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# The impact of syn- and post-extension prograding sedimentation on the development of salt-related rift basins and their inversion: Clues from analogue modelling

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Keywords: Analogue modelling, inverted salt-related rift basins, prograding sedimentation, Central High Atlas

Highlights (3 to 5 with 85 characters including spaces)

- Tectono-sedimentary interplay in salt-related rift basins is studied in seven models
- Different sedimentary patterns influenced the growth of diapirs at a basin scale
- Reactive–active, roller-like, passive, and extrusive diapirs developed in models

- Inversion caused flank steepening, close-up, and final welding of existing diapirs

## Abstract

Various studies have demonstrated the intrinsic interrelationship between tectonics and sedimentation in salt-related rift basins during extension as well as during their inversion by compression. Here, we present seven brittle–ductile analogue models to show that the longitudinal or transverse progradation of sediment filling an elongate extensional basin has a substantial impact on the growth of diapirs and their lateral geometrical variations. We use five extensional models to reveal how these prograding systems triggered diapir growth variations, from proximal to distal areas, relative to the sedimentary source. In the models, continuous passive diapir walls developed, after a short period of reactive–active diapiric activity, during syn-extensional homogeneous deposition. In contrast, non-rectilinear diapir walls grew during longitudinal prograding sedimentation. Both longitudinal and transverse post-extensional progradation triggered well-developed passive diapirs in the proximal domains, whereas incipient reactive–active diapirs, incipient roller-like diapirs, or poorly developed diapirs were generated in the distal domains, depending on the modelled sedimentary pattern. Two models included final phases of 6% and 10% shortening associated with basin inversion by compression, respectively, to discriminate compressional from purely extensional geometries. With the applied shortening, the outward flanks of existing diapir walls steepened their dips from  $8^{\circ}$ – $17^{\circ}$  to  $30^{\circ}$ – $50^{\circ}$ . Likewise, 6% of shortening narrowed the diapir walls by 32%–72%, with their fully closing (salt welds) with 10% of shortening. We compare our results with the distribution of salt walls and minibasins of the Central

High Atlas diapiric basin in Morocco, which was infilled with a longitudinally prograding mixed siliciclastic and carbonatic depositional sequence during the Early–Middle Jurassic with a minimum thicknesses of 2.5–4.0 km.

## 1. Introduction

Analogue modelling has become an essential tool for studying the mechanisms involved in tectonic processes such as lateral compression, extension, and strike-slip tectonics (e.g., Dooley and Schreurs, 2012; Graveleau et al., 2012; Koyi, 1997; McClay, 1996). Analogue modelling has been applied to the study of salt tectonics to obtain a better understanding of the mechanisms that trigger the onset of diapirs and the evolution of diapiric structures and minibasins (Brun and Fort, 2004; Nalpas and Brun, 1993; Schultz-Ela and Jackson, 1996; Vendeville and Jackson, 1992; Weijermars et al., 1993). The geometries generated by these models have been compared with the large number of high-resolution seismic lines imaging salt-related provinces in various parts of the world, such as along the Atlantic passive margins (Fort et al., 2004; Krézsek et al., 2007), the Gulf of Mexico (Dooley and Hudec, 2017; Ge et al., 1997a; Rowan and Vendeville, 2006), and the North Sea (Dooley et al., 2003, 2005; Ge et al., 1997b). Analogue modelling has allowed the relationship between extension and the development of coeval salt diapirs to be defined (e.g., Dooley et al., 2005; Koyi et al., 1993; Schultz-Ela and Jackson, 1996; Vendeville and Jackson, 1992, amongst others). Post-salt sedimentary layers deposited over an active basement graben filled with salt may stretch laterally by a combination of extension and bending. Continuous regional extension and the drag folding of post-salt strata may trigger reactive–active diapirism

followed by a phase of passive diapirism whose evolution depends on the thickness of the source layer, the thickness of the overburden, and the rate of sedimentation (Jackson and Vendeville, 1994; Vendeville and Jackson, 1992). In cases where the salt layer was deposited over a faulted basement, the extension above and below the salt can be partially coupled depending on the thickness of the salt layer and the velocity of extension (Dooley et al., 2003, 2005; Nalpas and Brun, 1993; Vendeville et al., 1995; Vendeville and Jackson, 1992), and therefore the basement geometry influences the extensional structures developed in the post-salt layers. Dooley et al. (2005) designed analogue models in extension with a complex faulted basement, some of them later inverted, generating extensional grabens in the overburden that influenced the location of the diapirs. In most previous models of salt-related basins, the sedimentation during extension and coeval diapiric evolution was homogeneous, with uniform sedimentation along the entire model device, that it may not always be realistic. Consequently, in the present study we designed diverse depositional patterns within a salt-related rift basin to evaluate the potential impact of such patterns on the underlying salt migration and on the evolution of the salt structures.

Together with the study of extensional regimes, modelling studies have been used to examine the impact of the progradation of a sedimentary system over a ductile layer. It has been shown that the progradation of a sedimentary wedge mobilises a ductile substrate into a variety of geometries, with the proximal part of the prograding system being characterised by the formation of half-grabens and the growth of diapirs, whereas the distal part undergoes the development of fold-and-thrust belts in which well-developed diapirs are also observed (Koyi, 1996; McClay et al., 1998, 2003; Talbot, 1992; Vendeville, 2005). Previously published analogue models have shown that the geometry and distribution of diapirs increase in complexity with: (i) the progradation of

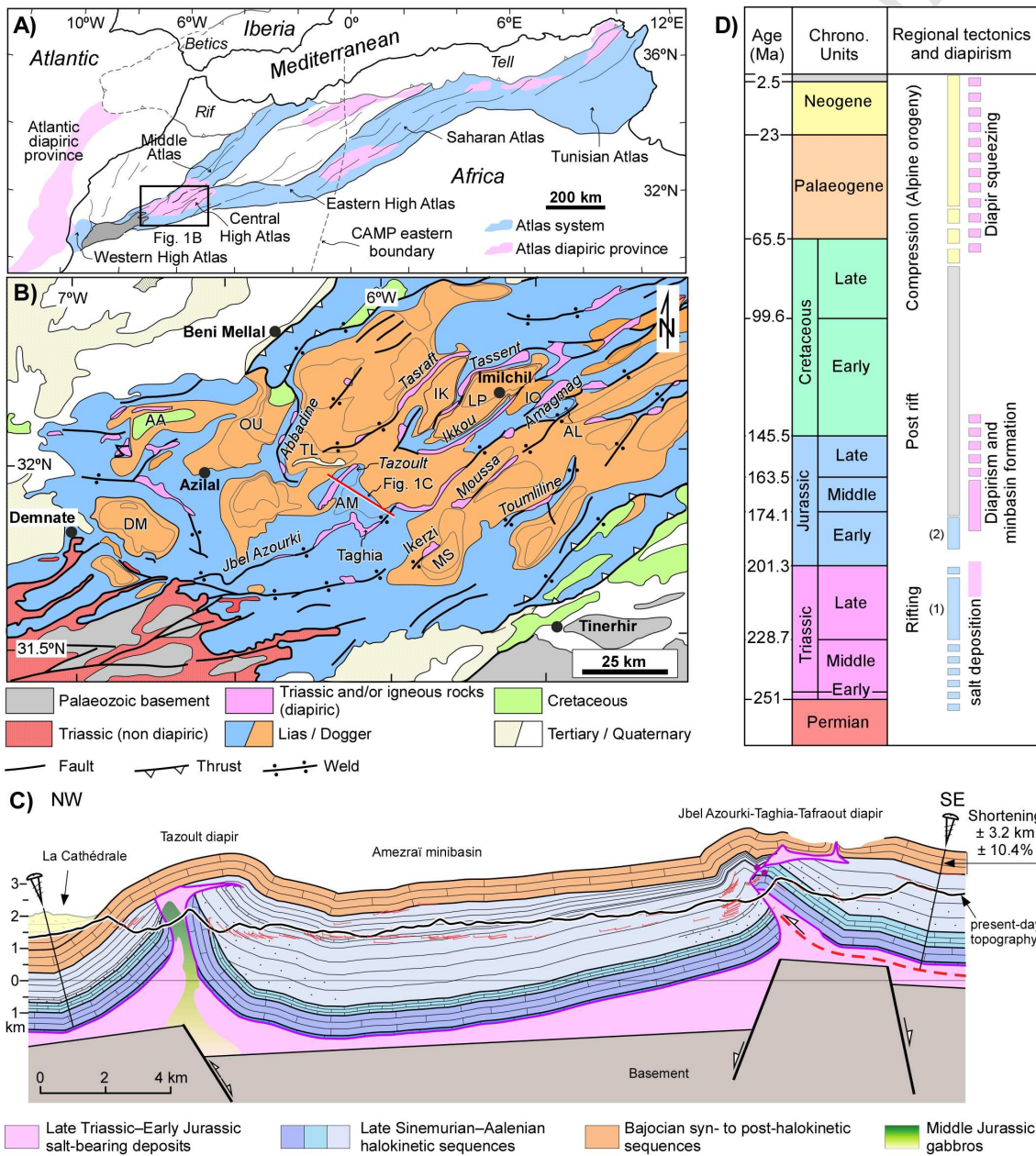
a sedimentary lobe, where the radial shape of the prograding system generates a complex network of polygonal or circular depocentres separated by salt ridges (Gauillier and Vendeville, 2005; Loncke et al., 2010); (ii) the topography of the rigid basement beneath the ductile layer and the consequent variation in the thickness of the mobile silicone representing salt (Dooley et al., 2017; Ge et al., 1997a); and (iii) the confluence of two sedimentary wedges (Guerra and Underhill, 2012).

Our experiments are based on natural examples from a case study in the Central High Atlas (Morocco), a Triassic–Jurassic rift basin inverted during the Alpine Orogeny. In the Atlas system, large basement faults limit thick Permo-Triassic half-grabens filled by siliciclastic deposits, volcanic lavas from the Central Atlantic Magmatic Province and Triassic–Jurassic evaporitic deposits (Courel et al., 2003; Oujidi et al., 2000). Permo-Triassic structures were buried under thick Early and Middle Jurassic successions displaying spectacular halokinetic deformation structures related to ENE–WSW-trending diapiric structures (e.g., Saura et al., 2014; Martín-Martín et al., 2017). Thus, our study aimed to characterise i) the extensional structures below and above the Upper Triassic salt-bearing deposits, ii) the effects of longitudinal versus transverse sedimentation on the growth of salt walls and on the development of minibasins in the Central High Atlas salt-related rift basin, and iii) the impact of the Alpine inversion on diapiric structures of the Central High Atlas. To attain these objectives, we present three sets of analogue models with an initial configuration of a single symmetric graben and designed them in accordance with the three main tectonic events defined in the study region (rift, post-rift, and inversion). Different depositional configurations are considered in each model during and after the rifting phase.

## 2. Geological setting

The Central High Atlas range is a doubly verging fold-and-thrust belt that formed as a result of the inversion of a Triassic–Jurassic rift basin during the Alpine Orogeny (Arboleya et al., 2004; Beauchamp et al., 1999; Frizon de Lamotte et al., 2000; Laville et al., 1977; Mattauer et al., 1977; Piqué et al., 2000; Teixell et al., 2003; Tesón and Teixell, 2008). In the central part of the range, the most common materials cropping out are Lower to Middle Jurassic deposits that form broad synclines or tabular plateaux separated by NE–SW-trending anticlines or thrust faults (Fig. 1). The cores of the anticlines, or ridges, are composed of Triassic evaporite-bearing shales and sandstones, commonly intruded by Middle to Late Jurassic gabbros (Frizon de Lamotte et al., 2008; Hailwood and Mitchell, 1971; Jossen and Couvreur, 1990; Laville and Harmand, 1982). The origin of these ridges has classically been attributed to various processes, including Jurassic compressional folding associated with reverse faults (Studer and Du Dresnay, 1980), Jurassic transpression along NE-trending sinistral strike-slip faults associated with en échelon opening (Laville and Piqué, 1992), the emplacement of Jurassic intrusions (Laville and Harmand, 1982; Schaer and Persoz, 1976), or the uplift of the borders of tilted blocks during extensional faulting (Jenny et al., 1981; Poisson et al., 1998). Only a few studies of local structures have proposed a diapiric origin for these ridges; for example, the Tazoult ridge (Bouchouata, 1994; Bouchouata et al., 1995), the Ikerzi ridge (Ettaki et al., 2007), the Tassent ridge (Michard et al., 2011), and the Toumliline diapir (Ayarza et al., 2005; Teixell et al., 2003). However, recent studies have interpreted the Central High Atlas as a complex diapiric basin in which more than ten elongated minibasins bounded by salt walls have been identified (Saura et al., 2014). Observed halokinetic stratal relationships in sediments flanking the ridges, as well as

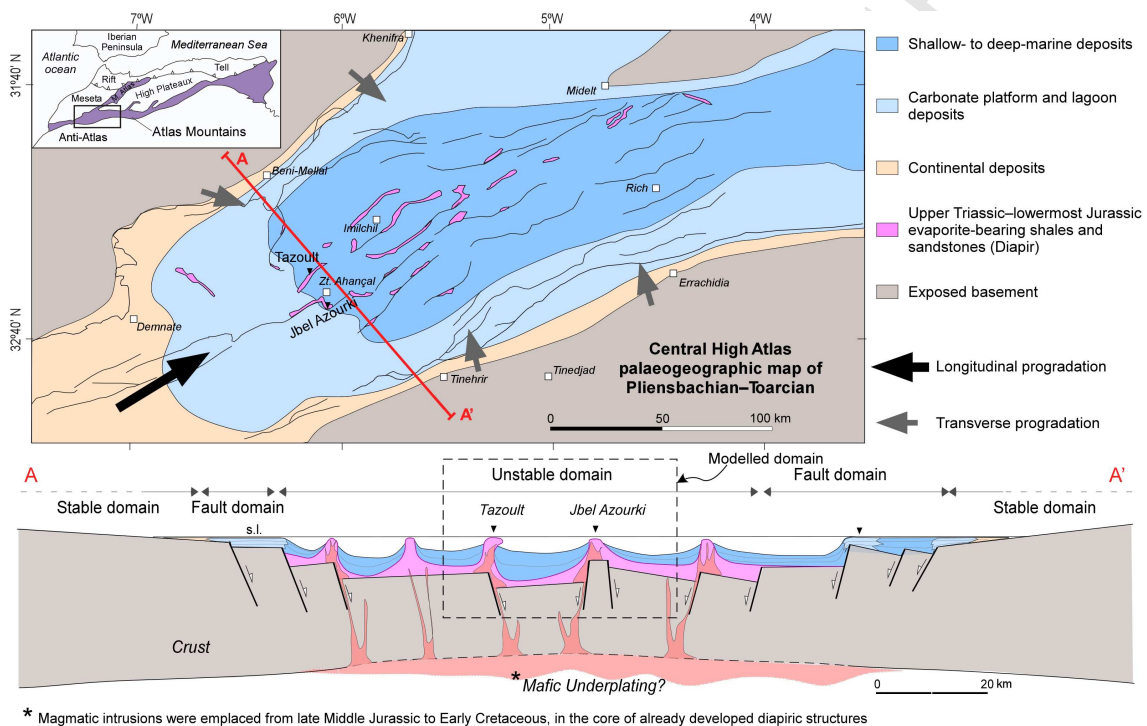
facies distribution analysis conducted in various minibasins throughout the Central High Atlas, show that diapir activity occurred from the Pliensbachian (and possibly earlier) to the Callovian (and possibly later) (Bouchouata, 1994; Bouchouata et al., 1995; Ettaki et al., 2007; Grélaud et al., 2014; Ibouh et al., 2011; Joussiaume, 2016; Malaval, 2016; Martín-Martín et al., 2017; Teixell et al., 2017; Vergés et al., 2017).



**Fig. 1** (A) Synthetic map of NW Africa showing the locations of diapiric areas (pink). Central Atlantic Magmatic Province labelled as CAMP (B) Geological map of the Central High Atlas showing the locations of diapiric ridges and associated minibasins labelled as follows: Aït Attab (AA); Almghou (AL); Amezraï (AM); Demnate (DM); Ikassene (IK); Ikkou (IO); Lake Plateau (LP); Ouaouizaght (OU); and Tiloughite (TL). (Modified from Martín-Martín et al., 2016 and Saura et al., 2014). (C) Balanced cross-section across the Amezraï minibasin from Martín-Martín et al., 2016. (D) Chart showing the age of geodynamic processes occurring in the Central High Atlas (modified from Vergés et al., 2017). Two rifting phases have been differentiated: (1) late Permian to Late Triassic and (2) late Sinemurian to late Pliensbachian.

After the Variscan orogeny, the Central High Atlas underwent two consecutive rifting phases (Fig. 1D). The first rifting phase spanned from the late Permian to the Late Triassic, but with greater activity during the Middle–Late Triassic (Frizon de Lamotte et al., 2009; Laville et al., 2004). This extensional event caused the development of half-grabens bounded by high-angle normal faults with a dominant dip-slip movement filled. These grabens were filled with Upper Triassic continental red beds up to 2 km thick and were associated with the local development of thick evaporitic successions (Baudon et al., 2009; Courel et al., 2003; Domènech et al., 2015; Oujidi et al., 2000). After a short period of tectonic quiescence from the Hettangian to the early Sinemurian (from 201.3 to 195 Ma), renewed extension occurred from the late Sinemurian to the Pliensbachian (from 195 to 182.7 Ma); this was followed by a long post-rift period extending through the onset of the basin inversion during the Alpine Orogeny, when shortening occurred at rates of about 0.3 mm/yr on average (Ellouz et al., 2003; Frizon de Lamotte et al., 2009; Moragas et al., 2016; Tesón and Teixell, 2008). Palaeoreconstructions for the Jurassic (Fig. 2) show that the Central High Atlas basin was open towards the east, with the main direction of prograding sedimentation being oriented along the basin from WSW to ENE with the leading edge perpendicular to the basin margins (Milhi et al., 2002; Pierre et al., 2010; Piqué et al., 2000; Souhel et al., 2000). Our physical models were

designed to simulate the processes occurring in the central part of the Atlas diapiric basin (Fig. 2), where the Tazoult and Jbel Azourki-Tafraout structures bounding the Amezraï basin are located (Fig. 1). Although the major depositional direction in the Central High Atlas during the Early and Middle Jurassic was longitudinal, we also present a model with transverse deposition to analyse the influence of both of these sedimentation patterns on diapir evolution and basin configuration.



**Fig. 2** Palaeogeographic map of the Central High Atlas basin for the late Pliensbachian to Toarcian period (modified from Moragas et al., 2016). Arrows indicate the direction of sediment input and progradation of the sedimentary systems in the basin. Cross-section A–A' shows an inferred configuration of the Atlas rift system tectono-sedimentary domains during the Early Jurassic and the interplay between normal faulting, diapir growth, and igneous intrusions above potential mafic underplating at the base of the crust.

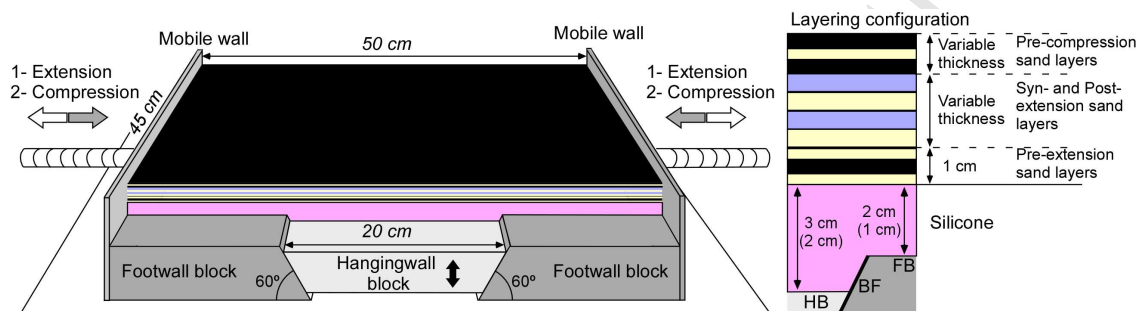
### 3. Modelling procedure and experimental design

We used the classical techniques applied in brittle–ductile analogue modelling experiments, with the use of sand and silicone (Faugère and Brun, 1984), developed by the Experimental Tectonics Laboratory of Géosciences Rennes (Université de Rennes 1, France). The scaling of models to nature was chosen from published laws (Davy and Cobbold, 1991; Hubbert, 1937; Ramberg, 1981; Weijermars et al., 1993). A length of 1 cm in the models represents 0.8 km in nature ( $1.2 \times 10^{-5}$ ), and a time of 1 hour represents 3 Myr ( $1.14 \times 10^{-10}$ ) (Fig. 3).

To simulate the brittle behaviour of sedimentary rocks with Mohr–Coulomb properties, dry Fontainebleau quartz sand (produced by SIBELCO, France) was used in the experiments, with a mean grain size of approximately 250  $\mu\text{m}$ , an internal friction angle of 30°–35° (Krantz, 1991), and a density ( $\rho$ ) of approximately 1500  $\text{kg/m}^3$  (Fig. 3). Sand layers of different colours but the same mechanical behaviour were used as passive markers to emphasise the resultant structures. Transparent silicone was used to simulate salt, as it has been shown to be a suitable analogue material for this rock type (Weijerman et al., 1993). The mean viscosity of the silicone was  $\sim 3 \times 10^4$  Pa·s at 20 °C, and its mean density was  $\sim 1000$   $\text{kg/m}^3$  (silicone polymer SGM 36, manufactured by Dow Corning, USA).

The modelling device used in this study was designed to simulate the Central High Atlas, where salt deposits were accumulated in the Triassic rift basin (Figs 1 and 2). The initial model set-up, comparable with but simpler than those presented in Dooley et al. (2005), simulated the inherited rift geometry from the Triassic rifting, with three rigid blocks representing the basement geometry and a central graben infilled with a thicker silicone layer and bounded by two faults dipping 60° (Fig. 3). The footwall blocks were fixed to mobile walls, whereas the central hangingwall block was not fixed, allowing it

to subside during extension. A constant velocity of 0.5 cm/h was applied to each mobile wall by a screw jack during both lateral extension and compression, resulting in a total deformation velocity of 1 cm/h (equivalent to 0.2 mm/yr in nature). The ductile silicone layer representing the syn-rift salt was covered by a thin pre-extension brittle layer of 1 cm of sand (Fig. 3), which represented a sedimentary succession deposited during a short period of tectonic quiescence prior to renewed extension (Fig. 1D).



Scaling parameters

	Length (L)	Gravity (g)	Dry quartz sand density ( $\rho$ )	Décollement layer density ( $\rho$ )	Décollement layer viscosity ( $\eta$ )	Time
Nature	800 m	9.81 m/s <sup>2</sup>	2.6 g/cm <sup>3</sup>	salt 2.2 g/cm <sup>3</sup>	salt $1 \times 10^{19}$ Pa · s	$3 \times 10^6$ yr
Model	1 cm	9.81 m/s <sup>2</sup>	1.5 g/cm <sup>3</sup>	silicone 1.0 g/cm <sup>3</sup>	silicone $3 \times 10^4$ Pa · s	1 h
Model/nature ratio	$1.2 \times 10^{-5}$	1	0.57	0.45	$3 \times 10^{-15}$	$1.14 \times 10^{-10}$

**Fig. 3** The experimental modelling apparatus, consisting of a single central graben bounded by two fault planes dipping inwards at 60°. The column on the right-hand side of the figure shows the configuration of the different layers in the model (HB: Hangingwall block, FB: Footwall block, BF: Basement fault). The silicone thickness values in parentheses correspond to the thicknesses used in Model A1. The table details the scaling parameters.

The evolution of salt-influenced rift basins was analysed by running three sets of models (Fig. 4) representing the three main post-Triassic tectonic phases of the Central High Atlas (Fig. 1D): Early Jurassic rifting (extension), post-rift (tectonic quiescence), and Alpine inversion (compression). The influence of different amounts of

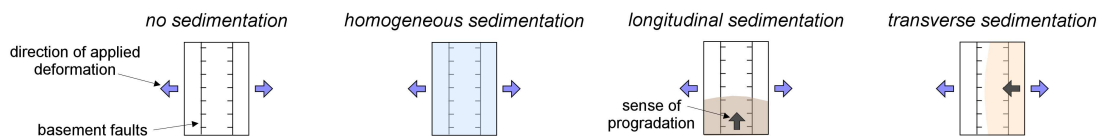
sedimentation on diapir geometry was determined by varying the sedimentation thicknesses and patterns during the extension and post-extension phases (Fig. 4). Model A1 represented the evolution of the system during extension (2 h, equivalent to  $6 \times 10^6$  yr, with a total extension of 2 cm, equivalent to 1.6 km) without the influence of sedimentation. Model set B (four models) represented the evolution of the system during extension, with sedimentation rates being equal to the accommodation rate (that combines the subsidence of the central block plus silicone expulsion), followed by a post-extension phase (3 h, equivalent to  $9 \times 10^6$  yr) with a sedimentation rate equal to or lower than the rate of diapir growth. Model set C (two models) represented homogeneous sedimentation during the extension and post-extension phases and included a final compression phase with shortening (perpendicular to the rifting tectonic trends) of 3 cm (6%) and 5 cm (10 %), respectively (Fig. 4). The aim of Models C1 and C2 (Fig. 4) was to simulate a compressional event after diapirism in order to analyse the resulting geometries and basin deformation. The extension and compression phases of Models C1 and C2 were run using the same deformation velocity (1 cm/h), the same time lengths for the extension and post-extension phases (2 and 3 h, respectively), and the same sedimentation pattern (i.e., homogeneous during the extension and post-extension phases). This was designed to make to ensure that any differences between the resulting geometries were caused by the compression phase (Fig. 4).

Photographs of the surface of each model were taken every 6 minutes, corresponding to 0.1 cm of deformation during the extension and compression phases. At the end of each experiment, a thick layer of post-kinematic white sand was sprinkled on top of the model and then humidified to induce sand cohesion, allowing vertical sections to be sliced through the model. These cross-sections were cut parallel to the extension/shortening direction (perpendicular to the basement faults) across the central

part of each model and labelled with numbers in order to localise them within the model.

	Model A1	Model B1	Model B2	Model B3	Model B4	Model C1 3 cm short.	Model C2 5 cm short.	
Time ↑	<b>compression</b> (1 cm / h)					no sed.	no sed.	
	<b>pre-compression</b>					homogen. sed. 1.5 cm thick	homogen. sed. 3.0 cm thick	
	<b>post-extension</b> (3 h)		homogen. sed. 1.5 cm thick	longitudinal sed. variable thickness	transverse sed. variable thickness	longitudinal sed. variable thickness	homogen. sed. 2.0 cm thick	homogen. sed. 2.3 cm thick
	<b>extension</b> (2 cm / 2 h)	no sed.	homogen. sed. 2.0 cm thick	homogen. sed. 2.2 cm thick	homogen. sed. 2.4 cm thick	longitudinal sed. variable thickness	homogen. sed. 3.3 cm thick	homogen. sed. 3.4 cm thick
	<b>pre-extension</b>	homogen. sed. 1.0 cm thick	homogen. sed. 1.0 cm thick	homogen. sed. 1.0 cm thick	homogen. sed. 1.0 cm thick	homogen. sed. 1.0 cm thick	homogen. sed. 1.0 cm thick	homogen. sed. 1.0 cm thick

#### Sedimentation patterns



**Fig. 4** Diagram showing the modelled tectonic phases through time (pre-extension, extension, post-extension, pre-compression, and compression), the applied sedimentation pattern (none, homogeneous, longitudinal, or transverse progradation), and the amount of shortening applied in the final phase of each model (for C1 and C2). The values (in cm) below the boxes indicate the mean thickness of the sedimentary layers for each phase.

#### 4. Experimental results: the interactions between extension, diapirism, sedimentation, and compression

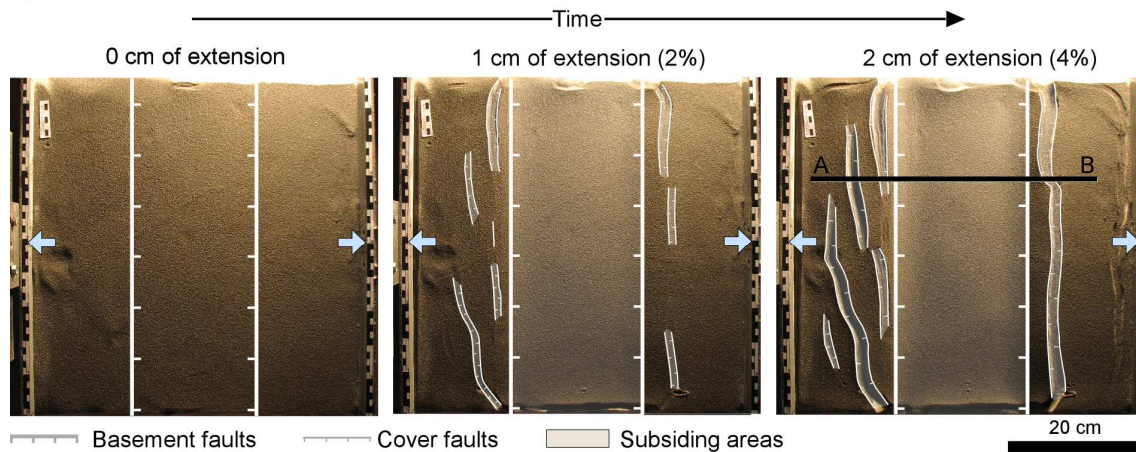
##### 4.1. Model A1: extension phase without sedimentation

Model A1 represented the evolution of a basin during extension (a total extension of 2 cm or 4%) without sedimentation (Figs. 4 and 5). The hangingwall of the graben subsided homogeneously during extension at a rate of 1 cm/h (0.2 mm/yr in nature). After a total extension of 1 cm, newly formed small grabens developed on the pre-extension layer in the footwall of the basement faults and subparallel to their strike. On both footwalls, the grabens were located either close to the tip line of the basement faults (border grabens) or in the footwall interior (lateral grabens) (Fig. 5). These grabens were still extending at the end of the extensional phase (2 cm of extension).

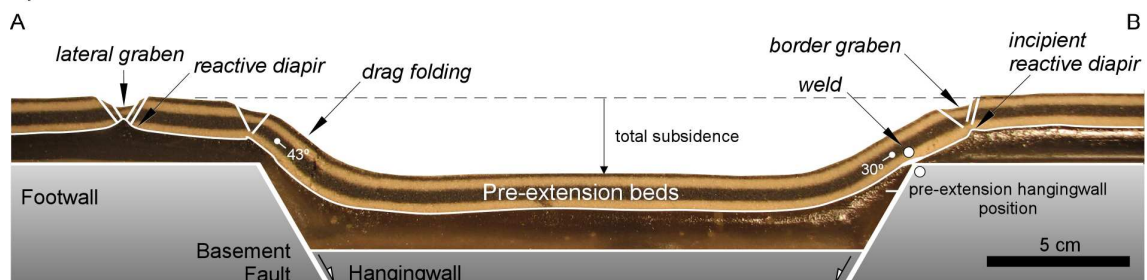
The cross-sections sliced after applying extension showed roughly symmetric grabens affecting the pre-extension sedimentary unit above the silicone layer (Fig. 5). Along the edges of the basement graben, small border grabens showed rotation towards the main graben basin because of the drag folding of pre-extension beds above the basement faults. Both border and lateral grabens exhibited nucleation of small triangular reactive diapirs in their basal parts, similar to other previous analogue models (e.g., Dooley et al., 2005; Hudec and Jackson, 2007; Nalpas and Brun, 1993; Rowan et al., 1999; Vendeville and Jackson, 1992). Along the tip lines of the basement normal faults, a necking effect produced either a thinning of the silicone layer (only 0.47 cm thick) or its final welding (Fig. 5). The thickness of the silicone in the initial phase of this model (1 cm thick in the footwall and 2 cm thick in the hangingwall block) was not thick enough to avoid the welding caused by the subsidence of the hangingwall block and the related drag folding during extension. This has also been observed in previously published analogue models, such as those of Nalpas and Brun (1993), Jackson and Vendeville (1994), Vendeville et al. (1995) (their fig. 5), Dooley et al. (2003) (their figs 8 and 10), Dooley et al. (2005), Loncke et al. (2010) (their fig. 10), and Burliga et al. (2012).

The total subsidence measured in the central part of the hangingwall was 2.5 cm. This subsidence resulted from a combination of 80% of tectonic subsidence, associated with 2 cm of vertical displacement of the hangingwall during extension (Fig. 5), and 20% of silicone migration.

A) Overhead views Model A1



B) Section Model A1



**Fig. 5** (A) Overhead views of Model A1 (total extension = 2 cm) showing the distribution of several newly formed grabens in the pre-extension beds. (B) Model A1 section showing the geometry of grabens in the pre-extension beds bounded by two opposing normal faults rooted in incipient reactive diapirs. Post-kinematic sediments have been removed in the section view.

#### 4.2. B-series models: extension and post-extension phases with variable sedimentation

Four models (B1, B2, B3, and B4 in Fig. 4) were developed using the same extension rate as that of Model A1 (1 cm/h for 2 h, with a total extension of 2 cm or 4%) and a post-extension phase of 3 h. The silicone layer in these models was thicker than in Model A1 (2 cm in the footwall and 3 cm in the hangingwall) to avoid welding on top of the footwall blocks, which would have hampered the formation and growth of diapiric structures. In models B1, B2, and B3, syn-extension homogeneous sediments were deposited in the main area of subsidence corresponding to the graben between the two conjugate basement faults. This sedimentation occurred at a rate equivalent to the accommodation rate (subsidence of central block plus silicone expulsion). During the post-extension phase, the models were subjected to different patterns of sedimentation (Fig. 4): homogeneous sedimentation for Model B1 (Fig. 6), longitudinal prograding sedimentation (parallel to the graben axis) in Model B2 (Fig. 7), and transverse prograding sedimentation (orthogonal to the basin axis) in Model B3 (Fig. 8). In contrast, Model B4 was run with longitudinal prograding sedimentation in both the extension and post-extension phases (Fig. 9).

Model B1 (Fig. 6), characterised by homogeneous sedimentation during both the extension and post-extension phases (Fig. 4), showed the development of elongated newly formed grabens in the footwall blocks, close to the tip lines of the basement faults, during the extension phase. Taking the results from Model A1 as a reference, these elongated grabens marked the locations of reactive diapirs beneath the pre-extension sand layers. This model, as well as other B-series models, showed how diapiric structures increased their level of maturation from reactive and active diapirism

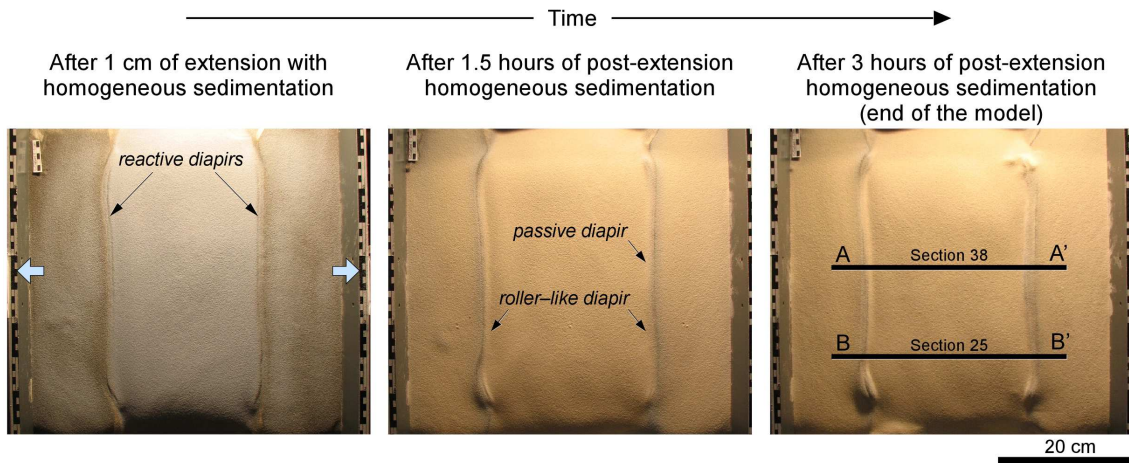
to passive diapiric phases. Thereby, during the post-extension phase of Model B1, the sedimentation in the central graben and the associated silicone migration enabled long and continuous reactive–active diapirs to transition to more mature diapiric structures with different geometries, as shown in cross-sections at the end of the post-extensional phase (Fig. 6B and C). The transition from reactive to passive diapirism through an active phase, which occurred when the pre-extension sand layers had thinned sufficiently to reach the threshold for forceful piercing (Vendeville and Jackson 1992), was recorded by the arching of both the pre-extension beds thinned during extension and the footwall domains (Fig. 6).

Diapirs on the left-hand side of Model B1 showed geometries similar to those of salt rollers (roller-like diapirs in Fig. 6). Salt rollers are immature structures developed in extensional settings and are defined as ridge-like diapirs, being asymmetrical in cross-section with a long gently dipping flank and a short steep scarp flank delineated by a normal fault (Bally, 1981; Brun and Mauduit, 2009; Jackson and Talbot, 1986; Quirk and Pilcher, 2012). The geometry of the diapir on the right-hand side of the model changed laterally from a roller-like diapir in section 25 (B–B' in Fig. 6) to a well-developed passive diapir with steep walls, a smooth basinward vergence, and incipient extrusion in section 38 (A–A' in Fig. 6). The silicone layer at the end of the experiment exceeded 2.2 cm in thickness in both footwall blocks, but only exceeded 1.7 cm along the central part of the hangingwall. The necking effect at the tip of the basement faults reduced the thickness of the silicone layer by up to 0.20–0.65 cm. The area balances of the silicone layer calculated in both sections 38 and 25 recorded similar reductions in silicone thickness (80% remaining) compared with the original silicone area. This indicates that silicone was expelled in the extension direction with little or no migration along strike. The total subsidence measured in Model B1, 3.7 cm, resulted from a

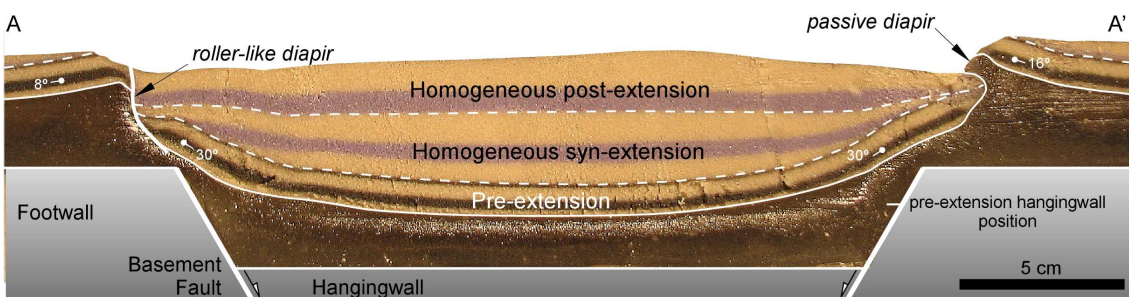
combination of tectonic subsidence and silicone expulsion (50% and 50%, respectively).

In Model B1, the migration of silicone towards the diapirs of the footwall domains tilted and folded the pre-extension sand layers, resulting in flaps flanking the diapiric structures (Dooley et al., 2005; Rowan et al., 2003; Saura et al., 2016; Schultz-Ela, 2003; Schultz-Ela et al., 1993) (Fig. 6). The outward flaps, in the footwall of both basement faults, had low dips that ranged from  $8^{\circ}$  to  $16^{\circ}$ . In contrast, in the hangingwall, the inward flaps were steeper, with a mean inclination between  $24^{\circ}$  and  $30^{\circ}$  close to the basement faults. In addition, the left-hand-side inward flap showed a 0.5-cm long uppermost segment dipping up to  $38^{\circ}$  in both sections 25 and 38 (Fig. 6C and Fig. 6B, respectively).

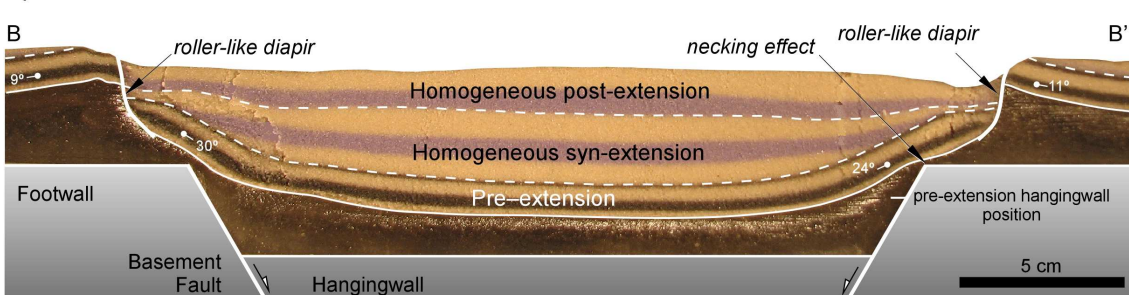
## A) Overhead views model B1



## B) Section 38 - Distal section



## C) Section 25 - Proximal section



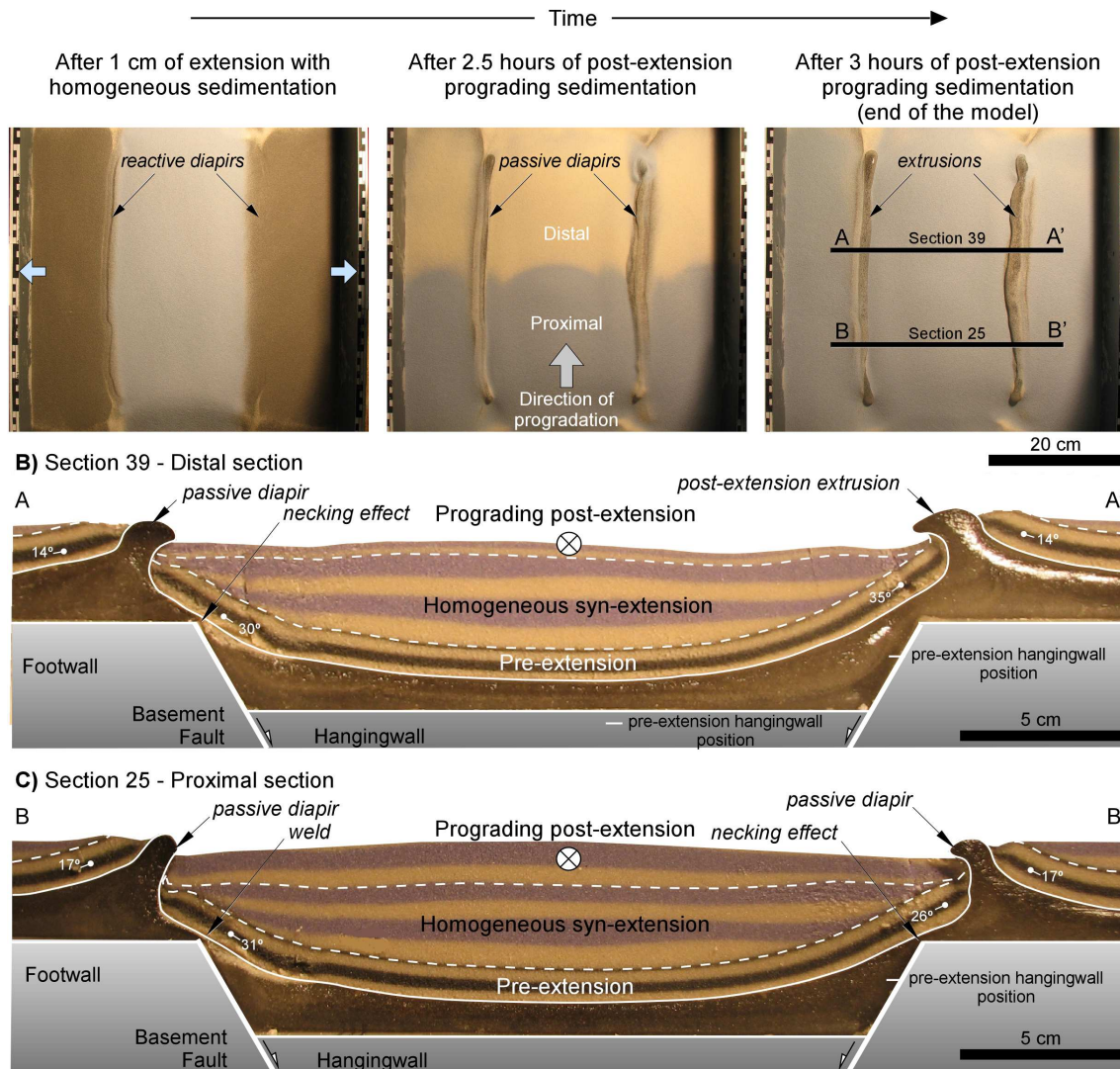
**Fig. 6** Overhead views (A) and sections (B and C) showing the geometries resulting from Model B1. This model was characterised by an extensional phase (total extension = 2 cm) and a post-extensional phase (3 h) with homogeneous sedimentation during both phases. The thick white line marks the diapir wall acting as a normal fault. Post-kinematic sediments have been removed in the section view.

Model B2 (Fig. 7) was run with syn-extension homogeneous sedimentation followed by post-extension longitudinal sedimentary progradation (Fig. 4). The model displayed

elongated newly formed grabens in the footwalls close to the main basement faults, with each of them being associated with reactive diapirs, as seen in Model B1 (Figs 6 and 7). The reactive diapirs developed during the extension phase, then evolved to become active and, finally, passive diapirs along the entire modelling device (Fig. 7). The proximal diapirs in Model B2 leaned inwards (Fig. 7C), with very steep walls and incipient silicone extrusions as the silicone migrated to the diapir. Towards the distal part of the model (Fig. 7B), the silicone extrusions and diapirs were wider and more mature, with proximal diapir walls being less than 0.9 cm wide, and distal diapir walls being more than 1.2 cm wide (Fig. 7). These differences in diapir width and silicone extrusion were partially due to the local thickness of the sedimentary pile and the silicone migration related to differential loading. The mean value of the syn-extension sedimentary succession along strike was around 2.6 cm, but the progradational post-extension sedimentary succession changed in thickness, being 3.5 times thinner in the distal domain than in the proximal domain (thicknesses of 0.45 cm and 1.5 cm, respectively).

At the end of the Model B2, the percentage of silicone preserved in the distal and proximal domains (67% and 60%, respectively) indicated a migration of silicone along the graben axis due to differential loading related to the longitudinal progradation of the post-extension sedimentary wedge. In this model, the welding of the pre-extension deposits against the tip line of the basement normal faults was strong in the proximal areas (0.2-cm-thick silicone compared with 0.6-cm-thick in the distal domain; Fig. 7), even reaching the complete welding stage. In Model B2, the outward flaps displayed dips ranging between  $14^{\circ}$  and  $17^{\circ}$  whereas the inward flaps dipped between  $26^{\circ}$  and  $35^{\circ}$  (Fig. 7), which were similar dip values to those observed in Model B1.

## A) Overhead views Model B2



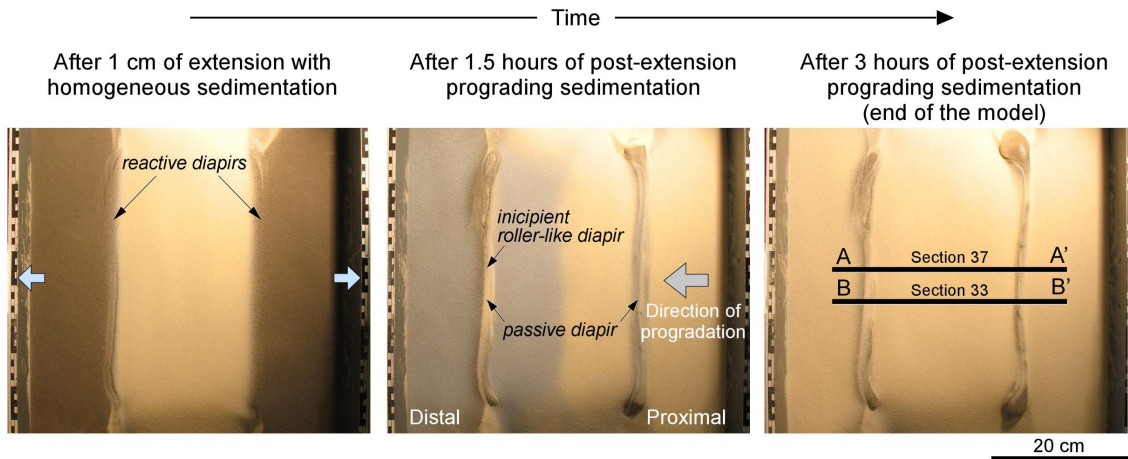
**Fig. 7** Overhead views (A) and sections (B and C) showing the geometries resulting from Model B2. This model included an extensional phase (total extension = 2 cm) with homogeneous sedimentation and a post-extensional phase (3 h) with longitudinal prograding sedimentation. The diapir geometries varied laterally from the proximal (section 25) to the distal domains (section 39). Post-kinematic sediments have been removed in the section view.

Model B3 was run with syn-extension homogeneous sedimentation followed by post-extension transverse sedimentary progradation with the proximal domain being on the right-hand side of the model device (Figs 4 and 8). The extension phase reproduced

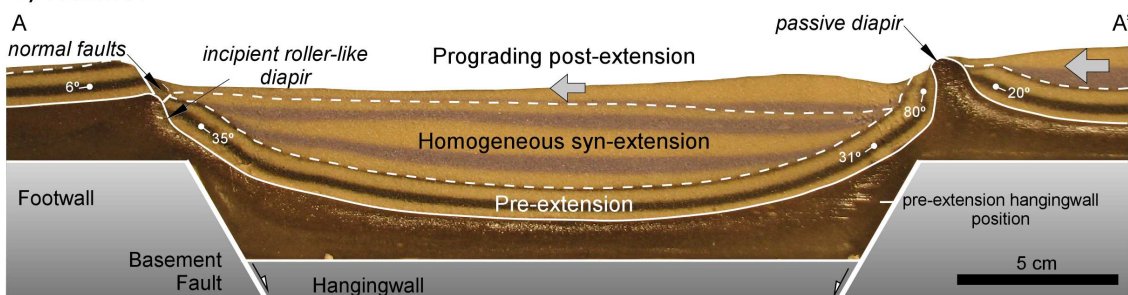
extensional geometries comparable with those of models B1 and B2 (Fig. 8A). Across strike, Model B3 developed an asymmetric geometry during the post-extension phase, in contrast to models B1 and B2. At the end of the experiment, passive diapir walls in the proximal domain, where post-extension deposits were three times thicker than in the distal domain, were mature and characterised by very steep flanks in both sections shown in Figure 8B and C. In contrast, the diapiric structures developed in the distal domain were incipient roller-like diapirs, with a predominant right-dipping normal fault, laterally evolving to silicone walls that were less developed than in the proximal domain.

The final geometry of the silicone layer in Model B3 was asymmetric across the model, with a thinner silicone layer in the proximal domain (1.22 cm thick compared with 1.76 cm in the distal domain) on account of the transverse progradational deposits of the post-extension units and the resultant differential loading. In this model, the necking effects on top of the tip lines of the basement faults were limited (0.6-cm-thick silicone layer). The steepest pre-extension layering of the model corresponded to the inward flaps of the proximal domain with dips ranging from  $80^\circ$  to overturned (Fig. 8), whereas the inward flaps of the distal silicone walls ranged from  $35^\circ$  to  $38^\circ$ . The overturning, which reached 1 cm in length, was clearly a diapiric contribution as the only tectonic mechanism operating in Model B3 was extension.

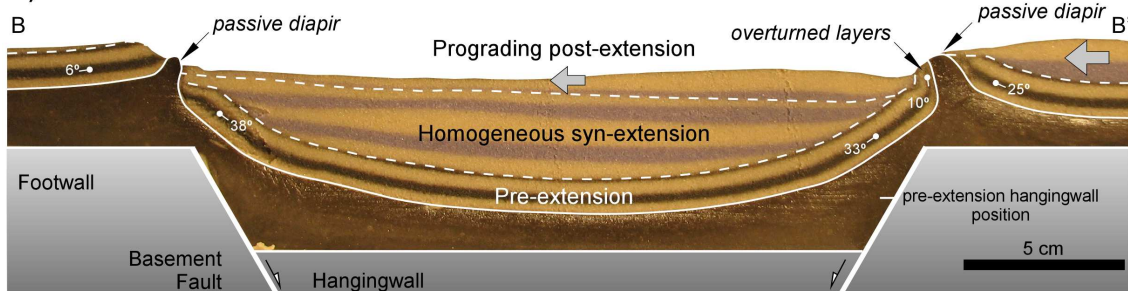
## A) Overhead views Model B3



## B) Section 37



## C) Section 33



**Fig. 8** Overhead views (A) and sections (B and C) showing the geometries resulting from Model B3. This model included an extensional phase (total extension = 2 cm) with homogeneous sedimentation and a post-extensional phase (3 h) with transverse prograding sedimentation. The diapir geometries of the proximal domains (right-hand sides of the model) differed from those of the distal domains (left-hand sides). Post-kinematic sediments have been removed in the section view.

Model B4 (Fig. 9) was run with both syn-extension and post-extension longitudinal sedimentary progradation (Fig. 4) and showed a final configuration that differed

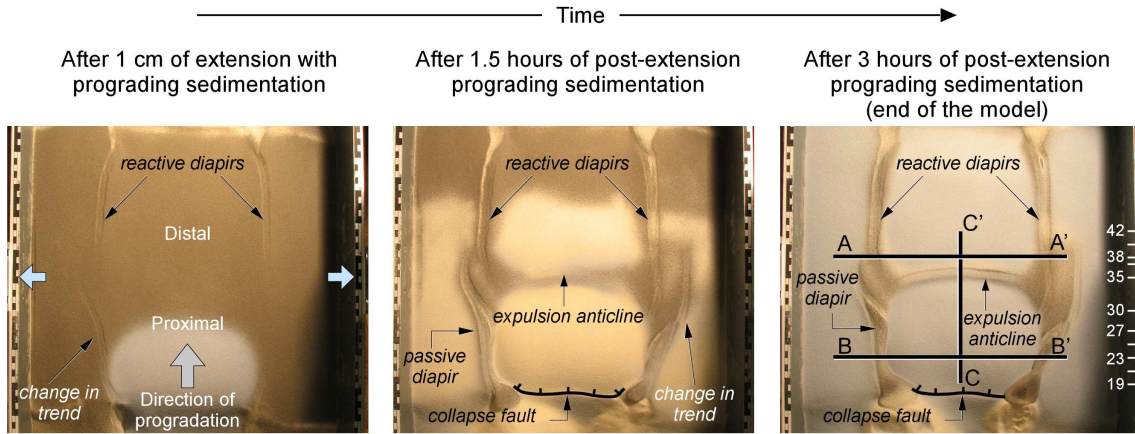
substantially from those of models B1, B2, and B3. During the extension phase, differential loading associated with longitudinal progradation caused the expulsion of silicone towards the distal domain and triggered the development of three distinctive features in Model B4 not observed in the other models (Fig. 9): i) the disruption of the two elongated grabens that characterised other B-series models, ii) the presence of a collapse fault developed in the proximal domain, and iii) the growth of an expulsion anticline (Fig. 9A).

At the end of the Model B4, after the post-extension phase characterised by the continuation of longitudinal progradation, the geometries of diapirs displayed marked differences when comparing the proximal and distal domains. The proximal domain of the model was characterised by <1-cm-wide well-developed extrusive diapirs oriented subvertically or slightly vergent towards the graben (Fig. 9C). The geometry of the silicone walls after the complete welding against the basement differed slightly from those of the other B-series models, with smooth folding of the inward flanks owing to local subsidence of the upper parts of the inward flanks within the silicone wall pedestal (Fig. 9C). In contrast, the distal region of the main basement graben was characterised by a much less evolved diapiric system, with stagnant reactive diapirs being developed during the extensional phase. The axial inflow of silicone to this area limited the subsidence of the hangingwall, preventing the required thinning of the pre-extension layers for the transition from the reactive to the active phase (Fig. 9B).

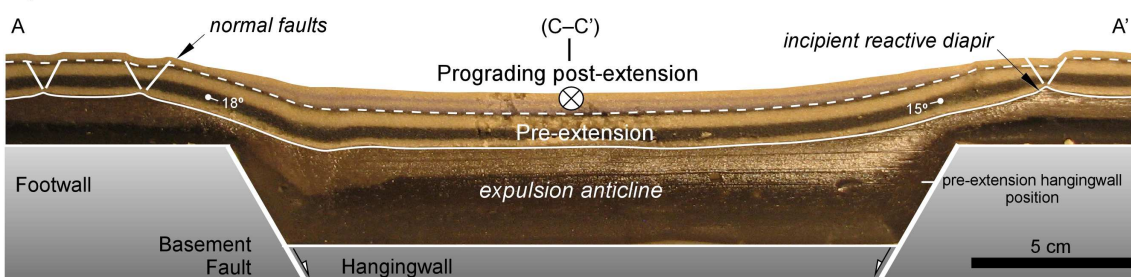
A comparison of the area balance of the silicone layer before and after the deposition of the prograding sequence showed a greater reduction in the proximal domain (about 47% remaining) than in the distal domain (nearly 70% remaining). These differences recorded the axial flow of the silicone from the proximal to the distal domain and the growth of an expulsion anticline from sections 35 to 38 (Fig. 9D). This expulsion

anticline, which was bounded by the along-strike silicone walls, produced a compartmentalization of the silicone domain separating growing minibasins as well as controlling their size and the direction of their major axes. It is possible to infer from these results that a continuous progradation along the axes of the minibasins infilling the graben could produce either several of these orthogonal silicone walls (given sufficient available silicone) or a migrating silicone inflation (implying the depletion of the former silicone wall over time and the formation of complex patterns of minibasin sedimentary infill).

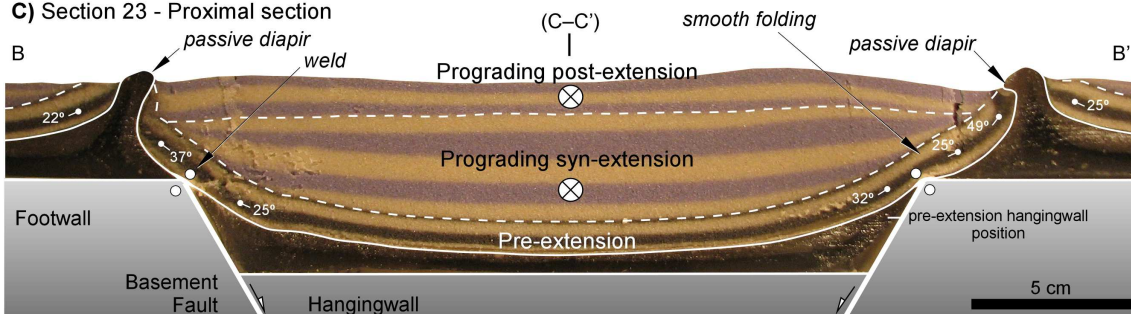
## A) Overhead views Model B4



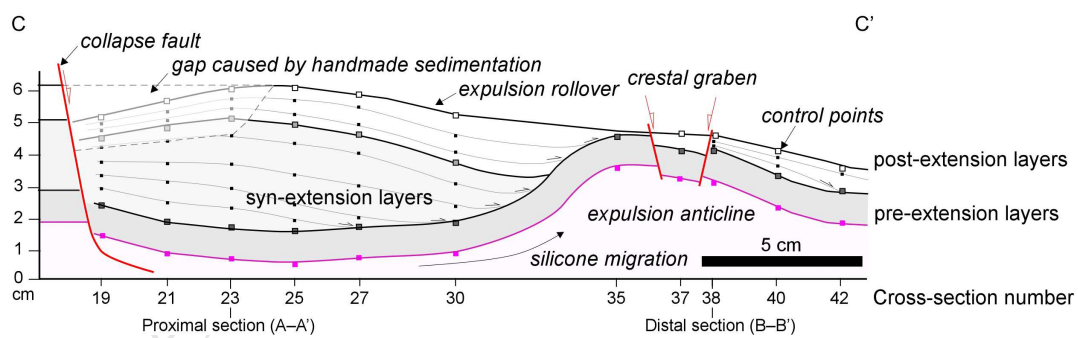
## B) Section 38 - Distal section



## C) Section 23 - Proximal section



## D) Interpreted longitudinal section



**Fig. 9** Overhead views (A) and sections (B and C) showing the geometries resulting from Model B4. This model included an extensional phase (total extension = 2 cm) and post-extensional phase (3 h) with longitudinal prograding sedimentation during both phases. Post-kinematic sediments have been removed in the section view. The longitudinal section from C to C' (D) was constructed by compiling the transverse sections from the model.

1 4.3. Models C1 and C2: extension and post-extension phases with  
2 homogeneous sedimentation and subsequent compression (6% and 10%  
3 shortening)

4

5 Models C1 and C2 (Fig. 10) used simple syn-rift and post-extension homogeneous  
6 sedimentation based on Model B1 (Fig. 4) and displayed compressional structures  
7 developed during basin inversion. Models C1 and C2 were performed with different  
8 amounts of post-diapiric and pre-compression sedimentation (1.5 and 3 cm,  
9 respectively) and with different applied shortening (3 and 5 cm, corresponding to 6%  
10 and 10%, respectively). The applied 6% of shortening (2.5 cm of uplift) did not fully  
11 invert the hangingwall in Model C1, whereas 10% of shortening (4 cm of uplift)  
12 produced a total positive inversion (pop-up structure) of the hangingwall in Model C2  
13 (Fig. 10). Owing to the presence of non-deformable basement in the model, the upward  
14 movement of the hangingwall during compression reopened previously welded areas  
15 above the tip lines of the basement faults and facilitated a renewed flow of silicone  
16 towards the diapiric structures and footwall domains.

17 Model C1 was characterised by the positive tectonic inversion of the central graben  
18 hangingwall, by the gentle folding of the entire pre-, syn-, and post-rift sedimentary  
19 successions within the extensional minibasin, and by the rejuvenation of subvertical  
20 diapiric ridges along both footwall blocks (section 37 A–A' in Fig. 10). An extreme  
21 thinning of the silicone layer occurred at the base of the extensional minibasin, giving  
22 an incomplete weld with a 0.1-cm-thick silicone remnant. The hangingwall pre-  
23 extension layers showed similar or slightly steeper dips of 30° to 49° compared with the  
24 purely extensional models (dips of 18° to 49°). However, the outward flaps of diapir

25 silicone walls along the footwall blocks were somewhat steeper (up to 30° on the left-  
26 hand side and up to 36° on the right-hand side of the model, respectively) than similar  
27 geometries formed in the B-series models (between 8° and 17°). The thin pre-  
28 compression layers in Model C1 were deformed and thrust on top of rejuvenated  
29 diapirs that showed steep to overturned flanks (Fig. 10). The compression event  
30 combined with moderately thick pre-compression beds permitted the complete breakup  
31 of the roof above the rejuvenated diapirs and the formation of syn-compression silicone  
32 extrusions. The vergence of silicone extrusion was away from the inverted central  
33 graben because of its higher elevation, in contrast to the extensional models, in which  
34 the topographic low was located in the subsiding central graben basin.

35 Model C2 was run with a pre-compression sand layer that was twice as thick as that in  
36 Model C1 (3 and 1.5 cm, respectively). After 5 cm of shortening, Model C2 showed  
37 positive inversion of the central graben and dome-shaped folding of the syn- and post-  
38 extension and pre-compression sedimentary successions above the pop-up basement  
39 structure (Fig. 10D). The thicker pre-compression layers in Model C2 were deformed  
40 mainly close to and above the rejuvenated diapirs. A large monocline, associated with a  
41 thrust fault, was observed in the pre-compression units on the left-hand side of the  
42 model, whereas the more evolved diapir on the right-hand side of the model produced a  
43 subvertical fold in the pre-compression layers (thrust box-fold geometry). Outer-arc  
44 extension resulted in roof-collapse normal faults along the crest of the rejuvenated  
45 diapir ridge. However, the most noteworthy results were the full welding of both border  
46 silicone walls, resulting in detached silicone stocks, and the formation of a primary weld  
47 of the minibasin against the uplifted basement of the central graben (Fig. 10). The  
48 secondary weld located on the right-hand side of the model was a thrust weld, with the  
49 right-hand-side footwall succession being thrust above younger rocks of the central

50 minibasin; the thrust weld might also have been generated by the inversion of roller-like  
51 diapirs (see Dooley et al., 2005). Compared with C1, the pre-extension layers in C2  
52 showed a steeper inclination that was achieved mostly around the time of or after the  
53 welding episode along the border silicone walls. The welding of the border diapirs and  
54 their subsequent buttressing resulted in an increase in the dips of the outward flaps (Fig.  
55 10), which were steeper than  $50^\circ$  where the flanks were subparallel to the welded diapir  
56 walls. The inward flanks of the silicone diapir walls, delimiting the central graben  
57 minibasin, showed dips of  $30^\circ$ – $40^\circ$  but in some cases were very steep ( $75^\circ$ ) to  
58 overturned in their uppermost short segments close to the diapir wall (uppermost short  
59 segment length about 0.3 cm on the left-hand-side flank of the minibasin and 0.5 cm in  
60 the right-hand-side flank; Fig. 10).

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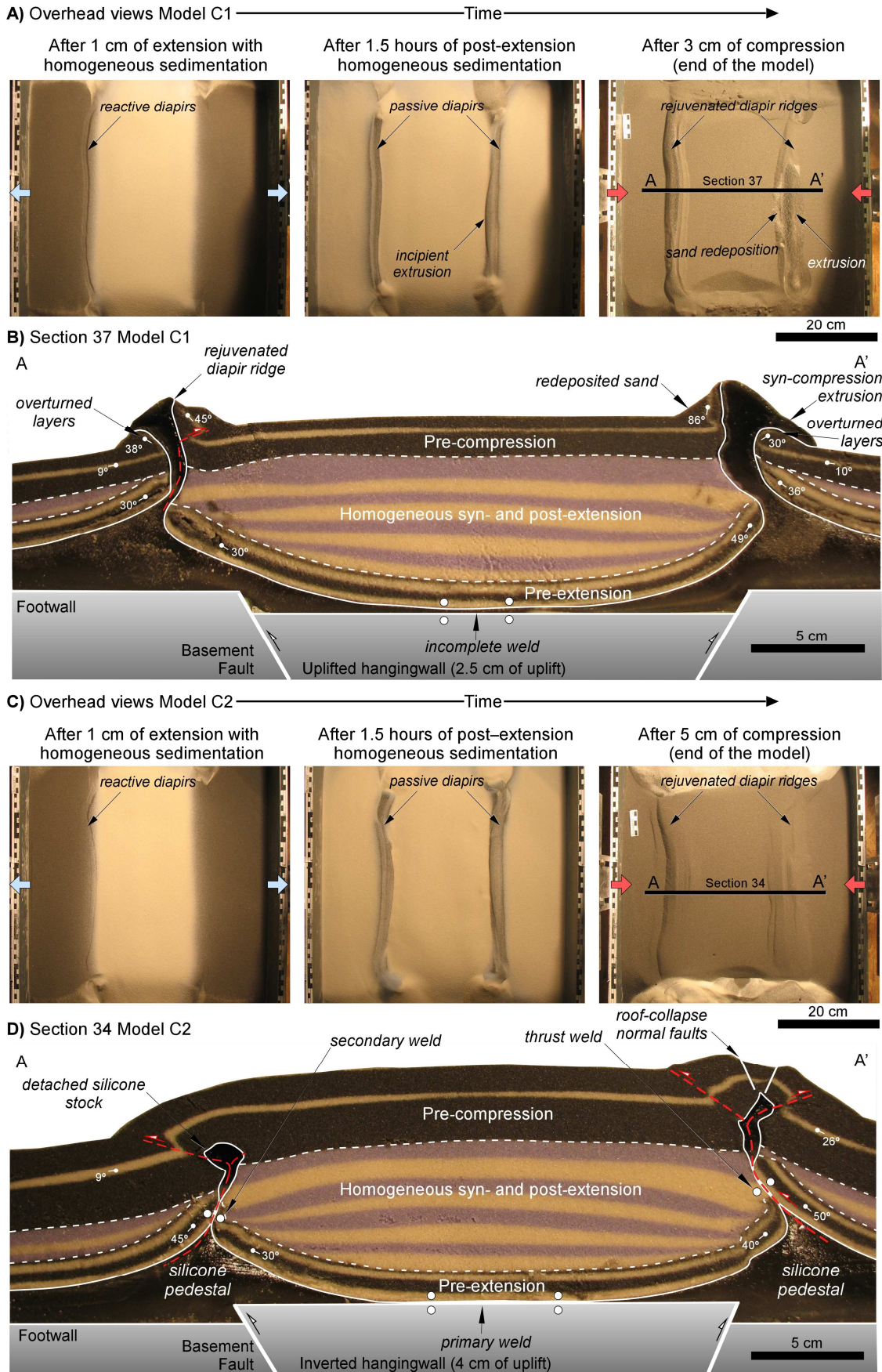
63

64

65 **Fig. 10** Overhead views (A and C) and sections (B and D) showing models C1 and C2. These  
66 models had an extensional phase (total extension = 2 cm), a post-extensional phase with  
67 homogeneous sedimentation, a pre-compression phase with homogeneous sedimentation of 1.5-  
68 cm thickness in Model C1 and 3-cm thickness in Model C2, and a final compression phase with  
69 a total amount of shortening of 3 cm (6%) in Model C1 and 5 cm (10%) in Model C2.

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## 74 5. Discussion

75 5.1. The sedimentary contribution to diapiric growth in salt-related rift  
76 basins

77

78 As has been demonstrated by the use of analogue models and case studies since the  
79 early 1990s, extension is a mechanism that can readily trigger reactive, active, and  
80 passive diapirism in salt-related rift basins (Dooley et al., 2005; Duffy et al., 2012;  
81 Jackson and Vendeville, 1994; Kane et al., 2010; Koyi et al., 1993; Lewis et al., 2013;  
82 Nalpas and Brun, 1993; Pascoe et al., 1999; Schultz-Ela and Jackson, 1996; Vendeville  
83 and Jackson, 1992; Vendeville et al., 1995; Withjack and Callaway, 2000). Syn- to post-  
84 rift sedimentation and the different sedimentary patterns used for infilling the basin,  
85 however, control the further evolution of passive diapirism in our models (see model  
86 configurations in Fig. 4).

87 Syn-extension homogeneous sedimentation applied in models B1, B2, and B3 causes  
88 the formation of well-developed rectilinear reactive diapirs along the footwall blocks of  
89 basement faults (extensional overhead views in Figs 6, 7, and 8) that evolve to passive  
90 diapirs resulting in the formation of rectilinear silicone ridges. The diverse post-  
91 extension sedimentation applied to these models control the final diapiric geometry.  
92 Diapirs in Model B1 and in the proximal domains of models B2 and B3 are well-  
93 developed steep passive diapirs owing to the equilibrium between sedimentation rate  
94 and diapir growth velocity (Giles and Lawton, 2002; Talbot, 1995; Vendeville et al.,  
95 1993). The distal domain of Model B2 (Fig. 7B), with longitudinal post-extensional  
96 prograding sedimentation, shows significant silicone extrusion along the passive diapirs.  
97 These allochthonous bodies are able to expand on top of the hangingwall deposits

98 because the post-extension sedimentation is too thin to compensate the growth of the  
99 diapirs (Koyi, 1998; Talbot, 1995; Vendeville and Jackson, 1992; Vendeville et al.,  
100 1993). The distal domains of Model B3 (Fig. 8B and C), with transverse post-  
101 extensional prograding sedimentation from right to left across the model, show poorly  
102 developed diapiric geometries, evolving from incipient roller-like diapirs to passive  
103 diapirs of smaller dimensions than those in the proximal domain.

104 Syn-extensional longitudinal prograding sedimentation applied in Model B4 (Fig. 9)  
105 causes the axial migration of silicone from the proximal domain to the distal domain,  
106 which, together with the limited amount of sand deposited during the late post-extension  
107 phase, limits the drag folding of pre-kinematic layers. Consequently, sufficient thinning  
108 of the pre-extension layers in the distal part of the model is not generated to allow a  
109 progression from incipient reactive diapiric structures to more evolved phases of  
110 diapirism. The longitudinal syn- and post-extensional prograding sedimentation of  
111 Model B4 also causes the growth of an expulsion anticline ahead of the prograding lobe  
112 that is perpendicular to the basement graben. Additionally, such sedimentation results in  
113 the development of discontinuous and non-rectilinear reactive diapirs that are diverted  
114 by silicone inflation (Fig. 9). We interpret the observed discontinuity as a result of the  
115 interaction between the regional extension and the local stress field associated with  
116 expulsion anticline growth caused by the thick and localised sedimentation of the syn-  
117 extensional sedimentary lobe (Gauillier and Vendeville, 2005; Sellier et al., 2013).

118 The salt migration resulting from the sedimentary load increases both the total and  
119 tectonic subsidence amounts, although its relative contribution is difficult to determine  
120 in subsidence analyses (Moragas et al., 2016). The comparison of models A1 and B1  
121 allows the value of this salt migration and its contribution to tectonic subsidence to be  
122 quantified, taking into account the initial configuration of the models and their silicone

123 thicknesses. In their respective final phases, Model A1 allows the tectonic (extension-  
124 related) subsidence at the end of extension to be isolated, whereas Model B1 allows  
125 both the tectonic (extension-related) and silicone migration (salt-migration-related)  
126 subsidence to be quantified for both extensional and post-extensional periods.

127 The total subsidence measured in the purely extensional Model A1 (i.e., without syn-  
128 extensional sedimentation) results from the sum of 80% of extension-related subsidence  
129 and 20% of salt-migration-related subsidence. These percentages change when both  
130 syn- and post-extensional sedimentation are applied (all B-series models). Calculations  
131 for Model B1 (Fig. 6C) yield 87% of extension-related and 13% of salt-migration-  
132 related contributions at the end of the extension. However, when the post-extension  
133 period of sedimentation is modelled, the contributions of the two mechanisms are very  
134 similar (ranging from 44% to 60% for the extension-related subsidence and from 40%  
135 to 56% for the salt-migration-related subsidence). This is because post-extension  
136 subsidence is strictly related to salt-migration in the models. These results should be  
137 applicable to subsidence analysis of sedimentary basins above thick salt layers, as in the  
138 Central High Atlas. However, one should take into account both the great differences in  
139 scale between the models and nature and the thermal subsidence associated with post-  
140 rift periods, as the reported in the Amezraï minibasin from the Central High Atlas  
141 (Moragas et al., 2016).

142 The results from the analogue modelling presented here show that sedimentation  
143 patterns are closely related not only to the evolution of diapiric structures and their  
144 geometries but also to the evolution of the diapiric basin as the configuration of the  
145 Central High Atlas reveals. Although transverse salt walls are exposed to a limited  
146 extent in the study area (see the geological map in Fig. 1), the combination of  
147 longitudinal syn- and post-extension prograding deposition generated a wide variation

148 in the geometries of basin infill along the axis of the salt-related rift basin. This  
149 variation resulted in the growth of transverse salt walls at depth that can be correlated  
150 with the polygonal array of minibasins in the Central High Atlas.

151

152

153 5.2. Learning from analogue models applied to the shortened diapiric  
154 structures of the Central High Atlas in Morocco

155

156 Our models including shortening (Models C1 and C2) show that pre-compression salt  
157 structures have a substantial impact on the final configuration of an inverted salt basin,  
158 with the majority of compression-related deformation taking place in locations with a  
159 thin overburden, such as over the top of diapirs, as extensively documented in many  
160 studies (Burliga et al., 2012; Callot et al., 2007; Callot et al., 2012; Dooley et al., 2005,  
161 2009; Letouzey et al., 1995; Nalpas et al., 1995; Rowan and Vendeville, 2006; Vially et  
162 al., 1994). As pointed out by previous studies, the anisotropy between the strong  
163 sedimentary pile and the weak silicone diapirs focuses the compressional strain in such  
164 diapiric structures.

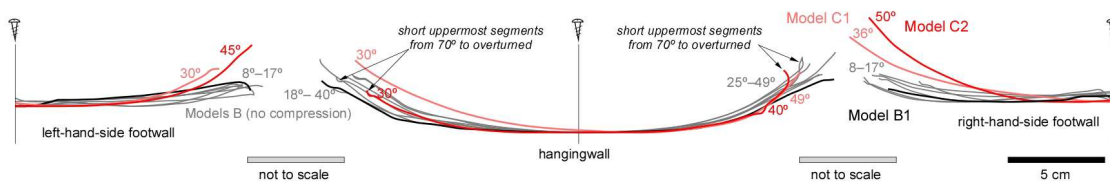
165 Two main problems concerning the geometry of inverted salt-related rift basins, as here  
166 with particular respect to the case of the Central High Atlas, can now be examined in a  
167 more refined way by applying the results obtained from our analogue models. The first  
168 of these problems is how to differentiate between purely extensional and compressional  
169 imprints of salt-related features. The second one is how to infer the positions of buried  
170 basement faults beneath the thick sedimentary overburden and salt-bearing layers, as  
171 discussed below.

172 The comparison of extensional models B1 to B4 with compressional Model C1 (6% of  
173 shortening) and Model C2 (10% of shortening) allows distinguishing the effects of  
174 extension on the geometry of silicone ridges from the effects of compression. This is  
175 especially clear when comparing Models B1, C1 and C2, all of which were run using  
176 pre-, syn-, and post-extension homogeneous sedimentation (Fig. 4). To compare the  
177 shapes and dips of the flanks of the silicone diapirs, we plotted them at the same scale  
178 using the pre-extension layers forming the walls of the diapirs as reference lines (Fig.  
179 11).

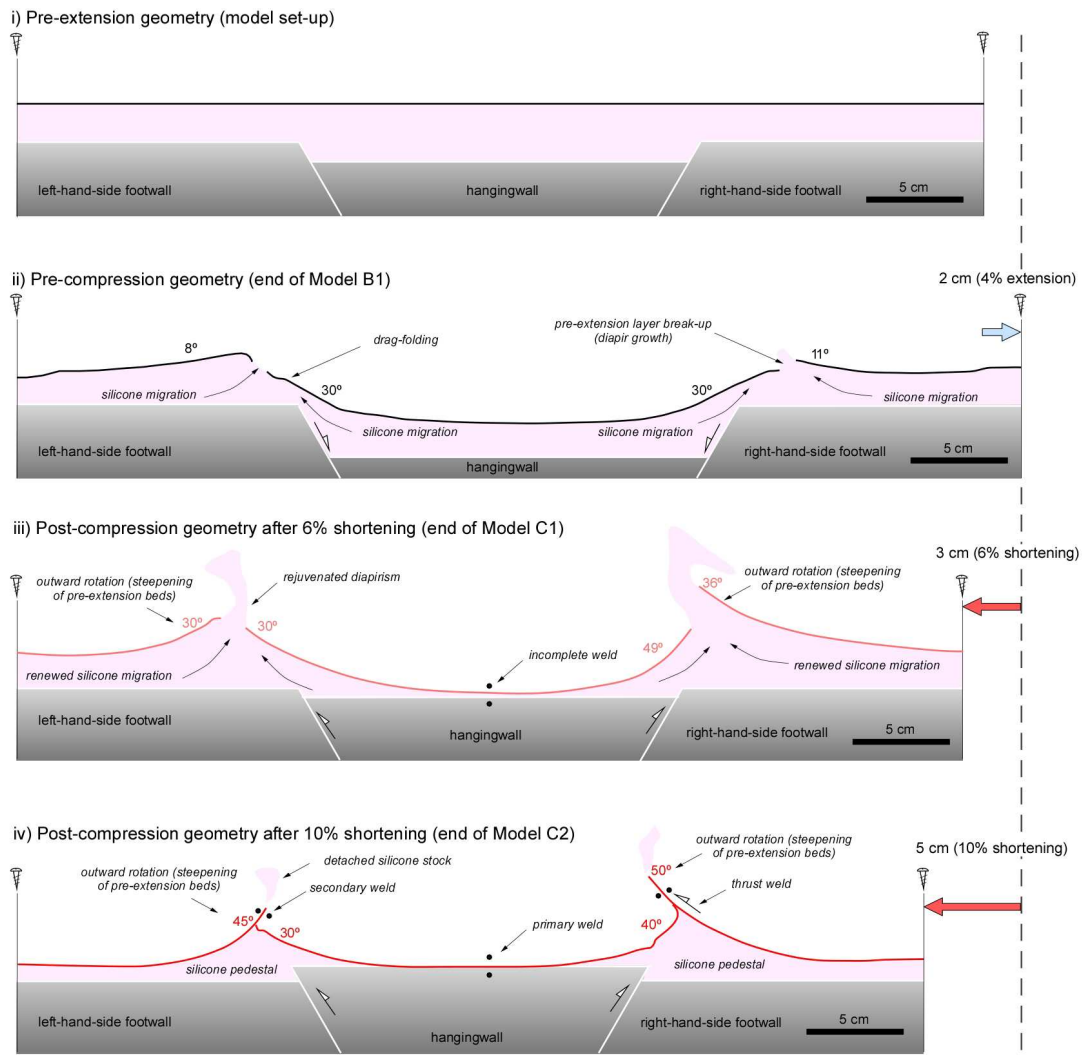
180 All purely extensional models display flaps with dips ranging from  $8^{\circ}$  to  $17^{\circ}$  in both  
181 footwall domains (outward flaps), whereas dips of the hangingwall flaps (inward flaps)  
182 are steeper, reaching  $18^{\circ}$ – $40^{\circ}$  and  $25^{\circ}$ – $49^{\circ}$  (Fig. 11). Data of Model B1 lay within these  
183 ranges, with  $8^{\circ}$ – $11^{\circ}$  in the footwalls and  $30^{\circ}$  for both sides of the hangingwall (Fig. 11).  
184 In contrast, the inverted models show an outward rotation of the outward flaps, with  
185 dips of  $30^{\circ}$  and  $36^{\circ}$  in Model C1 (6% shortening) and with the steepest outward flaps in  
186 Model C2 (10% shortening) reaching  $45^{\circ}$  and  $50^{\circ}$ . The evolution of the geometries of  
187 pre-kinematic beds in Fig. 11 shows how increasing shortening causes a steepening of  
188 extensional diapir flaps (mainly related to the footwall domains), whereas hangingwall  
189 flaps show dips in the same range as those of the extensional models, between  $18^{\circ}$  and  
190  $49^{\circ}$ . According to our models, which were set up with the minimum thickness of  
191 silicone needed to develop diapirs, all footwall flanks dipping more than  $17^{\circ}$  need the  
192 contribution of compression to develop. Thus, our models show that the tectonic  
193 inversion produces a greater steepening in the outward flaps and a smaller steepening  
194 along the flanks of the minibasin. The models also show short segments with  
195 overturning in the inward flaps, as observed in both flanks of the Tazoult salt wall in the  
196 Central High Atlas (Fig. 12). That being said, the inward flap of the Tazoult salt wall

197 shows a greater length and steeper inclination than that observed in any of our models,  
 198 including those with compression. A possible explanation for this could be that a greater  
 199 amount of salt migration would generate the  $\sim 80^\circ$  dip observed in the southern flap of  
 200 the Tazoult salt wall.  
 201

A) Comparison of pre-extension layer geometries from all extension and compression models



B) Pseudo-temporal evolution of pre-extension layers geometry



202

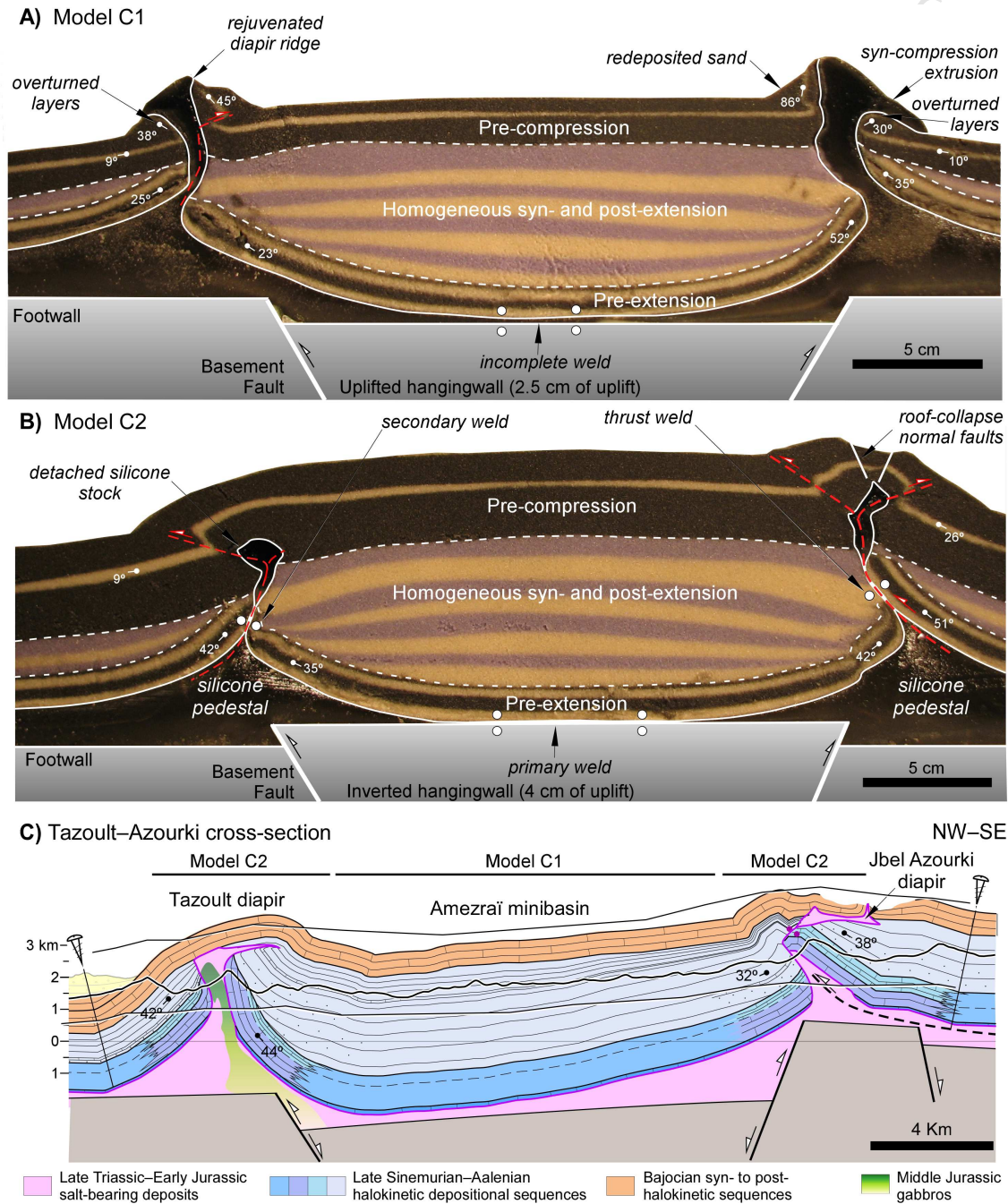
203 Fig. 11 (A) Comparison of pre-extension layer geometries from purely extensional models and  
204 models including 6% and 10% of compression. Compression causes the strongest steepening of  
205 pre-extension layers in the footwall domains. (B) Pseudo-temporal evolution of pre-extension  
206 layer geometry from extension through to 10% of shortening. The vertical scale equals the  
207 horizontal scale for all diagrams in both (A) and (B).

208

209 For the Tazoult and Jbel Azourki diapirs, the positions of basement faults in the cross-  
210 sections were inferred based on thickness variation between the two flanks of these  
211 diapiric salt walls (ridges). However, the positions and geometries of these faults at  
212 depth are imprecise, and the tip lines of the faults were placed directly under the diapir  
213 walls cropping out at the surface (Fig. 1; Martín-Martín et al., 2016). In contrast,  
214 according to our results and those of previously published models (e.g., Dooley et al.,  
215 2005), diapiric structures develop mostly in the footwalls of the basement normal faults  
216 and away from their tip lines. On the basis of our modelling, we propose a new cross-  
217 section in which the Tazoult and Jbel Azourki diapirs are located in the footwall  
218 domains of the basement normal faults that bound the Amezraï minibasin (compare Figs  
219 1C and 12C).

220 The final geometry of diapirs after the shortening overprinting is another outcome from  
221 modelling that can be applied to better discriminate between extensional and  
222 compressional geometries, as observed for Models C1 and C2 (Figs 11 and 12). Our  
223 results show that passive diapirs formed during extension are tightened during  
224 subsequent compression, with reductions in diapir width ranging from 32% to 72% in  
225 Model C1 with 6% of shortening. These pre-existing diapirs are completely welded in  
226 Model C2 with 10% of shortening. In addition, the tightening and welding of diapirs  
227 may create a thrust weld, moving the extensional footwall overburden above the thicker  
228 extensional hangingwall post-silicone sedimentary pile. These results are consistent

229 with field observations in the Tazoult and Jbel Azourki diapir walls, which, under a  
 230 regional shortening of 10.4% (Martín-Martín et al., 2016), are completely welded in  
 231 areas where there have been no magmatic intrusions or slivers of Hettangian carbonates.  
 232



233

234 Fig. 12 Sections from Model C1 (A) and Model C2 (B), showing the main sand units according  
235 to the timing of deposition (pre-, syn-, and post-extension, and pre-compression). (C) A newly  
236 proposed cross-section from the Tazoult–Jbel Azourki transect (modified from Martín-Martín,  
237 et al., 2016; Fig. 1) with diapirs relocated on the footwalls of basement normal faults and black  
238 bars on top referring to geometries comparable to the corresponding analogue models. The thin  
239 black lines indicate the currently exposed structural levels and topography. The lateral facies  
240 change from the oldest halokinetic depositional sequences represents the transition from  
241 shallow-marine (light blue) to deep-marine (dark blue) deposits as recently reported in the  
242 Central High Atlas diapiric structures by Jousseaume (2016), Malaval (2016), Martín-Martín et  
243 al. (2016), Teixell et al. (2017), and Vergés et al. (2017).

244

## 245 6. Conclusions

246

247 Our physical models were designed to refer to the main feature of the Central High  
248 Atlas, namely, with a minimum salt-layer thickness of around 1.5 km. The presented  
249 models indicate a close interaction between tectonics, sedimentation, and diapirism in  
250 salt-related rift basins. Once extension-driven diapirism in the models begins, governed  
251 by the geometry of the fault system, the syn- and post-extension sedimentary patterns  
252 have a major influence on the evolution of such diapirism and on the configuration of  
253 the modelled basin.

254 In the models, syn-extension homogeneous sedimentation enhances the mobilisation of  
255 silicone towards growing diapirs, resulting in the development of passive diapiric ridges  
256 along the footwalls of the basement faults. Post-extension longitudinal and transverse  
257 sedimentary progradations cause variations in the diapiric geometries along the  
258 direction of progradation. Well-developed passive diapirs with vertical walls grow in

259 the proximal domains, where sedimentation rates and diapir growth rates are in  
260 equilibrium. In contrast, silicone extrusions or poorly developed diapiric structures  
261 evolve in the distal domains, where sedimentation rates are lower than diapiric growth  
262 rates.

263 Syn- and post-extension longitudinal sedimentary progradation causes a substantial  
264 migration of silicone from proximal to distal domains of the prograding system,  
265 resulting in the growth of an expulsion anticline that hampers the progression of  
266 reactive diapirs to more evolved diapiric phases in the distal part of the models.

267 Tectonic inversion of the models triggers the following: i) the steepening of the outward  
268 flaps of pre-compression silicone diapir walls, with dips increasing from  $8^{\circ}$ – $17^{\circ}$  prior to  
269 compression to  $30^{\circ}$ – $50^{\circ}$  after compression; ii) the reopening of the silicone migration  
270 network by thrusting welds off the sub-silicone basement; and iii) the tightening of the  
271 silicone diapir walls and their final welding when shortening reaches  $\sim 10\%$ .

272 The modelling results provide insights into the natural example of the Central High  
273 Atlas diapiric province in Morocco, as follows: a) the polygonal array of minibasins of  
274 the Central High Atlas is inferred to be a result of the growth of transverse salt walls  
275 associated with the syn- and post-extension longitudinal sedimentary progradation of  
276 the Early and Middle Jurassic; b) the Alpine Orogeny contributed to the steepening of  
277 the outward flaps of the Tazoult and Jbel Azourki diapirs, whereas the flanks of the  
278 Amezraï minibasin appear to have been only slightly deformed during shortening; c) the  
279 inferred basement normal faults in the Tazoult–Jbel Azourki transect can be redrawn to  
280 adjust them with respect to the diapir walls located in their footwalls instead of directly  
281 above their tip lines; and d) tectonic subsidence in salt-related rift basins, such as the  
282 Central High Atlas, includes a salt-migration-related subsidence component as high as

283 40%–56% of the total tectonic subsidence, which hampers the calculation of extensional  
284 factors.

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286

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