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The rift to break-up evolution of the Gulf of Aden: Insights from 3D numerical lithospheric-scale modelling

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Abstract

The Gulf of Aden displays an ideal setting to study oblique rift processes since numerous structural data are available onshore and offshore, down to the ocean-continent transition (OCT).

We investigate key observations by means of 3D numerical thermo-mechanical experiments on lithospheric scale. The Gulf formed with a rift-normal of -15° azimuth and under the influence of a supposedly 25° trending far field extension. By reproducing oblique rifting with 40° , we study the evolution of the Gulf of Aden in terms of crustal fault geometries and stress patterns from rift initiation to break-up.

Our model suggests that intermediate faults dominate during the initial rift phase, followed by rift-parallel normal faulting at the rift flanks and strike-slip faults in the central part of the rift system. Upon break-up, displacement-orthogonal as well as intermediate faults occur. We compare our results to previous analogue experiments of oblique rifting on lithospheric scale as well as to the structural evolution of the Gulf of Aden.

The basic evolution is in accordance with the development of fault patterns in the analogue model and allow to propose further interpretation of the distal margin evolution of the Gulf of Aden. To large extent, this study supports previous hypotheses about the processes taking place during oblique rifting, however, it proposes different deformation

patterns during the possible exhumation of deep material in the OCT.

1. Introduction

Oblique rifts are evolving through a complex pattern of deformation, which changes through time. By nature, they cannot be assimilated to 2D structures since they display strong lateral variations. Therefore, they need to be studied in three dimensions and their understanding still depends on the ability to reproduce 3D processes.

The obliquity of the Gulf could be linked to the interaction between the laterally-evolving subduction of the Tethyan Ocean toward the north and the Afar hot spot in the south-west (Bellahsen et al., 2003). Field and seismic studies were conducted onshore and offshore Oman (Fournier et al., 2004; d'Acremont et al., 2005; Bellahsen et al., 2006) and in Yemen (Huchon and Khanbari, 2003), and allow to recognise three fault populations (extension-normal, rift-parallel, intermediate). They correspond to several directions of extension in Oman (Lepvrier et al., 2002; Fournier et al., 2004; Bellahsen et al., 2006), in Yemen (Huchon et al., 1991; Huchon and Khanbari, 2003) and on Socotra Island (Fournier et al., 2007). Offshore, near the Ocean-Continent Transition (OCT), both the faults and basins mainly strike perpendicular to the Gulf opening (d'Acremont et al., 2005). The OCT most likely exhibits exhumed serpentized mantle (d'Acremont et al., 2006; Leroy et al., 2010c; Leroy et al., 2012). The Gulf of Aden display also a high segmentation with first-order and second-order segmentations (Leroy et al., 2004; d'Acremont et al., 2005). Due to these structural data sets that were collected both onshore and offshore, the Gulf of Aden is an ideal area to study rifting with moderate obliquity.

Fault patterns of oblique rifts have been investigated during the last decades using analogue models on two different levels of complexity: (i) Crustal scale models simplify the rift system to a deforming crust influenced by a basal zone of extension that involves an oblique velocity discontinuity (Withjack and Jamison, 1986; Tron and Brun, 1991; McClay and White, 1995; Clifton et al., 2000; Mart and Dauteuil, 2000; Corti et al., 2001, 2003; Corti, 2004; Sokoutis et al., 2007). The advantage of this setup is that crustal strain patterns can be studied independently of mantle deformation, but this also limits the applicability to the first rift stage where isostatic balancing with the mantle and lithospheric necking can be neglected. Furthermore, the role of the basal discontinuity is

1 overestimated intrinsically. (ii) Analogue experiments on lithospheric scale have been
2 conducted recently and did successfully reproduce lithospheric thinning and its effect on
3 crustal fault patterns (Sokoutis et al., 2007, Agostini et al. 2009, Autin et al. 2010).
4 However, thermal effects or rheological changes occurring during rifting are not modelled
5 in these experiments and their absence remains a significant limitation of such analogue
6 models. They also do not show the progression from oblique rift initiation to plate
7 rupture.
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14 Within the recent decade, numerical models became more and more important in
15 understanding lithospheric deformation. In contrast to analogue models, state-of-the-art
16 geodynamic codes are capable of computing realistic temperature dependent viscosity as
17 well as complex elasto-visco-plastic rheologies. Despite these advantages, numerical
18 models of oblique rifting intrinsically require computationally expensive calculations in
19 three dimensions which is why few numerical models have been published so far: Van
20 Wijk (2005) pioneered this topic by investigating rift evolution using a relatively coarse
21 lithospheric-scale model. She showed that individual rift segments will cross an inherited
22 weak zone that is oblique to the direction of extension. Allken et al. (2011, 2012) studied
23 the influence of an offset between two rift segments on the structures of an extensional
24 system. While using a high resolution of 1.3 km, they restricted their models to crustal
25 scale. Brune et al. (2012a) showed by means of a simple analytical model that oblique
26 rifting is energetically preferred over rift-perpendicular extension, which they
27 corroborated by means of lithospheric-scale numerical experiments. This model has been
28 extended in order to investigate the influence of plume-related lithosphere erosion on the
29 dynamics of continental break-up (Brune et al. 2012b).
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46 In this paper, we show that lithospheric-scale numerical experiments are capable to
47 reproduce extensional structures from initial rifting to break-up. We thereby apply elasto-
48 visco-plastic rheology with laboratory-based flow laws for temperature/pressure-
49 dependent viscosity. We investigate the fault geometries during of oblique rifting based
50 on strain-rate and plastic strain patterns. Moreover, we exploit the fact that numerical
51 models provide direct access to the stress tensor at any numerical element, which allows
52 to evaluate fault patterns on a sub-shear zone level. We explicitly compare our
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1 experiments to previous analogue modelling results and relate them to present structural
2 knowledge of the Gulf of Aden.
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4 5 **2. Model description** 6

7 We consider a rectangular Earth segment that measures 249 km times 249 km
8 horizontally and 120 km vertically (Fig. 1a). We thereby use 275560 cubic elements with
9 a length of 3 km. The modelled Earth segment consists of a 20 km thick upper crustal
10 layer with a wet quartzite rheology (Gleason and Tullis, 1995), a lower crustal layer of 15
11 km thickness of granulite properties (Wilks and Carter, 1990), and a 45 km thick layer of
12 strong mantle material with dry olivine rheology (Hirth and Kohlstedt, 2003). We
13 introduce a chemical asthenosphere by applying the flow law of wet (i.e. 500 ppm H/Si)
14 olivine below 90 km depth (Hirth and Kohlstedt, 2003). All rheological parameters are
15 listed in Table 1.
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26 We impose a full extension velocity of 10 mm/yr through velocity boundary conditions
27 at the model sides facing in x-direction so that they move symmetrically with 5 mm/yr.
28 This velocity is equivalent to 10 km/My which results in a maximum extension of 200 km
29 after 20 My model time. We refer to the angle between boundary velocity and the
30 boundary itself as the angle of obliquity α . In this study, we use $\alpha=40^\circ$ which represents
31 the obliquity encountered in the Gulf of Aden. The front and back sides of the model are
32 connected via periodic boundary conditions which effectively realises an infinitely long
33 rift zone.
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43 We apply the implicit, finite element code SLIM3D (Semi-Lagrangian Implicit Model
44 for 3 Dimensions) to solve the thermo-mechanically coupled conservation equations of
45 momentum, energy and mass in three dimensions. Detailed description of the numerical
46 methods can be found in Popov and Sobolev (2008) and Brune et al. (2012a).
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52 In many cases oblique rifting arises because inherited lithospheric weak zones like
53 sutures are reactivated with an oblique extensional component (Ziegler and Cloetingh,
54 2004). We introduce a weak zone by implementing a small linear temperature
55 heterogeneity in the centre of the prospective rift (Fig. 1b). In doing so we anticipate a
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small amount of lithospheric necking that focuses the extensional deformation into the desired rift axis. This is one possible way of rift initialization. Alternative means are mechanical anisotropy (Tommasi and Vauchez, 2001), implementation of a weak plastic seed (Huisman and Beaumont, 2003), or crustal thickening (van Wijk, 2005). Note that after a small amount of extension, all of these techniques will result in lithospheric necking comparable to our initial condition.

Three weakening mechanisms are reproduced in the model. (i) Friction softening is introduced using a strain-dependent effective friction coefficient that decreases linearly from 0.6 to 0.06 for plastic strains between 0 and 1 while it remains constant at 0.06 for plastic strains larger than 1. (ii) Shear heating results in increased temperature proportional to stress times strain rate. (iii) Stress softening and strain rate softening are intrinsic to dislocation creep and reduce the local viscosity.

In most geodynamic codes (including SLIM3D), fault structures are represented by finite width shear bands that localize within a couple of elements. Here we utilize a technique that allows to use the principal stress components and their orientation in order to extend the geodynamic interpretation of our models (Brune et al., to be submitted): At each surface element we therefore evaluate the scalar Regime Stress Ratio (RSR) that indicates extension (RSR=0.5), strike-slip motion (RSR=1.5), and compression (RSR=2.5) on a continuous scale (Simpson 1997). Once the stress regime is computed for each surface element, we assume Andersonian faulting to infer the optimally oriented fault direction: For extensional and compressive stress regimes in isotropic and homogeneous materials, faults emerge with azimuths orthogonal to σ_3 . Strike-slip faults occur at $\pm 30^\circ$ from σ_1 . For reason of symmetry it is not possible to differentiate between sinistral and dextral strike-slip faulting based on stress tensor information only which is why we account for both conjugate fault populations. In the azimuth diagram, however, they are scaled with a factor of 0.5 so that the overall number of evaluated elements is not affected. Azimuth is measured as the clockwise angle from northward direction. This method delivers a preferred fault mechanism and orientation for any given stress tensor, even if no deformation takes place inside the element. Taking this into account, we restrict our stress analysis to a zone of tectonic activity where the strain rate is larger than 10^{-15} s^{-1} .

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2 The number of elements within the active region varies with time and is indicated in the
3 azimuth diagrams of Fig 3c in the upper left corner (#Elements).
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5 **3. Model results**

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7 The model resolution of 3 km inhibits the formation of faults in a strict sense, since
8 faults localize in nature on a much smaller scale. Instead, we observe finite-width shear
9 zones with a typical width of few elements (Fig. 2a). The spontaneous formation of shear
10 zones takes place within the first computational time steps. Fault spacing is directly
11 related to the brittle-ductile transition depth in the upper crust (e.g. Vendeville, 1987).
12 Due to strain softening, shear zones become weaker with accumulated deformation. Thus,
13 the individual small-scale shear zones compete and their number reduces with time.
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22 The largest amount of deformation is taken up by shear zone parts atop the lithospheric
23 necking peak. During the first 6 My large conjugate normal faults develop within the
24 shear zones that cut through the whole crust so that surface spacing is controlled by the
25 Moho depth (Fig 2e). Since the rift process decreases crustal thickness, the distance of the
26 conjugate faults at the surface is reduced with time until it vanishes and break-up takes
27 place at 14 My. This model does not account for petrophysical formation of oceanic crust
28 so that continental break-up is complete when the crust is broken and asthenospheric
29 material reaches the surface.
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39 Three fault azimuths will play a fundamental role during the discussion of the model,
40 i.e. rift-parallel (75°), displacement-orthogonal (115°), and intermediate (95°). These
41 directions result from a 25° -oriented direction of extension so that the global orientation
42 of the least principal stress σ_3 is 5° .
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48 The evolution of the numerical model can be divided in three main phases. Note,
49 however, that the transitions between phases are not abrupt but take place over one or two
50 My. Figures showing the evolution in steps of 1 My can be found in the supplementary
51 materials.
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56 **Phase 1.** (1-5 My): At 1 My, the strain rate pattern of Fig 2a shows small-scale shear
57 zones that strike at an angle of 95° azimuth that is intermediate between the extension-
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1 orthogonal direction and the rift orientation. Within few million years, they develop into
2 an en-echelon system with a wavelength of several tens of kilometres. The stress-inferred
3 fault mechanism (Fig. 3a), is of normal type everywhere and shows intermediate fault
4 orientation (Fig 3b,c).
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9 **Phase 2.** (6-13): Deformation of the en-echelon shear zones strongly localises towards
10 the lithospheric necking region. (Fig 2a). The normal fault azimuth map (Fig 3b) shows
11 rift-parallel and intermediate normal faulting at the rift flanks. In the rift centre, however,
12 extension-orthogonal faults occur together with strike-slip faults that delimit individual
13 shear zones (Fig 3a). The azimuth diagram (Fig. 3c) shows a shift from intermediate to
14 rift-parallel directions (the distribution starts to be asymmetric). At 10 My, rift-parallel
15 faults are dominant at the rift borders (Fig. 3b and 3c) and a strong localization of the
16 deformation occurs again.
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26 **Phase 3.** (14 My and after): Incipient break-up links up the individual shear zones. Only
27 intermediate and extension-orthogonal faults develop during break-up of the lithosphere.
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32 The final strain distribution (Fig 2b at 14 My) shows sigmoid deformation patterns. The
33 sigmoidal shape can be explained by successive rift localisation and the longevity of
34 individual shear zones: After formation of the initial en-echelon pattern, the central
35 portion of each shear zone gets stretched parallel in direction of extension which appears
36 as a counter-clockwise rotation (Fig 4). Since deformation localizes towards the rift
37 centre, the area where rotation occurs narrows with time. Thus, shear zones of the
38 proximal margin experience less rotation while distal margin shear zones are deformed
39 until they are nearly parallel to the direction of extension.
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48 Note that the model capabilities are limited in several aspects. Most importantly, magma
49 migration and dike formation that tend to decrease lithospheric strength perpendicular to
50 the direction of extension are not accounted for. Moreover, the limits of computational
51 power restrict our model resolution to 3 km which is still far from resolving individual
52 faults. Nevertheless, the presented model is one of the first that reproduces lithospheric-
53 scale rift evolution from initial deformation until break-up.
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4 Numerical model vs. Gulf of Aden analogue model

Before we compare the numerical model to the Gulf of Aden, we will explore differences and similarities to the previously conducted lithospheric-scale analogue model of Autin et al. (2010). Both experiments feature an obliquity of 40° . As the numerical model uses a prescribed weakness in the lithosphere, we compare it to an analogue model which also contains a pre-existing lithospheric weakness (Autin et al., 2010, model B).

The analogue model is constructed in order to reproduce oblique rifting by the way of shifted lateral discontinuities (see Autin et al., 2010 for details). Moreover, an oblique weakness trends parallel to the direction of obliquity imposed by the lateral velocity discontinuities, and joins them. It displays a four-layer type lithosphere strength profile (Fig. 1c) modelled using granular materials and silicone. This modelled lithosphere overlies a low viscosity, higher density glucose syrup that mimics the asthenosphere. Lateral dimensions of the setup (56 cm times 30 cm in the laboratory) scale to 750 km times 400 km in reality.

Autin et al (2010), as well, observed three main fault populations: rift-parallel (75°), intermediate faults (95°), and displacement-normal (115°). The main results of Autin et al. (2010) was the recognition of 3 main steps of development: (i) The fault populations, especially during the early stages of deformation, are composed of faults that in strike, are largely intermediate between rift-parallel and perpendicular to displacement. This fault population is characteristic of oblique rifts as observed in previous studies . (ii) In later stages, faults parallel to the rift become numerous. Autin et al. (2010) propose that buoyancy forces related to thickness variations in the lithosphere during rift localization play a significant role and control the initiation of rift-parallel faults (see also Bellahsen et al., this volume). (iii) During the final stages of extension, the small-scale deformation pattern is composed of displacement-normal faults in the deepest parts of the rift.

The three stages are very similar to the ones observed in the numerical models. Nevertheless, several differences exist: The numerical model indicates that during the beginning of Phase 2, rift-parallel and extension-orthogonal faults develop

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simultaneously. Contrarily to analogue models, Phase 3 of the numerical model displays extension-orthogonal as well as intermediate faults (Fig. 3c). In the numerical model, this stage is controlled by the ascent of the hot asthenosphere and subsequent plate cooling, which cannot be reproduced in analogue models. These processes induce a strong localization of the deformation, where oblique weakening combined with the far-field stress could lead to the intermediate fault development. The azimuth diagram (Fig 3c) shows clearly extension-orthogonal faults at 14 My, when the breakup occurs. It is slightly different from the analogue model, which suggests that they appear at earlier stages, during hyper-extended domain formation. It is noteworthy that clockwise rotations of the structural pattern start at 7 My in the numerical model. Rotations are observed at the boundaries of the analogue models, which are either clockwise or counter-clockwise depending on the rift border where it occurs. They are thought to be responsible of the initiation of second order transfer zones in oblique rifts.

Another fundamental similarity between both models is the fault repartition in the rift. In both the numerical and the analogue models, rift-parallel faults are always located along the rift borders. Rift borders is where the overall oblique thinning of the lithosphere creates the strongest thickness variations, inducing density variations which are thought to enhance the buoyancy forces, perpendicular to the oblique rift (Bellahsen et al., this volume). On the other hand, displacement-normal faults are always created in the rift centre, particularly in later stages of deformation, when no more stresses perpendicular to the rift occur in the rift centre.

5 Numerical model vs. Gulf of Aden natural rift

5.1 Final deformation pattern

The overall deformation pattern of the Gulf shows an en-echelon disposition of the syn-rift faults and grabens, on both side of the oblique OCT (Fig. 5a). When the Gulf is closed to the OCT (Fig 5b), the tertiary main depocentres show en-echelon sigmoid grabens. This pattern is similar to the plastic strain pattern observed in the numerical model (Fig. 2b). The modelled basin topography is also comparable with en-echelon sigmoid basins progressively linked together and finally separated when the final localization occurs at 10-11 My (Fig 2d). Nevertheless, this first order deformation occurs

1 at larger scale (ca. 100 km) than in the numerical model (40-50 km) and is partly
2 controlled by the mesozoic inheritance (Ellis et al., 1996; Granath, 2001; Leroy et al.,
3 2012; Autin et al., this volume). This wavelength difference is certainly due to the initial
4 pattern of the inherited, widely spaced, mesozoic basins, which have focused the
5 deformation during their reactivation, preventing the appearance of a more distributed
6 pattern as in the numerical model. Although, the main tertiary depocentres focus in these
7 inherited basins, deformation is also observed outside of them as for example in the
8 Ashawq graben (Fig 5c). Thus, it appears that the distributed deformation pattern of the
9 model can be recognized in the Gulf of Aden but is locally controlled and enhanced by
10 the inheritance overprinting the general pattern.
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20 As proposed above, the sigmoid shape in the model can be due to successive rotations
21 during ongoing extension of long-lived shear zones (Fig. 4). In the Gulf of Aden, the
22 inherited mesozoic basins have a sigmoid shape and thus could have experienced such
23 rotations. Novel analogue models that reproduce this inheritance during oblique rifting
24 show indeed that the inherited structure ends with a sigmoid shape (Autin et al., this
25 volume).
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32 **5.2 Chronology and localisation of the fault populations**

33 Another point of comparison is the distribution of the fault populations in the rift and
34 their chronology. As describe in Bellahsen et al. (this volume), the Western Gulf of Aden
35 displays a general fault organisation: (i) The external parts of the rift show intermediate
36 and extension-orthogonal faults; (ii) The steep slopes focus rift-parallel faults and; (iii)
37 The internal parts are mainly composed of intermediate (rather in the OCT) and
38 extension-orthogonal faults (ridge trend). If we consider that the deformation localizes
39 progressively towards the rift centre, then the most proximal structures are older than the
40 distal ones. This evolution is in accordance with the 3 step evolution of the numerical
41 model. Indeed, figure 3b and 3c show that (i) the intermediate faults are created first and
42 will then be located in the external parts of the rifts; (ii) rift-parallel faults form later and
43 are localised where the thinning is the strongest (and thus the slope the steepest), i.e. at
44 100 km and 150 km along the x-axis at 6 My or 110 km and 140 km at 10 My (Fig. 2e);
45 (iii) intermediate and extension-orthogonal appear in the rift centre during the final
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1 evolution of the model.

2 3 4 **5.3 Processes at work**

5 We propose that the numerical model of oblique rifting captures the main evolution of
6 the deformation through the Gulf of Aden and in particular in the Western part, where a
7 localising effect of the high thermal regime could be assimilated to the initial weakness
8 introduced in the models. In this frame, the deformation processes occurring during
9 oblique rifting proposed in previous works are partly supported by our model.
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16 For the Phase 1 (1 to 5 My), the stress-inferred azimuth diagrams show that the normal
17 faults have first an “intermediate” direction which results from the combination of the far
18 field stresses and the local stresses induced by the weakness zone as proposed by
19 Withjack and Jamison (1986).
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26 During Phase 2 (6 to 13 My), the progressive development of rift-parallel faults
27 seemingly indicates that deformation localises along the oblique trend. This localisation
28 could be linked to enhanced buoyancy forces that induce a rift-orthogonal extension
29 (Bellahsen et al., this volume). Buoyancy forces arise from density variations in the
30 lithosphere (Artyushkov, 1973; Fleitout and Froidevaux, 1982) and thus are perpendicular
31 to the major lithosphere thinning. They are thought to be driving forces for rift
32 localization even in orthogonal settings (Huismans et al., 2001; Davis and Kusznir, 2002;
33 Burov, 2007). This hypothesis is supported by the localisation of rift-parallel faults in the
34 rift borders above the maximal thinning area (Fig. 2e) and the emergence of rift shoulders
35 (Fig 2c). Moreover, this deformation is correlated with a strong localisation of the
36 deformation, which would confirm the localising effect of buoyancy forces, that are
37 thought to allow the distal margin formation.
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50 Deformation during the third step (from 14 My on) is localised in the rift centre. The
51 intermediate and extension-orthogonal faults indicate that the far-field extension
52 dominates. The rift centre is far enough from the thinning zones (rift borders) so that the
53 newly formed faults develop mainly in response to the far field stresses, as proposed in
54 Autin et al. (2010). Nevertheless, the presence of intermediate faults suggests that a local
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1 stress field is still active. The reason is probably that plate cooling takes place parallel to
2 the former rift zone. The ongoing evolution of the model shows that intermediate fault
3 proportion tends to decrease with time compared to the extension-orthogonal faults,
4 suggesting that the far-field is more and more dominant.
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9 The numerical model also coincides with a general conceptual model of passive
10 margins. Indeed, the strain rate shows clearly the transition from the distributed
11 deformation (stretching mode) to a more localizing rift pattern from 6 to 9 My (thinning
12 mode) and the final narrow oblique localization at 10 My and progressive exhumation of
13 the deep layers before the lithospheric break-up at 14 My (exhumation mode), correlating
14 with the general margin formation modes described by Lavier and Manatschal (2006) and
15 Péron-Pinvidic and Manatschal (2009). Exhumation of lower crust and mantle material
16 under a poor magmatic regime suggests the formation of magma-poor rifted margins in
17 the numerical model. Although, geophysical data allow to propose the presence of
18 exhumed serpentinitised mantle in the Eastern Gulf of Aden (d'Acremont et al. 2006;
19 Leroy et al. 2010a and b), such dataset is not available in the Western Gulf of Aden.
20 Nevertheless, in the eastern part of the Western Gulf, margins are thought to be magma-
21 poor as no magmatic structures as seaward dipping reflectors were recognised (e.g.
22 Bosworth et al., 2005). If so, our numerical model would suggest that such exhumation
23 will take place when the deformation pattern is dominated by a combination of numerous
24 intermediate faults and only few extension-orthogonal faults, i.e. at around 13 My.
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41 **6. Conclusion**

42 We identify a characteristic evolution of fault patterns: At first, faults develop with
43 orientations that are intermediate between the rift-direction and the displacement-normal.
44 Then rift-parallel normal faults occur at the rift flanks simultaneously with strike-slip
45 faults and extension-orthogonal faults in the central part of the rift system. Finally
46 displacement-orthogonal as well as intermediate orientation dominate during break-up.
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54 This evolution is in accordance with the emergence of the fault pattern in lithospheric
55 analogue models and allows to propose further interpretation of the distal margin
56 evolution of the Gulf of Aden. The comparison with the natural oblique rift confirms that
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this evolution is highly probable. The final pattern of the deformation, the distribution of the fault populations in the rift and their probable chronology are all compatible with the deformation steps.

As already proposed in other studies, we correlate these steps to the following deformation processes: (i) Interaction of the far-field stress and the local stresses induced by the weakness zone, which form intermediate faults. (ii) Buoyancy forces induced stress field that strongly localises the deformation and creates rift-parallel faults. (iii) A progressive return to far-field stress conditions as the thinning of the rift centre becomes important, creating intermediate and displacement-orthogonal faults. Our model suggests that the exhumation of lower crustal and mantle material could take place when intermediate faults are most likely to dominate the deformation pattern.

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24 25 26 27 28 29 30 31 **Table caption**

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34
35 **Table 1.** Model parameters. Dislocation creep parameters for upper crust: wet quartzite
36 (Gleason & Tullis 1995), lower crust: Pikwitonian granulite (Wilks & Carter 1990),
37 lithospheric mantle: dry olivine (Hirth & Kohlstedt 2003), asthenospheric mantle: wet olivine,
38 i.e. 500 ppm H/Si (Hirth & Kohlstedt 2003). Peierls creep parameters for mantle: (Kameyama
39 et al. 1999). *the friction coefficient decreases linearly by 90 % of the initial value when
40 plastic strain reaches 1, and remains constant for larger strains.
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50 **Figure captions**

51 **Figure 1**

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54 Model setup. (a) Extensional velocities are prescribed at the boundaries in x-direction.
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56 The angle of obliquity α is defined as the angular difference between extension velocity
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1 and rift normal. Periodic boundary conditions in y-direction realize an in principle
2 infinitely long rift zone. **(b)** A thermally weak zone initializes the rift by affecting the
3 lithospheric strength as shown in the yield strength profiles. **(c)** Yield strength profile of
4 the analogue model (Autin et al. 2010).
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9 **Figure 2**

10 Model evolution at 1 My, 6 My, 10 My, and 14 My (i.e. 10 km, 60 km, 100 km, and 140
11 km extension, respectively). **(a)** Initially, shear zones are parallel to the expected
12 intermediate azimuth of 95° . Localisation occurs towards the rift centre. **(b)** The central
13 shear zone part rotates successively to form sigmoidal deformation patterns. **(c)**
14 Topography shows rift shoulder uplift due to hot asthenospheric upwelling at 6 My
15 followed by subsidence due to lithospheric cooling. **(d)** Basin geometries is strongly
16 affected by shear zones. **(e)** Mid-model cross section shows successive localisation
17 towards the rift centre. Black lines indicate boundaries between material layers. Note that
18 figures showing the whole evolution in steps of 1 My can be found in the supplementary
19 materials.
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31 **Figure 3**

32 Stress-inferred fault evolution at 1 My, 6 My, 10 My, and 14 My. **(a)** Normal faulting is
33 the dominant mechanism except for a temporary strike-slip region in the rift center. **(b,c)**
34 Normal fault orientations are intermediate at the beginning, rotate towards rift-parallel at
35 the rift flanks until 10 My where after they show intermediate and extension-normal
36 orientation at break-up. Note that figures showing the whole evolution in steps of 1 My
37 can be found in the supplementary materials.
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47 **Figure 4**

48 Illustration how long-lived shear zones generate sigmoidal strain patterns (Compare to
49 Fig 2b). Note that the azimuth of final shear zone pattern depends on the distance from
50 the continent-ocean boundary.
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56 **Figure 5**

57 **(a)** Structural map of the Gulf of Aden with the main tertiary depocentres as well as the
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mesozoic inherited basins (after Bellahsen et al., this volume and Leroy et al, 2012). **(b)**
Reconstruction of the margins at the onset of the ocean-continent transition (OCT) based
on Leroy et al., 2012. SSFZ: Shukra El Sheik Fracture Zone, KAFZ: Khanshir Al Irquah
Fracture Zone, AFFZ: Alula-Fartak Fracture Zone.

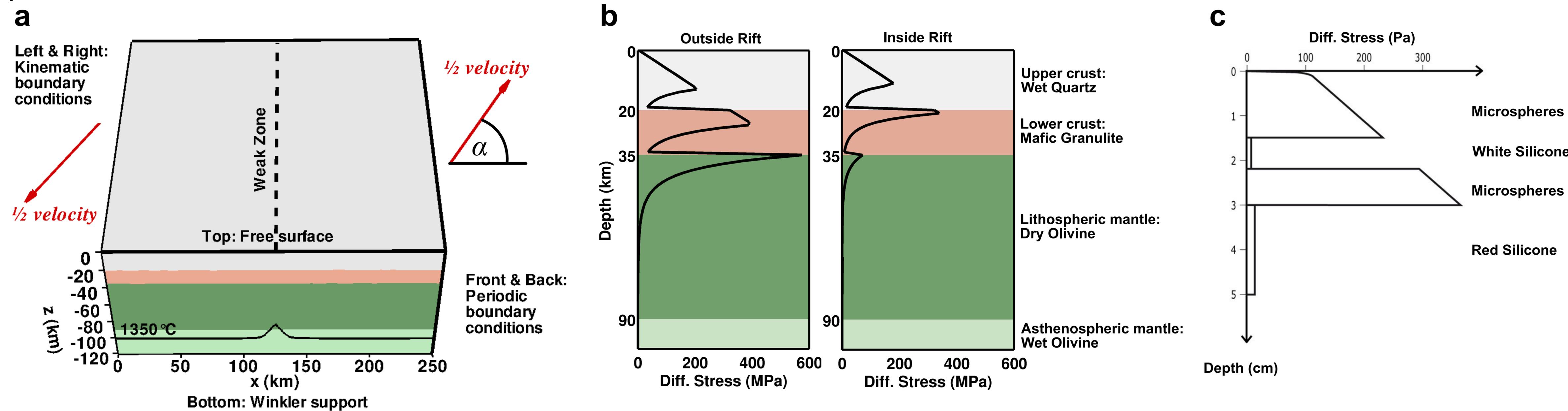
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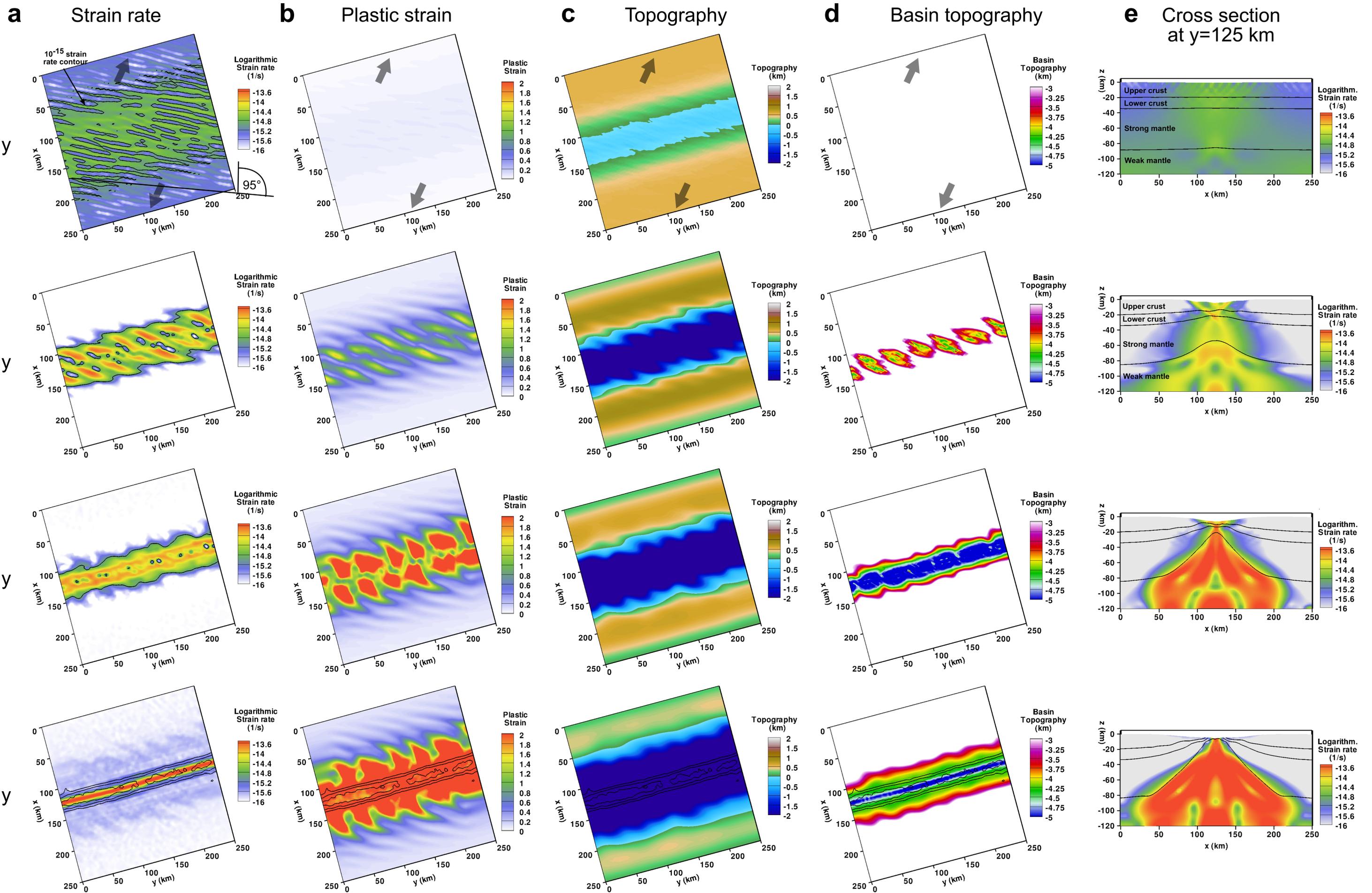
Table 1

[Click here to download Table: Table1.pdf](#)

Parameter	Upper Crust	Lower Crust	Strong Mantle	Weak Mantle
Density, ρ (kg m ⁻³)	2700	2850	3300	3300
Thermal expansivity, α_T (10 ⁻⁵ K ⁻¹)	2.7	2.7	3.0	3.0
Bulk modulus, K (GPa)	55	63	122	122
Shear modulus, G (GPa)	36	40	74	74
Heat capacity, C_p (J kg ⁻¹ K ⁻¹)	1200	1200	1200	1200
Heat conductivity, λ (W K ⁻¹ m ⁻¹)	2.5	2.5	3.3	3.3
Radiogenic heat production, A (μ W m ⁻³)	1.5	0.2	0	0
Initial friction coefficient, μ (-)	0.6	0.6	0.6	0.6
Maximum plastic friction softening*	90 %		none	none
Cohesion, c (MPa)	5.0	5.0	5.0	5.0
Pre-exponential constant for diffusion creep, $\log(B_{Diff})$ (Pa ⁻¹ s ⁻¹)	-	-	-8.65	-8,65
Activation energy for diffusion creep, E_{Diff} (kJ / mol)	-	-	375	335
Activation volume for diffusion creep, V_{diff} (cm ⁻³ / mol)	-	-	6	4
Pre-exponential constant for dislocation creep, $\log(B_{Disloc})$ (Pa ⁻ⁿ s ⁻¹)	-28.0	-21.05	-15.56	-15.05
Power law exponent for dislocation creep, n	4.0	4.2	3.5	3.5
Activation energy for dislocation creep, E_{Disloc} (kJ / mol)	223	445	530	480
Activation volume for dislocation creep, V_{Disloc} (cm ⁻³)	0	0	13	10
Pre-exponential constant for Peierls creep, $\log(B_{Peierls})$ (Pa ⁻ⁿ s ⁻¹)	-	-	11.76	-
Activation energy for Peierls creep, $E_{Peierls}$ (kJ / mol)	-	-	540	-
Peierls stress, $\tau_{Peierls}$ (GPa)	-	-	8.5	-

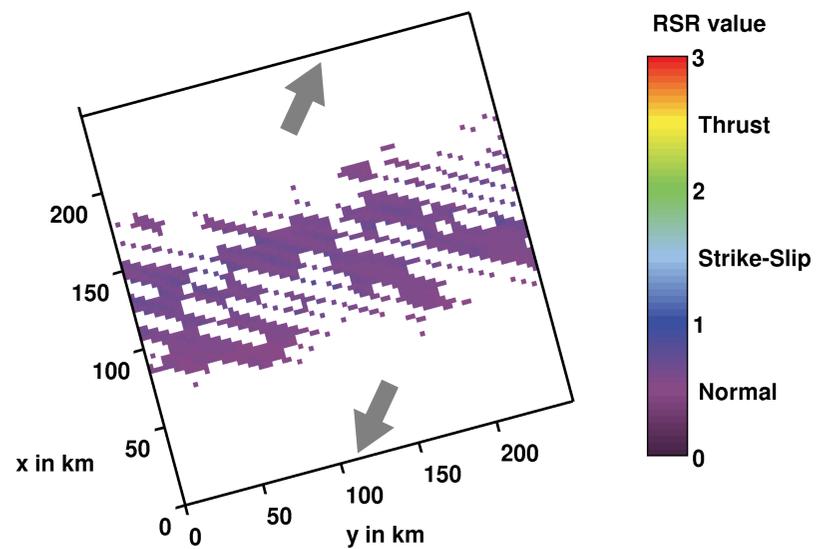
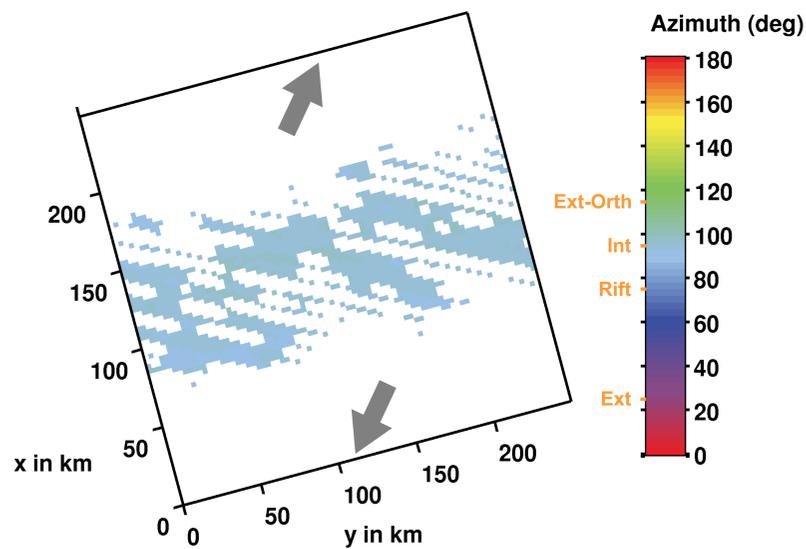
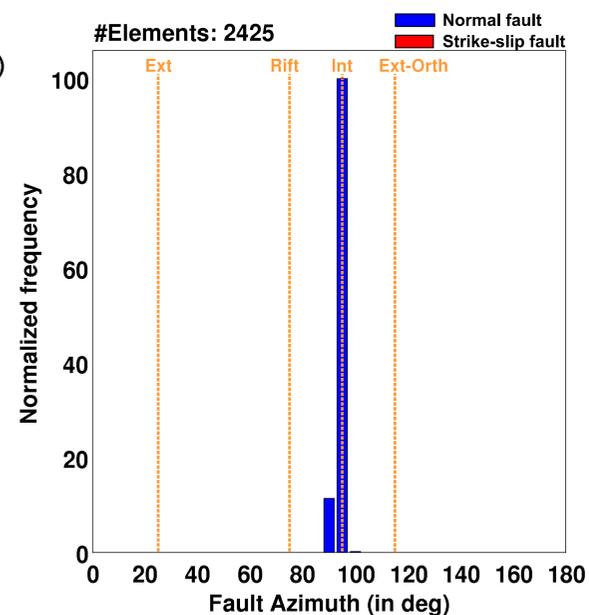
Figure 1



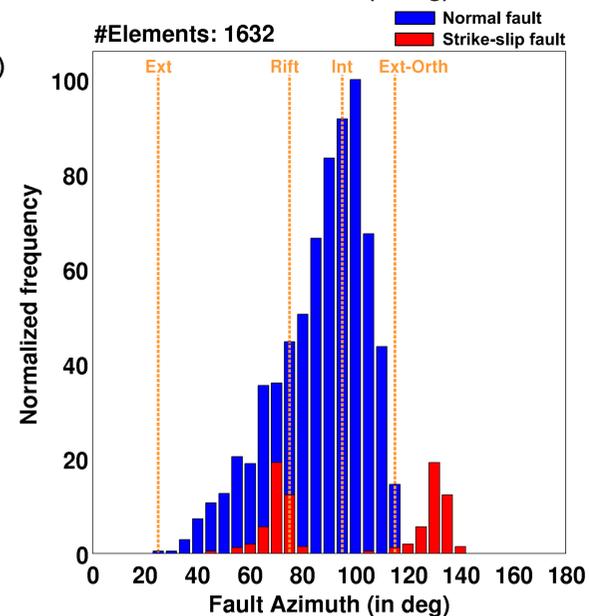
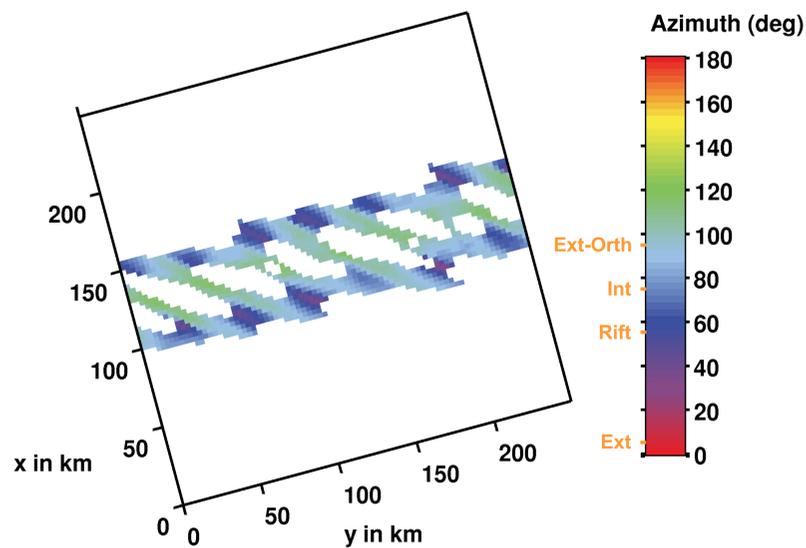
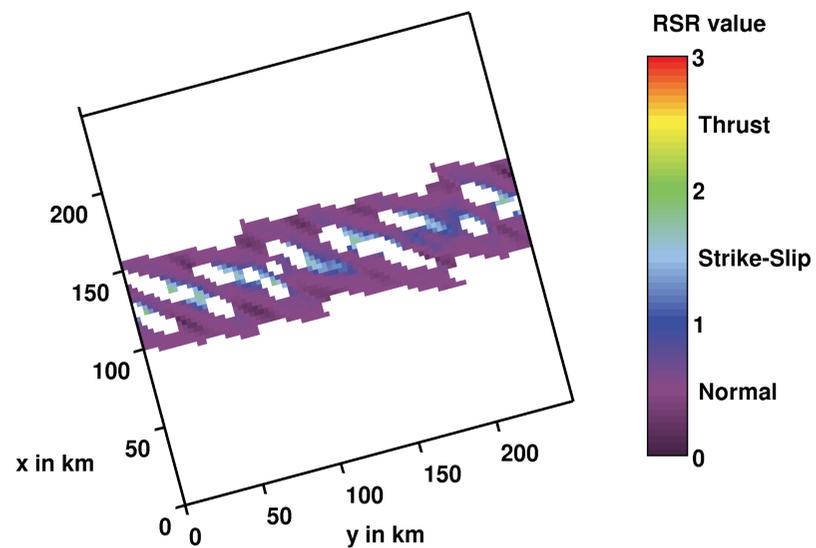


a Stress-inferred fault mechanism

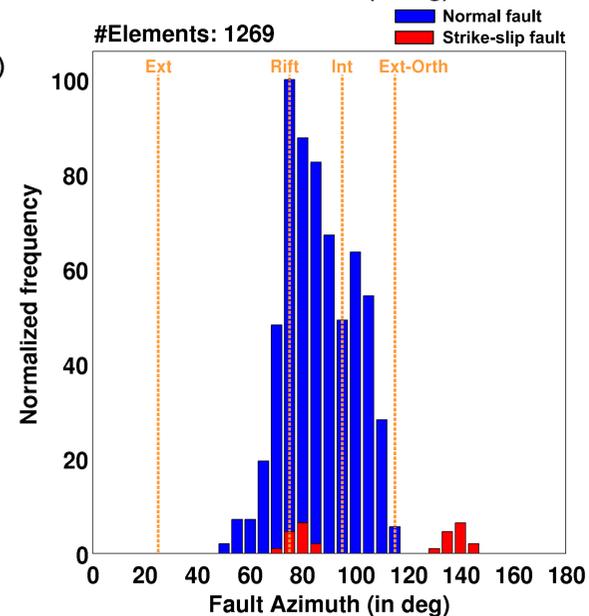
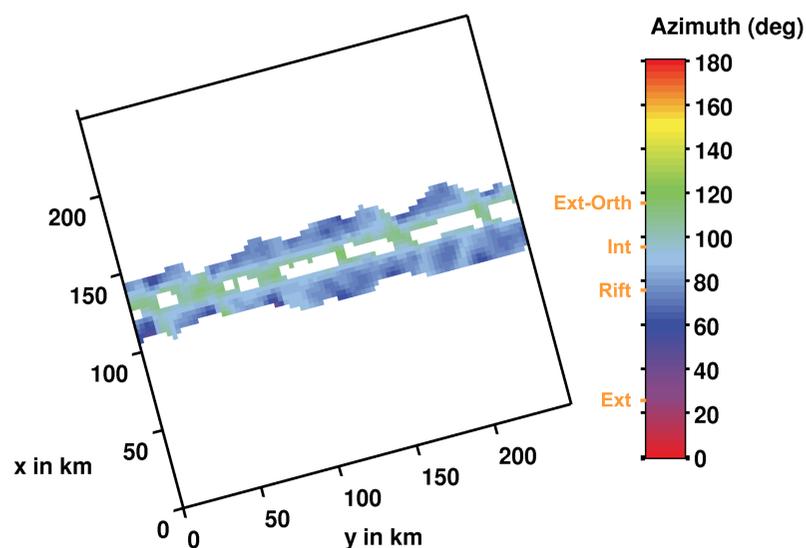
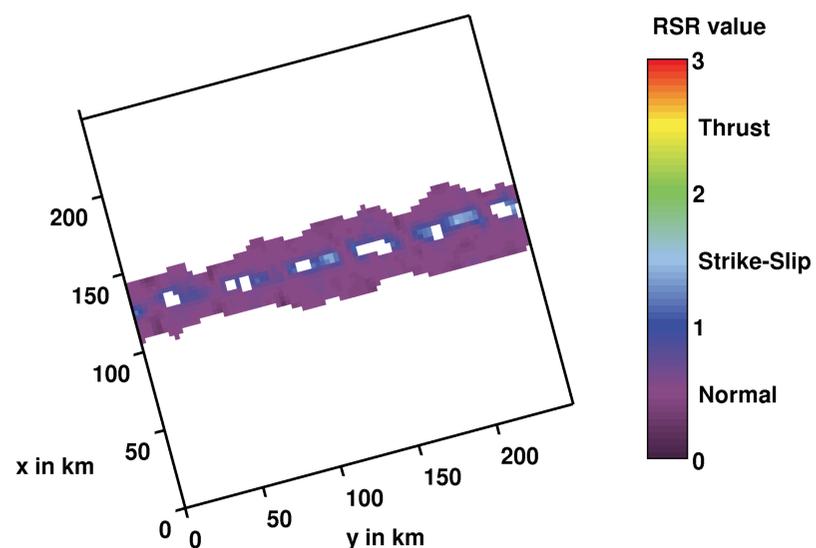
1 My

**b** Stress-inferred normal fault azimuth**c** Stress-inferred azimuth diagram

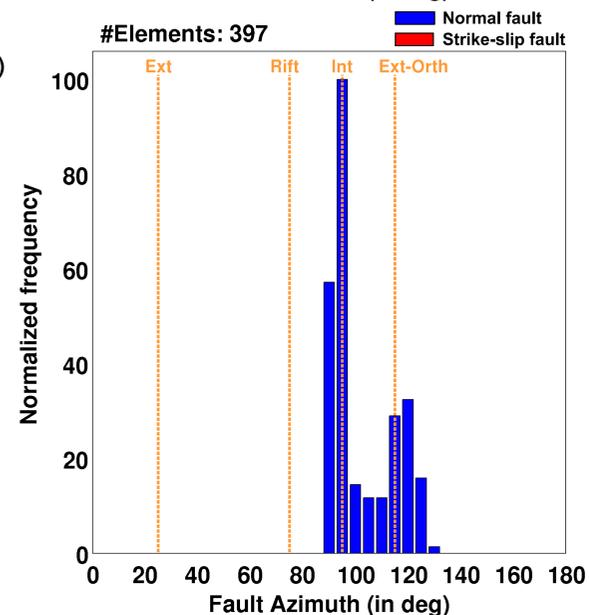
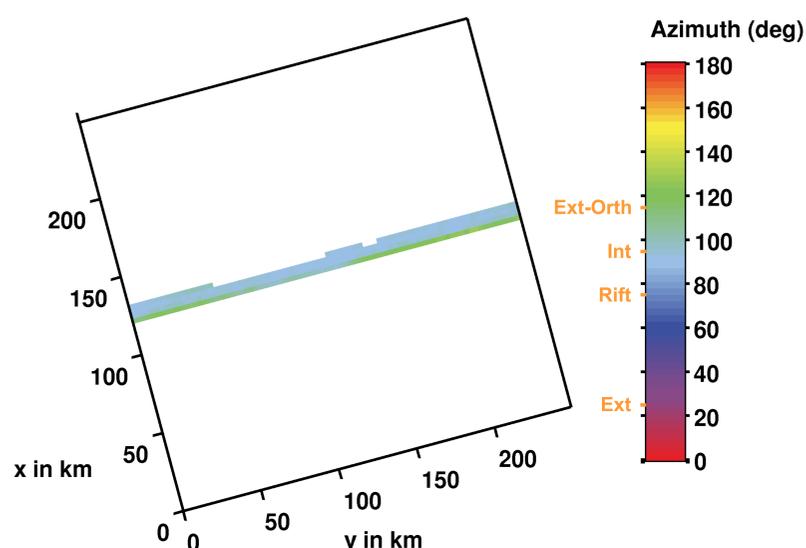
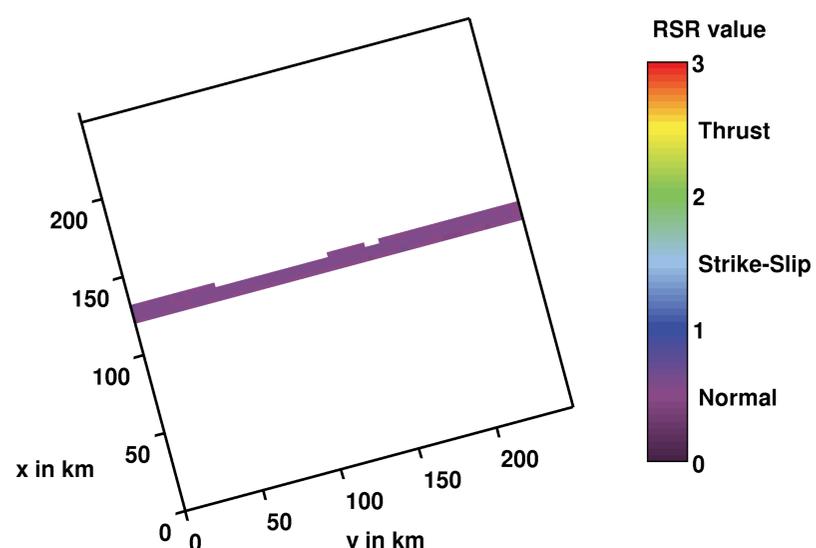
6 My



10 My



14 My



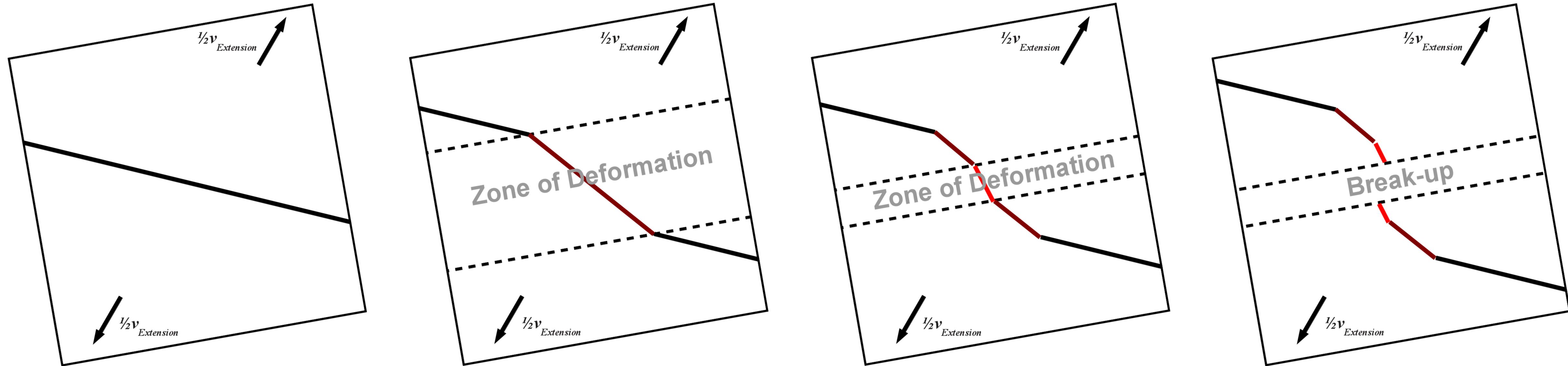
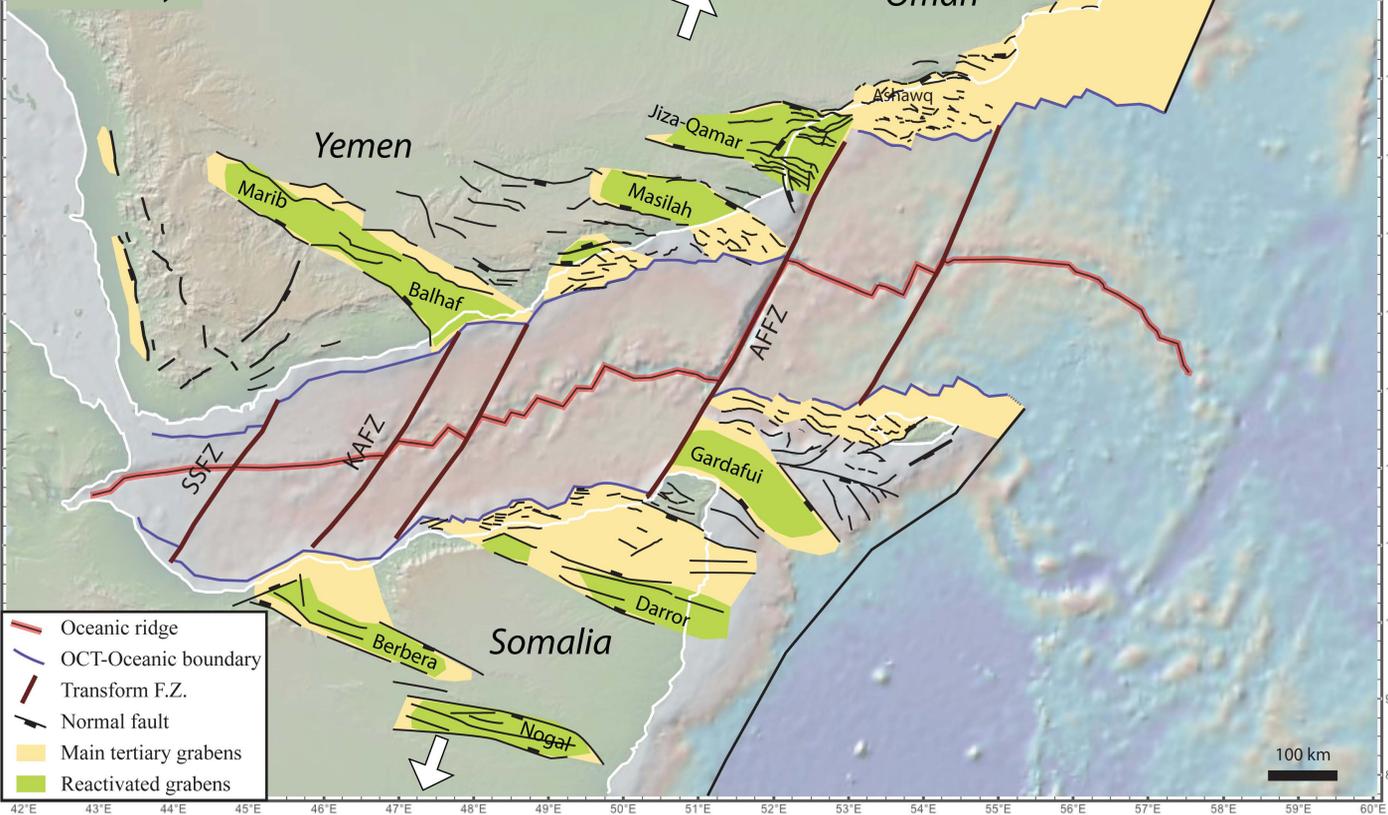


Figure 5

a) Present day



b) 20 Ma (OCT onset)

