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Selective inversion of rift basins in lithospheric-scale analogue experiments

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Abstract. Basin inversion is commonly attributed to the reverse reactivation of normal basin-bounding faults. This association implies that basin uplift and inversion-related structures are mainly controlled by the frictional behaviour of pre-existing faults and associated damage zones. In this study, we use lithospheric-scale analogue experiments of orthogonal extension followed by shortening to explore how the flow behaviour of ductile layers underneath rift basins promote or suppress basin inversion. Our experiments show that the rheology of the ductile lower crust and lithospheric mantle, modulated by the imposed bulk strain rate, determine: (1) basin distribution in a wide rift setting and (2) strain accommodation by fault reactivation and basin uplift during subsequent shortening. When the ductile layers deformed uniformly during extension (i.e., stretching) and shortening (i.e., thickening), all of the basins were inverted. When viscous deformation was localised during extension (i.e., necking) and shortening (i.e., folding), only some basins – which were evenly spaced apart – were inverted. We interpret this selective basin inversion to be related to the superposition of crustal-scale and lithospheric-scale boudinage during the previous basin-forming extensional phase.

1 Introduction

Ancient rift basins record more than just the extensional event during which they formed. The initial basin-forming phase is commonly followed by subsequent events associated with thermal equilibration of the lithosphere (Morgan and Ramberg, 1987) or a change in the driving far-field plate kinematics (Forsyth and Uyeda, 1975). Some rifts fail before continental breakup and remain as fossil features within continents, which are likely to be overprinted by younger geological features. There are many examples from around the world in which the initial rift phase is interpreted to have been succeeded by shortening that resulted in basin inversion (Williams et al., 1989; Beauchamp et al., 1996; Turner and Williams, 2004; Blaikie et al., 2017; Le Gall et al., 2005; Elling et al., 2021; Thorwart et al., 2021). Sustained shortening (i.e., collision between two continental plates or blocks) can also form orogenic belts; the characteristics of these belts may record the influences of pre-existing extensional basins (e.g., NW Argentinian Andes, Carrera et al., 2006; Chungnam Basin, Park et al., 2019; Cape Fold Belt, Paton et al.,

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2006). Modern examples of orogenic belts that were impacted by pre-existing basins include the European Alps and Apennines (Boutoux et al., 2014; Scisciani et al., 2019; Pace et al., 2022) and the Pyrenees (Mencos et al., 2015).

Zwaan et al. (2022) defines basin inversion as the uplift and/or exhumation of the sedimentary infill of a basin due to the reversal of subsidence. In this paper, we focus on "positive" inversion, which was defined by Williams et al. (1989) as the contraction of a region that previously underwent extension. Analogue modelling to date has focused on the role of crustal-scale extensional structures in accommodating strain during shortening, from the scale of the basin to that of individual basin-forming faults (e.g., Bonini et al., 2012; Molnar and Buiter, 2022); also see reviews by McClay, 1995 and Zwaan et al., 2022). Many analogue experiments on basin inversion have examined the influence of pre-existing normal faults or shear zones (e.g., McClay, 1989, 1995; Del Ventisette et al., 2006; Marques and Nogueira, 2008) and basin fill that is relatively weak compared to the extended crust (e.g., Panien et al., 2005) on deformation of the sedimentary layers within the basin. In these cases, specific assumptions are made on the behaviour of the viscously deforming crust and lithospheric mantle, and this behaviour is imposed as boundary conditions from the start of the experiments.

Complementary to analogue models, numerical experiments have focused on the drivers of basin inversion at the lithospheric-scale (e.g., Hansen and Nielsen, 2003; Sandiford et al., 2006; Buiter et al., 2009). They have examined the interactions between lithospheric-scale instabilities (e.g., necking), the thermal history of basins (including the post-rift phase), and sedimentation/erosion, all of which modulate the rheological stratification of the lithosphere. Experiments by Buiter et al. (2009) demonstrate that basin inversion is promoted mainly by: (1) mechanically weak basin fill (relative to the basement rocks), (2) strain-weakened, basin-bounding shear zones or normal faults, and (3) the erosion of sedimentary overburden once basin inversion begins, which facilitates isostatic uplift and further reduces the brittle strength of the crust. Such experiments show that during shortening, localised viscous deformation and isostasy contribute to strain localisation and uplift along pre-existing rift basins.

Lithospheric-scale analogue experiments can be a useful tool for investigating the influence of the lithosphere underneath rift basins, especially its mechanical stratification, in promoting or suppressing basin inversion (e.g., Gartrell et al., 2005; Cerca et al., 2010). Such models allow us to investigate how the interaction between brittle and viscous deformation drives inversion, from the scale of an individual basin to an entire system of basins. While isothermal analogue models do not specifically take into account the thermal structure and evolution of the studied system, model parameters can be chosen such that the experiments simulate first-order natural rift- and inversion-related processes (e.g., upwelling of mantle material under thinned lithosphere due to rifting).

In this paper, we introduce a series of isothermal, lithospheric-scale analogue experiments that simulate continental extension (before reaching the necking and break-up stages) followed by shortening. The aim of our study was to understand how distributed, pre-existing rift basins control basin inversion and orogenesis driven by tectonics (i.e., far-field shortening). We also evaluate the role of rheological layering in promoting or suppressing basement uplift. In any given inverted basin, we can assume that uplift of sedimentary infill is driven by uplift of the underlying basement (Figure 1). In our experiments, we observed the impact of shortening on the topography of the model surface. We refer to the normal fault-bounded, topographic

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lows that formed during extension as "basins". As we did not introduce sedimentary infill during and following extension, we assume that the model surface is analogous to the top of the basement of natural rift basins (i.e., pre-rift rocks). Therefore, we consider a basin to be inverted when the top surface of that basement is displaced upwards.

Our experimental setup is inspired by the Proterozoic basins of the North Australian Craton (northern Australia; Betts et al., 2006), which have long drawn the interest of the petroleum and mineral exploration industries. The mineral-rich lithologies of these basins and their multistage history have been associated with the formation of world class mineral deposits, including the world's single largest source of sediment-hosted Pb–Zn deposits (Mount Isa, Queensland; Betts et al., 2003; Large et al., 2005; Gibson et al., 2016; Gibson and Edwards, 2020), the planet's oldest oil deposits (Northern Mount Isa Basin, Queensland; McConachie, 1993), and conventional and unconventional gas (Greater McArthur Basin, Northern Territory; Cox et al., 2022). This distributed system of intra-cratonic basins in the North Australian Craton underwent multiple phases of extensional and compressional deformation driven by far-field plate boundary processes (Giles et al., 2002; Cawood and Korsch, 2008; Betts and Giles, 2006; Betts et al., 2008, 2011; Scott et al., 2000; Gibson et al., 2008). Our experiments are comparable to the initial basin-forming extensional phase (ca. 1800–1750 Ma; Jackson et al., 2000; Betts et al., 2006) and the shortening phase (ca. 1750–1710 Ma; Betts, 1999; Blaikie et al., 2017; Spence et al., 2021) that followed extension.

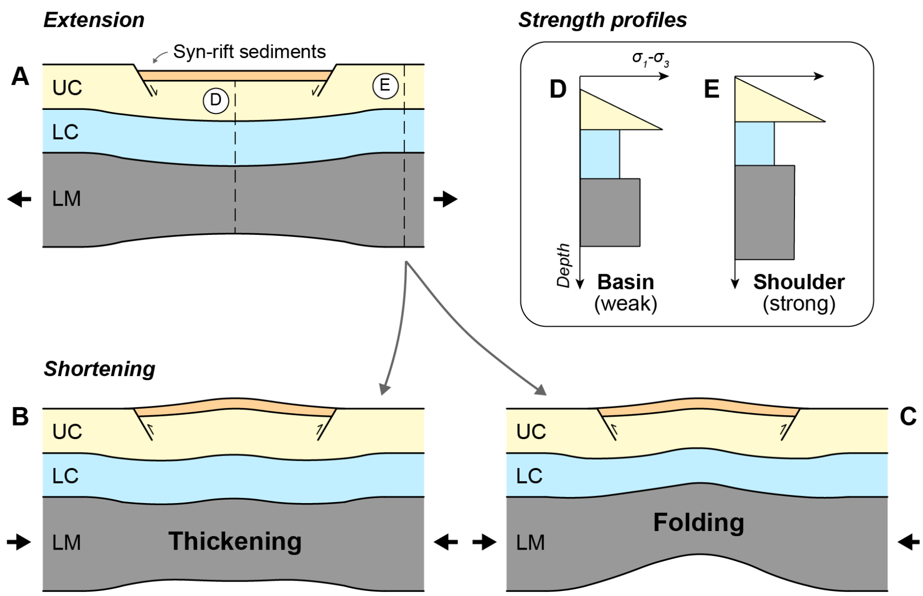

Figure 1: Hypothesised deformation of a lithospheric-scale three-layer analogue model (supported by a fluid asthenosphere, not pictured) during extension and subsequent shortening. The model lithosphere comprises a brittle upper crust (UC), weak ductile lower crust (LC), and strong ductile lithospheric mantle (LM). During extension (a), localised thinning of the strong lithospheric mantle correlates with normal faulting and rift basin formation in the upper crust. During shortening, basin inversion could potentially be driven by thickening (b) or folding and upwelling (c) of the ductile lower crust and lithospheric mantle (cf. Zwaan and Schreurs, 2023). This viscous deformation is accompanied by the reactivation of weakened, rift-related normal faults in a reverse sense (Marques and Nogueira, 2008; Buiter et al., 2009). (d and e) Comparison between strength profiles in the middle of a rift basin and at the rift shoulder.

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The experiments presented here highlight that the initial rheological layering of the models (which represents the thermal and compositional layering of the lithosphere) and the imposed kinematic boundary conditions (i.e., rate of rifting) influence rift evolution and the distribution and segmentation of rift basins. In turn, the distribution of these basins and the rheology of the model layers at the end of extension determines which of the basins are inverted during shortening. This selective uplift of basins in the (brittle) upper crust layer, which has not been observed in previous crustal and lithospheric-scale models of basin inversion, appears to be controlled by viscous deformation of the (ductile) lower crust and lithospheric mantle layers.

2 Experimental method

2.1 Boundary and initial conditions

The model layers comprise a granular "upper crust", ductile "lower crust", and ductile "lithospheric mantle". The ductile materials exhibit spatially continuous deformation at the scale of observation. They behave viscously under our range of experimental strain rates, simulating deformation in the viscous layers of the lithosphere (i.e., the lower crust and lithospheric mantle). The brittle-ductile layers are isostatically supported by a fluid that is analogous to the natural asthenosphere (Figure 2).

The yield strength profiles of the models resemble natural lithospheric strength profiles. The model strength profiles include a relatively strong upper crust as well as lower crust and lithospheric mantle layers of varying relative strengths (Figure 3; Table 1). As the experiments were designed to help us better understand Proterozoic craton-wide rifting in the North Australian Craton (Allen et al., 2015), we implemented a rheological layering that allowed extension to be relatively uniform across the entire model area and create a distributed system of basins (i.e., "wide rifting" *sensu* Buck, 1991; also see Brun, 1999 and Buck et al. 1999). Hence the model lithosphere is analogous to a natural thick lithosphere (with a thick crust) shortly after orogenesis or with a higher-than-normal heat flow (Buck et al., 1999). In models R1 and R2, the thicknesses of the crustal layers scale to 10 km and 40 km for the upper and lower crust, respectively; these were modelled after crustal thickness estimates for the Basin and Range Province (Gueydan et al., 2008), a well-known example of a wide rift (e.g., Hamilton, 1987; Parsons, 2006). The upper and lower crust layers in models R3, R4, and R5 have the same thicknesses, which is representative of the North Australian Craton (Betts et al., 2002; Kennett et al., 2011).

The models were extended at a velocity that scales to 1–2 cm/year in nature (Table 2), which is within the range of estimated rates of extension for the Basin and Range Province (Bennett et al., 1998; Snow and Wernicke, 2000; Hammond and Thatcher, 2004; Tetreault and Buiter, 2018). After ~20% bulk extension, the models were shortened in the reverse direction at the same rate until they reached their initial pre-extension length (Figure 2). Seven to 24 minutes elapsed between the end of extension and the start of shortening, which scales to a maximum of ~1.7 Myr in nature. Model R1 is an exception, as no shortening was applied after extension.

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Table 1: Summary of experimental parameters (UC = upper crust; LC = lower crust; LM = lithospheric mantle).

Exp	Layer thickness			Extension & shortening		Brittle-ductile thickness	Experimental strain rate	Layer strength		Pre-/post-service	Comment
	UC	LC	LM	Velocity	Duration			LC	LM		
	[cm]	[cm]	[cm]	[mm h^{-1}]	[h]		[s^{-1}]				
R1	0.4	1.6	1.2	6.2	14.0	0.1	5.4×10^{-5}	Weak	Weak	Pre	Extension only
R2	0.4	1.6	1.2	6.2	14.0	0.1	5.4×10^{-5}	Weak	Weak	Pre	Lubricated cut on moving side
R3	0.8	0.8	1.2	31.0	3.0	0.4	3.1×10^{-4}	Strong	Strong	Post	
R4	0.8	0.8	1.2	28.3	3.1	0.4	2.8×10^{-4}	Weak	Strong	Post	
R5	0.8	0.8	1.2	28.3	3.1	0.4	2.8×10^{-4}	Weak	Strong	Post	Lubricated cut on moving side

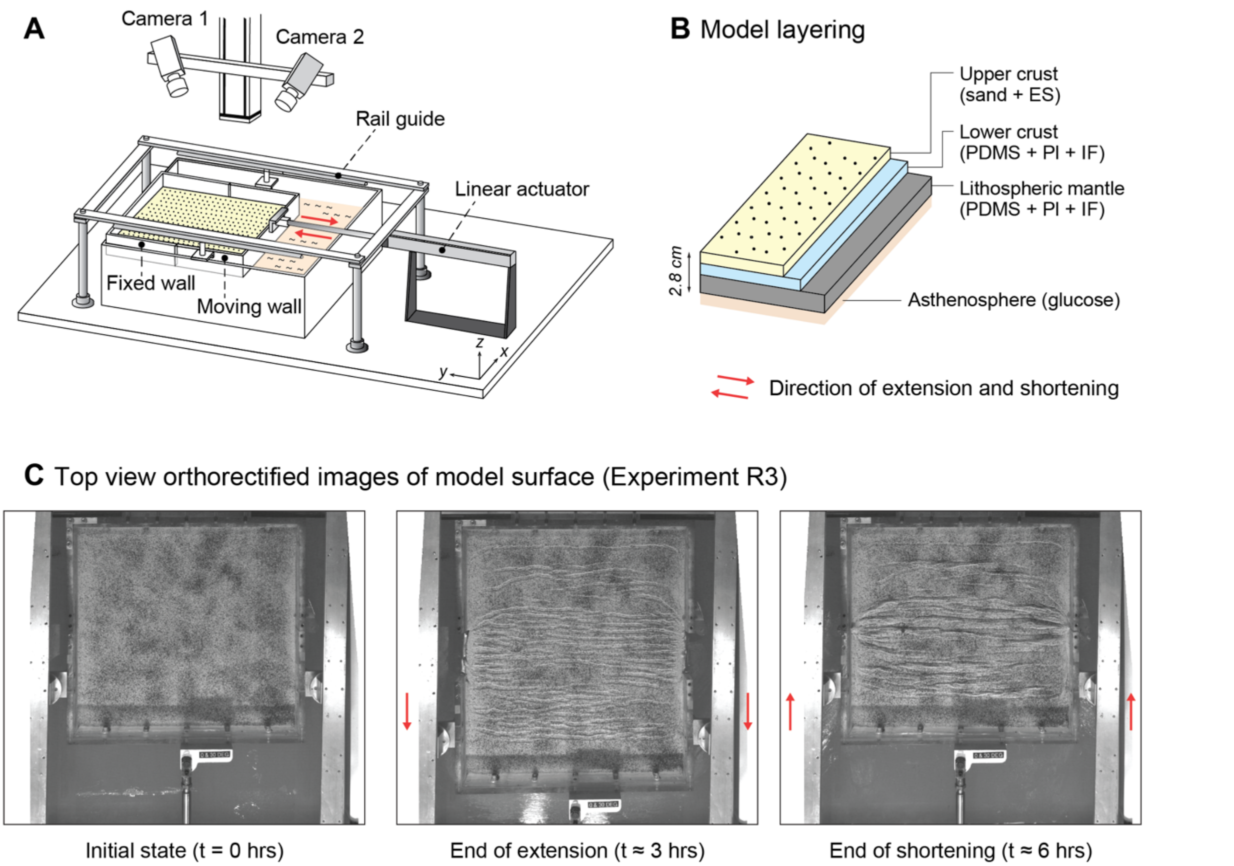

Figure 2: Experimental setup. The red arrows indicate the imposed extension and shortening directions. (b) Cross section of model layers. (c) Orthorectified top view images of the model surface before extension, after extension, and after shortening.

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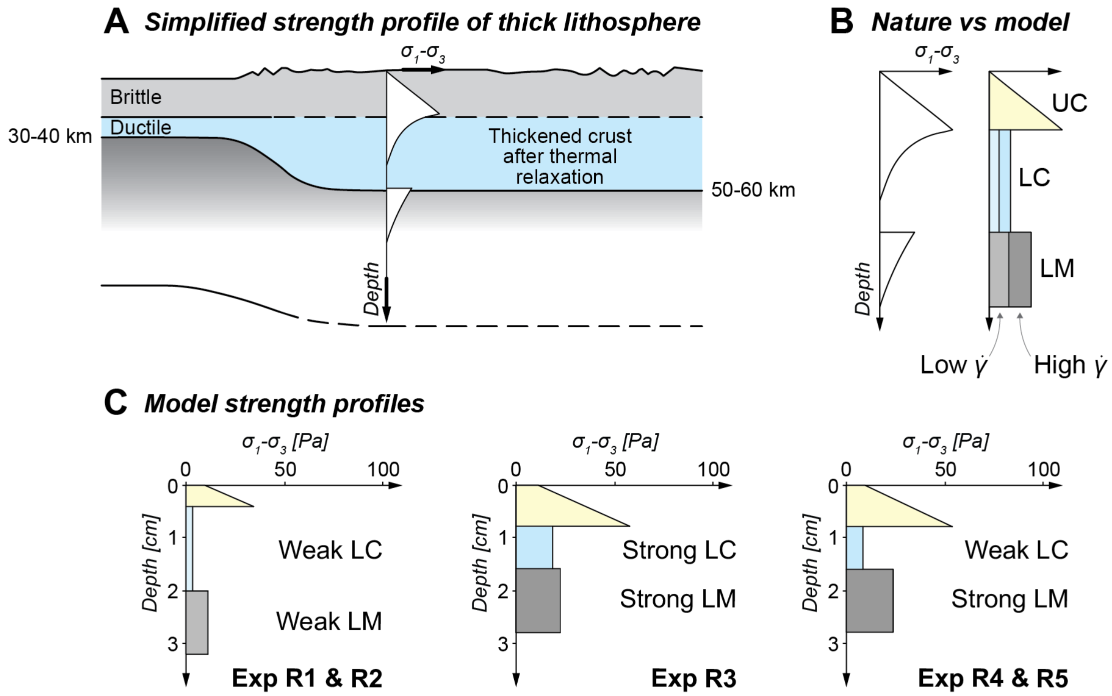

Figure 3: (a) Natural strength profile of a thickened lithosphere (including a thickened crust) after orogenesis and thermal relaxation, which forms widely distributed grabens upon extension (Brun, 1999). (b) Comparison between natural strength profile (same as a) and three-layer analogue model strength profile, where the strength of the ductile layer increases with strain rate $\dot{\gamma}$ (after Brun, 1999). (c) Initial strength profiles of models in this study.

2.2 Scaling parameters and rheology of model materials

The model lithosphere layers were created using granular and ductile materials similar to those used by Molnar et al. (2017) and Samsu et al. (2021). The brittle upper crust, the behaviour of which can be described using Mohr–Coulomb law (Byerlee, 1978), was modelled using a granular mixture comprising mainly dry quartz sand (Rocla 90 Fine Foundry Sand, Hanson Australia). Hollow ceramic Envirospheres® were added to the sand to ensure that the density of the model upper crust is lower than that of the lower crust (Table 2). For the calculation of differential stress in the brittle layer of the strength profiles, it was assumed that this granular mixture has very low cohesion (i.e., ~9 Pa; Molnar et al., 2017).

The ductile lower crust and lithospheric mantle layers were modelled using mixtures that mostly consist of polydimethylsiloxane (PDMS). For the lithospheric mantle, black Colorific® plasticine was added to the PDMS in order to increase its effective viscosity (Boutelier et al., 2008). On its own, PDMS is a Newtonian fluid (e.g., Boutelier et al., 2008), meaning that its viscosity is strain rate-independent. However, the PDMS-based mixtures used to model the lithospheric mantle in our experiments are non-Newtonian, as they exhibit strain rate-softening behaviour (stress exponent $n > 1$) (Table 2).

Iron filings were also added to the lower crust and lithospheric mantle material in order to increase their densities. For models R4 and R5, the addition of fine-grained iron filings (0.42–0.82 mm grain size, manufactured for Chem-Supply Australia) did

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not significantly affect the flow behaviour of PDMS during our experiments. However, in models R1, R2, and R3, the use of ultrafine-grained iron filings (manufactured for Mad About Science), which has a powder-like consistency, doubled the viscosity of the PDMS (Table 2). Hence the yield strength profile for R3 contains a "strong" lower crust, while models R4 and R5 contain a "weak" lower crust (Figure 3).

The scaling parameters (Ramberg, 1967) used in our experiments and the properties of the model layers are presented in (Table 2). These parameters were chosen so that model deformation is consistent with natural processes but occurs over a time scale that is convenient for laboratory experiments. The length scaling factor $L^* = L_m/L_p = 4 \times 10^{-7}$ meant that 0.4 cm in the model scales to 10 km in nature, whereby the subscripts *m* and *p* denote the model and natural prototype, respectively. Therefore, the model surface area of 44 cm x 40 cm corresponds to 1100 km x 1000 km in nature, and the model thicknesses of 2.8–3.2 cm represent lithospheric thicknesses of 70–80 km. The bulk extension and shortening rate scales to 1 cm/year in nature; this extension rate falls within the range of estimated divergence rates for the Basin and Range Province (Tetreault and Buiter, 2018). The extensional phase ended after ~8 cm of extension (~20% extension), after which the model was shortened at the same rate in the exact opposite direction until it reached its initial length, simulating orthogonal extension and shortening. The density scaling factor $\rho^* = \rho_m/\rho_p$ was set to 0.46 based on the density ratio between the model asthenosphere material (Queen Glucose Syrup; Schellart, 2011) and estimated asthenosphere densities between 3100 kg/m^3 and 3400 kg/m^3, consistent with previous lithospheric-scale analogue experiments (e.g., Molnar et al., 2017; Santimano and Pysklywec, 2020; Samsu et al., 2021). Similarly, the viscosity scaling factor $\eta^* = \eta_m/\eta_p$ was determined using the ratio between the effective viscosity of the model asthenosphere and that of the natural asthenosphere. As the experiments were conducted under normal gravitational acceleration, the scaling factor for acceleration due to gravity $g^* = g_m/g_p = 1$, resulting in a stress scaling factor $\sigma^* = \rho^* \times g^* \times L^*$ between 1.7×10^{-7} and 1.9×10^{-7}.

There are significant differences between the scaling parameters for models R1 and R2 and the rest of our models. For R1 and R2, we assumed that: (1) the viscosity of the natural asthenosphere is one order of magnitude lower than R3, R4, and R5; (2) the density of the natural asthenosphere (3100 kg/m^3) is lower than in R3, R4, and R5 (3400 kg/m^3); and (3) the time scaling factor ($t^* = \eta^* / (\rho^* \times g^* \times L^*)$) is an order of magnitude higher, and therefore the duration of extension and shortening for R1 and R2 (14 hours) is much longer than for R3, R4, and R5 (2.8 to 3.1 hours). This means that the duration of extension and shortening each corresponds to ~11 Myr each for R1 and R2 and ~22–23 Myr each for R3, R4, and R5.

The difference in t^* also means that the experimental strain rate for models R1 and R2 (~5.4×10^{-5} s^{-1}) is one order of magnitude lower than for R3 (~3.1×10^{-4} s^{-1}) and R4 and R5 (~2.8×10^{-4} s^{-1}). These strain rates were estimated by dividing the rate of extension (i.e., the velocity of the moving wall during extension) by the initial thickness of the model lithosphere (Benes and Scott, 1996). As the strength of the ductile layers increases with the applied strain rate (Ranalli, 1995; Brun, 1999), the lithospheric mantle in models R1 and R2 is weak compared to the lithospheric mantle in R3, R4, and R5 (Figure 3). For R1 and R2, the strength profile shows that the lower crust and lithospheric mantle are both weak compared to the upper crust, due to the slow strain rate applied to these experiments.

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The scaling parameters used in model R3 and models R4 and R5 are relatively similar, with the main difference being the effective viscosity of the lower crust in R3 being twice greater than in R4 and R5 (Table 2). As a result, in Model R3 the lower crust is almost as strong as the lithospheric mantle (Figure 3). In R4 and R5, the lower crust is much weaker than the lithospheric mantle. The relative strength of the lower crust with respect to the overlying and underlying layers affects the mechanical coupling between the upper crust and lithospheric mantle and therefore the strain distribution in the upper crust, as discussed further in Section 3.1.

Table 2: Scaling parameters for all experiments. Abbreviations of modelling materials: ESPH = Envirospheres, BPL = black Plasticine, IF_{uf} = ultrafine iron filings, IF_f = fine iron filings.

Models R1 & R2		**Thickness**		**Density**		**Viscosity**			
		Model	Nature	Model	Nature	Model	Nature	Stress	
		(mm)	(km)	(kg/m^3)	(kg/m^3)	(Pa s)	(Pa s)	exponent	Material
Upper crust	Brittle	4	10	1222	2650	-	-		Sand+ESPH
Lower crust	Ductile	16	40	1245	2700	6.0 x 10^4	2.2 x 10^{21}	1	PDMS+IF_{uf}
Lithospheric mantle	Ductile	12	30	1338	2900	3.6 x 10^5	1.3 x 10^{22}	1.36	PDMS+BPL+IF_{uf}
Asthenosphere	Fluid	-	-	1430	3100	520	1.9 x 10^{19}		Glucose
Scaling factors		L^* = 4.0 x 10^{-7}		ρ^* = 4.6 x 10^{-1}		η^*= 2.7 x 10^{-17}			
		t^* = 1.5 x 10^{-10}		g^* = 1		1 h in model ~ 0.8 Myr in nature			
		v^* = 2.7 x 10^3		σ^* = 1.9 x 10^{-7}		3.1 mm/h in model ~ 10 mm/yr in nature			
Model R3		**Thickness**		**Density**		**Viscosity**			
		Model	Nature	Model	Nature	Model	Nature	Stress	
		(mm)	(km)	(kg/m^3)	(kg/m^3)	(Pa s)	(Pa s)	exponent	Material
Upper crust	Brittle	8	20	1245	2700	-	-		Sand+ESPH
Lower crust	Ductile	8	20	1315	2850	6.0 x 10^4	2.2 x 10^{22}	1	PDMS+IF_{uf}
Lithospheric mantle	Ductile	12	30	1384	3000	2.1 x 10^5	7.8 x 10^{22}	1.36	PDMS+BPL+IF_{uf}
Asthenosphere	Fluid	-	-	1430	3100	5.2 x 10^2	1.9 x 10^{20}		Glucose
Scaling factors		L^* = 4.0 x 10^{-7}		ρ^* = 4.6 x 10^{-1}		η^*= 2.7 x 10^{-18}			
		t^* = 1.5 x 10^{-11}		g^* = 1		1 h in model ~ 7.8 Myr in nature			
		v^* = 2.7 x 10^4		σ^* = 1.9 x 10^{-7}		31 mm/h in model ~ 10 mm/yr in nature			
Models R4 & R5		**Thickness**		**Density**		**Viscosity**			
		Model	Nature	Model	Nature	Model	Nature	Stress	
		(mm)	(km)	(kg/m^3)	(kg/m^3)	(Pa s)	(Pa s)	exponent	Material
Upper crust	Brittle	8	20	1136	2700	-	-		Sand+ESPH
Lower crust	Ductile	8	20	1199	2850	3.0 x 10^4	1.1 x 10^{22}	1	PDMS+IF_f
Lithospheric mantle	Ductile	12	30	1304	3100	2.7 x 10^5	9.0 x 10^{22}	1.37	PDMS+BPL+IF_f
Asthenosphere	Fluid	-	-	1430	3400	520	1.9 x 10^{20}		Glucose
Scaling factors		L^* = 4.0 x 10^{-7}		ρ^* = 4.2 x 10^{-1}		η^*= 2.7 x 10^{-18}			
		t^* = 1.6 x 10^{-11}		g^* = 1		1 h in model ~ 7.1 Myr in nature			
		v^* = 2.5 x 10^4		σ^* = 1.7 x 10^{-7}		28 mm/h in model ~ 10 mm/yr in nature			

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2.3 Deformation monitoring and analysis

Digital image correlation (DIC) was applied to sequential images of the model surface in order to monitor deformation in the cover layer during the experiment. This technique allowed us to observe the strain and topographic evolution of the models. Strain maps and orthorectified photographs of the model surface were used to track the formation of rift basins and inversion structures at different stages of the experiments.

The image acquisition and DIC workflow is similar to that outlined in Molnar et al. (2017) and Samsu et al. (2021). The DIC system comprises two cameras at oblique angles to the model surface (Figure 2a). Images were recorded at five-minute intervals over 14 hours for experiments R1 and R2 and at two-minute intervals over approximately three hours for experiments R3, R4, and R5 for each extension or shortening phase. Surface strain and topography were computed using the StrainMaster module of the commercial image correlation software DaVis (version 10.1.2, LaVision). The software uses stereo cross correlation to compute the incremental displacement field from which the strain tensor components derived.

For the strain maps, the displacement vector fields obtained from DaVis were used to derive incremental and cumulative axial strain (e_{yy} and E_{yy}, respectively) in MATLAB. e_{yy} and E_{yy} are measures for normal strain parallel to the extension and shortening direction. E_{yy} was computed from the displacement gradient tensor. The displacement gradient tensor H comprises the components $\frac{\Delta D_i}{\Delta x_j}$, where D_i are the displacement components in the x- and y-direction (i.e., D_u and D_v) and x_i refer to the x- and y-axis of the coordinate system (with the y-axis being parallel to the extension and shortening axis). Using the Lagrangian finite strain tensor E (Allmendinger et al., 2011):

$$E = \begin{bmatrix} E_{xx} & E_{xy} \\ E_{yx} & E_{yy} \end{bmatrix} \quad (1).$$

The normal strain along the extension and shortening axis can be calculated with:

$$E_{yy} = \frac{1}{2}\left(\frac{\Delta D_u}{\Delta y} + \frac{\Delta D_v}{\Delta y} + \left(\frac{\Delta D_u}{\Delta y}\frac{\Delta D_u}{\Delta y} + \frac{\Delta D_v}{\Delta y}\frac{\Delta D_v}{\Delta y}\right)\right) \quad (2).$$

The incremental vertical displacement (d_w), cumulative vertical displacement (D_w), and the height (i.e., topography) of the model surface were also calculated in DaVis.

The incremental and cumulative displacement field data generated by DaVis were imported into MATLAB for post-processing for visualisation purposes. Post-processing of data included detection and replacement of spurious displacement vectors and interpolation of missing data. To this end, we used the DCT-PLS algorithm (Garcia, 2011). Topographic data was corrected by fitting a plane through the initially flat but possibly tilted model surface. Finally, the resulting linear correction parameters were applied to all subsequent digital elevation models. The MATLAB scripts used for post-processing and visualisation are available at https://github.com/TimothySchmid/PIV_postprocessing_2.0.

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3 Results

Here we present the results of five experiments titled R1 to R5 (Table 1). With the exception of R1, each experiment comprises an "extension" and subsequent "shortening" phase, discussed separately in the subsections below. The resulting fault strikes and basin long axes are roughly perpendicular to the extension and shortening direction, given the kinematic boundary conditions that simulate orthogonal rifting and shortening. When viewing the models in map view, the upper side of the image is referred to as "north", and the model is being extended towards the "south". Curvature of the deformation features near the western and eastern model edges results from friction between the model edges and the confining U-shaped walls. We therefore limit our analysis to the central area that is unaffected by these boundary effects.

3.1 Extension: Normal faulting and basin formation

In all of our experiments, the imposed bulk extension of the model resulted in an extension-orthogonal, E-W trending horst and graben system. Here we define "graben" as a depressed area bounded by parallel normal faults (Reid et al., 1913; Peacock et al., 2000) which accommodate the bulk extension imposed on the models (Figure 4,Figure 5, Figure 7). We also use the terms "graben" and "basin" interchangeably. The wide distribution of basins is analogous to natural examples of wide rifting, such as in the Proterozoic North Australian Craton (Allen et al., 2015; Betts et al., 2008), Basin and Range Province (Wernicke et al., 1988), Aegean Sea (Doutsos and Kokkalas, 2001), and East China Rift System (Tian et al., 1992).

In the experiments, rift evolution occurred in two main phases: (1) rift basin formation associated with normal fault nucleation, growth, and linkage; and (2) basin deepening and widening. In general, the rift basins formed as faults reached their final length. Progressive extension allowed the throw along the basin-bounding faults to increase, resulting in deepening of the grabens. Downward fault propagation was limited by the thickness of the upper crust, after which strain was accommodated by widening of the grabens.

The timing of the above processes with respect to amount of bulk extension varied with different model setups (see initial strength profiles and boundary conditions in Figure 3 and Table 1). The rift system evolved most quickly in models R4 and R5, followed by R3 and then R1 and R2. For example, the model topography shows that by 10% bulk extension, graben-bounding faults in R4 and R5 appear to have already reached their full lengths, while the faults in R1, R2, and R3 are still in the nucleation and growth stage (c.f., Figure 4 Figure 5).

The models also exhibit different degrees of strain distribution, in terms of the spacing of faults and grabens, at the end of the extensional phase (at 19–20% bulk extension; Figure 4 Figure 5). Grabens in R1 and R2 are evenly spaced across the model area. Their spacing is greater compared to R3, R4, and R5. They are highly segmented, bounded by normal faults with short and irregular (non-linear) fault traces. These grabens are also relatively shallow, due to the thin upper crust layer (Table 1).

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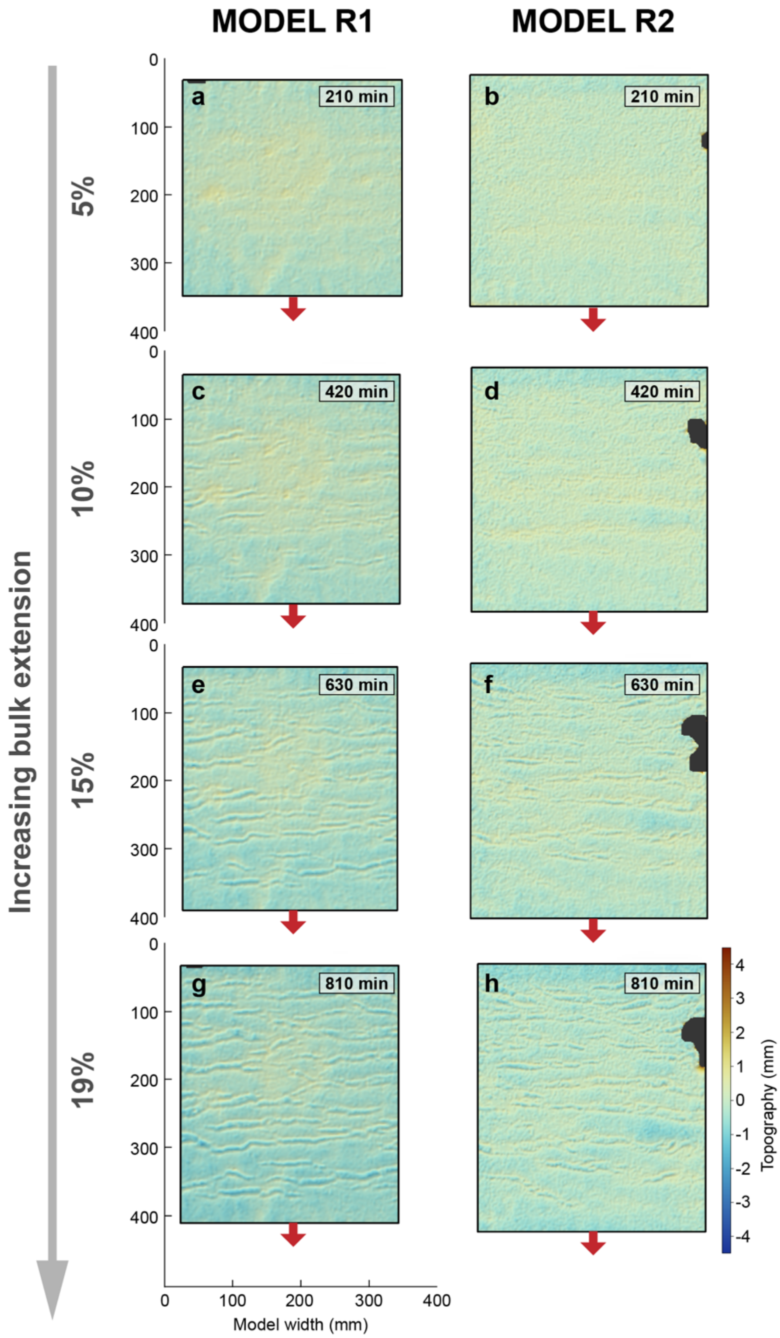

245 **Figure 4: Topography of models R1 and R2 at increasing durations and amounts of extension applied to the model (e.g., a and b correspond to 5% bulk extension, c and d correspond to 10% bulk extension, etc). Arrows show the direction of extension. The resulting rift basins are uniformly spaced along the y-axis.**

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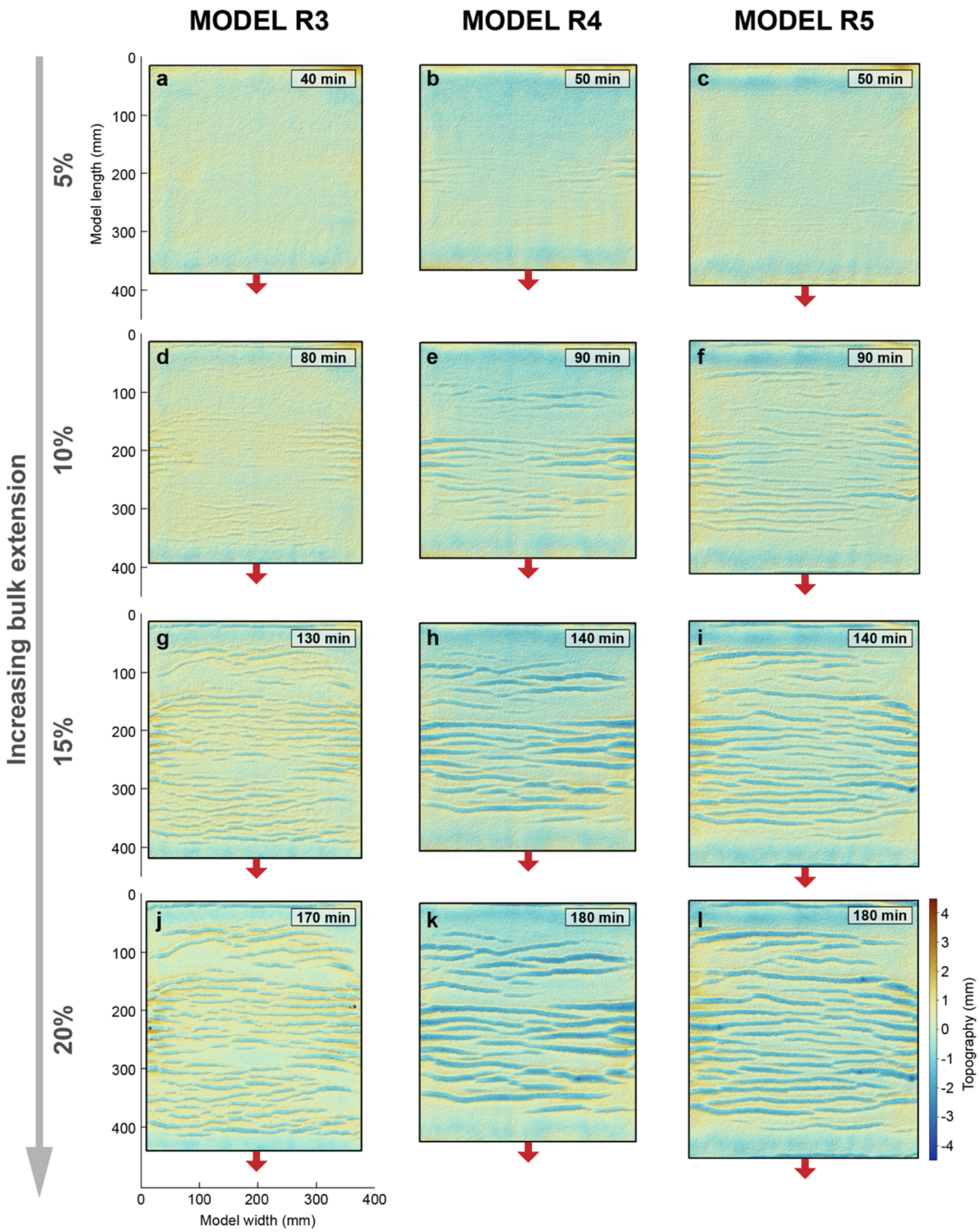

250 **Figure 5: Topography of models R3, R4, and R5 at increasing durations and amounts of extension applied to the model (e.g., a, b, and c correspond to 5% bulk extension, c, d, and e correspond to 10% bulk extension, etc). The rift basins are not uniformly spaced. Arrows show the direction of extension.**

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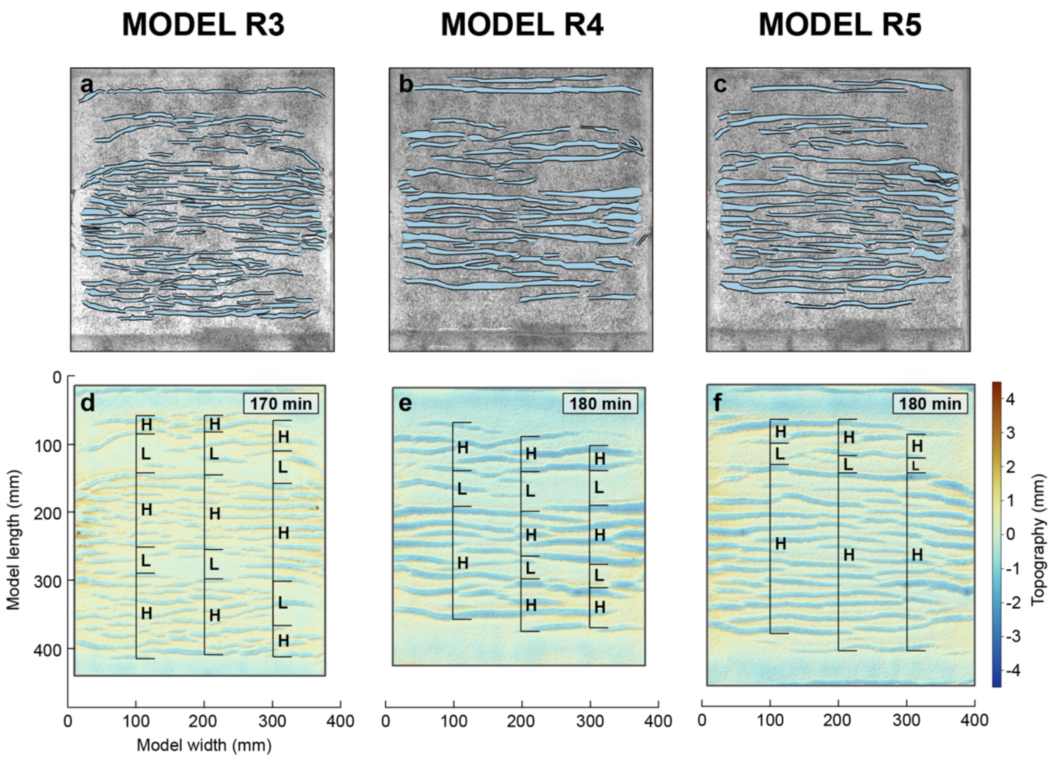

Figure 6: (a–c) Orthorectified top-view photos of the R3, R4, and R5 model surface at the end of extension, with overlay of the interpreted basins and basin-bounding fault traces. (d–f) High-strain (H) and low-strain (L) zones at the end of extension. Zones are marked along a profile at x = 100, 200, and 300 mm.

Fault traces in models R3, R4, and R5 are smoother than in R1 and R2. Grabens in Model R3 are narrower and more segmented (i.e., less laterally continuous along the graben axes) than those in models R4 and R5. Models R3, R4, and R5 exhibit clusters of grabens that make up so-called "high-strain zones" which are separated from each other by "low-strain zones" (Figure 5). Hence, basins are not distributed uniformly across the model area as they are in models R1 and R2. There is no strain in the northernmost and southernmost ~5 cm of the model, as these segments were attached to the confining plexiglass walls; these areas are not included in the subsequent analyses. In the north of the model, there is a narrow low-strain zone adjacent to a wide high-strain zone, the latter of which makes up approximately the southern two-thirds of the model area. In models R3 and R4, another low-strain zone is present within the wide high-strain zone (Figure 6).

3.2 Shortening

3.2.1 Strain localisation at the boundaries of pre-existing basins

During shortening, strain is localised along pairs of basin-bounding normal faults that formed during the preceding extensional phase (Figure 7). At lower percentages of bulk shortening, E_{yy} is localised along the edges of the grabens, suggesting that rift-related normal faults are reactivated in a reverse sense during shortening. In models R3, R4, and R5, high strain accumulation

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during shortening occurs in the high-strain zones formed during the extensional phase. In addition, deformation is more intense in the southern part of the model compared to the north (see Model R5 example; Figure 7). The high cumulative strain in the south may be related to the increasing displacement gradient from north to south.

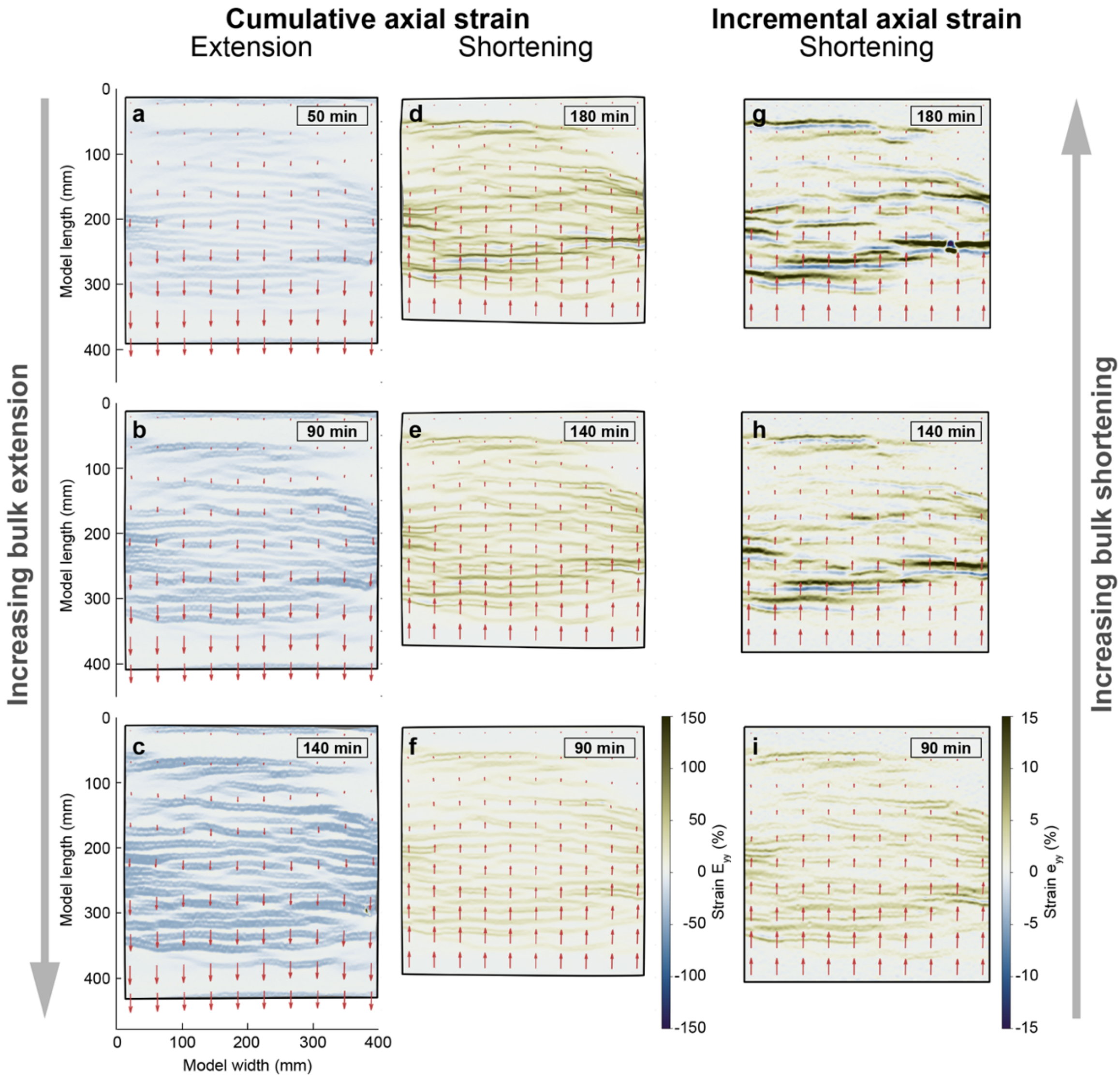

Figure 7: Evolution of cumulative axial strain (E_{yy}) during extension and shortening (a-f) and incremental axial strain (e_{yy}) during shortening (g-i) in Model R5. During extension, E_{yy} is localised along normal faults that form the edges of grabens (i.e., basin-bounding faults). During shortening, deformation is localised first along the same basin-bounding faults and then within the basins. Vector lengths represent relative amounts of displacement within the model.

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3.2.2 Correlation between axial strain and topography

High axial strain (E_{yy}) during shortening coincides with high vertical displacement (D_w) and high topography (Figure 8). We interpret these linear, high-topography features to be analogous to inverted basins after low amounts of shortening and orogens (i.e., mountain belts) after high amounts of shortening. The inverted basins and "orogens" on the surface of the model are underlain by uplifted (ductile) lower crustal material, which can be observed after the granular upper crustal material has been removed at the end of the experiments. A comparison of the topography of models R3, R4, and R5 at the end of shortening (~19–20% bulk shortening) shows that orogens are more laterally continuous (i.e., less segmented) when they form along laterally continuous, pre-existing grabens (i.e., models R4 and R5; Figure 9). In contrast, each elongate uplifted area in Model R3 correlates with several segmented pre-existing basins.

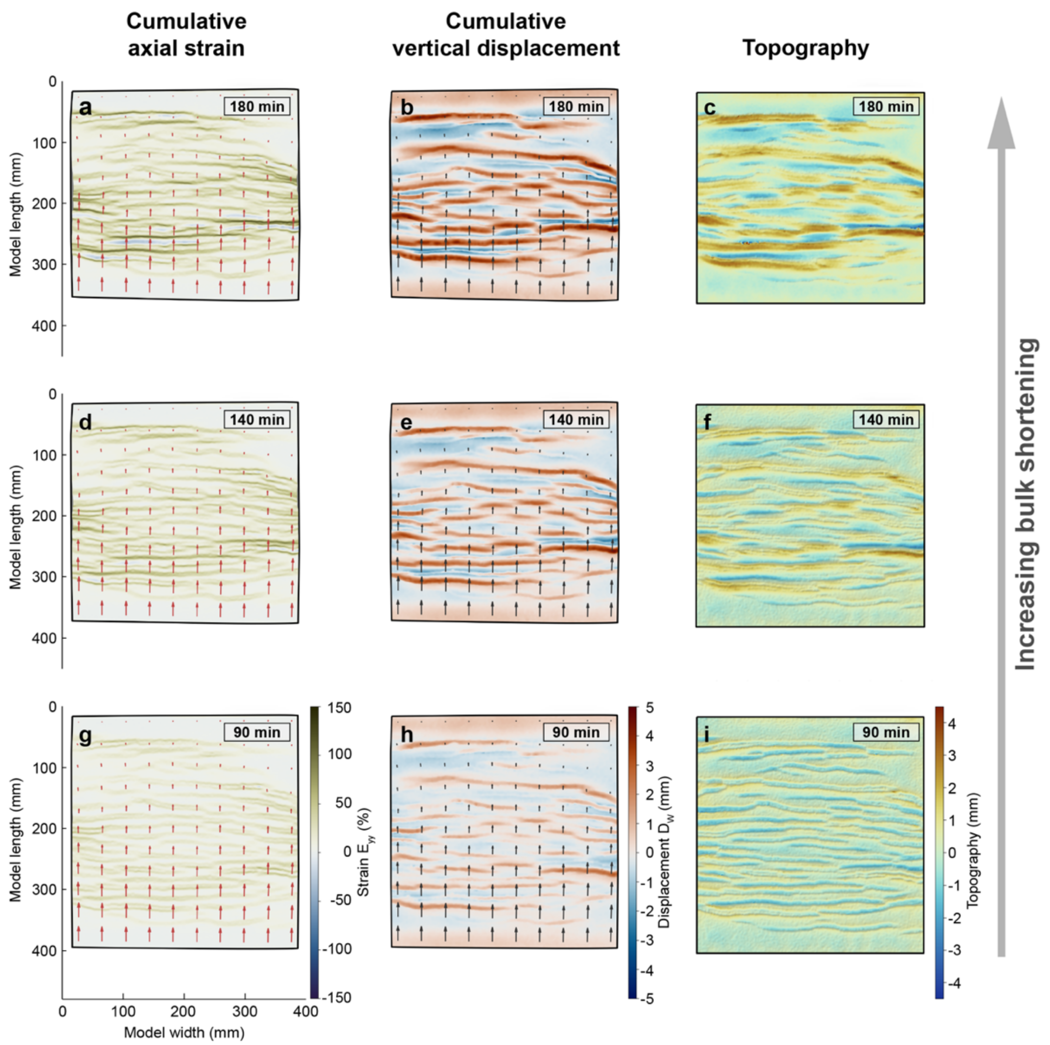

Figure 8: Evolution of cumulative axial strain (E_{yy}), cumulative vertical displacement (D_w), and topography during shortening in Model R5.

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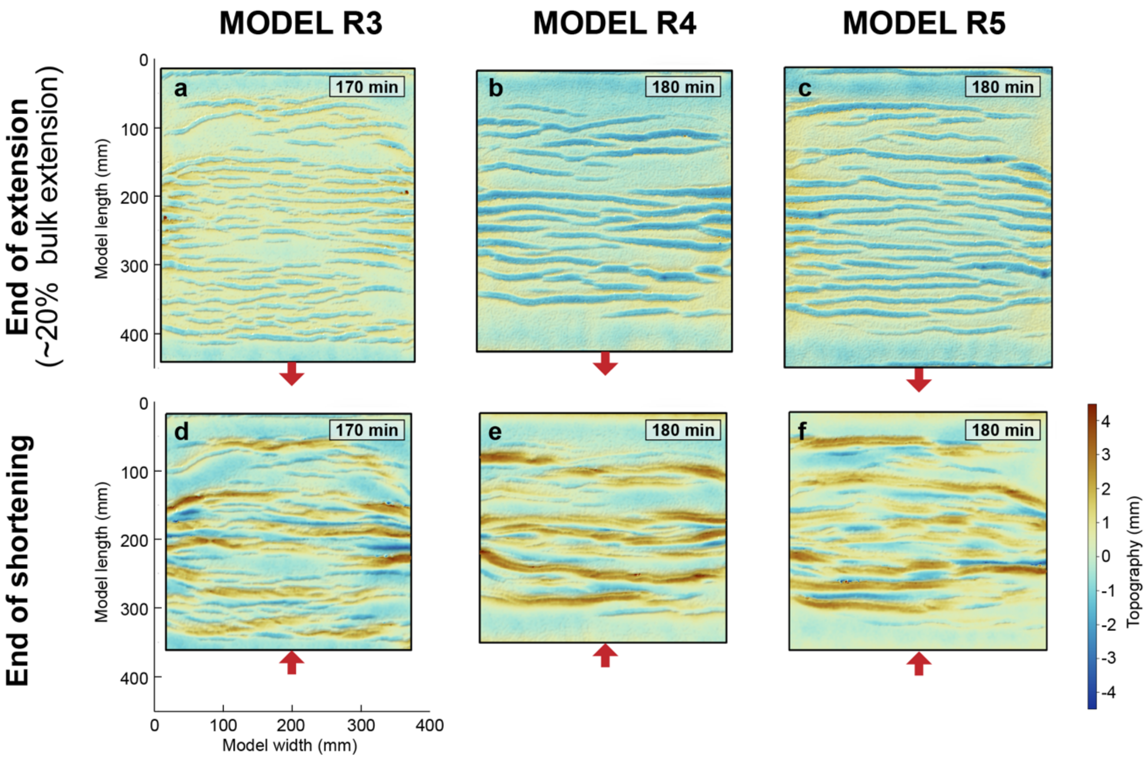

290 **Figure 9: Topography of models R3, R4, and R5 at the end of extension and shortening. Highly segmented extensional basins correlate with highly segmented orogens (R3). In contrast, laterally continuous extensional basins appear to localise laterally continuous orogens (R5).**

3.2.3 Selective uplift of basins

The topography of models R3, R4, and R5 at the end of the extension and shortening phases (Figure 9 and Figure 10) suggests that some basins evolve into high-topography areas during shortening, while others remain as topographic lows. This selective inversion of basins is emphasised in the topographic profiles of models R3 and R5 (Figure 11, Table 3 and Table 4). There appears to be a periodicity of uplift along the y-axis (or N-S axis) of the models, with regular spacing between basins that were eventually uplifted. The basins that remained as basins during shortening localised a high amount of axial strain in the direction opposite to shortening (negative axial strain in Figure 7). In contrast to models R3, R4, and R5, all of the basins in models R1 and R2 were uplifted during shortening. We discuss possible explanations for these uplift patterns in Section 4.2.

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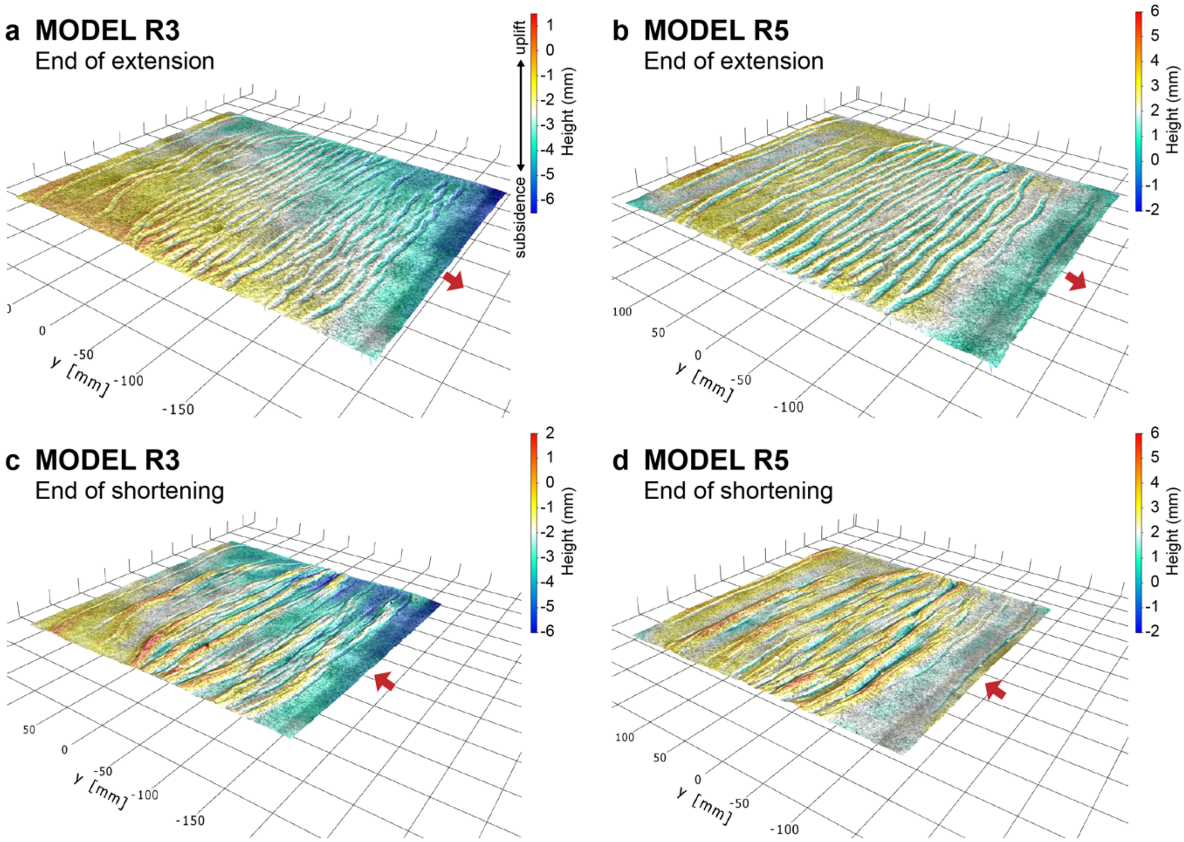

Figure 10: Oblique 3D view of topography of models R3 and R5 at the end of the extension and shortening phases. This 3D visualisation was done in DaVis prior to the postprocessing steps outlined in Section 2.3.

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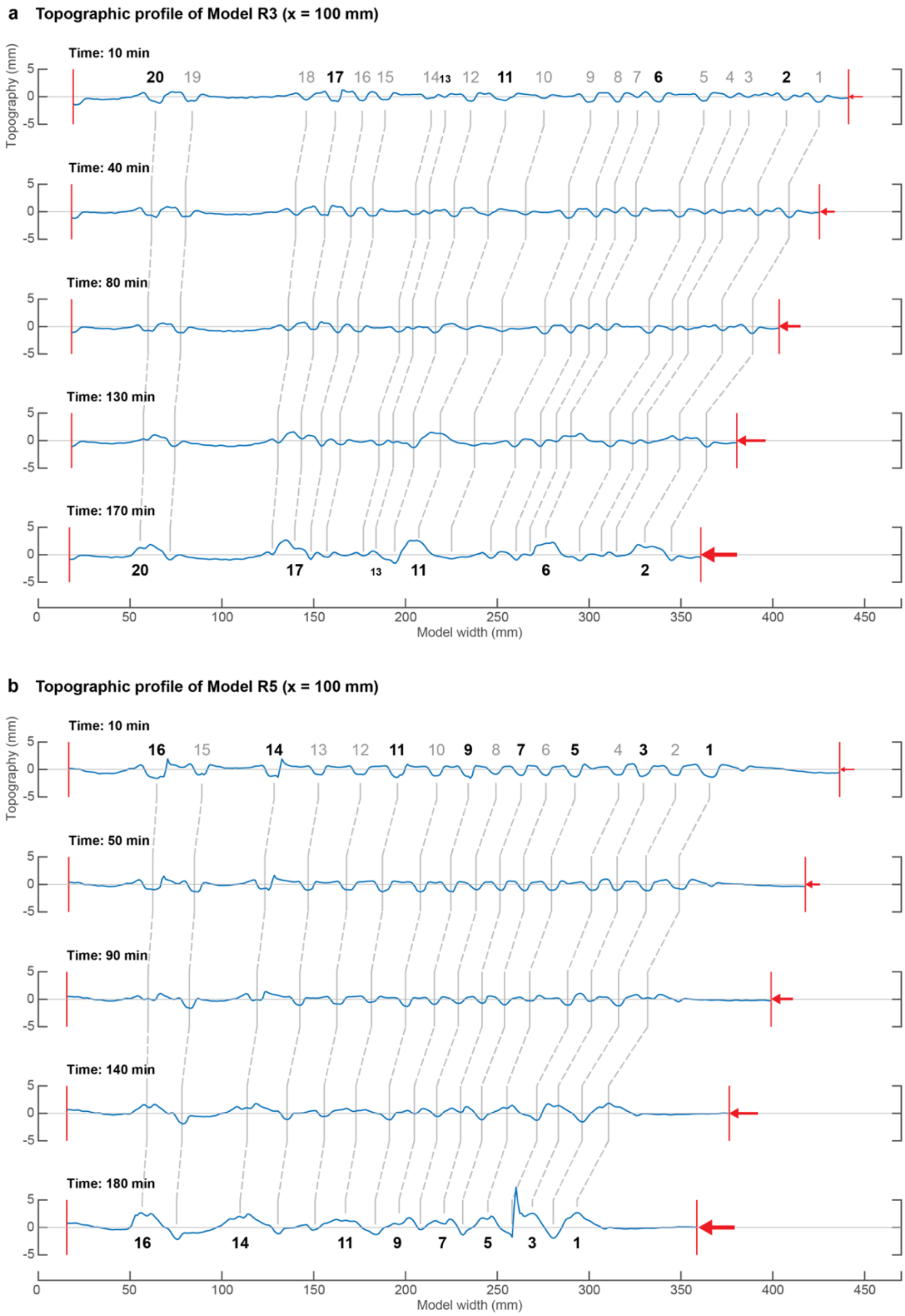

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Figure 11: Evolution of the N-S topographic profiles of models R3 and R5 during shortening, drawn along x = 100 mm (see location in Figure 5). The numbers denote basins which had formed by the end of the extensional phase; numbers in bold correspond to basins that were uplifted during shortening. The red arrows represent the direction and cumulative amount of bulk shortening.

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Table 3: Model R3 basin depths and positions 10 min after the start of shortening. Basins 2, 6, 11, 17, and 20 were uplifted by the end of shortening (Figure 11a); the even spacing between them suggests that the dominant wavelength for basin uplift (measured from the start of shortening) is between 69.2 mm and 98.1 mm. The increasing spacing between basins towards the north correlates with the lower displacement velocity towards the northern edge of the model.

Basin	y-position [mm]	Depth [mm]	Δy from previous basin [mm]	Δy from previous inverted basin [mm]
1	424.6	-1.0		
2	406.1	-0.6	18.5	
3	386.5	-0.1	19.6	
4	376.1	-0.2	10.4	
5	362.3	-0.7	13.8	
6	336.9	-0.9	25.4	69.2
7	325.3	-0.1	11.5	
8	315.0	-0.9	10.4	
9	300.0	-0.9	15.0	
10	274.6	-0.3	25.4	
11	252.7	-0.6	21.9	84.2
12	234.2	-0.8	18.5	
13	220.4	-0.1	13.8	
14	212.3	-0.3	8.1	
15	188.1	-0.5	24.2	
16	175.4	-0.7	12.7	
17	161.5	-0.7	13.8	91.1
18	145.4	-0.6	16.2	
19	83.1	-0.7	62.3	
20	63.5	-0.9	19.6	98.1

Table 4: Model R5 basin depths and positions 10 min after the start of shortening. Basins 1, 3, 5, 7, 9, and 11 were uplifted by the end of shortening (Figure 11b); the even spacing between them suggests that the dominant wavelength for basin uplift (measured from the start of shortening) is between 29.0 mm and 38.4 mm. The greater distance between 11, 14, and 16 correlates with the lower displacement velocity towards the northern edge of the model.

Basin	y-position [mm]	Depth [mm]	Δy from previous basin [mm]	Δy from previous inverted basin [mm]
1	366.9	-1.4		
2	347.2	-1.1	19.7	
3	329.6	-1.2	17.6	37.3
4	316.2	-1.0	13.5	
5	292.3	-1.3	23.8	37.3
6	276.8	-1.1	15.5	
7	263.3	-1.1	13.5	29.0
8	248.8	-0.9	14.5	
9	234.3	-1.2	14.5	29.0
10	216.6	-1.1	17.6	
11	195.9	-1.4	20.7	38.4
12	175.2	-0.9	20.7	
13	153.4	-1.0	21.8	
14	128.5	-1.1	24.9	67.4
15	89.1	-0.7	39.4	
16	64.3	-1.6	24.9	64.3

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4. Discussion

4.1 Rheological controls on rift basin distribution

Deformation of the extended lithosphere is accommodated by brittle faulting in the upper crust and viscous flow of the lower crust and lithospheric mantle. In our experiments, an initial period of distributed extension was followed by the localisation of deformation onto rift-related normal faults, which controlled the formation of rift basins (Figure 4, Figure 5, and Figure 7). The wide distribution of basins is consistent with previous extensional experiments of brittle-ductile models in which a rift seed was not implemented (e.g., Benes and Davy, 1996; Gartrell, 1997; Corti, 2005), which would have otherwise localised rifting from the onset of extension (i.e., "narrow rifting" in Buck, 1991). While all of our models demonstrate wide rifting, different degrees of mechanical coupling between the model layers appear to have influenced the details of rift evolution (i.e., timing of basin formation) and the overall distribution of faults and basins.

In a wide rift mode, regular spacing between basins reflects the characteristic wavelengths of periodic instabilities during extension. These instabilities require a strength or viscosity contrast between two or more layers, and they form uniformly spaced domains of greater thinning, known as boudinage or pinch-and-swell structures (Ramberg, 1955; Smith, 1977). The formation of rift basins is controlled by two different wavelengths of periodic instabilities, which occur at a smaller scale in the brittle upper crust (i.e., crustal boudinage) and at the whole-lithosphere scale (i.e., lithospheric boudinage) (e.g., Benes and Davy, 1996). It is possible that the characteristic wavelengths of deformation localisation in our experiments is a product of the superposition of crustal and lithospheric boudinage, given their brittle-ductile, multi-layer setup. In models R3, R4, and R5, zones of localised lithospheric necking may correspond with high-strain zones, while areas that underwent minimal stretching may correspond with low-strain zones (Figure 5 and Figure 6). Here, the distance between the centres of the high-strain zones may represent the characteristic wavelength of lithospheric-scale boudinage.

Observations on the relationship between characteristic wavelengths and layer thicknesses in our experiments are not directly comparable with previous analytical predictions based on two-layer (e.g., Fletcher and Hallet, 1983) and three-layer models (e.g., Montési and Zuber, 2003), given that our model lithosphere rests on and is likely to be affected by the viscously deforming asthenosphere underneath. In addition, our lithospheric mantle has a stress exponent $n \approx 1.36$–1.37, lower than the estimated stress exponent for the natural lithospheric mantle ($n \approx 3$) used in the calculations by Fletcher and Hallet (1983). However, some general observations can be made on how the spacing between basins varies between our experimental setups. For example, in Models R1 and R2, the spacing between basins is on the order of the lithospheric thickness (3.2 cm). In Models R3, R4, and R5, the spacing between basins is on the order of the crustal thickness (1.8 cm), which is more in line with observations from wide rift experiments by Benes and Davy (1996).

The coupling between the layers in analogue experiments is controlled by the relative strengths (i.e., effective viscosities) of the ductile layers, which is in turn influenced by the rate of extension (Zwaan et al., 2021; Brun, 1999). The rate of extension in models R1 and R2 was much slower than in the other models (Table 1), so that the lithospheric mantle was relatively weak and underwent uniform thinning during extension. The thick and also weak lower crust in R1 and R2 had sufficient time to

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flow during extension. As a result, both ductile layers thinned over a wide region, so that strain was distributed evenly in the overlying upper crust, resulting in evenly spaced basins from north to south (cf. Benes and Davy, 1996). The wide spacing between the basins could be attributed to the large ratio of upper to lower crustal strength (Figure 3), which predicts large spacing between basins, with each basin-bounding fault taking up a relatively large amount of strain (Wijns et al., 2005; Corti, 2005).

The faster rate of extension in models R3, R4, and R5 resulted in a relatively strong lithospheric mantle compared to R1 and R2 (Figure 3; cf. Brun, 1999; Nestola et al., 2015). This strong lithospheric mantle is overlain by a strong ductile lower crust in R3 and weak ductile lower crust in R4 and R5, which results in stronger coupling with the strong brittle upper crust in models R4 and R5. In previous experiments by Gartrell (1997), the brittle upper crust was underlain by a strong, high-viscosity ductile layer (i.e., a so-called stress guide) and weak ductile lower crust. In their experiments, necking instabilities developed in the strong ductile layer and localised deformation into rift basins in the directly overlying upper crust. Similarly, our Model R3 consists of a strong ductile lower crust that directly underlies the upper crust and is almost as strong as the lithospheric mantle. Hence, the tight spacing between basins in Model R3 may correspond to short-wavelength localisation instabilities in the strong lower crust and lithospheric mantle. In models R4 and R5, the weak lower crust acted as a decoupling layer between the strong upper crust and strong lithospheric mantle. While this decoupling by an intervening weak layer does not appear to significantly influence the spacing between basins, it may have contributed to the formation of more laterally continuous basins of Models R4 and R5, as opposed to the short and segmented basins of Model R3 (cf. Benes and Davy, 1996).

4.2 Strain accommodation and selective basin inversion during shortening

The evolution of axial strain in our models show that pre-existing rift basins exert a strong control on deformation related to far-field shortening. As normal basin-bounding faults formed in the upper crust during extension, they became zones of dilation within an otherwise undisturbed granular layer (Bellahsen and Daniel, 2005; Sassi et al., 1993; Mandl et al., 1977); these became pre-existing zones of weakness that were reactivated in a reverse sense in the early stages of shortening (Figure 7 and Figure 8). The reverse reactivation of weakened normal faults during basin inversion has also been observed in previous analogue (e.g., Marques and Nogueira, 2008) and numerical experiments (e.g., Buiter et al., 2009).

As shortening progressed, basins became narrower (Figure 8). These basins correspond with areas of previously thinned lithosphere (Figure 1), which would have been weaker than the rift shoulders. Continued shortening resulted in inversion of the basins, which we interpret to have been driven by anticlinal folding of the ductile layers, based on observations of uplifted lower crust underneath the inverted basins (following the removal of upper crustal material at the end of the experiments). We also interpret that the anticlinal folding was facilitated by upward buoyancy forces where the upper crust (and therefore the lithosphere) was thinnest, in order to achieve isostatic equilibrium (Figure 12). This upwelling of viscous material underneath thinned crust or lithosphere has been observed in previous analogue models of rifting (Zwaan and Schreurs, 2023). Numerical simulations by Sandiford et al. (2006) suggest that inversion is localised in the centre of the basin due to higher-than-average

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thermal gradients beneath the basin centre. Even though our experiments are isothermal, the basin centres in our models represent the thinnest parts of the crust/lithosphere in nature, where heat flow would be higher than in the adjacent areas.

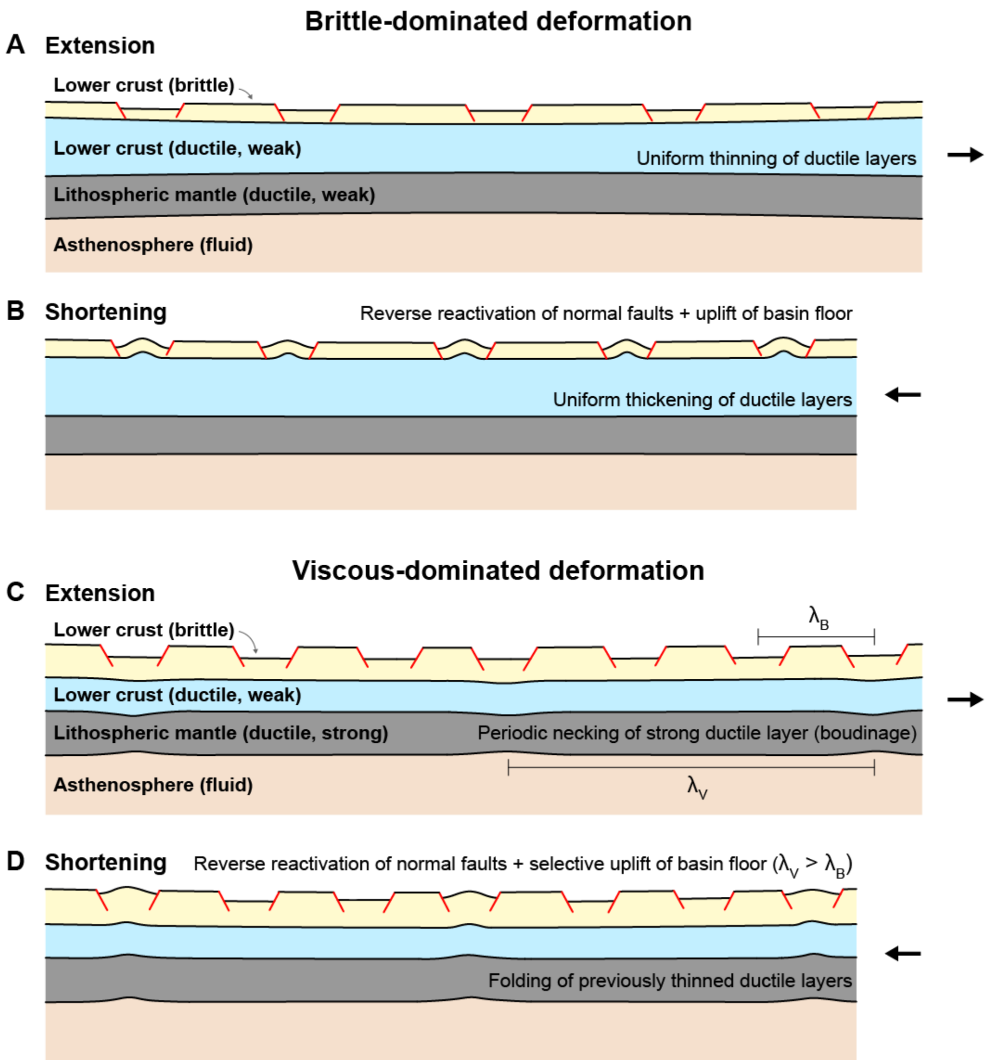

Figure 12: Conceptual illustration of brittle-dominated deformation (e.g., R2) and viscous-dominated deformation (e.g., R3, R4, and R5) during post-extension shortening. Brittle-dominated deformation occurs at slow strain rates, as deformation localisation is controlled by the frictional properties of the brittle upper crust (a and b). Viscous-dominated deformation corresponds with faster strain rates: During extension (c), lithospheric-scale boudinage occurs due to periodic instabilities (with a characteristic wavelength λ_V) in the strong lithospheric mantle. λ_B denotes characteristic wavelength of strain localisation in the upper crust (i.e., the spacing between basins). When the lithosphere is shortened (d), the previously thinned ductile layers undergo folding, and the basins above these areas are inverted.

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In models R3, R4, and R5, some of the pre-existing basins underwent positive inversion (i.e., uplift), while others remained as basins (Figure 9, Figure 10, and Figure 11). In contrast, all of the pre-existing basins in Model R2 were inverted. Here we discuss factors that may have influenced: (1) whether all or only some rift basins are inverted during subsequent shortening (i.e., comparing Model R2 with models R3, R4, and R5), and (2) the periodicity of selective basin inversion (i.e., comparing Model R3 with models R4 and R5).

Model R2 was extended and shortened at a rate that was five times slower than models R3, R4, and R5 (Table 1). Therefore, the ductile lower crust and lithospheric mantle layers of Model R2 were much weaker than the brittle upper crust. This meant that the ductile layers would have thinned and thickened uniformly during extension and subsequent shortening, respectively. With no strain localisation in the ductile layers, rift-related faulting and basin formation in the granular, brittle upper crust would have been controlled by the localisation of brittle deformation (Figure 12a and Figure 11b).

The periodicity of basin inversion is only apparent when the models were extended and shortened at a sufficiently fast rate (i.e., models R3, R4, and R5), which we interpret to have promoted localised viscous deformation as opposed to uniform thinning and thickening (i.e., models R1 and R2). As in models R1 and R2, the inverted basins in models R3, R4, and R5 were underlain by uplifted (presumably folded) ductile lower crust. We assume that these anticlinal folds which were spaced evenly apart (Figure 11) represent periodic instabilities with a characteristic wavelength λ_V (i.e., the distance between two anticlines; Figure 12c). Hence λ_V also corresponds to the distance between two inverted basins. Based on our models, we conclude that for a system of distributed basins where the distance between basins is shorter than λ_V, only some basins will be inverted (Figure 12d). While we acknowledge that the wavelength and amplitude of folds during layer-parallel shortening is controlled by the thickness and rheology of the folded layers (e.g., Schmalholz and Mancktelow, 2016), it is outside the scope of this work to further determine how λ_V is influenced by the model setup (e.g., layer thicknesses, viscosity ratios, bulk strain rates).

The selective inversion of rift basins in our models has not been observed in previous crustal- and lithospheric-scale analogue experiments. There are few other analogue experiments in which extension followed by shortening is applied to a brittle-ductile model during the same experimental run. Examples of such experiments include the work of Gartrell et al. (2005) and Cerca et al. (2010), where extension is followed by shortening in a direction that is oblique to the extension direction (by 10° and 15°, respectively). However, these experiments differed from ours in that extension was localised by initial zones of weakness. As a result, the rift basins were not distributed across the entire model area, and all of the extensional basins localised subsequent shortening and associated inversion structures.

Our experiments show the importance of conducting lithospheric-scale analogue experiments – with a brittle-ductile multi-layer model underlain by a fluid asthenosphere for isostatic support – to investigate rheological controls on basin inversion. Future investigations on selective basin inversion would need to take into account sedimentation, as the density of basin sediments may suppress folding and uplift of the ductile layers during shortening. Imaging of the ductile layers during experimental runs (e.g., using a CT-scanner; Zwaan and Schreurs, 2023) would allow us to better track viscous deformation, which, as we have shown, plays a significant role in promoting basin inversion.

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4.3 Comparisons with the North Australian Craton

4.3.1 1800–1750 Ma wide rifting and 1750–1710 Ma inversion of the Mt Isa terrane

The basins system of the North Australian Craton span more than 0.5 billion years of Earth history during the accretion and dispersal of the Paleo- to Mesoproterozoic Supercontinent Nuna (Betts et al., 2016; Gibson et al., 2018; Johnson, 2021; Kirscher et al., 2020; Zhang et al., 2012). There are numerous interpreted tectonic drivers for the basin evolution of the North Australian Craton (e.g., rifting *sensu stricto*: O'Dea et al., 1997; strike slip tectonics: Southgate et al., 2001). Several tectonic models agree that this series of basins formed in the overriding plate of one or more convergent plate margins (Scott et al., 2000; Giles et al., 2002; Betts and Giles, 2006; Yang et al., 2019). These basins have stratigraphy that can be correlated, even though they are dispersed across the entire craton over a distance on the order of 1000 km (Figure 13). At least four unconformably bounded Superbasins are resolved spanning 1840–1350 Ma. The oldest Superbasin is the Leichhardt Superbasin (ca. 1800–1740 Ma, Jackson et al., 2000).

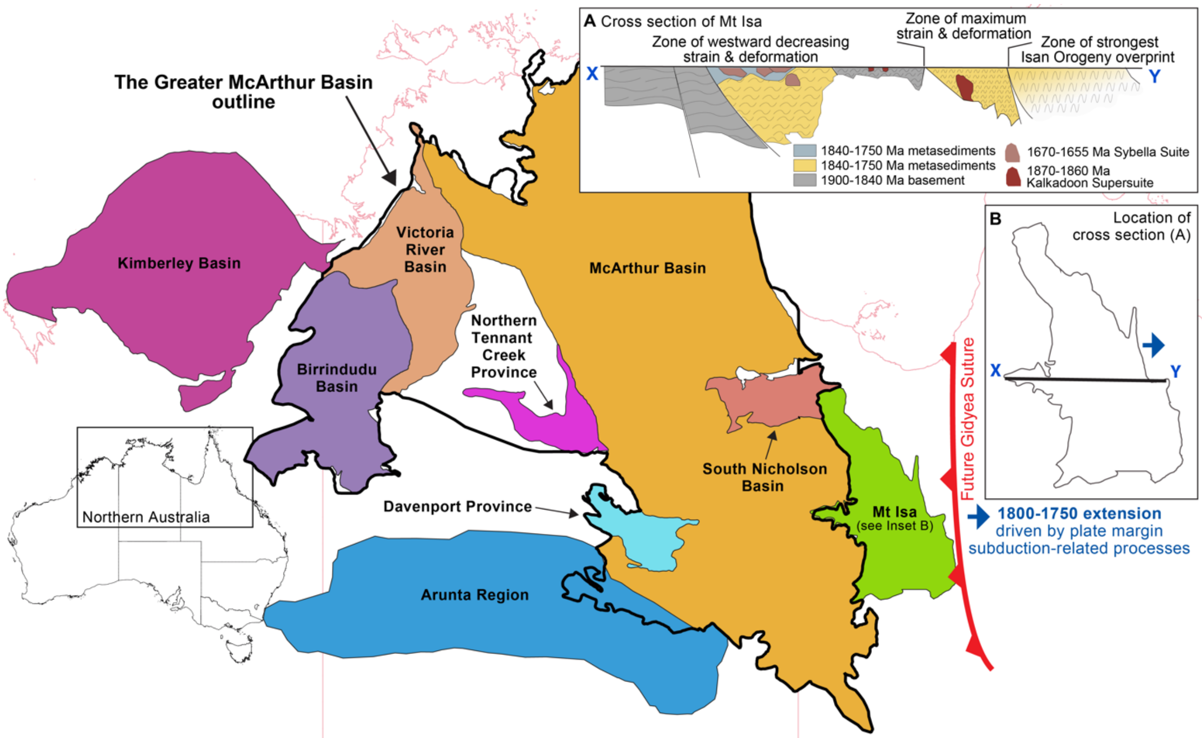

Figure 13: Map of Proterozoic basins in northern Australia (basin shapes obtained from Geoscience Australia portal: https://portal.ga.gov.au). (a) Cross section of with 1800–1750 Ma showing strain and deformation intensity related to compression during the 1750-1710 Ma Wonga Orogeny (Spence et al., 2022) which followed the 1800-1750 Ma extension. (b) Outline of the Mt Isa terrane, showing the location of (A).

The Superbasin sequences are usually separated by transient inversion events (Blaikie et al., 2017; Betts, 1999; Spence et al., 2021). Each new Superbasin phase is associated with the renewed reactivation of faults inherited from the previous, underlying basin (e.g., O'Dea et al., 1997b; Betts et al., 2004, 2006). The intensity of the inversion events varies across the North Australian Craton. Inversion is subtle in the interior of the craton (Bull and Rogers, 1996), whereas along the craton edges (e.g., Mount Isa) it is much stronger and dominates the structural grain of the region.

The ~2000 km-long, roughly north-south trending Mount Isa terrane lies in the eastern part of the North Australian Craton (Figure 13). This polydeformed terrane has been affected by multiple extensional and compressional episodes due to its interaction with Laurentia to the east between ~1800 and 1500 Ma (Betts et al., 2006; Betts et al., 2008; Gibson et al., 2018; Korsch et al., 2012; Olierook et al., 2022). The Leichhardt Superbasin is the oldest basin exposed in the region, cropping out throughout the entire Mount Isa terrane (Gibson and Edwards, 2020; Neumann et al., 2006). A ~20 Myr period of extensional activity was followed by a compressional event known as the Wonga Orogeny (Spence et al., 2022), possibly related to the accretion of the seismically-imaged Numil province to the east (Blaikie et al., 2017). The Wonga Orogen appears to decrease in intensity westwards (Figure 13a). The orogen has mostly been recognised in the eastern zone of the Mount Isa region within the Mary Kathleen Domain. Evidence of this compressional event becomes more scarce to the west, possibly due to overprinting by younger events and lack of exposure (Spence et al., 2022).

The experiments we presented here simulate one-sided extension and shortening, similar to continental extension and shortening resulting from far-field plate margin processes at the eastern margin of the North Australian Craton, i.e., Mount Isa. The wide distribution of rift basins in our experiments is analogous to the distributed basins system of the North Australian craton. During the extensional phase, models R3, R4, and R5 exhibit partitioning of strain into so-called low-strain zones (characterised by the absence of basins) and high-strain zones (populated by clusters of basins) (Figure 5 and Figure 6). The northern half of these models are dominated by low-strain zones, and the southern half by high-strain zones. During the shortening phase of our experiments, more strain is accommodated in the southern half of the models. We interpret that this strain partitioning is partly due to the displacement gradient imposed by the moving U-shaped wall at the southern end of the model (Figure 2). This displacement (and velocity) gradient and the resulting high strain in the southern part of our models is analogous to the stronger effects of plate margin processes (i.e., west-dipping slab roll-back; Betts et al., 2016), and therefore the intensity of deformation, towards the east of the Mount Isa region (Spence et al., 2022) (Figure 13a).

4.3.2 Selective basin inversion: Implications for metamorphism and Pb-Zn mineral systems in Mount Isa

The structures and the metamorphic facies distribution in the Mount Isa region predominantly reflect peak metamorphism during with the Isan Orogeny (e.g., Betts et al., 2006; Foster and Rubenach, 2006; Austin and Blenkinsop, 2008; Blenkinsop et al., 2008). This episode of metamorphism is associated with regional shortening which followed basin-forming rifting and the subsequent thermal sag phase (O'Dea et al., 1997a). The map-view pattern of the metamorphic facies distribution is characterised by north-south trending, amphibolite facies belts separated by zones of mainly greenschist facies rocks (Foster and Rubenach, 2006). This juxtaposition of high- and low-grade metamorphic rocks reflects steep upper crustal thermal

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gradients, the cause of which has been the subject of debate. McLaren et al. (1999) proposed that the source of heat contributing to the high geothermal gradient is the granitic Sybella batholith that was emplaced at shallow crustal levels during the initial basin-forming rift phase (O'Dea et al., 1997b). They further proposed that high-temperature metamorphism was facilitated by the trapping of heat (from the granitic batholith) within the upper crust by the insulating, overlying basin sediments during protracted rift-related subsidence. Their model takes into account that despite the spatial correlation between this granite batholith and high metamorphic-grade rocks, granite emplacement and peak regional metamorphism are separated by ~130 Myr (Connors and Page, 1995).

The results of our experiments suggest that steep thermal gradients in basin inversion settings could be attributed to: (1) strain localisation by the rift basins during extension prior to basin inversion and (2) the selective inversion of basins during subsequent shortening (models R3, R4, and R5). In nature, rift basins correspond to areas of crustal thinning, with which high geothermal gradients are associated; areas of active rifting correspond to high heat flow (.g., Lysak, 1987). During the shortening phase of our experiments, strain is localised by the rift basins, regardless of whether or not these basins were inverted (Figure 8). By the end of shortening, inverted basins correspond to high topography, while the basins that were not inverted remained as topographic lows. In nature, these low-topography areas would correspond to deeper units that are subject to higher-temperature (i.e., amphibolite-facies metamorphism). In contrast, high-topography areas (i.e., inverted basins) would be subjected to lower-temperature metamorphism (i.e., greenschist facies). These experimental observations align with: (1) field observations of the alternating high- and low-grade pattern of metamorphic facies at Mount Isa and (2) the interpretation that this metamorphism is associated with regional shortening. While our isothermal analogue experiments do not directly take into account the effects of and changes in temperature during extension and shortening, they provide some insight into the role of the rheological stratification (and by proxy thermal stratification) of the lithosphere during wide rifting and subsequent inversion. More complex future experiments could be designed to investigate the role of post-rift sedimentation in potentially suppressing basin inversion and provide a useful comparison to existing numerical models (e.g., Buiter et al., 2009).

Early models of Pb-Zn ore formation at Mount Isa suggested that mineralisation occurred during basin formation and was facilitated by fluid transport along active normal faults (Betts et al., 2003; Betts and Lister, 2001). However, more recent interpretations suggest that basin inversion strongly controlled the emplacement of Pb-Zn mineral systems as well as the development of petroleum systems in Mount Isa (Gibson and Edwards, 2020; Gibson et al., 2017). For the ca. 1575 Ma Century Pb-Zn deposit, Gibson and Edwards (2020) proposed that hydrocarbons and then a more metalliferous ore-forming fluid were consecutively trapped following their ejection from deeper stratigraphic levels during the 1620–1500 Ma Isan Orogeny shortening. They further suggested that Pb-Zn exploration strategies in this region should take into account the close temporal and spatial links between mineral and petroleum systems, the latter of which may consist of hydrocarbon traps associated with inversion structures (e.g., Turner and Williams, 2004).

While previous studies have shown how selective fault reactivation contributes to mineralisation (e.g., Sibson, 1995; Nortje et al., 2011), there has been little focus on the selective inversion of entire basins. Understanding the factors contributing to varying amounts of inversion (between basins) within the same basins system could assist in exploring for ore deposits that

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formed during pre-inversion extension. The amount of inversion has implications for orebody preservation, as uplifted areas are subject to erosion. For example, many of the Pb-Zn deposits in the Mesozoic basins of western Europe (i.e., France, Spain) formed during extension and then experienced inversion during the Alpine orogeny (Munoz et al., 2016). The young age of the inversion allowed for the preservation and therefore extensive exploration of these mineral systems. Similarly in northern Australia, the extended post-orogenic evolution of the Mount Isa Inlier is characterised by heterogeneous but regionally consistent slow cooling and exhumation (< ~0.5 mm/yr) driven mostly by diachronous fault movements (Li et al., 2020). If the ore deposits in this region had formed during pre-Isan extension (e.g., Betts and Lister, 2001; Betts et al., 2003), this slow uplift could have contributed to ore preservation. Further investigation into the spatial and temporal relationships between basin inversion and mineralisation, as well as the drivers of variable basin inversion in mineralised regions, could provide useful insights for future exploration programs.

5. Conclusions

The analogue experiments presented here demonstrate that basin inversion is driven by deep processes occurring beneath the brittle upper crust. For a distributed system of basins, comparable to a series of basins in a natural wide rift setting, basin inversion can occur selectively. In this scenario, basin inversion is facilitated by upward movement of the ductile lower crust or mantle. This viscous process occurs at a different scale to the reverse reactivation of upper crustal normal faults, which is a frictional process.

Selective basin inversion in our experiments could be related to the superposition of crustal-scale and lithospheric-scale boudinage that had formed during the previous basin-forming extensional phase. We show the important role of isostasy in analogue modelling of basin inversion, as folding or uplift of the ductile model layers (as interpreted from our experiments) would not have occurred to the same extent if the model lithosphere had not been supported by a fluid asthenosphere. Cross sectional or 3D imaging of the evolution of these structures during analogue experiments, facilitated by non-destructive monitoring methods, could help us better understand the interplay between crustal- and lithospheric-scale structures in facilitating or suppressing basin inversion.

Data availability

QGIS project files containing the interpreted fault traces and basins, as well as time series of orthorectified top-view images, strain and displacement maps, and topographic profiles are provided in Samsu et al. (20xx): [DOI will be added here].

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Author contributions

AS, WG, PB, and ARC contributed to the conceptualisation of the experiments. ARC, WG and PB acquired funding for this project and its publication. AS conducted the investigation with the assistance of FA and EM. AS, TCS, and EM analysed and visualised the data. The original draft was written by AS, WG, and TCS. All authors contributed to reviewing and editing the manuscript.

Competing interests

The contact author has declared that none of the authors has any competing interests.

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