Finite-difference numerical simulations have been carried out to model the formation and deformation of thrust wedges involving previously formed extensional basins in the sub-thrust region. The presence of weak inherited faults has been revealed as the main driver for reactivation of the extensional fault array, with a significant impact on the kinematic evolution and structural styles on thrust wedges. Weak faults favour the forelandward imbrication and tilting of syn-rift half-graben basins, as well as the forced folding of the sedimentary cover including the upper décollement level. Instead, weak syn-rift sediments simulated by the presence of locally high fluid pressures favoured hanging wall by-pass thrusting and the formation of pinched in synclines after the half-graben basins. The relative strength between the lower and the upper décollements has been revealed as a fundamental control on the structural styles of inverted basins; stronger upper décollement allowed for the inversion of syn-rift basins

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Figure List

Figure 1. Basin inversion concept and archetypal structural styles. **a)** Half-graben basin illustrating main features of a half-graben basin; the regional elevation of the top syn-rift is taken as reference datum. **b)** Inverted half-graben with mild reactivation of the extensional fault causing hanging wall uplift and shift of depocentres. The null point separates extensional offsets below from contractional offsets above. **c)** Inverted half-graben showing an imbricate fan of footwall shortcuts and back-thrusts; the syn-rift is internally deformed against the footwall block forming a buttress structure. **d)** Inverted half-graben with basement-involved hanging wall by-pass thrusts. **e)** Inverted half-graben with footwall shortcuts, back-thrusts and a pseudo-flower structure formed after reactivation of hanging wall extensional faults.

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Figure 9. Plots of averaged accumulated plastic/brittle strain in syn-rift sediments against model time depending on: **a)** Inherited fault strength; **b)** Syn-rift strata strength; **c)** Upper décollement strength.

Figure 10. Parameter study for fault reactivation and basin deformation based on results of 49 experiments after 7.5 Myr. **a)** Percentage of inherited fault reactivation plotted as a function of the inherited fault strength against rift basin fluid pressure ratio. **b)** Average accumulated strain in the syn-rift basin plotted as a function of the inherited fault strength against rift basin fluid pressure ratio. Grey lines indicate contours of the fraction of the coefficients of effective friction of faults and basin sediments (see Table 2).

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10 showing a fully inverted half-graben bound by surface-breaching faults, mostly controlled by weak inherited faults and a relatively strong upper décollement. See text for further explanations.

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