Insights Into the Effects of Oblique Extension on Continental Rift Interaction from 3D Analogue and Numerical Models

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Abstract

Continental rifts often develop from linkage of distinct rift segments under varying degrees of extension obliquity. These rift segments arise from rift initiation at non-aligned crustal heterogeneities and need to interact to develop a full-scale rift system. Here, we test the effects of 1) oblique extension and 2) initial heterogeneity (seed) offset on continental rift interaction with the use of an improved analogue model set-up. X-Ray computer tomography (CT) techniques are used to analyse the 3D models through time and the results are compared with additional numerical models and natural examples. The experimental results reveal that increasing extension obliquity strongly changes rift segment structures from wide rifts in orthogonal settings to narrower rifts with oblique internal structures under oblique extension conditions to narrow strike-slip dominated systems towards the strike-slip domain. We also find that both decreasing seed offset and increasing extension obliquity promote hard linkage of rift segments through the formation of continuous rift boundary faults at the surface. (Initial) soft linkage through the formation of relay ramps is more likely when seed offset increases or extension is more orthogonal. Rather than linking at depth, the rift boundary faults curve around each other at depth and merge towards the surface to form a continuous trough. Orthogonal extension promotes the formation of intra-rift horsts, which may provide hydrocarbon traps in nature.

1. Introduction

In the early stages of continental extension, inherited crustal heterogeneities often determine the location of initial rift segments (e.g., Morley et al., 1990; Nelson et al., 1992; Bonini et al., 1997; Corti, 2012). This inheritance is commonly formed by lithological contrasts or shear zones resulting from previous deformation stages (e.g. plate convergence and collision, Buiter and Torsvik, 2014) that can occur over zones hundreds of kilometres wide. As these inherited heterogeneities are often not properly aligned, early rift segments form separately and need to interact in order to develop a continuous rift system. Evolving rift segments may interact through a transfer zone in which rift boundary faults connect (e.g. by a discrete fault system), or through a distributed accommodation zone in which rift boundary faults do not link (e.g., Rosendahl 1987; Larsen, 1988; Childs et al., 1995; Faulds and Varga 1998, Fig 1c, d). Examples of rift segment interaction can be found in Iceland (Gudmundsson, 1995), the North Sea Viking Graben (Fossen et al., 2010), Utah Canyonlands (Trudgill and Cartwright, 1994; Fossen et al., 2010), offshore Brazil (Milani and Davison, 1988), the Gulf of Suez (McClay and Khalil, 1998), the Cenozoic European Rift System (Illies, 1997; Ustaszewski et al., 2005) and the East African Rift System (Morley et al., 1990; Corti, 2012, Fig. 1a, b). Understanding rift interaction processes is of importance since they influence the migration and trapping of hydrocarbons (Fossen et al., 2010; Paul and Mitra, 2013) and magma (Acocella et al., 1999; Corti et al., 2004).

Previous analogue studies of continental rift interaction indicate that the structural evolution is sensitive to various parameters such as the presence and geometry of inherited structural fabrics, melts and detachment layers (Elmohandes, 1981; Naylor et al., 1994; Acocella et al., 1999, 2005; Basile and Brun, 1999; Le Calvez and Vendeville, 2002; McClay et al., 2002; Paul and Mitra, 2013). The offset between initial seeds that represent inherited heterogeneities in nature has a first-order influence on rift interaction. Larger seed offsets produce narrower transfer zones (Mauduit and Dauteuil 1996; Acocella et al., 1999) or even no linkage at all (Le Calvez and Vendeville, 2002). When rift interaction occurs in analogue models, areas of curved normal and oblique-slip normal faults develop that connect the main rift segments. In the early stages of such experiments, accommodation zones and associated relay structures may develop (e.g. Acocella et al. 1999, 2000). As extension progresses, however, deformation typically localizes to form discrete transfer zones (Acocella et al., 2005). In certain models, strain partitioning generates strike-slip faults accompanying the curved (oblique-slip) normal faults. However, the presence of such strike-slip faults is largely limited to cases where the angle between the transfer zone and

extension direction is greater than 45° (Acocella et al., 1999). An important constraint on most of these models is the use of rigid base plate set-ups, which force deformation along the base plate edges and which may have a significant impact on the final model results (Morley, 1999; Le Calvez and Vendeville, 2002; Acocella et al., 2005).

Recent 3D numerical modelling of orthogonal continental extension also indicates that a range of structures may form between interacting rifts, with the style of rift interaction influenced by lower crustal viscosity and the interplay between the initial offset between rifts and magnitude of strain weakening in shear zones (Allken et al. 2011, 2012). The degree of brittle-ductile coupling controls the mode of rifting and is a function of crustal viscosity and strain rate. High viscosity and strain rate promote coupling, resulting in widespread rifting, whereas low viscosity and low strain rate reduce coupling, which leads to localized or narrow rifting (Brun, 1999; Buiter et al., 2008). This subsequently influences the rift interaction structures, as rift segment linkage is more likely when rifting is localized (Allken et al., 2012). Additionally, increasing rift offset prevents rift linkage, whereas increasing the rate of strain weakening has the opposite effect. The interplay of these factors determines what type of rift interaction structures forms during orthogonal extension (Allken et al., 2011, 2012). This type of models has the advantage that structures have more freedom to develop because the lower crust distributes deformation in contrast to the rigid base plate edges in most analogue set-ups.

These analogue and numerical experiments provide important insights in the structure and evolution associated with rift linkage. However, to date no studies exist that systematically examine rift interaction processes during oblique extension, a parameter which may strongly control the evolution of rift interaction structures (Trudgill and Cartwright 1994) and which promotes eventual continental break-up (Brune et al., 2012; Bennett and Oskin, 2014).

Here, we present a series of models designed to examine crustal-scale continental rift interaction as a function of oblique extension and rift offset. In contrast to most previous analogue studies that apply a rigid base plate set-up to study rift interaction, our models undergo distributed basal deformation that permits rifts more freedom to evolve (Le Calvez and Vendeville, 2002). We also apply X-ray computer tomography (CT) techniques that allow a detailed analysis of internal model geometry with time (e.g. Colletta et al. 1991; Schreurs and Colletta, 1998; Schreurs et al., 2002). The analogue model results are then compared with additional numerical models and natural examples.

2. Materials and methods

2.1 Material properties

We use a standard brittle-viscous layering of materials generally applied to crustal-scale modelling of continental rifting. Alternating 0.5 cm thick layers of fine quartz sand (ϕ : 60-250 μm , ρ : 1560 kg/m^3) and corundum sand (ϕ : 88-175 μm , ρ : 1890 kg/m^3) are sieved from ca. 30 cm height into the modelling apparatus to form a 2 cm thick layer that represents the brittle upper crust. We apply this sand alternation to create density contrasts within the brittle layer that are visible on CT-images. Granular materials typically show elastic-plastic behaviour where increasing stress first results in strain-hardening, followed by failure at peak strength and a subsequent decrease in strength until the system reaches a state of dynamic stable sliding (Lohrmann et al., 2003). The mechanical properties of our granular materials were determined with a ring-shear tester that was filled using the same physical handling technique as the one used to construct our analogue models, i.e. sieving from ca. 30 cm height. Angles of peak- and stable friction are 36.1° and 31.4° for our quartz sand and 37° and 32° for our corundum sand, respectively, with cohesion values in the order of several tens of Pa (Panien et al., 2006; Klinkmüller, 2011) (Table 1). As both types of sand have nearly similar mechanical properties, the alternation of these granular materials in our models does not change its characteristics significantly, and we consider that they represent a quite homogeneous upper crust. Note that we are aware of the fact that the sand in our models produces shear zones rather than discrete faults, but here we choose to use the terms “faults” and “shear zones” as equivalent. Further details of our model materials are given in Table 1.

A 2 cm thick mixture of silicone (polydimethylsiloxane, PDMS) and the same corundum sand as mentioned above represents the ductile lower crust. This mixture (weight ratio: 0.965 kg of PDMS mixed with 1.0 kg corundum sand) has a density of 1600 kg/m^3 , which is close to that of the overlying brittle sand layers (average density of ca. 1700 g/cm^3). We use this viscous mixture to create a more realistic density profile than the use of pure PDMS would provide and so prevent PDMS diapirism. Tests reveal shear-thinning behaviour, but the mixture can be regarded as near-Newtonian since its n -value is 1.05. The viscosity of the PDMS/corundum sand mixture is ca. $1.5 \cdot 10^5$ Pa·s for our strain rates of ca. $7 \cdot 10^{-6}$ /s. The corundum sand remains in suspension for the duration of the model runs and only sinks to the base of the PDMS medium after several weeks.

2.2. Experimental set-up

The experimental machine (Fig. 2) consists of two longitudinal walls, between which a series of 21 plexiglass and 20 foam bars (0.5 cm and 1 cm wide respectively, all 79 cm long) are tightly stacked, overlying two base plates. Before applying the model layers, the initially 30.5 cm wide plexiglass and foam base is 16 % compressed to a 25.5 cm width by driving the rigid sidewalls towards each other with computer-controlled motors. During a model run, the rigid sidewalls move apart so that the compressed foam expands, creating distributed extensional deformation with a constant velocity gradient at the base of the overlying model material (2e,f). This contrasts with deformation localizing along the edges of base plates in traditional set-ups (Fig. 2f).

One base plate can move laterally, producing a distributed strike-slip component in the overlying model. The combination of strike-slip and orthogonal extension produces oblique extension. Two pivoting bars along the short ends of the model control the lateral motion of the plexiglass and foam bars during oblique extension (Fig. 2a,d). Rubber sidewalls confine the overlying model materials at their short ends in order to ensure the same uniform extension as at the model base thus minimizing potential boundary effects due to relative motions between model materials and sidewalls.

Following Le Calvez and Vendeville (2002), semi-cylindrical seeds of the viscous PDMS/corundum sand mixture (ca. 0.3 cm high and ca. 0.5 cm wide) were applied on top of the basal viscous layer to define the geometry of inherited heterogeneities. These are segmented by a “staircase” type seed offset (offset angle $\phi = 90^\circ$ in Fig. 2b, after e.g. Corti (2004)), which leaves no space for seeds to under- or overlap each other. Deformation localises above these discontinuous seeds due to their strength contrast with the relatively strong overlying and adjacent sand layers (Fig. 2b, c, e). Note that the seeds we apply will not represent all natural settings involving inherited structure reactivation, in the sense that structural inheritance may have a pervasive or discrete character (Morley, 1999). In our models, the seeds clearly represent discrete inherited structures. Similar to Allken et al. (2011, 2012) seed offset (H) is described relative to the scale of the crust: a 1 H offset is equal to the thickness of the brittle crust (2 cm in this study), which together with the rift boundary fault angle controls rift geometry (Allemand and Brun, 1991). This way, we can easily compare our models with natural and experimental examples that have different brittle layer thicknesses. The seed offset is also varied to test the influence of rift proximity (Table 2). Instead of one seed offset configuration at a time, we apply three different seed offsets that represent three simultaneous experiments in each model run to increase our modelling efficiency (Fig. 2g).

According to the usual convention (e.g. Brune et al., 2012; Corti, 2012), extension obliquity is defined as the angle α between the extension direction and the normal to the rift or seed trend (Fig. 2b and g). In our model series, α ranges from 0° (orthogonal extension, perpendicular to seed strike) to 90° (dextral strike-slip movement parallel to seed strike) in steps of 15° (Fig. 2b). When applying oblique extension ($0^\circ > \alpha > 90^\circ$), the long sidewall and base plate motions are adjusted to produce a constant total extension velocity at the desired extension obliquity (Fig. 2g).

Selected models are analysed using a 64 slice Siemens Somatom Definition AS X-Ray CT-scanner. Because of time constraints and the limited amount of extension that the rubber sidewall and foam/plexiglass base can accommodate, the model run duration is set to 240 min. Several model set-ups are run multiple times and exhibit very similar results, indicating that the model outcomes are well reproducible (Appendix C, Fig. C1).

2.3. Analogue velocity calibration

The analogue model series consists of 13 model runs, representing a total of 39 different experiments. Out of this set, 18 experiments from 6 model runs provide associated X-Ray CT-derived imagery (Table 2). An initial set of model runs (EXP472-474, Appendix C, Fig. C2) was aimed at finding a practical extension velocity that does not lead to high brittle-ductile coupling (Brun, 1999) and associated patterns of distributed faulting that prevent rift linkage (Allken et al., 2012). We tested a range of extension velocities between 6 and 30 mm/h. The higher extension velocities result in high brittle-ductile coupling and associated distributed faulting throughout the model. At a velocity of 6 mm/h, however, deformation sufficiently focuses along the pre-defined weak zones to develop the single rift segments that allow us to investigate rift interaction processes under controlled conditions. An extension velocity of 6 mm/h, which is representative of natural extension velocities (see section 2.5 Scaling), is subsequently used in all additional models. After 240 min each models has undergone 2.4 cm (or 1.2 H) of extension (ca. 10 % total strain) in the selected extension direction.

2.4. Numerical set-up

As a complement to the analogue experiments, we also model rift interaction using the thermal-mechanical finite-element code SULEC (e.g. Buiter and Ellis; 2012; Ellis et al., 2011; Naliboff and Buiter, 2015). These experiments aim to determine whether deformation driven from sidewalls and for a range of material properties and length scales produce similar structures and processes to those in the analogue models. Our broad approach follows that

of Allken et al. (2012, 2013), where the crust is represented by a combination of linear viscous and brittle (Drucker-Prager) behaviour.

The numerical setup (Appendix A, Fig. A1) is aimed at investigating the brittle upper crust and spans 15 cm along the x- (width) and z- (length) axes, and extends 2 cm to the base of the brittle upper crust along the y-axis (depth). We focus on one seed offset per experiment and the narrow horizontal (x- and z-axes) extent of the model permits a relatively high numerical resolution (120 x 16 x 120 elements of 0.125 cm widths in all directions) that captures first-order processes between interacting rifts. Additional models (Appendix B) contain a viscous lower crust that extends to 4 cm depth or test length scales (km instead of cm) and material properties commonly assigned to the crust in numerical experiments.

The composition of the brittle layer follows the analogue models with alternating layers (0.5 cm) of quartz and corundum sands. Brittle deformation initiates along viscous seeds at the model base that extend 4 cm in the z-direction. In cross-section, seed size is 0.25 x 0.25 cm (along the x- and y-axes). Offsets between the seeds (rift offset) range from 2 to 4 cm (1-2 H). The properties (Appendix A, Table 1) of the brittle sand layers and viscous seeds follow the analogue material properties (Table 1).

Wider shear zones and lower shear zone dip angles (45° versus ca. 60°) in the numerical experiments compared with the analogue experiments produce wider rifts that intersect the surface approximately 1 cm closer to each other than those in the analogue models. The wider shear zones and lower dip angle directly reflect the numerical resolution, which is well established as a first order control on these properties in numerical shear band experiments (e.g. Lemiale et al., 2008; Kaus, 2010; Buiter, 2012). To achieve significantly thinner shear zones and dip angles approaching those in the analogue experiments (~60°) would likely require ~ 5x higher resolution in each spatial direction (125x more elements in total), which is well outside the limits of most numerical codes. The combined effect of wider shear zones at lower dip angles is that numerical seed offsets of 2-4 cm (1-2 H) equate to analogue seed offsets of 0.5-1.5 H. We clarify this by hereby referring to numerical model seed offsets in terms of the exact offset in cm and a term H^* , which represents the numerical seed offset adjusted to equivalent analogue model seed offsets.

Fixed outward velocities along the left and right walls drive extensional deformation, which is oriented at 0°, 30° or 60° (angle α) relative to the x-axis. The total velocity magnitude is fixed at ca. 3.5 mm/h (ca. 1.75 mm/h on each side), which produces equivalent velocity gradients to those in the wider analogue models. The equivalent velocity gradients reflect a linear

relationship between the applied velocity and model width within the numerical experiments, which were selected to match the observed analogue model velocity gradient. The time step is fixed at 225 seconds. Horizontal velocities along the front and back walls vary linearly between the left and right walls and remain fixed through time in order to maintain constant velocity gradients. The base and top of the model, respectively, maintain free-slip and zero stress (free) velocity conditions. As noted above, the numerical free-slip base contrasts with the mobile base of the analogue models, which drives deformation. Additional details of the numerical methods are given in Appendix A.

2.5. Scaling

We apply the standard equations to scale our models up to natural scales. For brittle Mohr-Coulomb type materials, the following equation for stress ratios (σ^* , convention: $\sigma^* = \sigma_m/\sigma_n$ with subscripts m and n representing model and nature respectively) holds (Hubbert, 1937; Ramberg, 1981):

$$\sigma^* = \rho^* h^* g^* \quad (1)$$

We use the following values to calculate the density ratio (ρ^*): $\rho_m = 1690 \text{ kg/m}^3$ and $\rho_n = 2800 \text{ kg/m}^3$, the length ratio (h^*): $h_m = 4 \cdot 10^{-2} \text{ m}$ and $h_n = 4 \cdot 10^4 \text{ m}$ and the gravity ratio (g^*): $g_m = g_n = 9.81 \text{ m/s}^2$. The stress ratio for the brittle part of the models is thus ca. $6 \cdot 10^{-7}$.

Scaling of viscous materials is more complex, since in contrast with brittle materials, their behaviour is time-dependent. The following formulas can be applied for fluids with (near-) Newtonian rheology (Weijermars and Schmeling, 1986):

$$\dot{\epsilon}^* = \sigma^*/\eta^* \quad (2)$$

Where $\dot{\epsilon}^*$ is the strain rate ratio as a function of the stress ratio and viscosity ratio (η^*). With $\dot{\epsilon}^*$ we can derive the velocity ratio (v^*) as well as the time ratio (t^*):

$$\dot{\epsilon}^* = v^*/h^* = 1/t^* \quad (3-4)$$

It follows that, as lower crustal viscosities are generally considered to range from ca. $1 \cdot 10^{19}$ to $1 \cdot 10^{23} \text{ Pa}\cdot\text{s}$ (e.g. Buck, 1991) 1 hour in our models represents 0.04 to 42 Ma ($\eta_m = 1.5 \cdot 10^5 \text{ Pa}\cdot\text{s}$). The model extension velocity of 6 mm/h translates to a velocity between 0.14 and

$1.4 \cdot 10^3$ mm/y in nature. Typical extension velocities observed in continental rifts (e.g. East Africa) plot in this range (Stamps et al. 2008, Fig. 1)

In order to secure dynamic similarity between our models and nature, we consider the ratio R_s between gravitational stress and cohesive strength (or cohesion: C) for the brittle domain in both our models and nature (Ramberg, 1981; Mulugeta, 1988):

$$R_s = \frac{\text{gravitational stress}}{\text{cohesive strength}} = \frac{\rho \cdot g \cdot h}{C} \quad (5)$$

Similarly we use the Ramberg number or R_m for the ratio between gravitational forces and viscous stress in the viscous domain (Weijermars and Schmeling 1986):

$$R_m = \frac{\text{gravitational stress}}{\text{viscous stress}} = \frac{\rho \cdot g \cdot h}{\dot{\epsilon} \cdot \eta} = \frac{\rho \cdot g \cdot h^2}{\eta \cdot v} \quad (6)$$

Finally the Reynolds number or R_e indicates what flow type occurs in the viscous layers (Weijermars and Schmeling, 1986):

$$R_e = \frac{\rho \cdot v \cdot L}{\eta} \quad (7)$$

Where L represents a distance covered at a certain velocity v . The values we use to calculate these dimensionless numbers are given in Table 3. The model R_m value of 25 fits well in the natural range (1 to $1 \cdot 10^4$) and R_s values are 13.8 and 9.2 in the models and nature respectively ($C_{\text{nature}} = 6 \cdot 10^7$, after Corti et al. (2004)). As the Reynolds numbers are $\ll 1$ (indicating laminar flow) for any reasonable value of L , the requirement of dynamic similarity is fulfilled.

3. Overview of deformation patterns

3.1. Analogue structural surface overview

In all models, faulting initiates above the seeds shortly after the onset of extension. After ca. 1.5 h, enough topographic relief develops above the faults to analyse discrete structures. Final surface deformation patterns after a model duration of 240 min (or 2.4 cm of extension) reveal that significant structural variations occur when the extension direction or seed offset are modified (Fig. 3, 4). Orthogonal extension produces wide rift basins above the initial

seeds ($1\frac{1}{4}$ H or 2.25 cm for $\alpha = 0^\circ$, Fig. 3a), whereas increasing extension obliquity results in narrower rifts (0.6 H or 1.2 cm for $\alpha = 60^\circ$).

For extension obliquity $\alpha \geq 45^\circ$ (Fig 4) narrow basins and strike-slip motion characterise the system. Strike-slip motion along the rift boundary faults can be well traced due to the offset in the surface grid, which increases with extension obliquity (Fig. 3b). From $\alpha = 15^\circ$ to 60° , the increasingly narrower rift basins contain oblique-slip normal boundary faults and internal en echelon oblique/strike-slip features. Riedel-type secondary structures develop at the edge of the basins in strike-slip dominated oblique extension ($\alpha \geq 75^\circ$), with surface strikes oriented $14\text{-}23^\circ$ oblique with respect to the rift trend. Fault branches that stem from the edges of the initial seeds are present in several models with 2 H seed offset and $\alpha \geq 30^\circ$. These faults are strike-slip dominated as shown by their narrow geometry and the surface grid offset (fig. 4).

The initial seed offset and extension obliquity both determine whether rifts connect through a transfer zone (hard linkage, as the rift boundary faults are continuous) or through an accommodation zone (soft linkage, as the rift boundary faults are discontinuous and overlap from rift to rift). All models with a seed offset of $\frac{1}{2}$ H and 1 H produce a connected rift system at the model surface (Fig. 4). In contrast, a 2 H seed offset prohibits direct hard linkage for orthogonal and oblique extension models up to 30° extension obliquity. For $\alpha \geq 45^\circ$, a continuous transfer zone develops that connects the rifts through hard linkage.

Some top view images show minor additional boundary effects away from the initial seeds and the main rift structures (Fig. 4). The $\frac{1}{2}$ H seed offset models develop en echelon rifts striking perpendicular to the extension direction so that with increasing extension obliquity, these small rifts develop more obliquely with respect to the seeds. Similar structures occur in the 1 H seed offset models ($\alpha = 15^\circ$ and 45°), although there is less correlation with extension obliquity. Only the $\alpha = 15^\circ$ example contains faulting away from the main rift that develop perpendicular to the extension direction. When extension obliquity is 45° , they have the same orientation as from the experiments with $\alpha = 15^\circ$. Limited normal faulting along the long sidewalls in all models is another secondary feature. However, neither boundary effect interferes with the rift linkage structures.

3.2. Analogue internal deformation

Combining horizontal and vertical CT-sections of the final model stage provide 3D insights into the internal model structure as a function of extension obliquity and seed offset (Fig. 5 and 6). The horizontal CT-scan images are taken in the sand ca. $\frac{1}{4}$ H (0.3 cm) above the

silicone base layer at the end of the model run and show how the rifts propagate away from the initial seeds. The horizontal CT-sections cut the initial seeds, but also show where the viscous base layer flows upwards in regions away from the initial seeds (Fig. 5a). This vertical displacement reflects decreased loading under the propagating rifts, as the overlying sand layer thins. Vertical CT-sections are focussed on zones of rift segments interaction in order to accurately resolve their structural features. The selected models represent 0°, 30° and 60° extension obliquity for ½ H, 1 H and 2 H seed offsets.

All set-ups produce overlapping rift propagation structures stemming from the ends of the initial seeds. The rifts grow towards each other in a direction that tends to become more oblique with increasing extension obliquity (i.e. increasing α). This effect is best seen in the 2 H seed offset models, where rift propagation is oriented perpendicularly to the extension direction (Fig. 5c, f, i, i'). The rift segments in models with smaller offsets are likely too close to each other to solely illustrate the influence of extension obliquity. The propagating rift branches seem not to connect at depth in any of the models.

The relationship between structures at depth with those at the model surface (Fig. 5) is visible on vertical CT-sections (Fig. 6). Moving from the back to the front of most models, a single rift segment is initially present (e.g. Fig 6a, section 1). Towards the middle, a second rift segment appears (Fig 6a, sections 2-3) and systematically accommodates more deformation towards the opposite side where the first rift segment subsequently dies out (Fig 6a, sections 4-5). The rift segments do not connect at depth, but rather run along and curve around each other. Both rift segments grow wider towards the surface and consequently merge to form the continuous, well-defined troughs seen on top-view images (Fig. 4). In between the interacting rift segments, a horst structure develops. This general structure is typical for models in which the rift segments propagate close to each other (Fig. 6a, b, d). With increasing extension obliquity, some variations occur. The rift boundary faults tend to steepen and the trough between the linking rift segments becomes narrower and less deep, from maximum 0.45 H to 0 H in orthogonal extension and strike-slip settings respectively (Fig. 3 and 10, see also section 4.1). The trough appears therefore less pronounced on both surface images (Fig. 4) and vertical CT-sections (Fig. 6).

The fault branches in the models with high extension angles that propagate along strike from the edges of the initial seeds (Fig. 4) are well resolved on the CT-images (Fig. 5f, g, h, i). These structures are relatively narrow and steep with respect to the main rifts and represent oblique-slip normal faults (Fig. 6c) or strike-slip faults (Fig. 6d), which reflect strain partitioning between the main rifts and these additional transverse structures.

Some boundary effect faults are evident in the CT-images, which mostly occur in the models with higher extension obliquity and which are situated at the lower left acute angles of the set-up. The appearance of these boundary effects might be due to a scissoring effect in the lower left corner of the set-up, but they are not considered to have a significant effect on the main model structures.

3.3. Numerical structural overview and internal analysis

As observed in the analogue experiments, increasing extension obliquity (α) promotes rift linkage (Fig. 7), albeit over a comparatively limited range of seeds offsets. For an initial seed offset of $0.5 H^*$ (2 cm) propagating shear zones overlap (Fig. 7a-c, Fig. 8_{top}) to produce a connected rift system for all extension angles. Clear rotations of the topographic structures and shear zones towards the extension direction, however, occur with increasing extension obliquity. Notably, an extension obliquity of 60° (Fig. 7c) generates significantly less topographic development within rift zones as deformation transitions to dominantly strike-slip behaviour.

Increasing the seed offset to $1.0 H^*$ (3 cm) fails to produce rift linkage during orthogonal extension (Fig. 7d) at $\varepsilon = 10\%$, although relatively low topography at the rift intersection reflects interaction between the adjacent inner shear zones. This topographic structure (Fig. 7d) separating the rifts is similar in nature to the horsts observed in the orthogonal analogue models. An extension obliquity of 30° fully links the rift systems as the adjacent shear zones connect through a well-defined transfer zone (Fig. 7e, Fig. 8_{middle}). As with a $0.5 H^*$ (2 cm) offset, increasing the extension obliquity to 60° significantly inhibits development of the rift systems and also fails to produce rift linkage. A seed offset of $1.5 H^*$ (4 cm) (Fig. 7g-i, Fig. 8_{bottom}) fails to link in all cases, although increasing extension obliquity produces similar topographic and shear zone rotations as in Fig. 7b-c,e-f.

The results above reveal that rift linkage only occurs when the rift bounding shears zones, which initiate at the two seeds, initially intersect (Fig. 7a-c, Fig. 8_{top}) or nearly intersect at moderate extension obliquity ($\sim 30^\circ$, Fig. 7e, Fig. 8_{middle}). This contrasts with the analogue experiments, where rift linkage occurs over a wide range of seed offsets and extension obliquities (Fig. 4). While the inclusion of a lower crust layer (Appendix B, Fig. 1) in models with a 4 cm seed offset ($1.5 H^*$) nearly doubles topographic gradients across the rift zones, they still exhibit similar kinematic behaviour to the models without a lower crust. Similarly, numerical experiments with parameters modified to values typically associated with the

continental crust exhibit similar rift interaction for equivalent seed offsets (Appendix B, Fig. 2).

These findings indicate that the numerical results are robust at our resolution. However, the analogue experiments undergo distinct internal deformation behaviour not captured in the numerical experiments. We speculate that this may be caused by the limited numerical resolution that does not fully capture the details of the plastic behaviour of sand. Our resolution of 0.125 cm (or 1.25 km) and total number of elements (230,400 or 374,400) is in line with recent 3D numerical brittle-viscous experiments (e.g., Allken et al., 2011, 2012; Brune et al., 2012; see Table 5 in Thieulot, 2014), but is substantially lower than the maximum resolutions typically achieved in high-resolution 2D experiments modelling deformation at the lengths scales examined here.

4. Analogue structural evolution

4.1. Rift segment evolution

The top view images of final deformation patterns in the previous sections (Fig. 4) already show that away from the rift interaction zones, orthogonal extension creates standard rifts along the initial seed. In contrast, oblique extension models produce rift-internal oblique en echelon structures. Horizontal CT-sections taken in the sand $\frac{1}{2} H$ (1 cm) above the basal silicone layer provide additional insights (Fig. 9). Symmetric rifts develop under orthogonal extension conditions (Fig. 9a), while a 30° extension obliquity produces limited en echelon normal faults, oriented more or less perpendicular to the extension direction. The en echelon normal faults eventually link to form a continuous rift structure (Fig. 9b). For $\alpha = 60^\circ$ the initial structures are oblique-slip normal- or even strike-slip faults oriented ca. 30° to the extension direction. Subsequently steep (ca. 70°) and therefore probably oblique-slip normal rift boundary faults develop to create a continuous rift basin (Fig. 9c). For 30° and 60° extension obliquities, the en echelon structures are preserved until the end of the model run. A 60° extension obliquity system (Fig. 9c) however, develops significantly more en echelon structures than the 30° extension obliquity equivalent (Fig. 9b).

3D internal analysis also allows the quantification of rift geometry evolution (Fig. 10). Vertical CT-sections show a changing final rift segment geometry with increasing extension obliquity (Fig. 6). The degree of oblique extension also appears to control the initial boundary fault dip angle and initial rift width (Fig. 10a). Fault dip angles increase from ca. 67° under orthogonal extension to ca. 90° in strike-slip settings, while initial rift width decreases from ca. 0.75 H to ca. 0 H respectively. The relation between rift width (W), fault dip angle (β) and thickness of

the brittle (sand) layer (T) closely follows the formula $W = 2T/\tan \beta$ (after Allemand and Brun, 1991). Vertical CT-sections show that fault dip angles tend to decrease with time, as the model evolves (Fig. 10b). In orthogonal models, the initial ca. 67° boundary fault dip angle decreases to ca. 57° after 240 min. This effect is less pronounced when α is 30° and almost absent when α is 60° (fault dip angles of ca. 72° to 63° and ca. 80° to 77° respectively, Fig. 10b).

4.2. Analysis of 3D surface evolution

Figures 11-13 show 3D model topography maps that visualise the structural evolution for models with 0°, 30° and 60° extension obliquity and ½ H, 1 H and 2 H seed offset (Fig. 11-13). In all models surface structures are visible after 60 min and become more pronounced with time. Several observations from the top view images of the final model stages (Fig. 4) are also well resolved on the 3D model topography images. Wider rifts form in orthogonal settings and narrower rifts occur towards the strike-slip domain. Strike-slip faults stem from the initial rift segments when $\alpha \geq 30^\circ$. Additional rift structures in areas away from the seeds are visible and some boundary effects are present.

While poorly resolved on the previous top view images (Fig. 4), topography maps (Fig. 11-13) clearly show the development of relay ramps and subsequent formation of transfer zones. These structures occur in the orthogonal extension models with ½ H and 1 H seed offset (Fig. 11a and Fig. 12a) and in the 1 H offset model with $\alpha = 30^\circ$ (Fig. 12b). The relay ramps represent an initial soft-linkage mode of rift interaction, in contrast with the hard-linking transfer zones with continuous boundary faults that form in early stages in other models (e.g. Fig. 12c). The relay ramps, however, are breached towards the end of the model run as the rift boundary faults connect to form a continuous transfer zone. The final structure closely resembles the transfer zones in the other models (compare the evolution of Fig. 11b with Fig. 12b).

A horst structure between both rift segments develops in various models (Fig. 6a, b, d, 11a, Fig. 12a and Fig. 14). The horsts decrease in significance with increasing extension obliquity, as the rift segments propagate more closely to each other and leaving less space for a horst in between.

5. Discussion

5.1. Analogue rift segments

A schematic overview (Fig. 14) summarizes the range of structures observed in the analogue models. Model rift segment geometries closely follow the formula of Allemand and Brun (1991): $W = 2T / \tan \beta$ (Fig. 10a, insets in Fig 14a, d, g), which indicates that the measurements of rift characteristics are internally consistent. With increasing extension obliquity (α) in the models, rifts become narrower, bound by steeper faults and shallower as basin depth decreases from maximum 0.45 H to 0 H (Fig. 3b). These structural changes reflect the increasing strike-slip component in the system, as pure strike-slip faults are normally steep (vertical) structures, while normal faults generally dip ca. 60°. However, the fault dip angle also decreases with time, which is due to both flow within the viscous seed at depth and sand collapse along the rift faults at the surface.

Curved en echelon rift-internal structures that develop with increasing extension obliquity are similar to structures observed in both analogue models and nature (McClay et al., 2002; Philippon et al., 2015; Corti et al., 2003 and references therein). Our models show less structural detail due to their smaller physical size, however, hindering a more thorough comparison between our results and previous models and nature. The frequency of en echelon structures increases with increasing extension obliquity, which reflects the increasing amount of distributed strike-slip motion the system must accommodate. These structures tend towards normal fault characteristics at lower extension obliquities and more towards an oblique-slip normal fault or strike-slip fault character for models with higher extension obliquities.

The additional Riedel structures that originate from the rift segments in almost strike-slip settings ($\alpha \geq 75^\circ$, Fig. 4) have the same strike-slip nature and orientation as the Riedel faults observed in previous oblique extension models using a similar set-up, but without the initial seeds applied in this study (Schreurs and Colletta, 1998). As our seeds force faults to nucleate above them, the Riedel fault system seen in the models by Schreurs and Colletta (1998) do not develop and Riedel features only form along the rift boundary faults.

5.2. Analogue rift interaction zones

Initial seed offset is a decisive factor controlling whether rift segments initially link through transfer zones or accommodation zones (Le Calvez and Vendeville 2002; Allken et al., 2012). Small seed offsets promote the formation of transfer zones, as rifts inherently connect more easily when they are close to each other, whereas larger seed offsets tend to produce accommodation zones (Fig. 14). This is because the rifts develop independently from each other when seed offsets are 2 H or larger. However, formation of initial relay ramps in small seed offset models shows that soft linkage can also occur in small seed offset settings (Fig.

14 a, b, e). These specific relay ramps are subsequently breached to form a continuous transfer zone as deformation progresses, similar to the observations by Acocella et al. (2005). Therefore, the absence of relay ramps does not rule out their previous existence.

For a seed offset of 2 H (Fig. 14c, f, i), the extension obliquity significantly affects the rift propagation direction. As the propagating parts of the rift segments are not bound to any inherited heterogeneities, they will orient themselves according to the model stress field and propagate perpendicularly to the extension direction (Fig. 14f). Similar behaviour is observed in the normal fault experiments of Hus et al. (2005). As the extension obliquity increases, propagating rift segments migrate towards each other, facilitating hard linkage (Fig. 14i). In contrast, during orthogonal extension rifts propagate in a parallel fashion and fail to connect. Instead, an accommodation zone develops (fig. 14c). This process is also the reason why initial relay ramps form in orthogonal extension models with the low seed offsets (Fig. 14a, b). In the $\alpha = 30^\circ$, 1 H seed offset model, the rift segments propagate towards each other, but are also far enough away from each other to initially prevent hard linkage (Fig. 14e). This is contrary to the $\frac{1}{2}$ H equivalent model that directly produces a transfer zone (Fig. 14d). These variations in early rift linkage styles indicate the delicate interplay between seed offset and extension obliquity in low seed offset settings.

A similar influence of extension obliquity on rift propagation direction can be observed when additional minor rifts develop away from the main rift structures (Fig. 4). These rifts also open orthogonally to the extension direction and in the process create an en echelon series of basins. The latter effect only occurs when the extension velocity (and lower crustal viscosity) causes enough brittle-ductile coupling to initiate (limited) wide or distributed rifting (Brun, 1999; Allken et al., 2012).

The majority of models produce some type of horst structure at the centre of the rift interaction zone (e.g. Fig. 14e, inset). The most pronounced examples are formed in the 1 H models under 30° extension obliquity conditions (Fig. 6a, b, Fig. 11a, b and Fig. 12a, b). This is a result of the interplay of seed offset and initial rift width, which causes the rifts to propagate along and curve around each other at depth, but merge toward the surface. This creates a relatively large horst between the connecting rift segments as well as the formation of relay ramps as previously discussed, features also observed in previous orthogonal extension models by Corti (2004). Higher degrees of extension obliquity (increasing α) leads to direct rift segment connections, which prohibits a well-developed horst (Fig 6d). These intra-rift horsts can provide structural traps for hydrocarbons (Fossen et al., 2010) and are observed in this study during orthogonal and oblique (up to 30°) extension. Towards 60°

extension obliquity, such horsts are more likely to be poorly developed. Fossen et al. (2010) show that such horsts may also develop when the boundary faults curve away from each other in such a setting (Fig. 14b, option 2), which is also observed to a limited extent in our models (Fig. 4, experiment with $\alpha = 0^\circ$ and offset = $\frac{1}{2} H$).

Oblique extension also invokes strain partitioning in the form of additional strike-slip faults stemming from and oriented more or less parallel to the rift segments. Similar to the general rift segments, these secondary structures are narrower with more oblique extension as they have to accommodate more transverse motion (ca. 0.3 H and 0.1 H with 30° and 60° extension obliquity respectively). The internal strain partitioning between rift boundary faults and an additional set of strike-slip faults in the transfer zones themselves, as observed by Acocella et al. (1999) is not reproduced in our models, although the general structure with continuous rift boundary faults is quite similar. This is possibly due to the smaller size of our models (2 cm thick brittle sand layers vs. 7 cm thick brittle material), which might prevent us from observing more detailed internal structures, if present.

5.3. Numerical models

The numerical models show similar first-order linkage patterns as the analogue experiments and confirm that increasing extension obliquity aids rift linkage. Our tests of model thickness, numerical resolution, and model scaling also show that the numerical results are robust. However, in detail, the series of numerical experiments does not reproduce many of the analogue structures. This discrepancy almost certainly reflects the numerical resolution, which is limited due to the computational requirements of 3D models. While the 3D numerical rift interaction experiments in orthogonal extension of Allken et al. (2011, 2012) produce a range of structures similar to those observed in our analogue experiments, these structures developed over much larger total strains and in some case much larger seed offsets. For the small total strains (10 %) examined here, the complex rift linkage structures fail to develop in part due to the resolution-controlled broad width of the numerical shear zones, their 45° dip angle and the limited number of elements between offset rifts (~ 10 -20). We speculate that spatial resolutions at least an order of magnitude higher in areas of focused deformation are needed to resolve the analogue features. Future numerical comparisons to 3D analogue experiments will likely require adaptive-mesh refinement techniques (Choi et al., 2013) to resolve fine-scale structures while maintaining reasonable computational requirements. Future high-resolution numerical models should also consider the role of initial imperfections (i.e. random variations in strength) in analogue models, which may help facilitate the development of fine-scale structures not present in the comparatively homogeneous numerical experiments.

5.4. Comparison with nature

The structures observed in the early stages ($t = 120$ min, or 5 % extension) of the analogue $\frac{1}{2} H$ and $1 H$ offset orthogonal models and the $1 H$ offset, 30° extension obliquity model bear resemblance to the graben stepovers in the Utah Canyonlands and the North Sea Viking Graben (Fig. 15a). Between the analogue models and the natural examples, rift offsets are comparable and relay ramps and a horst structure between the rifts are present (Fig. 6a, 11, 12, 15a, b, e, f). A similar horst structure occurs in certain numerical models (Fig. 7d and B2b). The similarity between our models and the natural examples indicates that oblique extension (up to 30°) could have been involved in the evolution of these natural examples, especially in the Viking Graben (e.g. Zanella and Coward, 2003).

In the Western Branch of the East African Rift, the Kivu-Rusizi accommodation zone provides another natural example where rifts grow into each other without merging faults (Fig. 15d). More to the north, the accommodation zone (Pakwach Basin) between Lake Albert and Rhino-Camp Basin has a clear oblique extension setting. (Fig. 15c, relative locations shown in Fig. 1). Extension directions are moderately oblique with respect to the main rift trends of Lake Albert and the Rhino-Camp Basin ($\alpha = \text{ca. } 7^\circ$ and $\text{ca. } 20^\circ$ respectively). Rift offset is $\text{ca. } 1.5 H$, not enough to isolate both rift segments, as the sediment-filled Pakwach Basin connects both rift segments. We expect the structures to show similarities to our $1 H$ orthogonal extension model and our $\alpha = 30^\circ$ model (Fig. 15e, f). The relay ramps form both in the analogue model and nature, but they are most well developed at the southern tip of the Rhino-Camp basin, where extension obliquity is lowest. At the northern tip of the Lake Albert rift, where extension obliquity is higher, multiple en echelon normal faults (striking approximately perpendicular to the extension direction) disrupt and otherwise well-developed relay ramp. Such relay ramp breaching can be observed in our $\alpha = 30^\circ$ model (Fig. 15f), the formation of individual en echelon faults is however best expressed in our $1 H$, 60° extension obliquity model (Fig. 13c).

Breaching of relay ramps has also been reported in the Utah Canyonlands (Trudgill and Cartwright, 1994) as well as along the boundary faults of Lake Albert (Dou et al., 2004) and in the Mesozoic North Sea rifts (Roberts et al., 1990). In addition, rift segments curving around each other are present in the East African Rift System area and could ultimately lead to the creation of micro-plates (Koehn et al., 2008). Furthermore, the maximum subsidence of $0.45 H$ in our models represents a maximum of 9 km subsidence in nature, which is well in the range of subsidence values reported in continental rifts (e.g. Corti et al., 2007 and references therein).

Our models produce accommodation zones and transfer zones with rift-connecting boundary faults that strongly resemble natural examples. However, discrete transfer faults do not develop in our models. The 10 % maximum extension that we apply is possibly not sufficient to transform the initial structures into discrete transfer faults (Roberts, et al., 1990; Acocella et al., 2005).

6. Conclusions

We present 3D analogue and numerical experiments that investigate the influence of extension obliquity and initial offset between inherited heterogeneities on rift interaction in continental settings. We conclude that:

- Increasing extension obliquity (i.e. increasing angle α) strongly changes rift structures. Orthogonal settings produce wide rifts, while oblique extension settings result in narrower rifts with oblique internal structures. Towards the strike-slip domain, the system is dominated by strike-slip structures;
- Rift linkage can occur through the formation of continuous boundary faults at the surface (hard linkage mode), or through the formation of relay ramps (soft linkage mode). In most cases, initial relay ramps are breached and removed with increasing deformation;
- Transfer zone formation is promoted by both decreasing offsets between initial heterogeneities and increasing extension obliquity, which reflects that rifts tend to align in a direction that is perpendicular to the extension direction;
- Rift bounding faults in the analogue models do not link at depth, but rather curve around each other and merge towards the surface to form a continuous trough. We predict that associated intra-rift horsts, which may provide hydrocarbon traps in nature, are most likely to develop in more orthogonal extension settings;
- We find agreement in rift linkage patterns between analogue and numerical experiments to the very first order, highlighting the variability in structures that can be obtained from using different modelling methods. Future experiments could test techniques for achieving a substantially higher numerical resolution in the rift linkage zones and perhaps different plasticity (brittle behaviour) models.
- Several of the characteristic structures we observe in our orthogonal and oblique extension models are also present in natural rift settings such as the North Sea, the Utah Canyonlands and the East African rift system.

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8. Appendix A. Numerical methods and approximations

We model 3D viscous-plastic deformation with the particle-in-cell (Eulerian-Lagrangian) finite element code SULEC (e.g. Ellis et al., 2011; Buiter and Ellis, 2012; Naliboff and Buiter, 2015). SULEC solves for incompressible slow viscous flows through the equations for conservation of mass (A1) and momentum (A2):

$$\nabla \cdot \mathbf{u} = 0 \quad (\text{A1})$$

$$\nabla \cdot \boldsymbol{\sigma}' - \nabla P + \rho \mathbf{g} = 0 \quad (\text{A2})$$

Here, $\mathbf{u}$ is velocity, $\boldsymbol{\sigma}'$ is the deviatoric stress tensor, P is dynamic pressure (mean stress), ρ is density and $\mathbf{g}$ is gravitational acceleration (-9.81 m/s^2 in the vertical direction and otherwise 0). Determining velocity and pressure with an iterative penalty method (e.g. Pelletier et al. 1989) transforms equations (A1) and (A2) into

$$\nabla \cdot \boldsymbol{\sigma}'^{(i)} - \nabla P^{i-1} + \nabla(K \nabla \cdot \mathbf{u}^i) + \rho \mathbf{g} = 0 \quad (\text{A3})$$

$$P^i = P^{i-1} - K \nabla \cdot \mathbf{u}^i \quad (\text{A4})$$

The term K represents the compressibility factor (here $1 \cdot 10^{16} \text{ Pa}$), while i represents the pressure iteration number. In order to avoid pressure instabilities (i.e. checker boarding) associated with fixed x- and z- velocities along all vertical walls and Q1P0 elements (linear in

velocity, constant in pressure), we fixed the pressure at lithostatic values (P_L), modifying equations (A3) and (A4) to

$$\nabla \cdot \sigma' - \nabla P_L + \nabla(K \nabla \cdot u) + \rho g = 0 \quad (A5)$$

$$P = P_L \quad (A6)$$

Numerical instabilities associated with numerical overshoots across large density contrasts, mainly at the free surface, are stabilised with the method of Kaus et al. (2010) and Quinquis et al. (2011).

Material strength reflects a combination of viscous and brittle behaviour, with viscous flow following a linear creep model relating the effective stress (σ'_{eff}), viscosity (μ'_{eff}) and strain rate ($\dot{\epsilon}'_{eff}$):

$$\sigma'_{eff} = 2 * \mu'_{eff} \dot{\epsilon}'_{eff} \quad (A7)$$

Values of μ'_{eff} for sand and PDMS (if present) are, respectively $1 \cdot 10^9$ Pa·s and $1.5 \cdot 10^5$ Pa·s. The effective stress and strain rate are:

$$\sigma'_{eff} = \left(\frac{1}{2} \sigma'_{ij} \sigma'_{ij} \right)^{1/2} \quad (A8)$$

$$\dot{\epsilon}'_{eff} = \left(\frac{1}{2} \dot{\epsilon}'_{ij} \dot{\epsilon}'_{ij} \right)^{1/2} \quad (A9)$$

Brittle material behaviour follows a Drucker-Prager yield criterion, which defines an effective stress limit:

$$\sigma'_{eff} = P \sin \phi + C \cos \phi \quad (A10)$$

Where C is the cohesion and ϕ is the angle of internal friction. We use cohesion values (C) of 21 Pa and 39 Pa for quartz sand and corundum sand respectively. Values of ϕ for quartz (35° - 31°) and corundum (37° - 32°) sand weaken (Table 1) linearly between plastic strain values of 0.5 and 1.5 (as measured by the second invariant of the strain tensor). Viscosity is limited by imposed maximum ($1 \cdot 10^9$ Pa·s) and minimum ($1 \cdot 10^4$ Pa·s) values. Material properties are tracked through tracers, which initially number 27 per element. Tracer

injection and deletion keeps tracer density between 8 and 54 tracers per element throughout the model evolution.

Appendix B. Additional numerical experiments

[Fig. B1 and B2]

Appendix C. Additional analogue images

[Fig. C1 and C2]

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Table Captions

Table 1. Material properties. Quartz sand and viscous material characteristics were measured at GFZ Potsdam. Corundum sand characteristics from Panien et al. (2006) and Klinkmüller (2011).

Table 2. Summary of analogue model characteristics.

Table 3. Parameters used for model scaling.

Figure Captions

Fig. 1. Examples of rift interaction zones. (a) East African Rift System. Modified after Ebinger (1989) and Dou et al. (2004). RC Basin = Rhino-Camp Basin; (b) Western Branch of the East African Rift System. Location is shown in (a). Modified after Stamps et al. (2008) and Corti (2012). Hatched faults represent normal faults, the character of faults without motion indications are not specified in the original publications. (c-d) Block diagrams depicting the differences between accommodation and transfer zones. (c) Accommodation zones (soft linkage) in which rift boundary faults do not connect, but die out laterally and overlap. Examples are present in the East African Rift System; their locations are shown in (b). (d) Transfer zones (hard linkage) in which the rift boundary faults are continuous from rift to rift (e.g. the Selenga Accommodation Zone in Lake Baikal (Scholz and Hutchinson, 2000)) or in which a single transfer fault connects both basins (e.g. the Gulf of Suez or Thailand (Acocella et al., 1999; Morley et al., 2004)).

Fig 2. Set-up of the analogue experiments. a) Cut-out view of the experimental apparatus; b) Surface view showing the definition of extension obliquity (angle α) and seed geometry; c) Compositional layering and associated strength profile. The dotted line indicates the strength profile at the seed, where the brittle layer is thinner. Calculated after Brun (2002); d) Schematic surface view showing the initial set-up and subsequent deformation of the empty apparatus; e) Example of a CT-derived cross-section (n-n', see (d)), showing the initial model layering and seed geometry. White arrows indicate the velocity gradient at the model base; f) Velocity profile (perpendicular to extension direction, see (e)) in current and traditional set-ups; g) Left: Surface view of the PDMS/corundum sand mixture with seed geometry before adding the sand cover. Note that the seeds are not connected. Middle: applied extension obliquity. Right: Top view of the example model showing final surface deformation of three simultaneous experiments.

Fig 3. (a) Final rift segment width and strike-slip motion along rift boundary faults as a function of extension obliquity (independent of seed offset), measured on model surface views (Fig 4). (b) Final rift depth over extension obliquity, measured on CT-sections. In both graphs, error bars indicate the ranges of measurements and the yellow dots indicate typical values (mode). All trend lines are 2nd order polynomial, $R^2 = 0.97, 0.93$ and 0.99 , respectively.

Fig 4. Overview of final surface deformation patterns in the analogue models as a function of extension obliquity (angle $\alpha = 0^\circ$ - 75°) and seed offset, after 240 min or 2.4 cm extension. Colour coding shows rift linkage mode: green for hard linkage (transfer zones) and pink for soft linkage (accommodation zones).

Fig 5. Overview of final horizontal deformation patterns inside the analogue models as a function of extension obliquity and seed offset. The horizontal CT-sections cut the sand ca. $\frac{1}{6}$ H (0.3 cm) above the basal silicone layer and are taken after 240 min or 2.4 cm extension. The dark grey linear features represent the original seed and newly developed structures. Striped patterns in the middle of the images (mostly in g-i) are CT-scanning artefacts.

Fig 6. Final horizontal deformation patterns (left column) and corresponding vertical cross-sections (right column) for a selection of analogue models. The horizontal CT-sections cut the sand ca. $\frac{1}{6}$ H (0.3 cm) above the basal silicone layer and are taken after 240 min or 2.4 cm extension. The locations of the vertical sections are shown in the left column. Striped patterns in the middle of the horizontal sections are CT-scanning artifacts.

Fig. 7. Overview of the final stages of the numerical models. Surface elevation after 240 min of extension (10 % strain) for extension directions of 0° , 30° or 60° and seed offsets of 2, 3 or 4 cm (0.5, 1.0, 1.5 H^*). The entire surface extent of the model domain (15 cm x 15 cm) is plotted at a 1:1 scale. The vertical axis (y) is exaggerated by a factor of 10 to illustrate deformation structures. Note the annotated intra-rift horst (panel d).

Fig 8. Final deformation stages of the numerical models. Left: Surface deformation after 240 min of extension (10 % strain) for an extension direction of 30° and rift offsets of 2.0 cm (0.5 H*), 3.0 cm (1.0 H*) and 4.0 cm (1.5 H*). Dashed lines indicate the location of plotted cross-sections. Right: Deformation as a function of depth for cross-sections 1-5.

Fig 9. Evolution of horizontal deformation patterns of rift segments away from the rift interaction zones in the analogue models. The CT-sections cut the sand $\frac{1}{2}$ H (1 cm) above the silicone layer. Extension obliquities (angle α) are 0°, 30° and 60°. Striped patterns are CT-scanning artefacts.

Fig. 10. Measurements of the rift boundary fault dip angles in the analogue models. a) Dip angle of the initial rift boundary fault and rift width as a function of extension obliquity. W^* represent the ideal rift width according to the equation $W = 2T/\tan \beta$ (Allemand and Brun, 1991), using the measured fault dip angles (β) and brittle sand layer thickness $T = 0.75 H$ (instead of $1H$, since the brittle layer is thinner above the seeds); b) Decreasing fault dip angle with model evolution for different extension obliquities. The values shown are measured over the entire fault depth along rift boundary faults on representative CT-sections throughout the model. Trend lines are 2nd order polynomial, R^2 values are > 0.9 , except for the “angle $\alpha = 60^\circ$ ” linear trend line in (b) with $R^2 = 0.83$.

Fig. 11. CT-derived 3D surface topography images, showing the evolution of surface deformation patterns in the analogue models. Seed offset is $\frac{1}{2} H$ and extension obliquities (angle α) are 0°, 30° and 60°.

Fig 12. CT-derived 3D surface topography images, showing the evolution of surface deformation patterns in the analogue models. Seed offset is $1 H$ and extension obliquities (angle α) are 0°, 30° and 60°.

Fig 13. CT-derived 3D surface topography images, showing the evolution of surface deformation patterns in the analogue models. Seed offset is $2 H$ and extension obliquities (angle α) are 0°, 30° and 60°.

Fig 14. Graphic overview of observed structures in the analogue models as a function of extension obliquity and seed offset. Note that the vertical cross-sections are idealized (shown without the seed).

Fig. 15. Comparison of analogue model results with natural examples. (a) Northern part of the right-stepping Viking Graben system (left), shown together with the Devils Lane stepover (Canyonlands, Utah, right). Note the high areas and relay ramps at graben stepovers, associated with oil and/or gas fields in the case of the Viking Graben (GF, GFS, HU, KB, VS and VA). m.b.s.l. = meters below sea level. Modified after Fossen et al. (2010), AAPG © 2010 and reprinted by permission of the AAPG whose permission is required for further use. (b) CT-derived surface image showing the topography of an orthogonal extension (angle $\alpha = 0^\circ$), 1 H offset model after 120 min or 0.6 H (1.2 cm) extension. One numerical model develops a similar structure (Fig. 7d). (c) Albertine Rift structure with the accommodation zone between Lake albert and the Pakwach and Rhino-Camp Basins to the north. AN = Albert Nile; BBF = Bunia Border Fault; NTBF = North Toto Bunyoro Fault; RK = River Kafu; TF = Tonya Fault; VN = Victoria Nile. Image modified after Stamps et al. (2008) and PEPD (2012), copyright by the Petroleum Exploration and Production Department, Uganda. (d) Block diagram of the Kivu-Rusizi accommodation zone. Location is shown in Fig. 1b. Image modified after Ebinger (1989). (e,f) CT-derived surface images (zoom-ins of Fig. 12, EXP484 and EXP498 respectively) showing the model topography after 120 min.

Fig. A1. Compositional layering and boundary conditions of 3D numerical models. Light and dark tan layers, respectively, represent alternating layers (0.5 cm) of quartz and corundum sand. Rectangular viscous seeds (green) are separated by 2-4 cm ($0.5-1.5 H^*$) along the x-axis. Fixed velocities along the model sides drive deformation. The extension obliquity is 0° (light grey), 30° (grey) or 60° (dark grey).

Fig. B1. Surface elevation of numerical experiments after 240 min (10 % strain) for extension obliquities of 0° , 30° or 60° and a 4.0 cm ($1.5 H^*$) seed offset. In contrast to figures 7 and 8, these experiments contain a 2 cm thick viscous (PDMS) lower crust. The entire extent (15 x 15 cm) of the horizontal domain (x- and z-axis) is plotted at a 1:1 scale. The vertical axis (y) is exaggerated by a factor of 10 to illustrate deformation structures.

Fig. B2. Surface elevation after 1.5 Ma (10 % strain) for models containing properties commonly used in numerical studies of continental extension. The surface elevation is shown for extension obliquities of a) 0° , b) 30° or c) 60° and a 40 km ($1.5 H^*$) seed offset. As in figures 7 and B1, the vertical axis (y) is exaggerated by a factor of 10 to illustrate deformation structures. The model spans 150 x 150 km in the x- and z-directions, with a 20 km thick brittle upper (2800 kg/m^3) and a 20 km thick linear viscous lower crust (2900 kg/m^3). Grid spacing in the x- and z-directions is fixed at 1.25 km, while the vertical resolution ranges from 1.25 km (upper crust) to 2 km (lower crust). Seed dimensions are 2.5 km (x), 2.5 km (y) and 40 km (z). The crust extends at 10 mm/yr (5 mm/yr on each side) with a fixed time step of 25,000 years. The friction angle and cohesion in the brittle upper crust, respectively, weaken from 20° - 10° and 20-10 MPa over a finite plastic strain interval of 0.5-1.5. The upper and lower crust viscosities are, respectively, 10^{26} and $10^{21} \text{ Pa}\cdot\text{s}$.

Fig. C1. Surface views of multiple analogue model runs with identical starting conditions (angle $\alpha = 30^\circ$) result in similar surface structures, indicating that our model results are reproducible.

Fig. C2. Decreasing the overall extension velocity from 30 to 6 mm/h shows that lower extension velocities result in well localized faulting along the pre-defined seeds. Accordingly, we use an extension velocity of 6 mm/h throughout this study.

Table 1

Granular materials		Quartz sand	Corundum sand
Grain size distribution (weight %)			
Grain shape ^a	- surface texture	-	conchoidal fractures
	- roundness	angular	angular
	- form (elongation) ^b	-	2.00
	- sphericity	low	low
Density (sieved)		1560 kg/m ³	1890 kg/m ³
Coefficient of internal peak friction		0.73	0.65
Coefficient of dynamic-stable friction		0.61	0.63
Angle of internal peak friction		36.1°	37°
Angle of dynamic-stable friction		31.4°	32°
Strain softening ^c		16%	16%
Cohesion		9 ± 98 Pa	39 ± 10 Pa
Viscous material		PDMS/corundum sand mixture	
Weight ratio PDMS : corundum sand		0.965 kg : 1.00 kg	
Mixture density		ca. 1600 kg/m ³	
Viscosity ^d		ca. 1.5·10 ⁵ Pa·s	
Type		near-Newtonian (n = 1.05) ^e	

a Quartz grain shape determined after Powers (1953)

b Elongation: ratio between longest and shortest diameter

c Strain softening is the difference between peak strength and dynamic-stable-strength, divided by peak strength

d The viscosity value holds for model strain rates < 10⁻⁴ s⁻¹

e Stress exponent n (dimensionless) represents sensitivity to strain rate

Table 2

Model run	Extension obliquity (angle α)	Extension velocity (mm/h)	Seed offset (H = 2 cm)			CT-scanned	(Parts of) models presented in figures
			I	II	III		
EXP472	0°	30	1 H	2 H	3 H		Fig. C2
EXP473	0°	20	1 H	2 H	3 H	X	Fig. C2
EXP474	0°	6	1 H	2 H	3 H	X	Fig. C2
EXP483	0°	6	½ H	1 H	2 H		-
EXP498	0°	6	½ H	1 H	2 H	X	Fig. 4-6, 9, 11-15
EXP479	15°	6	½ H	1 H	2 H		Fig. 4
EXP478	30°	6	½ H	1 H	2 H		Fig. C1
EXP482	30°	6	½ H	1 H	2 H		Fig. C1
EXP484	30°	6	½ H	1 H	2 H	X	2, 4-6, 9, 11-15, C1
EXP480	45°	6	½ H	1 H	2 H		Fig. 4,
EXP476	60°	6	½ H	1 H	2 H	X	Fig. 4-6, 9, 11-14
EXP481	75°	6	½ H	1 H	2 H		Fig. 4,
EXP477	90°	6	½ H	1 H	2 H	X	-

Table 3

	General parameters			Brittle upper crust		Ductile lower crust		Dimensionless values		
	Gravitational acceleration g (m/s ²)	Crustal thickness h (m)	Extension velocity v (m/s)	Density ρ (kg/m ³)	Cohesion C (Pa)	Density ρ (kg/m ³)	Viscosity η (Pa·s)	Ramberg number R _m	Brittle stress ratio R _s	Reynolds Number R _e
Model	9.81	0.04	1.7·10 ⁻⁶	1690 [*]	24 ^{**}	1600	1.5·10 ⁵	25	13.8	<<1
Nature	9.81	4·10 ⁴	ca. 1·10 ⁻¹⁰	2800	7·10 ⁷	2900	1·10 ^{19 to 23} ^{***}	1·10 ^{0 to 4}	9.2	<<1

*

Average density of both sand types

**

Average cohesion of both sand types

Figure 1

Figure 2

Figure 3

Figure 4
[Click here to download high resolution image](#)

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Extension obliquity

$\alpha = 0^\circ$
(EXP498)

0 min

120 min

180 min

240 min

$\alpha = 30^\circ$
(EXP484)

$\alpha = 60^\circ$
(EXP476)

initially oblique faults
develop along the seed

oblique-slip normal faults
connect initial faults, creating
a continuous structure

Figure 10

Figure 11

Seed offset = $\frac{1}{2} H$

(for all models shown
in this figure)

60 min

120 min

180 min

240 min

a)

$\alpha = 0^\circ$

EXP498

b)

$\alpha = 30^\circ$

EXP484

c)

$\alpha = 60^\circ$

EXP476

Figure 12

Figure 13

Seed offset = 2 H
(for all models shown
in this figure)

Figure 14

Figure 15

Figure A1

Figure B1

Figure B2

Figure C1

Figure C2

Set-up models 472-474 (shown without sand cover)

EXP 471 (divergence: 30 mm/h) $t = 1:00\text{h}$; $e = \text{ca. } 11\%$)

EXP 472 (divergence: 20 mm/h) $t = 1:30\text{h}$; $e = \text{ca. } 11\%$)

EXP 474 (divergence: 6 mm/h) $t = 4:00\text{h}$; $e = \text{ca. } 10\%$)