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Tectonics

RESEARCH ARTICLE

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Key Points:

- Combined analog and numerical geodynamic modeling of the Turkana region elucidates the impact of tectonic inheritance on rift architecture
- The anomalously wide Turkana rift zone is due to a transversal region of thinned crust resulting from Mesozoic rifting
- Kenyan and Ethiopian rifts avoid a straight link through the Turkana region due to inherited mantle lithospheric strength variations

Supporting Information:

- Supporting Information S1
- Movie S1
- Movie S2
- Movie S3
- Movie S4
- Movie S5
- Movie S6

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Controls of inherited lithospheric heterogeneity on rift linkage: Numerical and analog models of interaction between the Kenyan and Ethiopian rifts across the Turkana depression

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Abstract Inherited rheological structures in the lithosphere are expected to have large impact on the architecture of continental rifts. The Turkana depression in the East African Rift connects the Main Ethiopian Rift to the north with the Kenya rift in the south. This region is characterized by a NW-SE trending band of thinned crust inherited from a Mesozoic rifting event, which is cutting the present-day N-S rift trend at high angle. In striking contrast to the narrow rifts in Ethiopia and Kenya, extension in the Turkana region is accommodated in subparallel deformation domains that are laterally distributed over several hundred kilometers. We present both analog experiments and numerical models that reproduce the along-axis transition from narrow rifting in Ethiopia and Kenya to a distributed deformation within the Turkana depression. Similarly to natural observations, our models show that the Ethiopian and Kenyan rifts bend away from each other within the Turkana region, thus forming a right-lateral step over and avoiding a direct link to form a continuous N-S depression. The models reveal five potential types of rift linkage across the preexisting basin: three types where rifts bend away from the inherited structure connecting via a (1) wide or (2) narrow rift or by (3) forming a rotating microplate, (4) a type where rifts bend towards it, and (5) straight rift linkage. The fact that linkage type 1 is realized in the Turkana region provides new insights on the rheological configuration of the Mesozoic rift system at the onset of the recent rift episode.

Plain Language Summary The Turkana depression in the Kenya/Ethiopia borderland is most famous for its several million years old human fossils, but it also holds a rich geological history of continental separation. The Turkana region is a lowland located between the East African and Ethiopian domes because its crust and mantle have been stretched in a continent-wide rift event during Cretaceous times about 140–120 Ma ago. This thin lithosphere exerted paramount control on the dynamics of East African rifting in this area, which commenced around 15 Ma ago and persists until today. Combining analog "sandbox" experiments with numerical geodynamic modeling, we find that inherited rift structures explain the dramatic change in rift style from deep, narrow rift basins north and south of the Turkana area to wide, distributed deformation within the Turkana depression. The failed Cretaceous rift is also responsible for the eastward bend of the Ethiopian rift and the westward bend of the Kenyan rift when entering the Turkana depression, which generated the characteristic hook-shaped form of present-day Lake Turkana. Combining two independent modeling techniques - analog and numerical experiments - is a very promising approach allowing to draw robust conclusions about the processes that shape the surface of our planet.

1. Introduction

The evolution and architecture of continental rifts is controlled by the interaction among several parameters, including the rate of plate divergence, the thermal state of the lithosphere, and the presence and volumes of magmatic products [e.g., Ziegler and Cloetingh, 2004; Brune, 2016]. However, almost all continental rifts form in predeformed, and thus already structured, anisotropic lithosphere and reactivate preexisting weak zones, such as mobile belts, while avoiding stronger regions, such as cratons [e.g., Dunbar and Sawyer, 1988; Versfelt and Rosendahl, 1989]. The prerift lithospheric rheological structure and in particular its along-axis variations play one of the most important roles during continental extension.

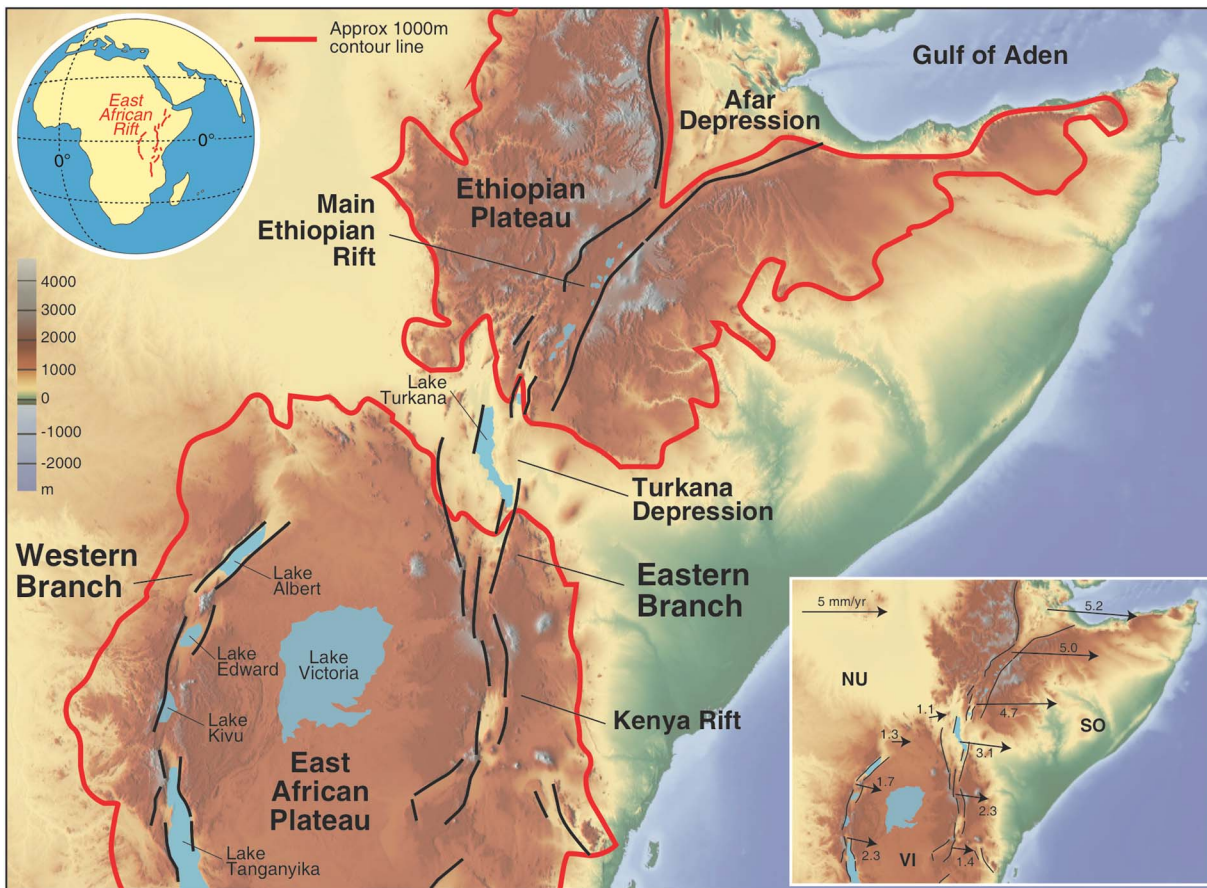


Figure 1. Tectonic setting of the East African Rift. Schematic fault pattern and present-day plate kinematics (inset). Black arrows show relative motions with respect to a stable Nubian reference frame according to the best fit model of *Saria et al.* [2014]. Values besides arrows indicate motion in mm/yr. NU: Nubian plate; SO: Somalian plate; VI: Victoria microplate.

The East African Rift (EAR) is a classical example where rifts are guided by inherited structures. The rift system developed within a region that has experienced several deformation events, from different phases of collision during the Precambrian to Mesozoic extension [Chorowicz, 2005]. This complex tectonic history has given rise to significant lateral variations in the rheological structure of the lithosphere [Ebinger et al., 1997], which in turn have played a major role in rift evolution. This is clearly testified by the localization and propagation of major rift segments within weak Proterozoic mobile belts, which surround the undeformed Tanzania Craton [Chorowicz, 2005, and references therein]. Linking and mechanical interaction between adjacent rift segments has typically occurred in correspondence to transverse preexisting fabrics, where structurally complex areas (transfer zones) have allowed significant along-axis variations in subsidence of grabens and elevation of uplifted flanks [Rosendahl, 1987; Morley et al., 1990; Faulds and Varga, 1998; Morley, 1999].

One of these complex areas is the Turkana depression, the region of interaction between the Ethiopian and Kenyan rifts, which is characterized by anomalous morphology and architecture with respect to the rift valleys in Kenya and Ethiopia. In this paper we test the hypothesis that these anomalies result from the presence of a preexisting rift, transverse to the trend of the rift valleys, and characterized by thin crust and lithosphere. To this aim, we complement crustal-scale centrifuge analog models with lithospheric-scale numerical modeling.

2. Tectonic Setting

The Turkana depression is a lowland located between the uplifted East African and Ethiopian domes (Figure 1). The tectonic setting of this area is controlled by long-term divergence between the Nubian and Somalian plates, which is currently occurring in a roughly E-W direction at rates of a few

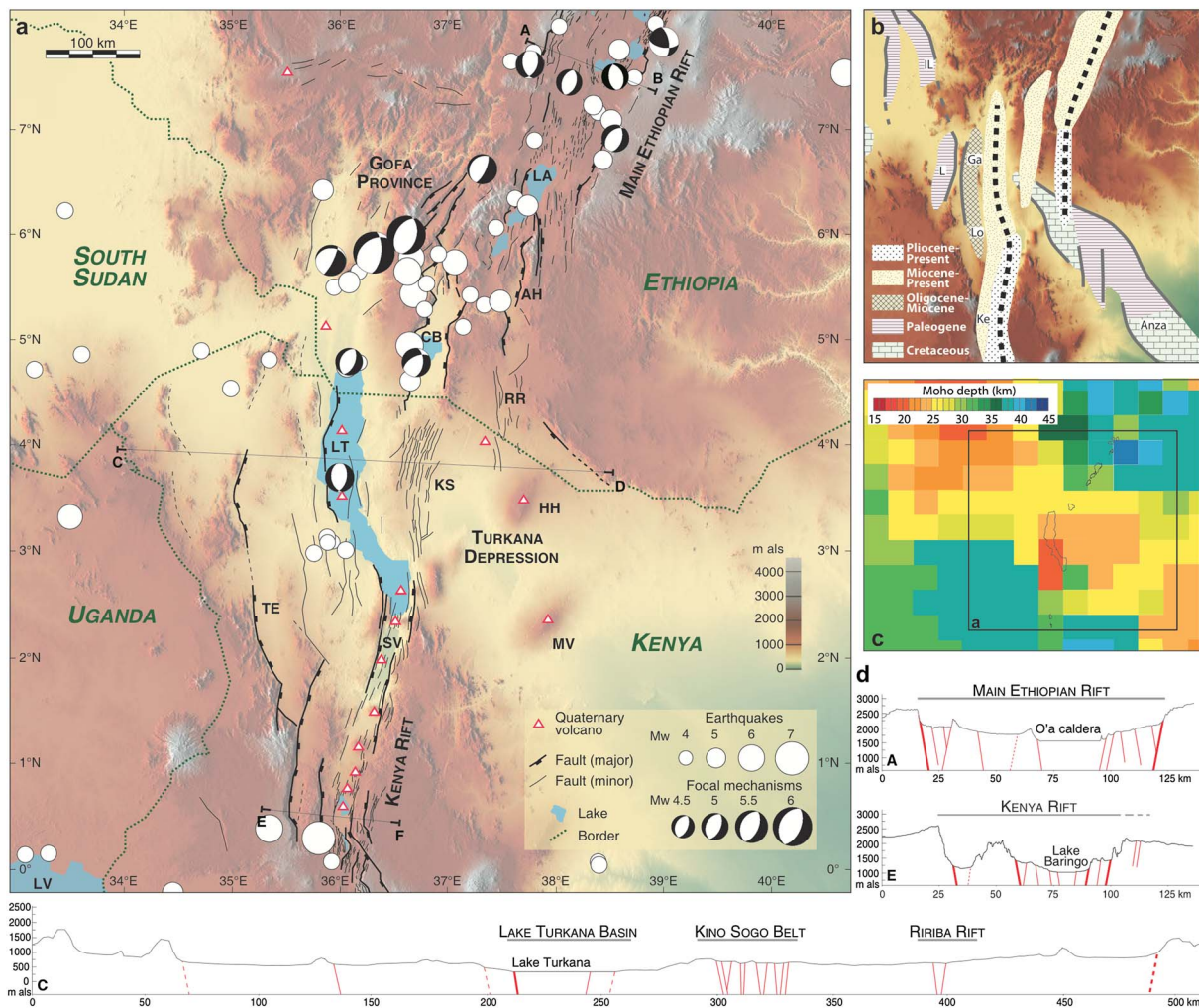


Figure 2. Detailed tectonics of the study area. (a) Quaternary faults, seismicity, and Quaternary volcanoes in the Turkana depression and surrounding regions superimposed on a SRTM (Nasa Shuttle Radar Topography Mission) digital elevation model. Seismicity from U.S. Geological Survey National Earthquake Information Center catalog (1900 to present); focal mechanisms from the global centroid moment tensor project; volcanoes from the Smithsonian Institution, Global Volcanism Program database. Fault pattern modified from Moore and Davidson [1978], Ebinger et al. [2000], Hautot et al. [2000], Morley [2002], Vetel et al. [2005], Melnick et al. [2012], and Philippon et al. [2014]. AH: Amaro Horst; CB: Chew Bahir basin; HH: Hurri Hills; KS: Kino Sogo belt; LA: Lake Abaya; LT: Lake Turkana; LV: Lake Victoria; MV: Marsabit volcano; RR: Ririba Rift; SV: Suguta Valley; TE: Turkwell Escarpment. (b) Ages of basins within the Turkana depression and simplified patterns of interaction between the Kenyan and Ethiopian rifts (black dashed lines indicate the rift axis). Basins are labeled as follows: Ga, Gatome; IL, Ilubabor; Ke, Kerio; L, Lokitipi; and Lo, Lokichar. Dark gray lines represent Cretaceous and Paleogene faults. (c) Map of crustal thickness in the region [from Benoit et al., 2006]. (d) Simplified cross sections highlighting the different style and distribution of deformation in the Turkana depression with respect to the Kenyan and Ethiopian rifts.

millimeters per year [Saria et al., 2014; Stamps et al., 2014]. Recent plate motion models [Jaffaldano et al., 2014; DeMets and Merkouriev, 2016] suggest that this direction of extension has likely remained steady during rift development (i.e., in the last 10–12 Myr).

The Main Ethiopian Rift (MER) and the Kenya rift are part of the Eastern Branch of the EAR (Figure 1) [Chorowicz, 2005]. North of the Turkana region, in the MER and southern portion of the Afar depression, extensional deformation is characterized by a typical narrow rift morphology exhibiting large fault escarpments with vertical displacement of >1 km, which bound a roughly 80–100 km wide rift valley and separate two uplifted plateaus [Mohr, 1983]. South of the Turkana depression, deformation in the Kenya rift is, similarly to the MER, characterized by large marginal fault systems, which separate a narrow rift valley (about 60–80 km in width) from surrounding uplifted plateaus. The MER and Kenya rift interact within the Turkana depression, which is characterized by a style of deformation where faulting, seismicity, and Quaternary-Holocene volcanic centers are widespread over a width of more than 300 km (Figure 2). This testifies a wide region of ongoing rift-related tectono-magmatic activity, which is in striking contrast with the two narrow

rifts to the north and south. Similarly, the Turkana region lacks the typical rift valley morphology dominated by large fault escarpments. Instead, extension is accommodated by numerous faults with limited vertical offset clustered within different subparallel deformation domains (Figures 2a and 2d) [see also *Ebinger et al.*, 2000], where the main domains are (i) the Turkana Basin, (ii) the Kino Sogo fault belt, and (iii) the Ririba Rift. The Turkana Basin corresponds to the northwestward propagation of the Kenya rift via the Suguta Valley, whereas the Ririba Rift (and the Hurri Hills and Marsabit volcanic lineaments) is believed to represent the southeastward propagation of the Ethiopian rift [*Ebinger et al.*, 2000; *Bonini et al.*, 2005]. The Kino Sogo belt, a 30 km wide system of small horsts and grabens located between the Turkana Basin and the Ririba Rift, accommodates a minor part of the extensional deformation [*Vetel et al.*, 2005]. To the north, this fault system links with the Chew Bahir basin, which is part of the complex Broadly Rifted Zone of South Ethiopia [*Ebinger et al.*, 2000; *Vetel and Le Gall*, 2006].

The deformation events that preceded the main Cenozoic rift phase have been suggested to have largely controlled the structural pattern and the physiography of the area, both at a regional scale [*Ebinger et al.*, 2000; *Benoit et al.*, 2006] and at the scale of individual faults and fault arrays [*Vetel et al.*, 2005; *Vetel and Le Gall*, 2006]. In particular, the area has been affected by a significant Mesozoic rift event, which created a system of NW-SE grabens (such as the Anza graben in Ethiopia) at a high angle to the Cenozoic Ethiopian/Kenyan rift valleys (Figure 2b). Geophysical data provide evidence for significant crustal thinning beneath the Turkana depression, where a crustal thickness <25 km has been imaged (Figure 2c) [*Woldetinsae and Götze*, 2005; *Benoit et al.*, 2006; *Sippel et al.*, 2017]. The inherited thin crust related to this Mesozoic event was suggested to cause the low topography of the region with respect to the surrounding uplifted plateaus [*Benoit et al.*, 2006]. As a consequence, the Turkana depression is one of the few places with elevations below 1 km that belongs to the region of anomalously elevated topography characterizing eastern and southern Africa.

The complex, pre-Miocene tectonic history within the Turkana depression is testified by the continuation of sedimentation in the Anza graben during the Paleogene and by the roughly coeval development of N-S trending sedimentary basins (e.g., Lotikipi and Lokichar) west of Lake Turkana (Figure 2b) [*Morley et al.*, 1992; *Morley*, 1999; *Ebinger et al.*, 2000]. These basins started developing during the Oligocene-early Miocene [*Morley et al.*, 1992; *Hendrie et al.*, 1994; *Morley*, 1999] before the beginning of Nubia-Somalia divergence and after the earliest manifestation of flood magmatism at ~45 Ma in the Broadly Rifted Zone of South Ethiopia [*Davidson and Rex*, 1980; *George et al.*, 1998]. During early Miocene times, extension started in the Chew Bahir basin, contemporaneous with a first, minor phase of extensional deformation in the Southern MER [*Bonini et al.*, 2005; *Pik et al.*, 2008]. Basin development began in the Lake Turkana region at around 15 Ma, followed by activation of large boundary faults at around 12 Ma in the Broadly Rifted Zone of South Ethiopia and at 8–10 Ma in the Amaro Horst region of the Southern MER [*Balestrieri et al.*, 2016]. During the Pliocene-Quaternary, the volcano-tectonic activity focused in the Lake Turkana Basin, now a direct continuation of the Kenya rift, whereas the extensional deformation related to the MER propagated southward into the Eastern Turkana depression (Ririba Rift, Hurri Hills, and Marsabit volcanic lineaments) [*Bonini et al.*, 2005].

Overall, a progressive eastward migration of Oligocene-recent deformation within the Turkana depression has been suggested [*Ebinger et al.*, 2000, and references therein]. This migration, together with a northward propagation of rifting in the Kenya rift and a southward propagation of the Main Ethiopian Rift to the Ririba and Marsabit volcanic lineaments, is considered responsible for the anomalous breadth of the zone of deformation in the region [*Ebinger et al.*, 2000]. Strain hardening between episodes of extension [*Morley et al.*, 1992] or deep-seated asthenospheric processes such as plume impingement at the base of the lithosphere [*Ebinger et al.*, 2000] has been considered responsible for the progressive eastward migration in the Turkana depression.

3. Initial and Boundary Conditions for Modeling

We conduct both analog and numerical experiments in order to investigate the interaction between the Ethiopian and Kenyan rifts across the Turkana depression. These modeling approaches complement each other since analog models are powerful in analyzing brittle deformation and the surface fault pattern at a very high resolution (here: ~800 m) but are limited in that they normally cannot account for complex

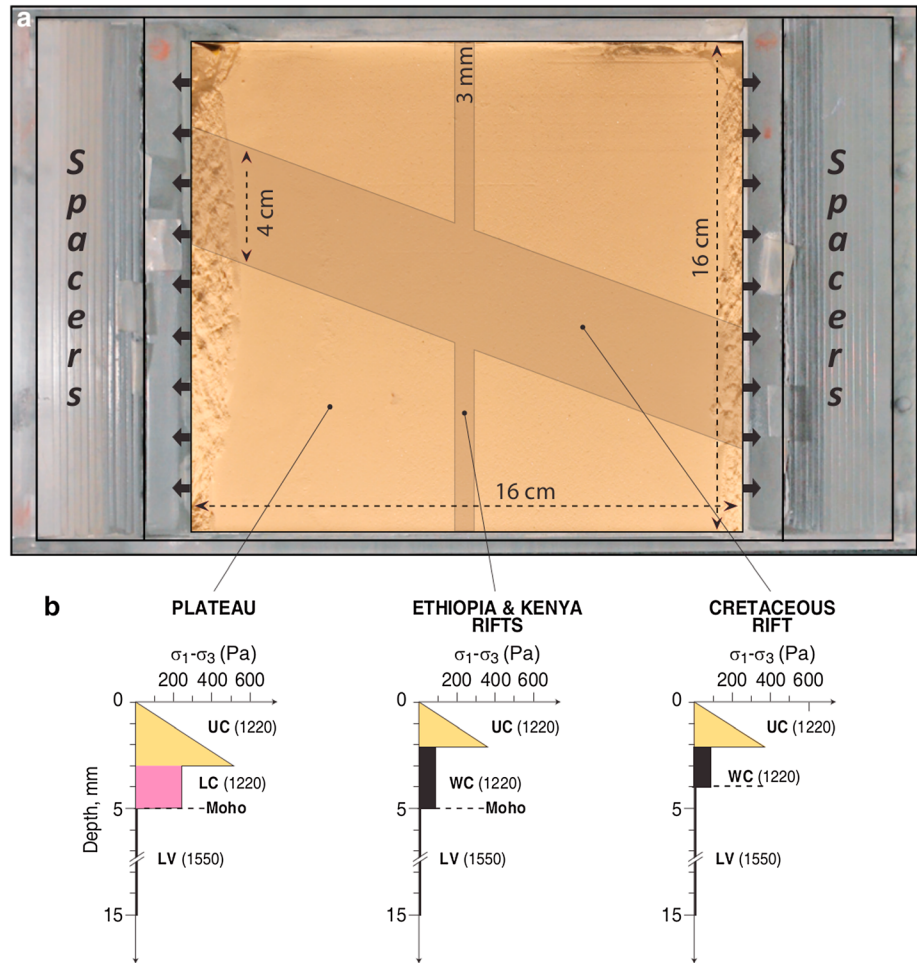


Figure 3. Analog modeling setup. (a) Top view of the experimental apparatus (see text for details). (b) Strength profiles in the different model domains. LC: lower crust; LV: low-viscosity mixture providing isostatic support to the overlying crust; UC: upper crust; WC: weak lower crust.

rheologies and thermal variations during deformation and related processes. These complexities are instead well reproduced and analyzed in numerical models, which, however, are more limited in model resolution (here: 7 km). In the following we outline the general model framework, while the specific setup for analog and numerical models is discussed in sections 4 and 5, respectively.

We capture the first-order regional plate tectonic setup by subdividing the region into three different domains, namely, Turkana depression, Kenya/Ethiopia rifts, and plateaus outside of any rift (Figure 3). The main starting point for our modeling is the presence of thin crust beneath the NW-SE trending domain of Mesozoic extension (section 2). We set the initial crustal thickness in this area to 30 km, which represents the initial Moho depth before the onset of Neogene rifting. This value corresponds to the thickness of crystalline crust within parts of the Anza graben that have not been reactivated (Figure 2c) [see also *Benoit et al., 2006; Sippel et al., 2017*]. The stable plateaus and the rift valleys are characterized instead by an initial thickness of about 40 km, which is the average Moho depth of today's plateaus in both domains [*Keranen et al., 2009; Sippel et al., 2017*]. For the thickness of the felsic upper crustal layer at the onset of Neogene rifting we use a value of 20 km. This is justified through Kenyan and Ethiopian seismic refraction surveys that exhibit v_p larger than 6.5 km/s at depths of 20 km alongside the main grabens [*Keller et al., 1994; Maguire et al., 2006*]. These velocities further indicate that the lower crust is composed of mafic lithologies [*Christensen and Mooney, 1995*]. We have also assumed a strength contrast between plateaus and Turkana depression and Kenyan/Ethiopian rift valleys. Both domains have been likely influenced by channeling and ponding of plume material in the Oligocene/Miocene

[Ebinger and Sleep, 1998; Ebinger et al., 2000], resulting in an elevated lithosphere-asthenosphere boundary (LAB) with respect to the surrounding plateaus. In the numerical models this is represented directly through different thermal LAB depths in each domain, while in the analog model we implement the strength contrast through different materials and layer thicknesses.

According to recent geodetical estimates [Saria et al., 2014; Stamps et al., 2014], extension in the models has been considered to be orthogonal to the Ethiopian/Kenyan rift valleys, i.e., roughly E-W (Figure 1). Note that our models are focused on the Eastern Branch of the EAR and do not consider the transfer of deformation from the latter to the Western Branch. The amount of total extension, the velocity of plate motion in the past, and the timing of initiation of Nubia-Somalia motion are not well constrained, as different plate motion models predict different values. Plate tectonic models that integrate seafloor spreading reconstructions into plate circuits indicate that divergence between the major plates started before 16 Ma [DeMets and Merkouriev, 2016], likely at ~20 Ma [Jaffaldano et al., 2014], with maximum values of extension of approximately 80 km [Jaffaldano et al., 2014]. Geological data, however, seem to point to a later rift activation (at 10–12 Ma) [see Balestrieri et al., 2016, and references therein] and lower amount of total extension (~30–40 km) [Corti, 2009, and references therein]. In this study we incorporated 15 Myr of symmetric extension at rates of ~4 mm/yr, resulting in a total amount of extension of 60 km.

4. Analog Modeling

4.1. Model Setup

The analog models were performed in an artificial gravity field of ~18 g using the large-capacity centrifuge available at the Tectonic Modelling Laboratory of the Institute of Geosciences and Earth Resources (National Research Council of Italy) and the Department of Earth Sciences of the University of Florence. The models simulated a simplified rheological layering composed of brittle-ductile crust floating above a low-viscosity material (Figure 3). The use of the centrifuge technique to perform experiments is based on the principle that the centrifugal force plays the same role in the models as does the force of gravity in natural geologic processes [Ramberg, 1981]. In the centrifuge, the models thin vertically and expand laterally in response to the centrifugal body force field that act during a centrifuge run. More specifically, the models, with dimensions of $16 \times 16 \times 1.5$ cm, were built inside a rectangular Plexiglas box and were laterally confined by rectangular blocks (spacers) that were removed before running the model for a given time interval in the centrifuge (Figure 3a). During the centrifuge run, the models expanded to fill the empty space at their sides, simulating extensional tectonics in nature; sequential removal of spacers during successive runs in the centrifuge allowed controlling the amount and rate of extension. Top view photos and laser scans of the models were taken after the end of each centrifuge run in order to monitor the evolution of surface deformation. The models were frozen before taking a number of cross sections to study their internal geometry. This modeling technique has been extensively tested and employed in different settings [see, e.g., Ramberg, 1981; Corti et al., 2003; Corti, 2012].

4.2. Rheological Layering, Materials, and Geometry

The two-layer models are composed of an upper brittle layer simulating the upper crust that, as in several previous centrifuge experiments [e.g., Agostini et al., 2009; Corti, 2012], was modeled by using a K-feldspar powder characterized by a linear increase in strength with depth to reproduce natural brittle behavior (Figure 3b). The underlying ductile lower crust in the plateau domains was modeled with a mixture of plasticine (Pongo Fantasia modeling dough, distributed by FILA) and PDMS-Polydimethylsiloxane (Dow Corning, SGM36), with proportions of 100:80% in weight. The lower crust beneath the rift valleys and the Turkana domain was reproduced by using a similar weak mixture of silicone (Dow Corning DC3179), corundum sand, and oleic acid (100:20:5% in weight). The models involved a thinner crust beneath the Turkana depression with respect to the surrounding domains. Considering the absence of the underlying lithospheric mantle, this makes the Turkana depression slightly weaker than the Ethiopia/Kenya rifts. However, we tested other conditions in which the Turkana domain had the same strength or was slightly stronger than the rifts (see below and supporting information Figure S1). All these crustal layers rested on a low-viscosity mixture of DC3179, corundum sand, and oleic acid (100:80:15% in weight), which provided isostatic support to the extending, overlying crust (Figure 3).

The geometry of the models mimicked that of the natural prototype at a scale of $\sim 1.3 \times 10^{-7}$, such that 1 cm in the experiments corresponded to ~ 80 km in nature. This allowed modeling ~ 60 km of total extension. Dynamic-kinematic similarity of gravitational, viscous, and frictional stresses acting in the system [Ramberg, 1981] ensured that the velocity of lateral displacement in the models $\sim 2.5 \times 10^{-5}$ m/s scaled to natural values of ~ 3 – 5 mm/yr.

4.3. Analog Modeling Results

Upon extension, deformation localized within the rift domains, giving rise to major fault escarpments bounding a subsiding narrow depression (Figure 4). These major faults formed at the rift/plateau boundaries, whereas systems of internal normal faults with more limited vertical displacement affected the rift floor. In both Kenyan/Ethiopian rift domains, deformation was entirely localized within the narrow depressions, i.e., in an area, which was ~ 1.4 cm (~ 110 km) wide after 4 mm (~ 30 km) of bulk extension (Figure 4a). The Turkana domain was instead characterized by a much wider region affected by deformation (up to ~ 7 cm—i.e., ~ 550 km—in width). Within this region, extension was accommodated by a much higher number of normal faults characterized by a limited vertical displacement and a general N-S trend.

During progressive extension, this overall architecture did not change significantly. Owing to an increase in slip on the boundary fault systems, the rift depression progressively increased its subsidence; the width of the rift valley increased to ~ 1.8 cm (~ 140 km), indicating that extension was entirely accommodated within it. In contrast, deformation within the Turkana depression was still distributed over a wide region (~ 8 cm—i.e., ~ 640 km—in width) and accommodated by a large number of small-offset normal faults (Figure 4). These faults were locally characterized by a zigzag pattern and by a trend that—although generally N-S—displayed local deviations to a more NNW-SSE orientation. Reactivation of the NW-SE-trending boundary between the weak Turkana depression and the strong surrounding plateaus occurred locally, as observed at the NW termination of the Kenyan rift valley (Figure 4).

Observation of the fault pattern and the topography of the model surface at the end of deformation (Figure 4b) indicates that entering the Turkana domain, the region of maximum subsidence (i.e., highest fault activity) related to the Ethiopian domain propagated toward the SE, whereas that of the Kenyan domain propagated toward the NW. Instead of linking directly to form a throughgoing N-S depression, the rift valleys propagated away from each other within the Turkana depression giving rise to a right-lateral step over of interacting rifts.

In different models we have tested different rheologies for the Turkana domain, rendering it slightly stronger or characterized by a similar strength with respect to the Kenyan/Ethiopian rifts. The model results are at a first order similar to the one described above in terms of distribution and characteristics of deformation (see supporting information Figure S1), underlining the robustness of our results.

5. Numerical Modeling

In addition to the analog experiments presented in section 4, the numerical model captures also the thermal evolution of the system taking into account radiogenic heating of the crust and thermal equilibration of the lithosphere. Hence, the brittle layer thickness and thus the degree of crust-mantle coupling is a dynamically evolving component of the model.

5.1. Model Setup

In the following we apply the finite element code SLIM3D [Popov and Sobolev, 2008], which solves the conservation equations of momentum, energy, and mass using an Arbitrary Lagrangian-Eulerian formulation involving the particle-in-cell technique. The deformation of materials is accommodated via an elastovisco-plastic rheology, which self-consistently reproduces diverse deformation processes like faulting, flexure, and lower crustal flow. SLIM3D has been extensively benchmarked and applied to model lithospheric-scale processes in divergent [Brune *et al.*, 2012, 2013; Brune and Autin, 2013; Brune, 2014; Heine and Brune, 2014; Koopmann *et al.*, 2014; Clift *et al.*, 2015; Brune *et al.*, 2016, 2017], convergent [Quinteros *et al.*, 2010; Quinteros and Sobolev, 2012; Duesterhoeft *et al.*, 2014], and transform [Popov *et al.*, 2012; Brune, 2014] plate boundaries and in a centimeter-scale study of localization dynamics [Cyprych *et al.*, 2016].

The model has a free surface at the top boundary, while at the bottom boundary we apply an isostatic Winkler boundary condition, where inflow and outflow of material are accounted for during remeshing. We use a

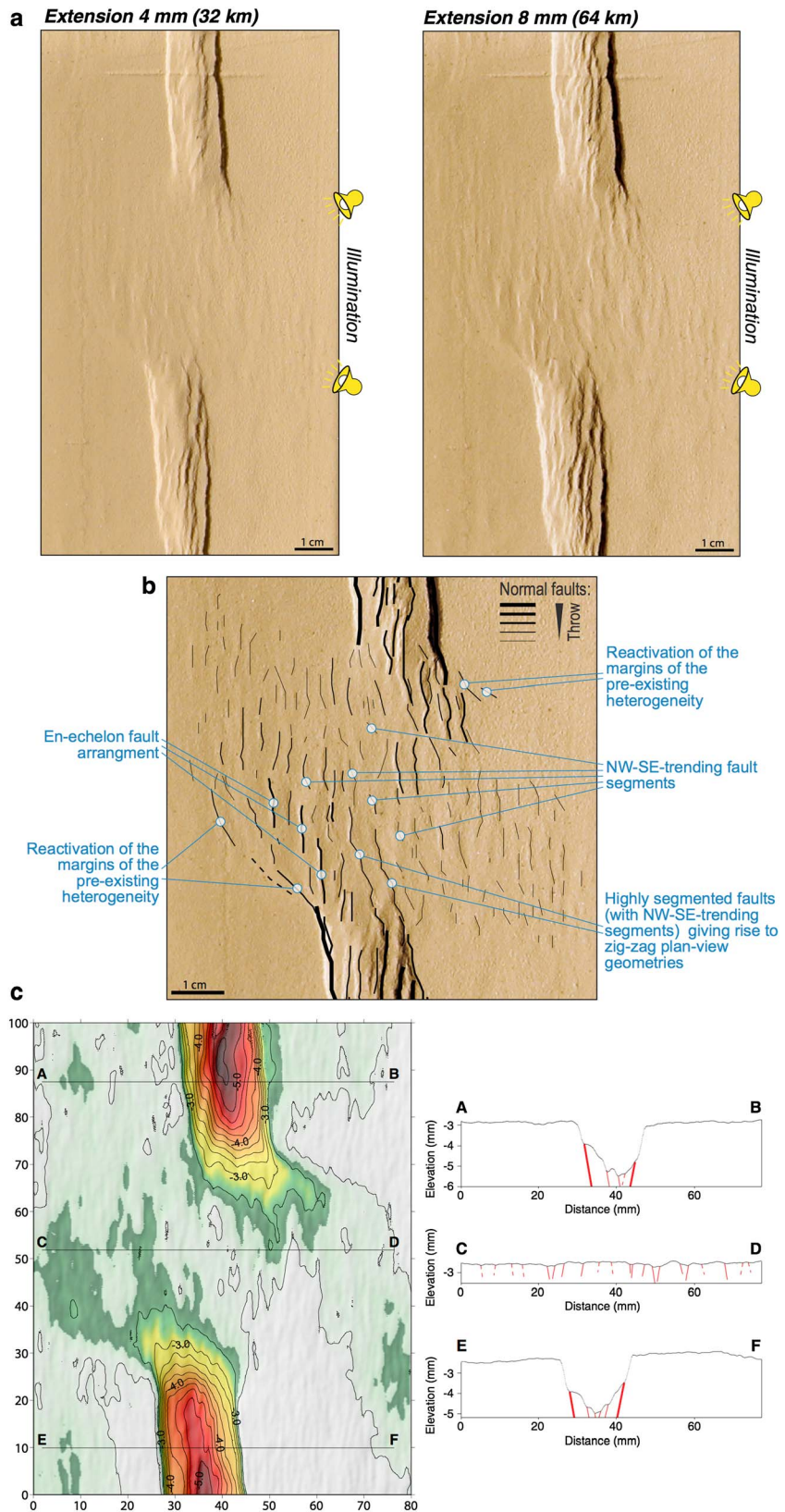


Figure 4. Analog modeling results. (a) Top view photos of the analog model for 4 mm and 8 mm bulk extension. (b) Line drawing of structures with examples of deformation features affecting the central part of the model. (c) Digital elevation model of the surface and topographic profiles at the end of deformation (8 mm bulk extension).

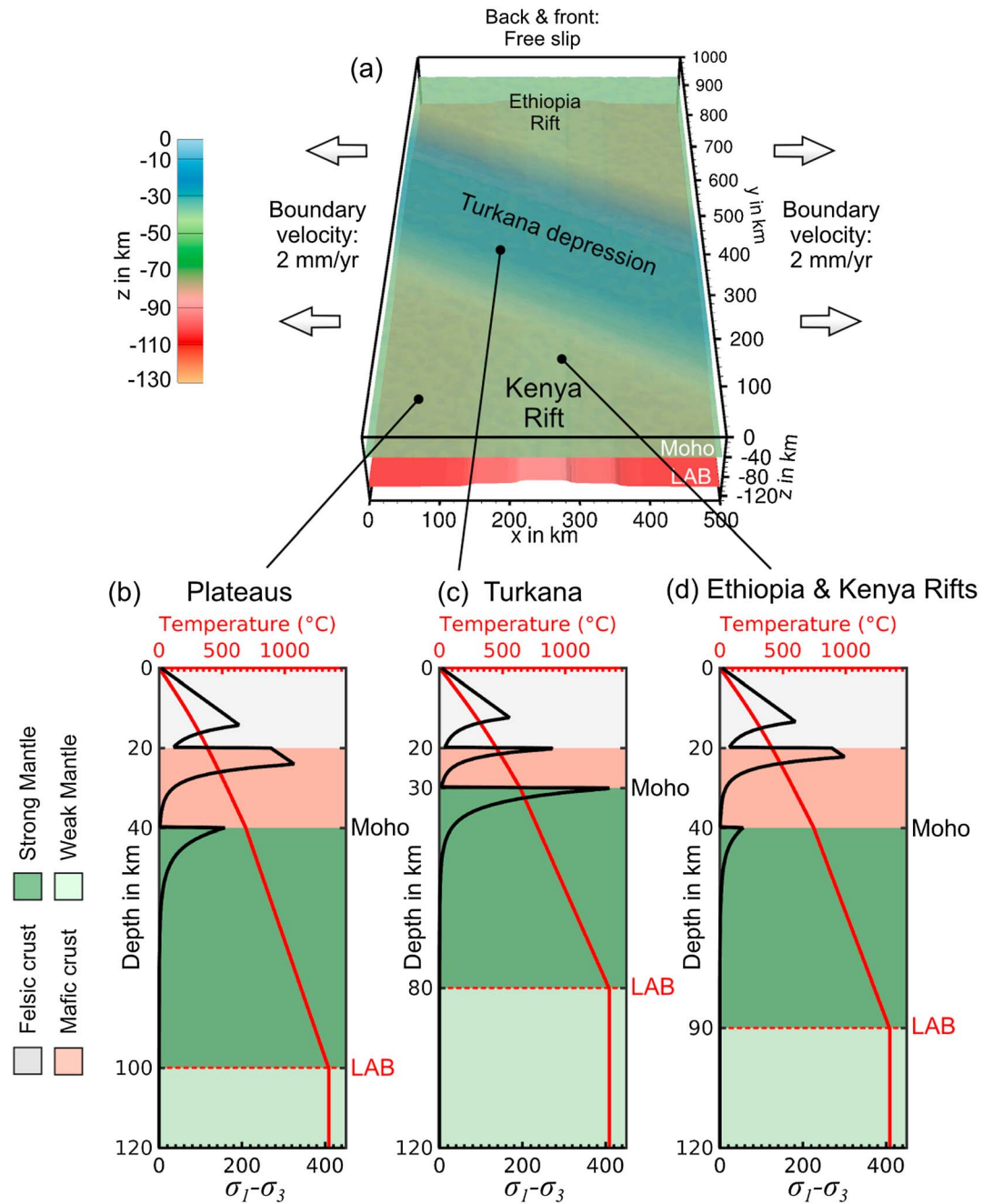


Figure 5. Numerical model setup. (a) Boundary conditions and domain locations, (b–d) Thermal profile and yield strength profiles computed for the initial bulk extension rate of the models, i.e., $2.7 \cdot 10^{-16}$ 1/s. For numerical parameters see supporting information Table S1.

constant rift velocity of 4 mm/yr (full rate) so that a 2 mm/yr velocity is prescribed orthogonal to both model boundaries facing in x direction (Figure 5). Front and back boundaries feature free slip. The numerical model comprises a domain of $500 \times 1000 \times 130$ km in x (cross rift), y (along strike), and z direction (depth) and a resolution of ~ 7 km in all directions.

Temperature boundary conditions are as follows: the surface temperature is held constant at 0°C and the bottom temperature at 1350°C . Lateral boundaries are thermally isolated. At the model start, the temperature field results from the thermal equilibrium governed by the boundary conditions, the crustal radiogenic heat contribution, and by the initial depth of the LAB, defined through the 1300°C potential temperature isotherm,

which locates at 100 km depth under the plateaus, 80 km depth beneath the Turkana depression, and 90 km depth beneath the Ethiopia and Kenya rifts (Figure 5) unless otherwise indicated. The slightly shallower LAB beneath the rifts is used to seed strain localization without imposing crustal fault locations. Note that due to radiogenic heat production, domains with thin crust generate less heat than those with thick crust. For simplicity, we focus on lithospheric processes and do not implement heterogeneities in the deeper parts of the mantle such as the Afar plume [Ebinger and Sleep, 1998; Hansen and Nyblade, 2013] or the northward shallowing of the LAB under East Africa [Fishwick and Bastow, 2011].

The model accounts for four material layers: felsic crust, mafic crust, strong mantle, and weak mantle (Figure 5b). For the quartz-dominated felsic upper crust, we use the wet quartzite flow law of Gleason and Tullis [1995]. The mafic lower crustal layer is represented by a wet anorthite flow law [Rybacki and Dresen, 2000]. We use dry olivine rheology [Hirth and Kohlstedt, 2003] to model deformation of the strong, depleted, subcontinental mantle, while a wet (i.e., 1000 ppm H/Si) olivine flow law [Hirth and Kohlstedt, 2003] is applied for the weak, asthenospheric mantle. All rheological parameters are listed in supporting information Table S1. The model involves linearized frictional strain softening, a standard strategy to parameterize fault weakening [Bos and Spiers, 2002] in numerical models [Huismans and Beaumont, 2011; Naliboff and Buitert, 2015; Le Pourhiet et al., 2017]. Here we linearly reduce the friction coefficient from 0.5 to 0.05 for brittle strain between 0 and 0.5. For strains larger than 0.5, it remains constant at 0.05. Similarly, we reproduce viscous strain softening [Bürgmann and Dresen, 2008] by increasing the preexponential factor of the ductile flow law by 10 times between viscous strains 0 and 0.5, which results in a moderate viscosity reduction of ~ 0.5 .

5.2. Numerical Modeling Results

In numerical model A1 the rift exhibits a distinct northern, central, and southern segment (Figures 6a–6c). After 5 Myr model time, high tectonic strain rates of more than 10^{-15} 1/s are encountered in the northern and southern segments in a narrow band of roughly 100 km width. At the transition to the central segment, the rift arms bend away left laterally from the transversal central region, giving rise to a right-lateral step over geometry. Simultaneously, the rifts connect with each other such that the lateral extent of the rift in the central domain increases almost to the entire model width of 500 km. During subsequent model evolution, the northern and southern segments localize toward the basin center finally reaching a width of ~ 70 km at 15 Myr model time, while the N-S trending deformation domains inside the central segment are distributed with a width of about 300 km.

Next we focus on the final structure of the model after 15 Myr model time. The northern and southern rifts feature a deep basin of up to 100 km width (Figure 6g), whereas the central domain consists of a wide plain with N-S oriented horst and grabens and relatively low relief. The final topographic pattern primarily results from the isostatic response to crustal thinning, which is why surface topography mirrors Moho depth (Figure 6f) except for local horsts and grabens. Both topography and Moho depth exhibit left-lateral bending where the north/south rifts merge into the central domain. However, the LAB topography does not show this pattern indicating the lateral strength contrast in the shallow lithosphere as the main control for this pattern. The surface velocity field clearly shows the transition from narrow to wide rifting in the x component of velocity (Figure 6i). The right-lateral step over from the northern to the southern segments creates a specific tectonic strain field that leads to northward velocities in the eastern part of the Mesozoic rift basin and southward movement in its western part (Figure 6j) predicting minor counterclockwise rotation of the central domain. These velocities of less than 0.4 mm/yr are a magnitude smaller than the overall extension velocity and do not generate any indication of strike-slip deformation. However, the direction of motion compares well with observations and models of other extensional plate boundaries with step over geometries, such as in the western branch of the EAR [Koehn et al., 2008].

In the previous analog and numerical models we find a left-lateral bend (resulting in a right-lateral step over) of the narrow rifts at the transition to the transversal central domain. In the following we investigate the rheological configurations that enable this type of rift connectivity. To this aim, we modify layer thicknesses and thereby the thermal profile within the inherited Mesozoic rift (see supporting information Table S2) while the other model domains remain unchanged. Note that the thicknesses of crust and mantle at the onset of Cenozoic rifting cannot be deduced directly. Hence, we define a range of plausible layer thicknesses for the Turkana region by considering those parts of the Mesozoic African rift systems that have not been reactivated. One such example is the Anza graben in northeast Kenya with 25 to 30 km thick crust [Sippel et al.,

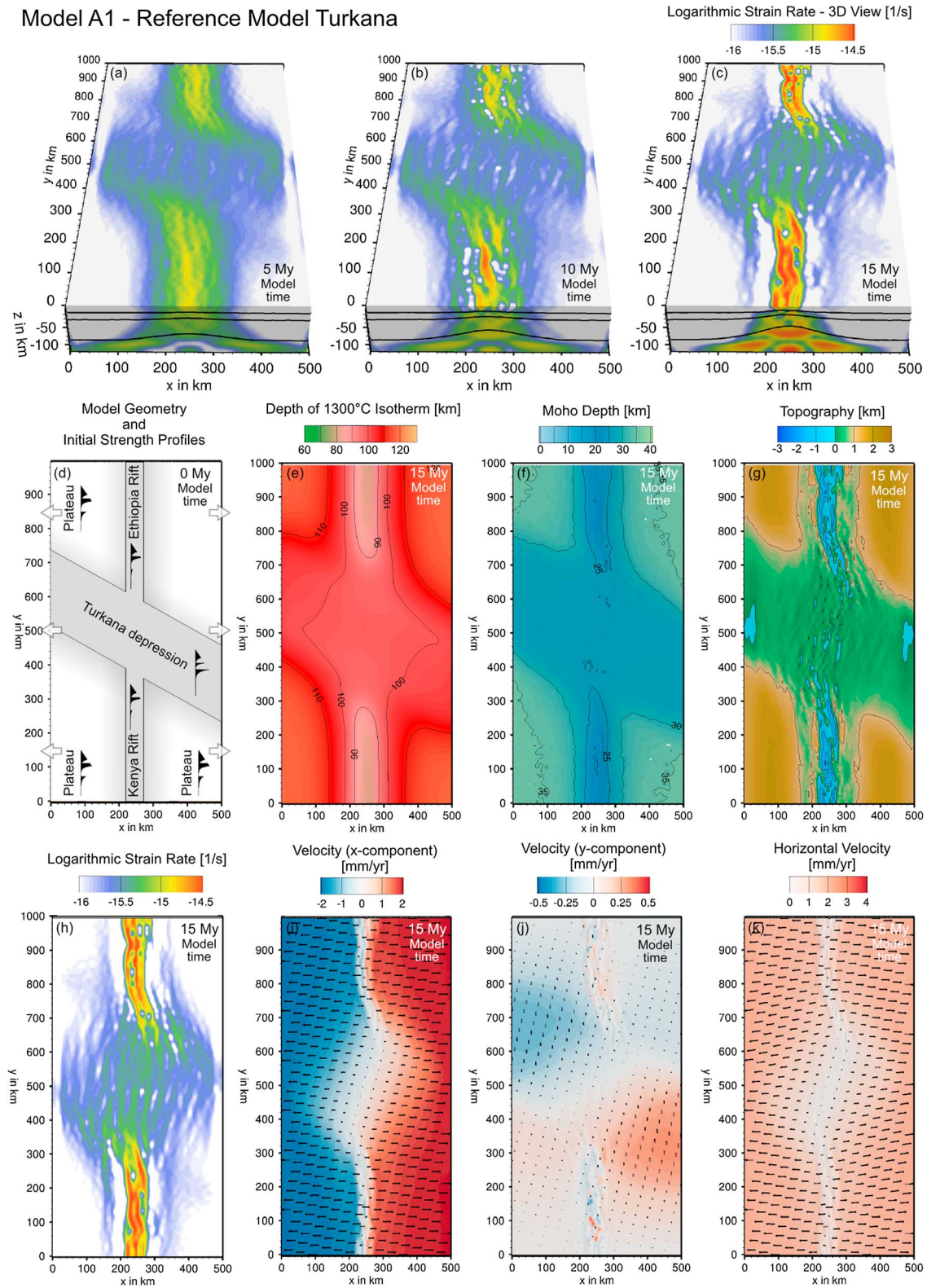


Figure 6. Numerical reference model. (a–c) Evolution of deformation pattern (second invariant of strain rate) in 3-D view. Numerical simulations at 15 Myr are comparable to the present-day structure of the study region. Note the left-bending rift tips (and thus right-stepping rift segments) where the northern and southern rifts interact with the central domain. (d) Initial model geometry in map view. (e) Final lithosphere-asthenosphere boundary (LAB). (f) Final Moho depth. (g) Final surface topography. (h) Final deformation pattern in map view. (i, j) N-S and E-W directed components of surface velocity. (k) Full horizontal velocity field. See rendering supporting information Movie S1 for detailed model evolution.

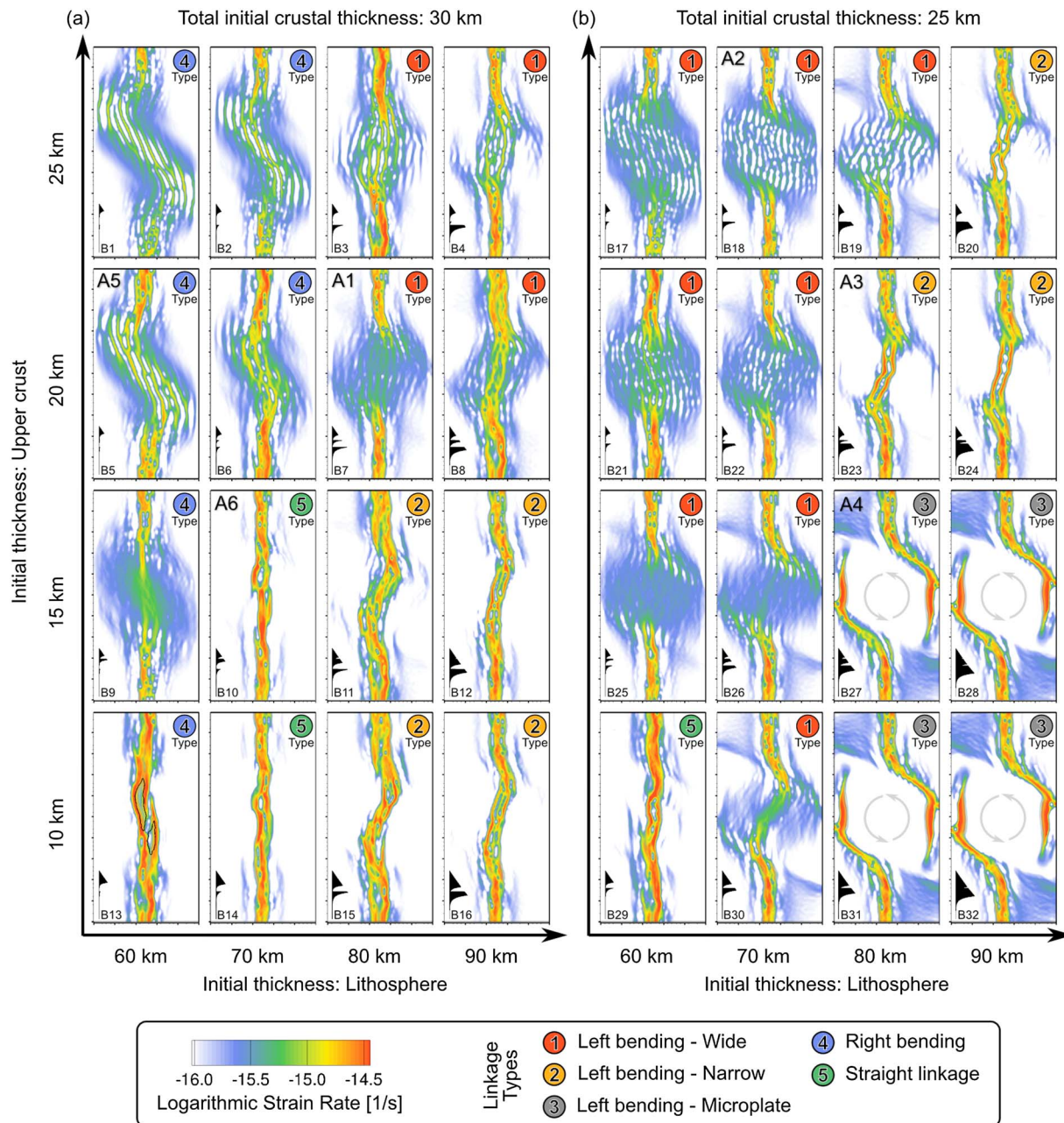


Figure 7. Alternative model results. We vary the initial conditions by scanning through different configuration of initial thicknesses of lithosphere, full crust, and upper crust within the transverse inherited domain of the Turkana region. (a) Initial crustal thickness of 30 km. (b) Initial crustal thickness of 25 km. All other parameters and layer thicknesses are kept identical to model A1. Images show strain rate patterns after 15 Myr of stretching, equivalent to 60 km of extension. In the lower left corner of each image we plot the yield strength profile of the transverse central region at the onset of rifting. Lithospheric strength is controlled by the interplay of lithospheric thickness and radiogenic heat production within the upper crust. The initial strength controls rift localization and leads to five different types of rift linkage after 15 Myr indicated in the upper right corner of each image. Initial strength of each layer within the central domain of all models is also provided in supporting information Table S2. Animations of six characteristic models representing key aspects of rift linkage are given in the supporting information as Movies S1 to S6.

2017] and a lithosphere thickness of 60 to 90 km [Fishwick, 2010]. Similar present-day thicknesses have also been found in the Central and West African Rift System for both lithospheric [Fishwick and Bastow, 2011] and crustal [Pasyanos et al., 2014] layers.

The resulting deformation patterns after 15 Myr model time are depicted along with initial vertical strength profiles of the central transversal domain in Figure 7. We find a large variety of rift patterns that are solely due

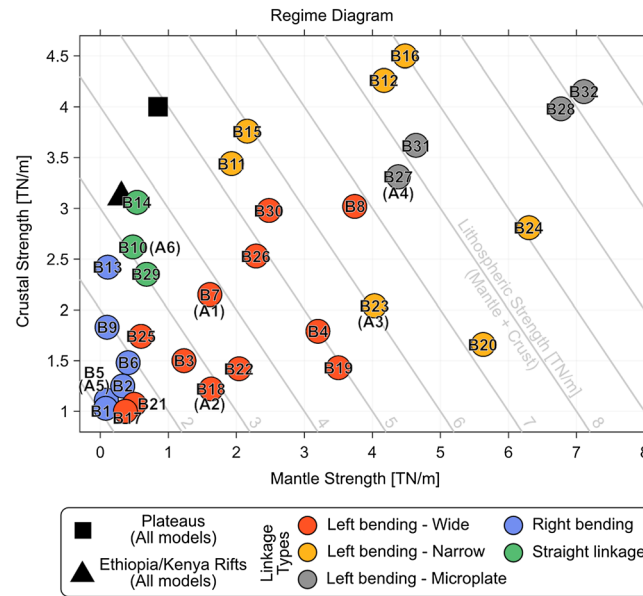


Figure 8. Regime diagram. All models of Figure 7 are represented in terms of initial crustal and mantle strength within the Turkana depression. Initial strength of other model domains such as plateaus, and Ethiopia/Kenya rifts are shown as black square and triangle, respectively. The five linkage types are identified by colored circles. Gray, diagonal lines indicate total lithospheric strength. For characteristic models S1 to S6 we provide supporting information movies of the model evolution.

by forming the left-lateral bend near the central segment (Types 1–3). Hence, the left-lateral bend results from minimizing the mechanical work that is required to deform the central domain. This further indicates that mantle strength exerts first-order control on surficial deformation pattern by controlling the location of lithospheric necking. Wide rifting (Type 1) dominates for integrated lithospheric strengths below ~4–6 TN/m (gray diagonal lines in Figure 6c), while narrow rifting (Type 2) and microplate formation (Type 3) prevail for higher lithospheric strengths. This can be understood when considering that in this high-strength regime brittle deformation leads to more rapid localization so that lithospheric necking occurs faster which promotes narrow rifting. In fact, left-bending wide rifts (Type 1) eventually localize into narrow rifts (Type 2), while the initial stages of narrow rifts (Type 2) feature wide rift characteristics (see supporting information Movies S1–S3); hence, the difference between these types is time dependent. Microplate formation occurs for high-strength end-members within the left-bending regime. In these scenarios, the central domain behaves like an almost undeformable block that diverts strain to the lateral model boundaries. In agreement with continental and oceanic microplate rotations, the right-lateral step over geometry generates a counterclockwise rotating microplate [Katz *et al.*, 2005; Koehn *et al.*, 2008]. Right-lateral bending (i.e., left stepping) of the northern and southern rifts (Type 4) occurs if the mantle strength in the central domain is less or similar to that of these segments. In this case the mantle beneath the central rift is so weak that it governs the necking process and forces crustal localization to bend right laterally into the central segment. This case is reminiscent to recent analog models of propagating rotational rifts interacting with linear rheological heterogeneities cutting the rift at moderate and high obliquity [Molnar *et al.*, 2017]. If the rheological configuration of the central domain is very similar to the rift valleys to the north and south, the model virtually does not notice the inherited heterogeneity and straight linkage (Type 5) occurs. While these explanations hold at first order, the strength diagram also shows that the transition from one regime to the other does not occur as an abrupt boundary.

5.3. Model Limitations and Robustness

The numerical model includes only the lithosphere and shallow asthenosphere; hence, it does not account for mantle-related topographic changes caused by asthenospheric flow [Emishaw *et al.*, 2017], small-scale convection [van Wijk *et al.*, 2008], or whole-mantle convection [Flament *et al.*, 2013; Rubey *et al.*, 2017].

to the rheological configuration of the central domain. The patterns can be classified into five distinct types: a group of three left-lateral bending types (i.e., right-stepping rift segments) where rifting in the central domain occurs (1) as a wide rift like in the Turkana region and as in model A1, (2) in narrow rift mode, and (3) through microplate formation. Further, we identify (4) a right-lateral bending type and (5) a type with straight segment linkage. Representative model animations for the evolution of each linkage type can be found in the supporting information.

The existence of these five types can be explained at first order by considering the strength contrast between crust and mantle in each model (Figure 8). In most cases where the strength of the mantle in the central domain is larger than that of the plateaus outside of the rift zones, the northern and southern segments avoid a direct connection

These dynamic topography contributions add to the isostatically compensated topography, especially above regions of focused upwelling. For the Ethiopian dome, independent studies attributed amplitudes of ~ 1.5 km to this kind of dynamic mantle support [Hoggard *et al.*, 2016; Sembroni *et al.*, 2016], which is why we chose this value for our initial topographic baseline.

When comparing numerical results to regional geophysical models of Moho and LAB depth, one has to keep in mind that our numerical model locally conserves crustal mass. This means that we do not account for magmatic underplating or basin sedimentation leading to replenishment of crustal thickness. The contributions of these processes to crustal rethickening beneath EAR rift valleys are up to 3 km for sedimentation [Ebinger *et al.*, 1999; Sippel *et al.*, 2017] and about 4 km for underplating in the Kenya rift [Thybo *et al.*, 2000].

Values of experimentally determined creep parameters for the chosen compositions vary slightly in the literature. We therefore tested the influence of alternative flow laws for upper crust [Rutter and Brodie, 2004], lower crust [Rybacki *et al.*, 2006], and mantle [Karato and Jung, 2003] and found that despite minor structural differences, our conclusions were robustly reproduced. The model results did not change when we simultaneously applied dislocation and diffusion creep within the mantle [Hirth and Kohlstedt, 2003] instead of only dislocation creep as in the models above. We used 1 mm and 10 mm grain size, and even though the mantle viscosity dropped from a minimum of 10^{19} Pa s to below 10^{18} Pa s, the strain rate field at the surface remained identical. We also tested asymmetric boundary conditions, where we applied the full extension rate to one model side while keeping the opposite side fixed. Since there are no shear tractions allowed across the bottom, top, front, and back boundaries, the results are invariant with respect to asymmetric boundary conditions.

Localization dynamics of faults and shear zones depend on the included weakening mechanisms and the employed numerical resolution [Buiter *et al.*, 2006, 2016]. Reducing the element size to 10 km and/or switching off strain softening for model A1 indeed changes the temporal evolution of the rift. However, after sufficient model time the resulting deformation patterns are qualitatively similar to the reference model (Figure 6). We conducted models where we used rift velocities of 3 mm/yr and 5 mm/yr (full rate) covering the plausible range from plate reconstructions and geological indicators (cf. section 3). The deformation pattern after the same amount of extension featured some small differences, but qualitatively, the structures remained the same. Finally, we varied the obliquity of the inherited Mesozoic rift between 0 and 60° (measured between the domain trend and the extension normal), while all other parameters were identical to model A1. For higher obliquities, the effect that the rifts avoid cutting into the transversal domain is enhanced. For low obliquities, the effect is less pronounced and at no obliquity it disappears (supporting information Figure S2).

6. Discussion

6.1. Along-Strike Differences in Deformation Style

The two independent modeling approaches, crustal-scale analog experiments and numerical lithospheric-scale models, were able to reproduce the significant along-axis differences in style and distribution of deformation between the Ethiopia/Kenya rift valleys and the Turkana depression. In particular, both modeling approaches showed a transition from localized deformation in the rift valley domains characterized by large marginal escarpments to a diffuse deformation within the Mesozoic rift, accommodated by a large number of minor faults with limited vertical displacement. Our results therefore highlight the important role of the transversal preexisting heterogeneity with thinned crust and lithosphere in the Turkana region on the style and distribution of deformation during Cenozoic rifting.

Our findings confirm previous works [Ebinger *et al.*, 2000] in that the width of deformation within the Turkana depression has no similarity to patterns in broad rift zones that developed in relatively hot lithosphere, such as the Basin and Range and Aegean extensional provinces [Buck, 1991, 1999; Labrousse *et al.*, 2016]. Instead, the anomalously wide rift zone is caused by the N-S propagation of the Main Ethiopian and Kenya rift systems into a region of thinned crust, with mantle lithosphere that is stronger than that next to the inherited Mesozoic heterogeneity.

Highly oblique rifts are typically characterized by a wider deformation zone than more orthogonal rifts [e.g., Corti, 2012; Brune, 2014]; since the reactivation of the NW-SE trending heterogeneity under

roughly E-W extension results in a high obliquity of 60° , this may indicate a control exerted by rift obliquity on the anomalously wide deformation domain in the Turkana depression. However, highly oblique rifts feature complex fault patterns that are typically dominated by faults which are not orthogonal or suborthogonal to extension but are instead oblique or highly oblique to this direction [see, e.g., *Agostini et al.*, 2009]. These patterns are very different from what is observed in the Turkana depression, where deformation is characterized by subparallel fault domains trending roughly N-S (i.e., roughly extension-orthogonal). This indicates that faults in the Turkana depression respond directly to the roughly E-W Nubia-Somalia motion and that the high rift obliquity does not play a major role in controlling the style and distribution of deformation. Moreover, the very good correspondence between our model results and the distribution and characteristics of deformation in the Turkana depression suggests a negligible influence exerted by the Western Branch on the interaction between the Kenyan and Ethiopian rifts. This is also supported by the lack of significant volcano-tectonic activity in a large region close to the border between South Sudan and Kenya, west of Lake Turkana.

The numerical models predict a progressive narrowing of deformation within the preexisting Mesozoic rift, which is in agreement with the eastward migration of deformation from basins in the western portion of the Turkana depression to more central domains (e.g., Lake Turkana itself and surrounding areas) [*Ebinger et al.*, 2000, and references therein]. However, the sedimentary basins west of Lake Turkana, which started developing prior to ~ 25 Ma, likely predate Nubia-Somalia divergence [*Iaffaldano et al.*, 2014; *DeMets and Merkuriev*, 2016] and therefore were not strictly related to the Tertiary rifting event in a strict sense. Consequently, other processes, such as deep-seated asthenospheric processes (plume impingement at the base of the lithosphere) [*Ebinger et al.*, 2000] or strain hardening between episodes of extension [*Morley et al.*, 1992], likely contributed to the observed evolution.

6.2. Characteristics of Rift Linkage Within the Turkana Depression

Both our analog model and Types 1–3 of the numerical models depict the tendency of the rift valleys to propagate preferentially to SE (Ethiopia) or to the NW (Kenya) instead of linking up directly. As explained in section 2, this reproduces a key natural observation. In fact, extensional deformation related to the Kenya rift is shifted to the NW to form the narrow depression hosting Lake Turkana; similarly, the Ethiopian rift is progressively shifted to the SE forming the Ririba Rift and Hurri Hills [*Ebinger et al.*, 2000; *Bonini et al.*, 2005].

What is the tectonic reason for diverting the Ethiopian and Kenya rifts toward the east and west, respectively? Our models show that this pattern occurs during the very beginning of deformation (Figure 6a), so it cannot be related to evolving variables such as strain softening or crustal thinning, as evoked for 2-D rift migration [*Kuszniir and Park*, 1987; *van Wijk and Cloetingh*, 2002; *Brune et al.*, 2014; *Svartman Dias et al.*, 2015; *Jammes and Lavier*, 2016]. We find that strength variations in crust and mantle play a first-order role in controlling the regional rift architecture. However, mode selection is not controlled by the vertically integrated lithosphere strength. If this was the case, models where the central domain is weaker than the plateaus outside of the rift zones, such as the analog model and numerical models A1, B3, B17, B18, B21, B22, B25, and B26, would not feature the left-lateral bend at the segment boundary. Therefore, we suggest that there is a hierarchy of dominating processes:

1. On crustal level, localization is solely controlled by brittle localization dynamics and stress focusing due to rift propagation into the transversal preexisting heterogeneity. This agrees with previous modeling [*Acocella et al.*, 1999] that in conditions of orthogonal rifting (i.e., the extension direction is perpendicular to the rift trend) when the angle between the rift margins and the margins of the preexisting heterogeneity is $>90^\circ$, reactivation of its boundaries is favored; conversely, when the angle is $<90^\circ$, reactivation is less probable (Figure 9).
2. If the existence of lithospheric and asthenospheric mantle is accounted for, the crustal processes are largely overruled by lithospheric thinning. In this case, it is the mantle strength that has first-order control on the transition from left-bending type to straight linkage and right-bending type, while second-order deviations of this rule are due to crustal processes.

When applying our results to nature, we suggest that both processes facilitate localization of deformation at the NW termination of the Kenya rift and at the SE tip of the Ethiopian rift (Figure 9), promoting the propagation of this latter to the SE and that of the Kenyan domain toward the NW.

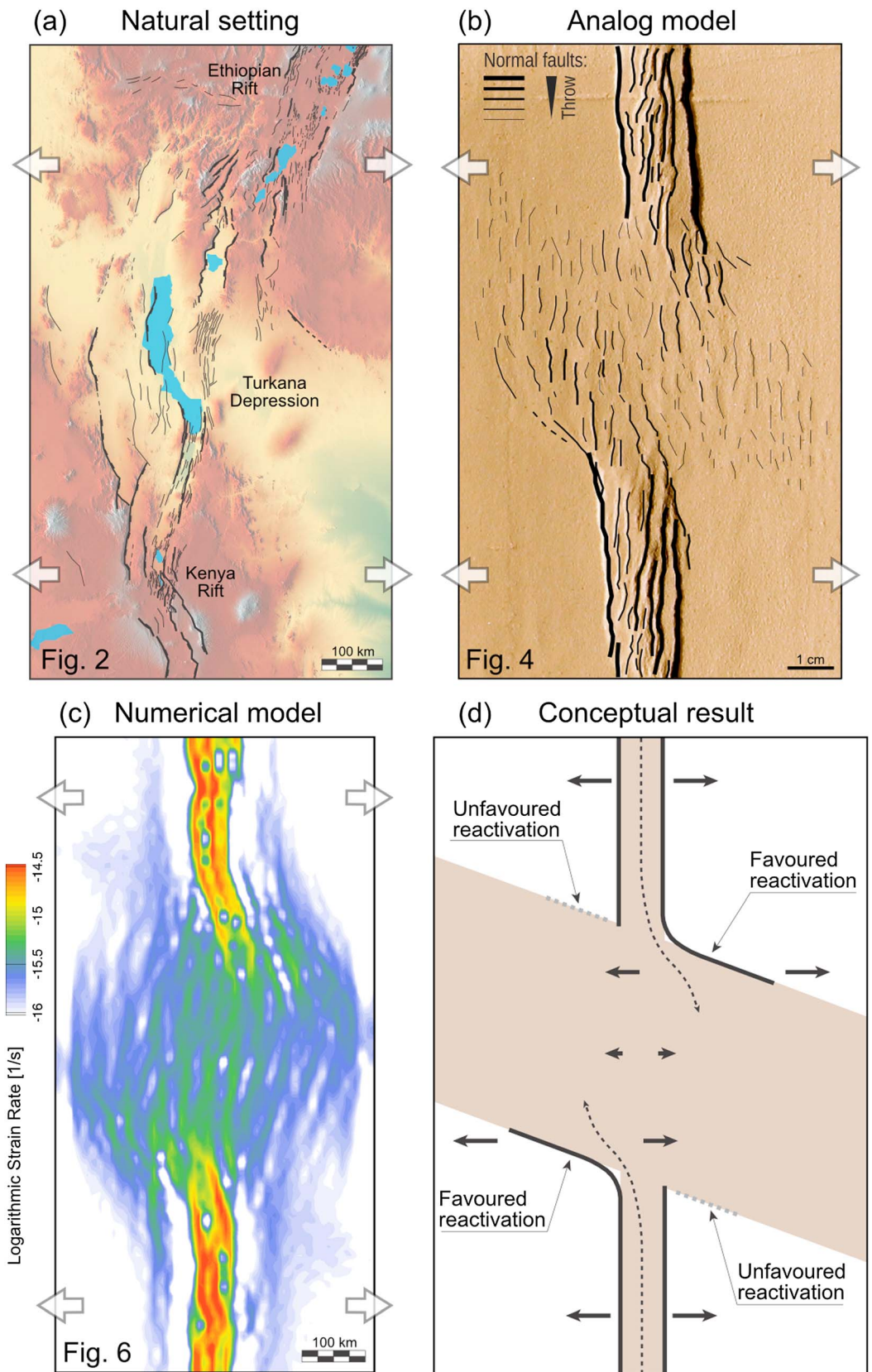


Figure 9. Summary plot. (a) Major faults of the study area. (b) Key features of the analog model. (c) Final deformation pattern of numerical reference model A1. (d) Summarizing interpretation (see text for details).

6.3. Influence of the Preexisting Lithospheric Structure on the Local Fault Pattern

Besides capturing the large-scale picture of rift interaction, our analog modeling indicates that the preexisting variations in strength of the different domains have an important control on the fault pattern at a more local scale. For instance, the models indicate significant reactivation of the boundaries between the Turkana domain and the strong surrounding plateaus, giving rise to NW-SE faults at the margins of the preexisting transversal heterogeneity, which are a continuation of the boundary faults of the main rift valleys. A similar feature can be hypothesized in nature at both the southern tip of the Ethiopian rift (where NW-SE faults characterize the margins of the preexisting Anza graben, east of the Ririba Rift) and the NW termination of the Kenya rift (where NW-SE to NNW-SSE trending faults give rise to the large Turkwell escarpment; Figures 2 and 9). In this latter case, the NW-SE faults typically curve at their tip to acquire a more N-S to NNE-SSW trend, a characteristic that is observed in the analog models. This feature resembles the horse-tail splays at the termination of transcurrent faults and likely result from the component of strike-slip motion at the margins of the Turkana depression imposed by the relative orientation between the boundary of the preexisting transversal heterogeneity and the extension direction (Figure 9).

Another interesting feature within the fault pattern is the presence of numerous NW-SE trending minor faults or fault segments, which develop with a high obliquity to the extension direction but subparallel to the transversal preexisting domain (Figure 4b). In places, these segments tend to connect N-S structures, resulting in highly segmented NNW-SSE trending normal faults with a typical zigzag plan view geometry. In turn, these fault systems bound highly segmented basins, which result from the coalescence of an echelon N-S trending subbasins connected by NW-SE segments. This geometry strikingly resembles the typical “staircase” pattern of the basin hosting Lake Turkana. Although this architecture has been suggested to be controlled by the reactivation of preexisting, discrete upper crustal fabrics [Vetel and Le Gall, 2006], our laterally homogenous models indicate that this architecture may result from the complex stress field developing within the area of wide deformation, which is controlled by rift propagation, the geometry of the preexisting domains, and the direction of extension (Figure 9). We cannot, however, exclude that on a local scale preexisting discrete heterogeneities may have controlled fault development and architecture [Vetel and Le Gall, 2006].

7. Conclusions

In order to elucidate the interaction between the Ethiopian and Kenyan rift valleys within the Turkana depression, we have integrated crustal-scale, isothermal analog experiments with lithospheric-scale, thermomechanical numerical models. This integration and the comparison of the results with nature lead us to draw the following conclusions:

1. The along-axis transition from narrow rift valleys in Ethiopia/Kenya to a distributed deformation within the Turkana depression results from inherited lithospheric strength variations. In particular, the anomalously wide rift zone is controlled by the presence of a NW-SE region of thinned crust resulting from Mesozoic rifting.
2. Within the Turkana region, the Kenyan and Ethiopian rift valleys are deflected left laterally away from one another, avoiding a direct link to form a throughgoing N-S depression. We find that this is primarily linked to the mantle lithospheric strength of the transversal preexisting heterogeneity and also influenced by stress focusing effects and upper crustal fault dynamics.
3. Local-scale characteristics of the fault pattern, such as the occurrence of horse-tail splays at fault terminations or the presence of faults with zigzag plan view geometry giving rise to basins with a staircase pattern as in the case of Lake Turkana, may result from a minor component of strike-slip motion controlled by relative orientation between the geometry of the preexisting domains and the direction of extension.

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