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Special Section:

Frontiers in lithospheric dynamics: bridging scales through observations, experiments, and computations

Key Points:

- Normal fault growth is poorly understood due to the lack of spatially and temporally constrained observational datasets
- Three-dimensional numerical modeling reveals that fault patterns are rapidly established (<100 kyrs) and active deformation is highly transient
- Fault interaction and strain migration are key elements of fault network growth and related seismic hazard

Supporting Information:

Supporting Information may be found in the online version of this article.

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Bridging Spatiotemporal Scales of Normal Fault Growth During Continental Extension Using High-Resolution 3D Numerical Models

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Abstract Continental extension is accommodated by the development of kilometer-scale normal faults, which grow during meter-scale slip events that occur over millions of years. However, reconstructing the entire lifespan of a fault remains challenging due to a lack of observational data with spatiotemporal scales that span the early stage (<10⁶ yrs) of fault growth. Using three-dimensional numerical simulations of continental extension and novel methods for extracting the locations of faults, we quantitatively examine the key factors controlling the growth of rift-scale fault networks over 10⁴–10⁶ yrs. Early formed faults (<100 kyrs from initiation) exhibit scaling ratios consistent with those characterizing individual earthquake ruptures, before evolving to be geometrically and kinematically similar to more mature structures developed in natural fault networks. Whereas finite fault lengths are rapidly established (<100 kyrs), active deformation is transient, migrating both along- and across-strike. Competing stress interactions determine the distribution of active strain, which oscillates between being distributed and localized. Higher rates of extension (10 mm yr⁻¹) lead to more prominent stress redistributions through time, promoting episodic localized slip events. Our findings demonstrate that normal fault growth and the related occurrence of cumulative slip is more complex than that currently inferred from displacement patterns on now-inactive structures, which only provide a space- and time-averaged picture of fault kinematics and related seismic hazard.

Plain Language Summary When the Earth's tectonic plates are stretched, they crack along fractures called faults. We have a general understanding that faults grow over millions of years by slipping during earthquakes. However, we don't yet know how earthquake slip on the scale of meters accumulate to form faults that are up to a few hundred kilometres-long, and which offset the Earth's surface by several kilometres. We developed numerical models that simulate tectonic stretching and fracturing of the Earth's plates. The models produce fault patterns similar to those seen in the Earth. Importantly, early modeled timesteps reproduce the same geometries associated with individual earthquake slip. We show that the style in which slip accumulates is more complex than that inferred by fault growth models, which only provide a space- and time-averaged picture. Stress-interactions between individual faults are critical, causing different faults to slip at different times. This leads to time periods where slip occurs along many small faults, and other periods of time when slip is focused on large faults. Variations in the number and size of faults that are slipping have implications for assessing earthquake hazard. Fault interaction may better our understanding of complex, often unexpected earthquakes that rupture along multiple faults.

1. Introduction

Recent advances in geodetic measurements allow for high-resolution surface observations of crustal deformation (e.g., Elliott et al., 2016). Seismological and geodetic data from an individual earthquake can be inverted using modeled fault geometry to infer slip distribution and magnitude (e.g., Walters et al., 2018). These data show that individual earthquake rupture patterns are variable and complex, with events typically temporally and spatially clustered (e.g., Nicol et al., 2006; Wells & Coppersmith, 1994). Seismological, geodetic, and geomorphological (i.e., field) data also show that rupture lengths are often considerably shorter than finite fault lengths, and multiple segment ruptures during a single event can trigger surprisingly high-magnitude, hazardous earthquakes (7.9 M_w Kaikoura, New Zealand; Hamling et al., 2017; 7.2 M_w El Mayor-Cucapah, Mexico; Fletcher et al., 2014) that challenge the models underpinning seismic hazard assessments (Field et al., 2014).

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Active seismogenic rifts (e.g., across the central Italian Apennines, Gulf of Corinth, Malawi, and Basin and Range) allow for slip rates to be constrained over intermediate (10^4 – 10^5 yr) timescales. For example, paleoearthquake data may be obtained by using radiometric dating (e.g., of, e.g., ^{14}C ; Froggatt & Lowe, 1990) of material preserved in sedimentary layers offset by faults and exposed in trenches (e.g., the Taupo Rift, New Zealand; Berryman et al., 2008; Villamor et al., 2007). Faults may also form within and offset instantaneously deposited volcanic units (e.g., lavas) of known (radiometric) age, allowing the magnitude and rate of fault slip to be constrained (e.g., the East African Rift; Baker et al., 1972; Muirhead et al., 2016). The age of fault surfaces exposed in the field may also be directly dated using cosmogenic nuclides, such as ^{36}Cl , again providing information on temporal variations in fault slip rate and magnitude (e.g., the Italian Apennines; e.g., Benedetti et al., 2013; Shillington et al., 2020; Wedmore et al., 2020). Increasingly, active rift studies show that fault activity is episodic and slip rates are spatially and temporally variable (Cowie et al., 2017; Friedrich et al., 2003; Schlagenhauf et al., 2008). Studies focused on Holocene deformation show that meters of displacement accumulate rapidly in higher-than-average slip rates over several thousand years, separated by periods of lower-than-average slip rates and relative quiescence (Benedetti et al., 2013; Goodall et al., 2021; Mechernich et al., 2018). The active rift record is thus short and provides only a temporal snapshot of the long, geological timescales over which faults grow. The way in which slip accumulates to produce the fault displacement patterns observed over these geological time-scales thus remains poorly understood (e.g., Mouslopoulou et al., 2009).

Large (e.g., tens of kilometres long, several kilometres of displacement) normal faults grow by accumulating meter-scale co-seismic slip during earthquakes, and slow aseismic growth (creep) between earthquakes. Geometrical observations of faults are commonly used to understand how faults grow, typically using the empirical relationship between displacement (D) and length (L) (e.g., Cowie & Scholz, 1992; Dawers et al., 1993; Walsh et al., 2002). More specifically, the broadly linear, positive correlation between these two parameters, as well as theoretical and mechanical considerations (Cowie & Scholz, 1992) support the “propagating fault model”, which argues fault length and displacement synchronously increase with time (Walsh & Watterson, 1988). The analysis and age-dating of growth strata preserved adjacent to ancient faults imaged in 3D seismic data have led to support for the “constant-length” model, where faults establish their near-final lengths early in slip history and displacement accumulates on faults of near-constant length (Meyer et al., 2002; Walsh et al., 2002). Displacement rates and earthquake recurrence intervals appear relatively stable when averaged over $>10^6$ yrs (Mouslopoulou et al., 2009; Nicol et al., 1997, 2005, 2006) although this appears at odds with the variability observed in active rifts and earthquake behavior, which are found to be irregular in terms of slip rate, slip magnitude, location and occurrence (e.g., Weldon et al., 2004). It is challenging to reconstruct fault growth with respect to how their length and displacement increase over shorter to intermediate timescales ($<10^6$ yrs) using observational data alone, particularly for ancient rifts, principally due to: (a) limited seismic reflection data resolution in the subsurface; (b) the lack of exposures of hanging wall growth strata in the field; and (c) a lack of age-constraints on syn-kinematic (growth) strata in both subsurface and field data (Jackson et al., 2017). Although there are some exceptions (e.g., Taupo Rift; Berryman et al., 2008), even young active rifts often lack the 3D subsurface data coverage required to really capture early fault growth processes in time and space.

Due to the lack of observational datasets of the appropriate temporal and spatial scale, it remains unclear how complex incremental slip patterns ultimately relate to the longer-term, often relatively simple and stable pattern of displacement accumulation that seemingly characterize large, mature faults (e.g., Mouslopoulou et al., 2009; Nicol et al., 1997, 2005, 2006). Here, we use high-resolution thermal-mechanical 3D simulations of continental extension to examine the evolution of normal fault networks across spatiotemporal scales poorly sampled by observational datasets (i.e., over 10^4 – 10^5 yr increments). Using novel image processing techniques, the active length, strain, and cumulative strain are extracted from large fault populations across multiple timesteps and compared to natural D-L observations from active and ancient rifts, providing insights into the kinematics that constrain fault growth, and the underlying dynamics governing their evolution.

2. Numerical Modeling of Continental Extension

2.1. Model Setup

We use the open-source, mantle convection and lithospheric dynamics code ASPECT (Heister et al., 2017; Kronbichler et al., 2012) to model 3D continental extension (Figure 1a). The governing equations (outlined in Section 2.2) are solved on a 3D gridded domain that spans 500 by 500 km across the horizontal plane (X, Y) and

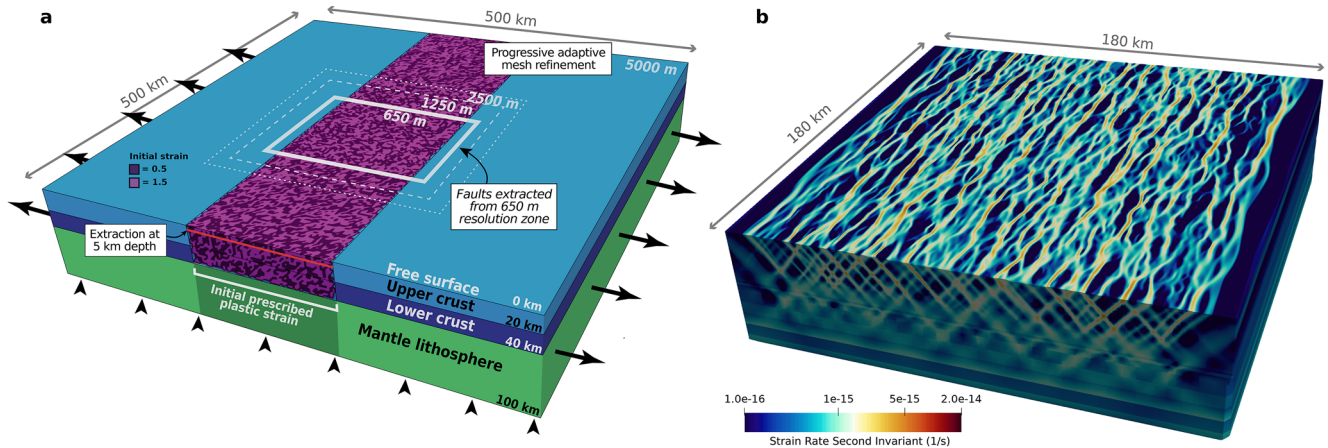


Figure 1. (a) Continental extension model setup outlining initial boundary conditions, compositional layers and prescribed initial strain. Resolution is progressively refined to higher levels up to 650 m in the model center (180 × 180 × 20 km region), where we perform fault extraction analysis at 5 km beneath the initial model surface (i.e., 95 km above the model base). (b) Strain rate second invariant (1/s) after 100 kyrs, cropped to the centered 650 m resolution zone. The deformation patterns reveal that the randomised initial plastic strain field produces rapid localisation of a well-defined normal fault network.

100 km in the depth (Z) direction (Figure 1a). The grids are coarsest (5 km) on the sides and base of the model domain and are successively reduced using adaptive-mesh refinement, increasing the resolution to 625 m over a central region measuring 180 × 180 × 20 km (Figure 1a). Aside from mesh deformation related to free surface evolution, the numerical resolution otherwise stays constant through time with no further adaptive refinement steps. Broadly, this approach provides “natural” boundary conditions for the formation of a distributed fault network within the upper crust. Only the (625 m resolution) central portion of the model is used for fault analysis in our results—See the Supporting Information S1 and Supplementary Files for the full, lower resolution extent of the model.

Orthogonal extension is driven by prescribed outflow velocities on the left and right sides, with inflow at the model base exactly balancing the outflow. The top of the model is a free surface (Rose et al., 2017), which is advected using both the vertical and horizontal components of the velocity field. We model extension rates (i.e., the prescribed outward velocity on each of the left and right boundaries) of 2.5, 5, and 10 mm yr^{−1}, which correspond to the range of values widely used in prior numerical simulations of continental extension (e.g., Naliboff et al., 2020; Van Wijk & Cloetingh, 2002) and are comparable to those characterizing active, natural rifts (e.g., Argus & Heflin, 1995; Bell et al., 2011).

2.2. Governing Equations

ASPECT solves the conservation of momentum, mass and energy equations, combined with advection-diffusion equations, which are outlined below. Velocity and pressure are solved for using the incompressible extended Boussinesq approximation, where the Stokes equations are defined as:

$$\nabla \cdot \mathbf{u} = 0 \quad (1)$$

$$-\nabla \cdot (2\mu \dot{\epsilon}) + \nabla p = \rho \mathbf{g} \quad (2)$$

Where $\mathbf{u}$ is the velocity, μ is the viscosity, $\dot{\epsilon}$ is the deviatoric strain rate tensor, p is pressure, ρ is density, and $\mathbf{g}$ is gravitational acceleration.

Temperature evolves through a combination of advection, heat conduction, shear heating, and adiabatic heating:

$$\rho C_p \left(\frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) - \nabla \cdot (\kappa \rho C_p \nabla T) = \rho H + 2\eta \dot{\epsilon}(\mathbf{u}) + -\alpha \rho T \mathbf{u} \cdot \mathbf{g} \quad (3)$$

Where C_p is the heat capacity, T is temperature, t is time, κ is thermal diffusivity, α is the linear thermal expansion coefficient, and H is the rate of internal heating. Respectively, the terms on the right-hand side correspond to internal heat production, shear heating, and adiabatic heating.

Density varies linearly as a function of the reference density (ρ_0), linear thermal expansion coefficient (α), reference temperature (T_0), and temperature:

$$\rho = \rho_0 (1 - \alpha (T - T_0)) \quad (4)$$

2.3. Initial Conditions

The model domain contains three distinct compositional layers, representing the upper crust (0–20 km depth), lower crust (20–40 km depth), and lithospheric mantle (40–100 km depth). Distinct reference densities (2700, 2900, 3300 kg m⁻³) and viscous flow laws for dislocation creep (wet quartzite; Gleason & Tullis, 1995, wet anorthite; Rybacki et al., 2006, dry olivine; Hirth & Kohlstedt, 2003) distinguish these three layers, which deform through a combination of nonlinear viscous flow and brittle (plastic) deformation (e.g., Glerum et al., 2018; Gouiza & Naliboff, 2021; Naliboff et al., 2020). Table S1 in Supporting Information S1 contains the specific parameters for each flow law.

The initial temperature distribution follows a characteristic conductive geotherm for the continental lithosphere (Chapman, 1986). We solve for the conductive profile by first assuming a thermal conductivity of 2.5 W m⁻¹ K⁻¹, a surface temperature of 273 K, and a surface heat flow of 55 mW/m², and constant radiogenic heating in each compositional layer (Table S1 in Supporting Information S1) that we use to calculate the temperature with depth within each layer. The resulting temperatures at the base of the upper crust, lower crust, and mantle lithosphere are 633, 893, and 1613 K, respectively. The temperature remains fixed at the top and base of the model, while the sides are insulating. The values of the compositional fields are only fixed at the base of the model.

2.4. Rheological Formulation

Rheological behavior combines nonlinear viscous flow with brittle failure (see Glerum et al., 2018). Viscous flow follows dislocation creep, formulated as:

$$\sigma'_{II} = A \frac{1}{n} \frac{1}{\dot{\epsilon}_{II}^n} e^{\frac{Q + P V}{n R T}} \quad (5)$$

Above, σ'_{II} is the second invariant of the deviatoric stress, A is the viscous prefactor, n is the stress exponent, $\dot{\epsilon}_{II}$ is the second invariant of the deviatoric strain rate (effective strain rate), Q is the activation energy, P is pressure, V is the activation volume, T is temperature, and R is the gas constant.

Brittle plastic deformation follows a Drucker Prager yield criterion, which accounts for softening of the angle of internal friction (ϕ) and cohesion (C) as a function of accumulated plastic strain:

$$\sigma'_{II} = \frac{6C \cos \phi + 2P \sin \phi}{\sqrt{(3 + \sin \phi)}} \quad (6)$$

The initial friction angle and cohesion are 30° and 20 MPa respectively, and linearly weaken by a factor of 2 as a function of finite plastic strain, which is derived from the second invariant of strain rate in regions undergoing deformation. Following our recent work (Gouiza & Naliboff, 2021; Naliboff et al., 2020), strain weakening occurs between plastic strain values of 0.5 and 1.5. Rather than a single weak seed (e.g., Huismans & Beaumont, 2007; Lavier et al., 2000), or randomised distribution at each grid point (e.g., Duclaux et al., 2020; Gouiza & Naliboff, 2021; Naliboff et al., 2020), initial plastic strain is partitioned into 2.5 km² blocks that are randomly assigned a strong (1.5) or weak (0.5) value, which correspond to the end and start of the strain weakening interval (i.e., 0.5–1.5). Additional block sizes of 1.25, 2.5, 5 and 10 km² were tested at lower (1250 m) resolutions (see Supplementary Information). This results in statistically random but pervasive damage, which from a geological perspective may reflect the structural heterogeneity observed in many natural systems, where deformation exploits inherited weaknesses such as pervasive fabrics such as metamorphic foliations (e.g., Phillips et al., 2016), or the margins of strong zones (e.g., ancient cratons; e.g., Dunbar & Sawyer, 1989). Critically, the use of a randomised initial plastic strain field decoupled from the numerical resolution produces significantly faster localisation of a well-defined normal fault network (Figure 1b) relative to recent studies that defined randomised plastic strain at each grid point, and which used a continuous randomised value range rather than a

binary distribution (e.g., Gouiza & Naliboff, 2021; Naliboff et al., 2020). The same initial plastic strain field is used for each of the model runs.

The viscosity is calculated using the viscosity rescaling method, where if the viscous stress exceeds plastic yield stress, the viscosity is reduced such that the effective stress matches the plastic yield stress (Glerum et al., 2018; Moresi & Solomatov, 1998). This standard approach for plasticity yields mesh-dependent shear band widths, which can be largely removed using a recently proposed viscoplastic regularization method (Duretz et al., 2020). We tested the effect of this approach using a plastic damper viscosity of 10^{21} Pa s, which produced similar first-order trends, but predictably more diffuse deformation patterns within and near faults (see Supporting Information). Nonlinearities from the Stokes equations are addressed by applying defect correction Picard iterations (Fraters et al., 2019) to a tolerance of $1e-4$ with the maximum number of nonlinear iterations capped at 100. The first timestep typically requires 50–75 nonlinear iterations, with subsequent time steps normally converging in 1–2 iterations. A stricter tolerance of $1e-5$ was tested up to 20 timesteps (see Supporting Information) for the 5 mm yr^{-1} model, and yielded variations in spatiotemporal faulting, but quantitatively similar trends and processes. The maximum numerical time step is limited to 20 kyrs and the solution is output every 100 kyrs (i.e., every 5 timesteps) for the 5 mm yr^{-1} model. In total, 50 timesteps are generated for each model. As the model time decreases with higher extension velocities, the maximum timestep is 10, 5, and 2.5 Myrs for extension rates of 2.5, 5, and 10 mm yr^{-1} , respectively.

3. Automated Fault Extraction Workflow

To quantitatively analyze the geometry and kinematics of faults throughout the entire model, automated fault identification and extraction is required. Previous studies (Duclaux et al., 2020; Naliboff et al., 2020) use conventional image processing techniques such as thresholding (where data is removed at a defined cut-off value) and labeling (where pixels connected to neighboring pixels are labeled as a number e.g., Dillencourt et al., 1992; Figure 2). However, manually defined thresholds may introduce biases to a fault's length and do not capture the complex fault interactions forming the focus of this study (Figure 2).

Here, the strain rate field is sampled on a horizontal plane located 5 km beneath the initial model surface as the extent of the fault network is captured at this depth (Figure 3a). The strain rate field is differentiated along the Y- (rift axis) direction, then histogram equalization is applied (e.g., Hum et al., 2014; Gonzalez, 2009) that spreads the intensity of the differentiated field in order to capture small-scale faulting and allow for a consistent global extraction across other timesteps (Figure 3b). The derivative of Figure 3b is taken along the X- (rift orthogonal) direction (Figure 3c) and extracted using a percentile threshold to produce a binary image (Figure 3d), which is labeled (Dillencourt et al., 1992) such that each identified fault has a unique, accessible index (Figure 3e). Noise is filtered out by removing fault labels that contain areas less than 1.5 km^2 , which, combined with our workflow approach whereby only sufficiently localized deformation is extracted, may lead to variable detection of smallest (c. 4.5 km long) faults. Overall, this approach largely recovers detailed interactions between distinct active fault strands, where extraction remains continuous along consistent strain rate magnitudes, and segments at abrupt decreases in strain rate (Figure 3e).

Geometric attributes such as fault length and displacement are extracted by iterating through each label. Labels are thinned to a pixel length and fault length is computed using the cumulative euclidean distance (Figure 3f). On a few occasions the calculated euclidean distance is abnormally high as the faults splay along-strike, therefore the localities of erroneous distances are used to split labels into discrete fault segments at the point of bifurcation (the jupyter notebook in the Supplementary Files demonstrates this issue). The locality of the fault label is masked over the cumulative strain field to extract the maximum strain occupied by that label. Strain is converted to displacement by multiplying by 2x the grid resolution (625 m), as on average, shear bands typically span two elements. The extracted faults are termed “active fault”, where an active fault represents a fault strand that was actively deforming during a single timestep (e.g., determined by the active deformation field). A “fault” represents a larger geometrically coherent structure that forms over the entire lifetime of the model (e.g., determined by finite strain).

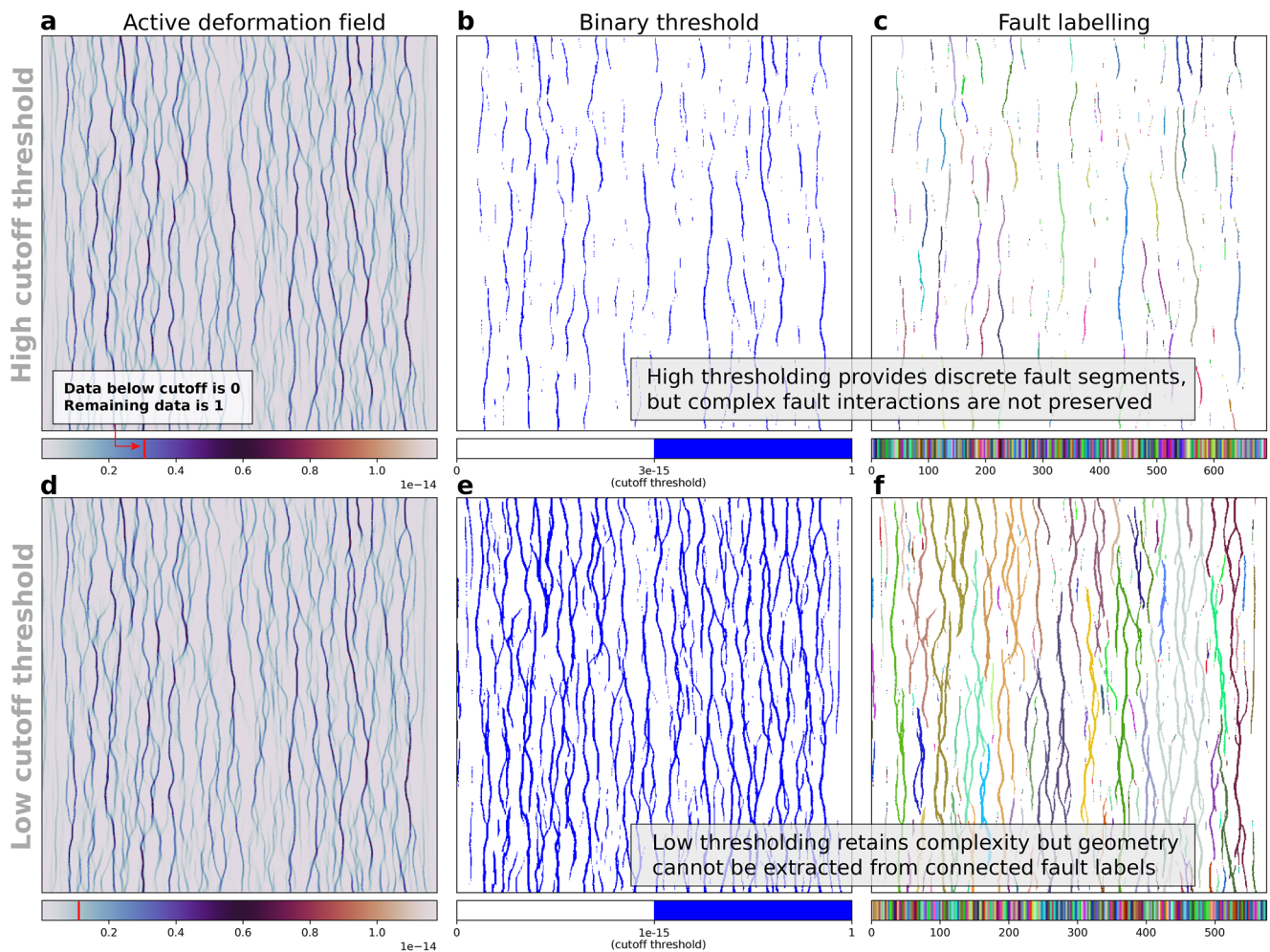


Figure 2. Conventional image processing techniques, which here are derived from the strain rate field 5 km below the initial model surface. This approach performs a binary threshold, where values below a specified cutoff point ($3\text{e-}15$ and $1\text{e-}15$ l/s) are assigned 0 and values above are assigned 1 (b and e). Regions which are connected in the binary threshold are grouped into components labeled as integers (c and f). However, a high cutoff value does not capture small scale strain and makes fundamental assumptions on the resulting fault length (c). In contrast, a low cutoff value results in labels containing multiple fault segments, in which length cannot be extracted (f). Details and examples of this fault extraction procedure are provided in the supplementary materials.

4. Model Results

The evolution of the modeled fault network for a 5 mm yr^{-1} extension rate is shown in Figure 4. The network contains faults of varying lengths (c. $4.5\text{--}190\text{ km}$), which often exhibit along-fault changes in strike, and that may splay and link with adjacent structures (Figure 4). A key observation is that fault patterns, and specifically length, are established during the first modeled timestep (Movie S1). For the remainder of extension, strain accumulates such that fault strain patterns (i.e., displacement) exhibit strain maxima that are typically located near the map-view centers of individual fault systems (e.g., Cowie & Shipton, 1998; Kim & Sanderson, 2005; Segall & Pollard, 1980, Figure 4, Movie S1). The along-strike variability in fault displacement is comparable to that observed in active natural systems in the field (e.g., Faure Walker et al., 2021; Mildon et al., 2017; Nixon et al., 2016) and in ancient rifts studied using seismic reflection datasets (e.g., McLeod et al., 2000; Reeve et al., 2015). The strain rate (Figures 5a–5c) and extracted active fault locations (Figures 5d–5f) for models with extension rates of 2.5 , 5 , or 10 mm yr^{-1} reveal active deformation accommodated on complex fault networks. Our results show that the spatial pattern of the fault network appears similar across the investigated models; however in models with faster extension rates, the overall magnitude of strain rate increases and is accommodated across increasingly diffuse zones of deformation (Figures 5a, 5c and 5e).

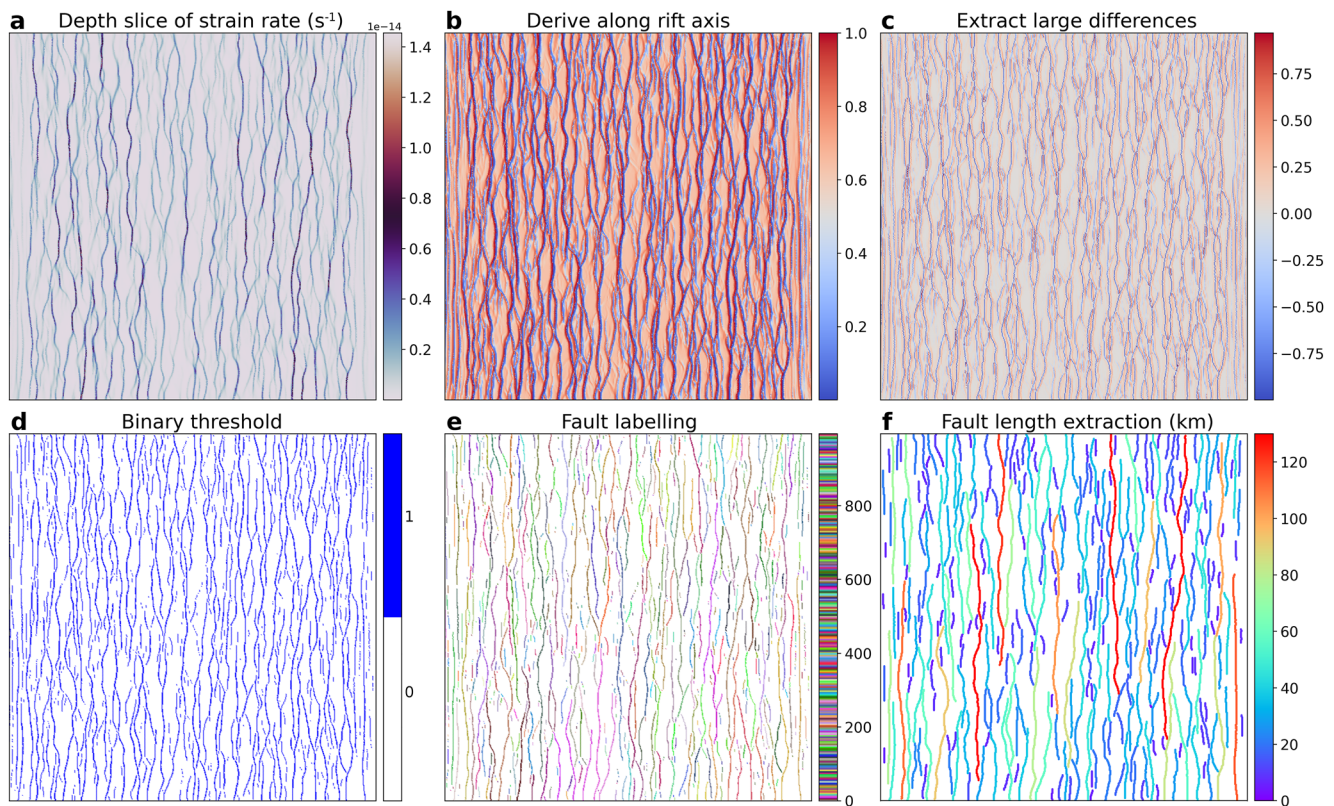


Figure 3. Novel fault extraction workflow. The active deformation field (e.g., second invariant of the strain rate) is extracted 5 km below the initial model surface (a). The spatial derivative of the second invariant of the strain rate across the Y- (along-strike) direction is calculated, and the resulting field is equalized to 0–1 showing that values of 0 and 1 lie on the sides of fault localities (b). An across-strike derivation (c) identifies these localities, in which a binary threshold is applied (d). Spatially connected regions of the binary array are labeled as integers (e) and noise is removed. This method results in the extraction of discrete fault segments, such that active fault lengths may be derived (f). Details and examples of the novel fault extraction are provided in the supplementary materials.

The final displacement-length (D-L) ratios for individual faults within the modeled network are geometrically similar to those identified in natural systems (e.g., Kim & Sanderson, 2005; Walsh & Watterson, 1991) (Figure 6). However, during the earliest stages of rifting, that is, within the first resolvable timestep (c. 200, 100 and 50 kys for extension rates of 2.5, 5, or 10 mm yr⁻¹, respectively), active deformation is accommodated along distributed fault networks (Figures 4a and 4b; Movie S1 and S2) that in terms of their plan-view pattern, change little during the model run (Figures 4c and 4d). During this earliest timestep (<100 kys) the faults are seemingly under-displaced compared to geological D-L datasets, instead plotting on the slip-length ratio associated with individual earthquakes ($D_{\max}/L = 0.00005$ from Baize et al., 2020; Li et al., 2012; Manighetti et al., 2009; Wells & Coppersmith, 1994; Wesnousky, 2008; Figure 6). The modeled D-L relationships show significant scatter (over 2–3 orders of magnitude) throughout each model timestep, similar to magnitudes observed from ancient and active fault D-L studies. The scatter may be attributed to the process of fault growth, rather than measurement errors or variation in mechanical properties (Cartwright et al., 1995; Pan et al., 2022), and may reflect how individual earthquakes accumulate displacement over intermediate (10³–10⁶ yr) timescales (Mouslopoulou et al., 2009; Nicol et al., 2009, 2010, 2020).

Time-series of the total number of active faults (Figure 7a), and the average active fault length in a given population (Figure 7b), reveal significant fluctuations throughout time. All three models (with extensions rates of 2.5, 5, 10 mm yr⁻¹) show an increase in fault number and average fault length within the first c. 10 timesteps (Figure 7), corresponding to a transition from an initially diffuse fault pattern that rapidly localizes (i.e., reduces in element width) within the first few timesteps (see Movie S1, S2, S4–S6). Both the total number of faults and mean length continue to fluctuate throughout the remainder of extension, reflecting oscillations between localized and distributed active deformation throughout the crust (Figure 7). In comparison to the 2.5 and 5 mm yr⁻¹ models, the oscillations in fault number and active lengths (Figure 7) in the 10 mm yr⁻¹ model are greater, and

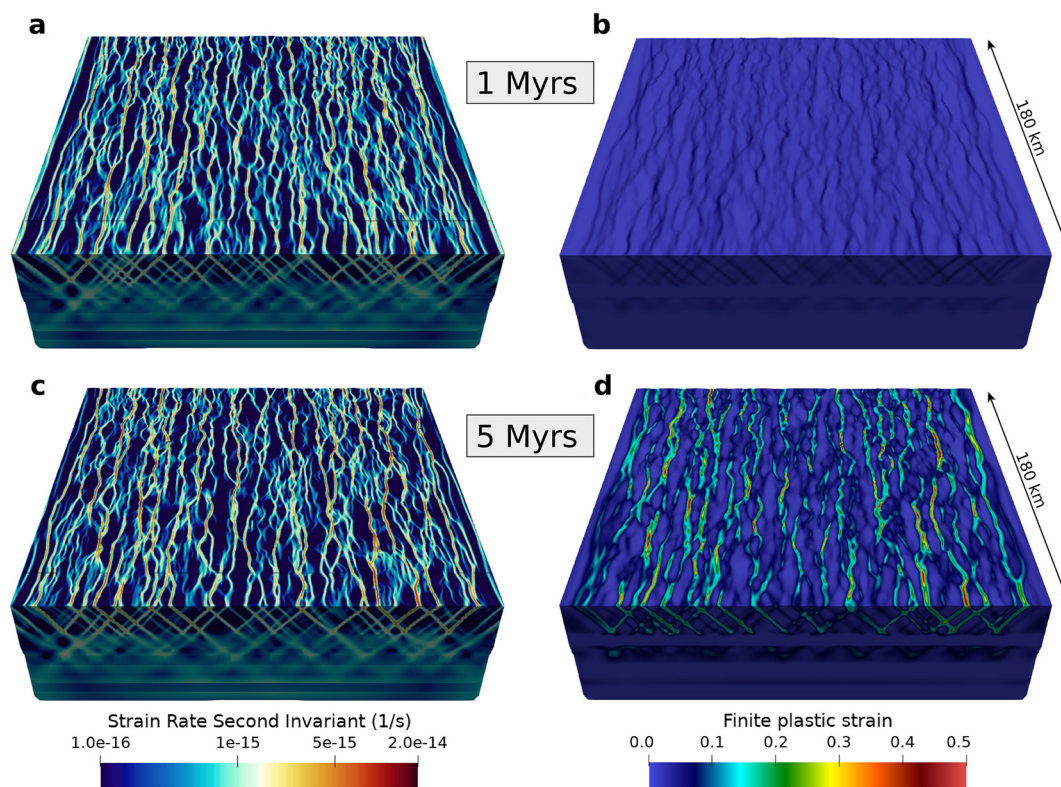


Figure 4. Active deformation (left) and accumulated brittle plastic deformation (right) for the model extending a constant rate of 5 mm yr^{-1} , after (a and b) 1 Myrs and (c and d) 5 Myrs of extension. The second strain rate invariant ($1/s$) on the left, documenting active deformation, is limited to values above $1e-16$ here to better reveal and compare the main active structures. The accumulated deformation on the right refers to non-initial plastic strain, where the initial strain field has been subtracted to better reveal fault patterns. At 1 Myr, deformation is relatively distributed across faults whereas at 5 Myrs structures are localized onto the largest faults. See Movie S1 for animation across all modeled timesteps.

its notably different pattern of active deformation (Figures 5e and 5f) indicates that increases in extension rate consequently produce more prominent stress redistributions through time. After an initiation phase of 0.5 Myrs in the 10 mm yr^{-1} model, the mean active lengths episodically reach highs of c. 19 km and lows of c. 13 km across recurrence intervals of c. 350 kyrs (Figure 7). Although transient behavior continues for the remainder of extension, the overall number of active faults decreases across all models and the average fault length slightly increases (Figure 7), demonstrating that large-scale localisation occurs as strain is concentrated onto fewer, larger fault systems (e.g., Cowie, 1998).

Transient deformation occurs both along- and perpendicular to the rift axis, which we view in a regional model subset (Movie S4–S6). This subset shows along-strike migration of deformation (Movie S7) consistent with the preferential along-strike propagation direction of (eventual) plate rupture. Across-strike migration of deformation is observed in all models (Movie S4–S6). In detail, however, at lower extension rates (2.5 and 5 mm yr^{-1}), the along-strike migration of deformation appears most prominently at along-strike bends (note areas A and B stay relatively static in Movies S4–S5), whereas straight active segments also migrate across-strike for 10 mm yr^{-1} extension rates. Both along- and across-strike strain migration are reflective of competing stress-interactions between faults in the near field (e.g., Cowie, 1998) and as documented in Figure 7, produce end-member behaviors characterized by localized, continuous slip (Figures 8a–8c) and distributed, segmented slip (Figures 8d–8f). The end-member behavior applies most to the fastest extension rate modeled (10 mm yr^{-1}) due to increased active strain migration and interaction.

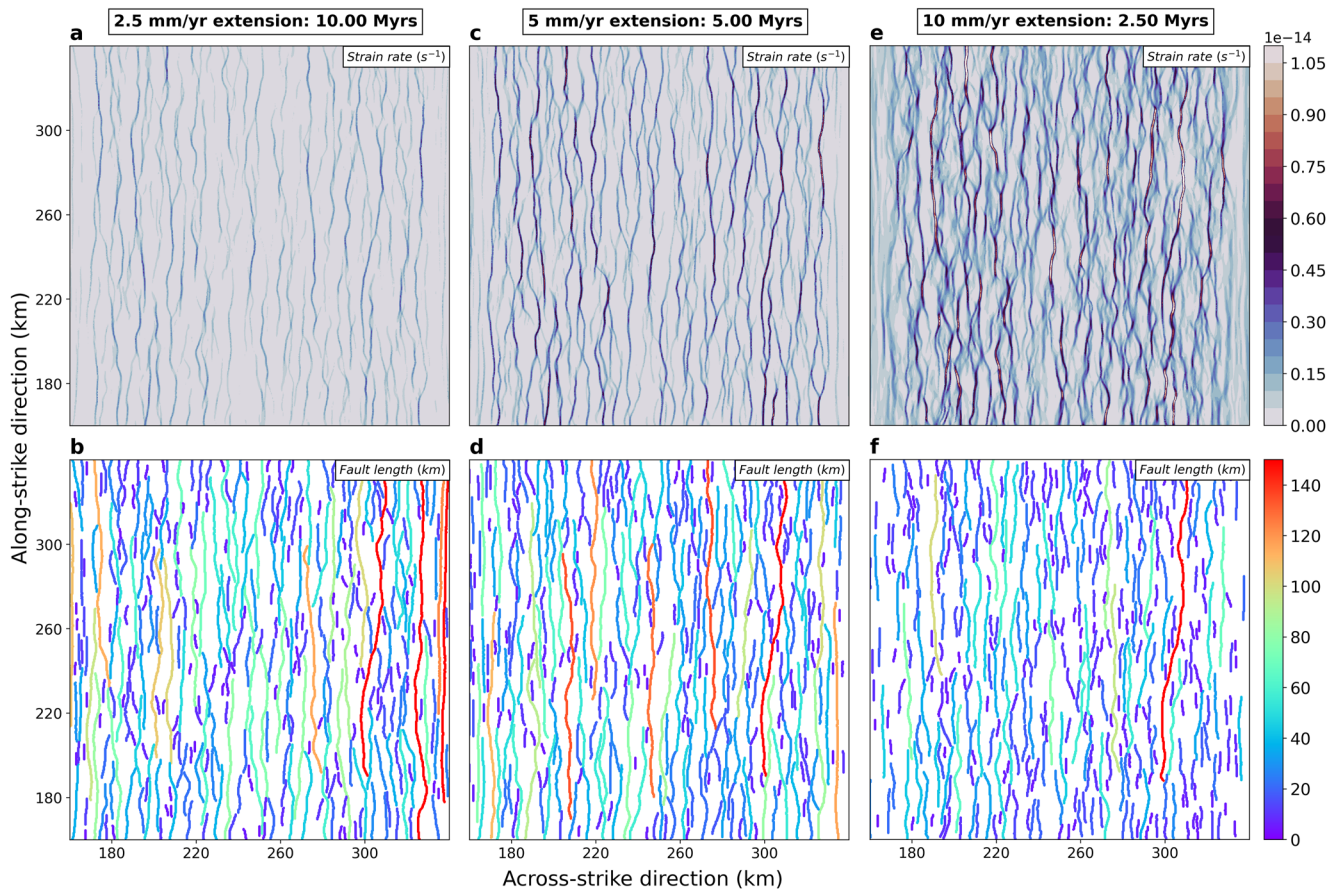


Figure 5. Comparison across models with extension rates of 2.5, 5, and 10 mm yr⁻¹. The top row shows the strain rate invariant (s⁻¹) in the upper crust (extracted at 5 km depth), documenting active deformation patterns within the last (50th) timestep representing 10, 5 and 2.5 Myrs extension, respectively. The bottom row shows their corresponding fault length extracted from the active deformation field. See Movie S2 for animation across all modeled timesteps.

5. Discussion

The numerical models presented here exhibit establishment of near-final fault lengths from the onset of extension, consistent with the constant-length model (e.g., Rotevatn et al., 2019; Walsh et al., 2002). Modeled faults therefore predominantly accumulate displacement (rather than length) through time and move upwards in D-L space, behavior which clearly diverges from the linear relationship derived from observational D-L data that has historically underpinned the propagating fault model (Figure 6; Cowie & Scholz, 1992; Mansfield & Cartwright, 2001; Walsh & Watterson, 1988; Walsh et al., 2002). Our results show that active fault lengths are established an order of magnitude (<100 kyrs) earlier than currently inferred from seismic reflection analysis of ancient (c. 1.3 Myrs, NW Shelf, Australian; Walsh et al., 2002) and active (c. 700 kyrs, Whakatane Graben, New Zealand; Taylor et al., 2004) rifts. Figure 6 shows that the D-L scaling relationship characterizing the early fault networks in our numerical models plot outside of and more critically, below the scatter observed in natural datasets (where D/L = 0.001), only beginning to lie within this scatter after 1 Myrs (Movie S3). This suggests that while the modeled faults rapidly formed their lengths within the first 100 kyrs, the initial 1 Myrs of fault lengthening may not be captured due to the resolution threshold of observational data used (i.e., seismic, field).

The constant-length model was initially proposed for faults that reactivated underlying pre-existing weaknesses (Meyer et al., 2002; Walsh et al., 2002). Our models indicate that the rapid establishment of complex fault patterns produced here reflect both the relative strength contrast between the binary blocks of initial strain (0.5 and 1.5) field, as well as the mechanical and kinematic interactions between adjacent faults, and does not necessarily require reactivation of a well-developed, pre-existing fault network. Here, the modeled fault network produces active faults with lengths limited to <5 km (observe the clear cut-off in datapoints on Figure 6); this is attributed

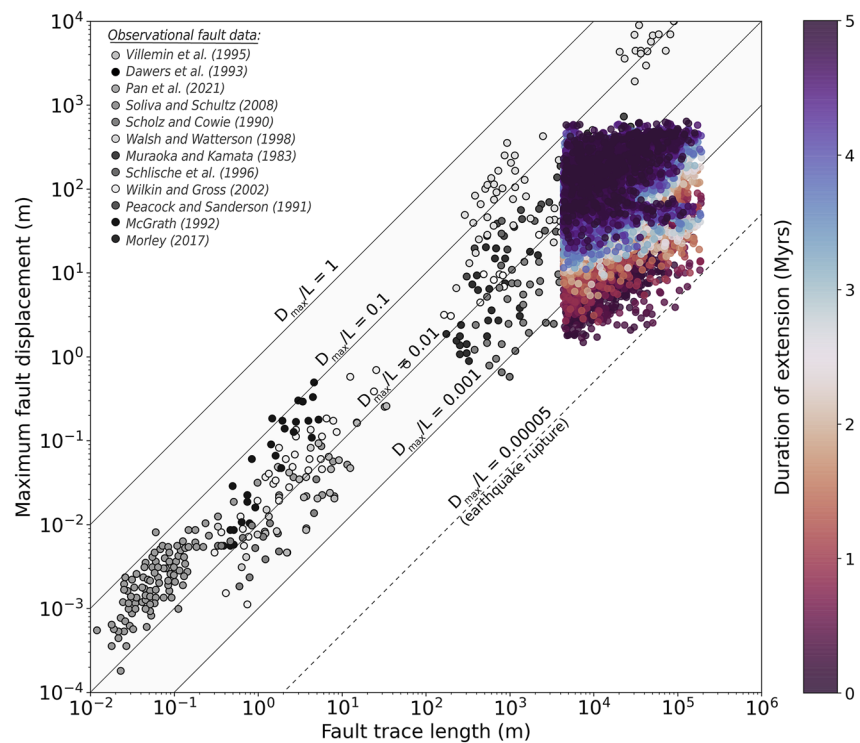


Figure 6. Fault D-L evolution for the modeled fault network extending at a rate of 5 mm yr⁻¹. Observational datasets are plotted in gray, where different shades correlate to references therein, and the modeled data are in color. Note that faults with areas <1.5 km² are removed in a post-processing step to reduce noise, and has led to a clear cut-off in datapoints at c. 4.5 km. See Movie S3 for animation across all modeled timesteps.

to the prescribed initial strain field consisting of “strong” and “weak” blocks (2.5 km²). This implies that the range of fault lengths that span eight orders of magnitude in D-L plots may reflect multiple scales of heterogeneity present in the lithosphere (e.g., anisotropy, lower crustal or mantle heterogeneities). In contrast to our study, Naliboff et al. (2020) use a range of initial strain across the same range (0.5–1.5) at every model grid point, and resulted in significantly longer time (i.e., >10⁶ yrs) to localize deformation, which, when combined with our results, indicates that varying initial strain distributions may promote other modes of normal fault growth. Our results suggest that inherited lithospheric heterogeneity (due to pervasive damage as explored in this study, and/or due to the presence of discrete, pre-existing faults), would play a major role in the rapid development of faults, as documented in the East African Rift (e.g., Klerkx, 1998; Laó-Dávila et al., 2015), South Africa (Paton, 2006), and the North Sea (Whipp et al., 2014), but that the style and distribution of deformation (and whether faults are reactivated or not) is dependent on crustal and lithospheric variations and their relative strength contrast at both local and regional scale (e.g., Brune, 2016; Buck, 1991; Buck et al., 1999; Corti, 2012; Corti et al., 2003, 2018; Sokoutis et al., 2007; Ziegler & Cloetingh, 2004).

Distinguishing between currently debated fault growth models has direct implications for the nature of earthquake slip and potential maximum moment magnitude. As earthquake magnitudes are proportional to their rupture length, the propagating fault model would require a progressive temporal increase in the maximum earthquake magnitude and a decrease in recurrence for constant slip rates, whereas the constant-length model with constant slip rates can arise from invariant earthquake magnitude and recurrence (Nicol et al., 2005). Whereas our results demonstrate that finite fault lengths were rapidly established (i.e., constant-length model), they do not explicitly support either of the two slip models, instead showing that active deformation is temporally and spatially variable (i.e., earthquake slip is variable, not uniform, in magnitude and location), an observation consistent with episodic slip patterns characterizing active fault networks (e.g., Benedetti et al., 2002; Friedrich et al., 2003; Mitchell et al., 2001; Nicol et al., 2006; Oskin et al., 2007; Schlagenhauf, 2008, 2010, 2011).

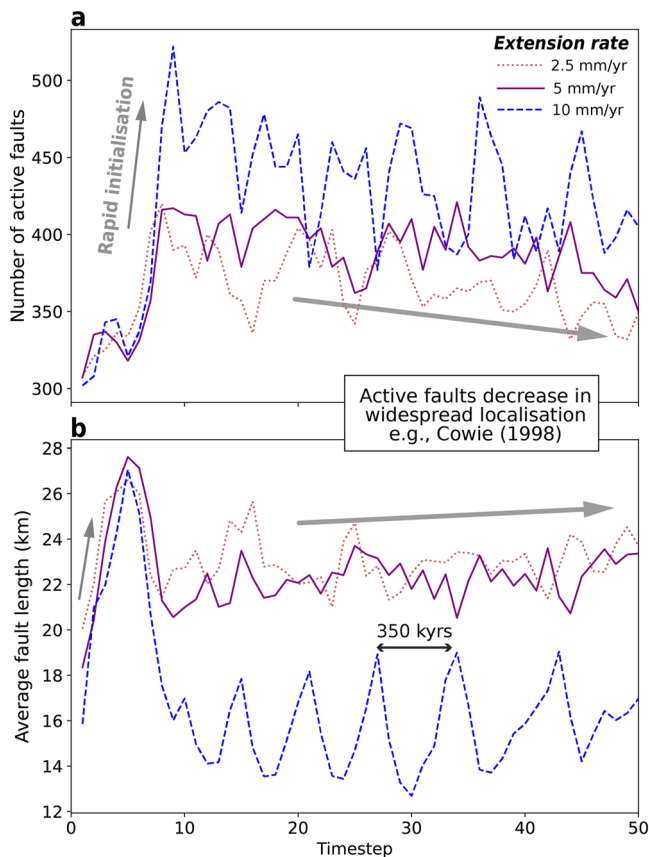


Figure 7. Geometric statistics of time-dependent fault properties. (a) The number of active faults through time. (b) The average active fault length through time. Note that while the plot contains all output timesteps for each model (50 total), the total duration for models deformed at 2.5, 5 and 10 mm yr⁻¹ are 10, 5 and 2.5 Myrs, respectively. As such, each timestep along the horizontal axis corresponds to an equivalent amount of total extension experienced by each model, rather than the same model time.

The across-strike migration of strain is spatially correlated to along-strike bends (Movie S4–S6), supporting observations from active settings that earthquakes commonly occur at segment boundaries (DuRoss et al., 2016), and that relays ramps may be associated with throw rate enhancements (Faure-Walker et al., 2009; Iezzi et al., 2018). This transient behavior evolves without explicitly modeling the earthquake cycle via a rate or rate-state friction type rheology (e.g., Van Dinther et al., 2013), suggesting that the recurrence of large, clustered slip (e.g., Figure 8a) can be produced by fault interaction and stress transfer within fault networks deforming at constant rates of tectonic extension. Faster extension rates (10 mm yr⁻¹) result in greater fluctuations between the aforementioned endmembers (Figure 7), resulting in sudden, large, recurring through-going slip events (Figure 8). The end-member behavior exhibited from the “localized” and “distributed” active field is consistent with spatiotemporal clustering of earthquakes promoted by stress interactions between neighboring faults (Stein, 1999).

While the transient nature of active deformation exhibits short-term variability, the finite plastic strain field visually appears to accumulate strain at near-constant rates (i.e., long-term stability). Note that the finite plastic strain field appears identical at 2.1 and 2.2 Myrs (Figures 8b and 8e) even though incremental slip (Figures 8a and 8d) occurred. We suggest that short-term variability and long-term stability of strain accumulation depend on how deformation is spatially and temporally averaged. As the area in which deformation is summed increases, the strain rate profile along the rift axis becomes increasingly uniform as strain deficits in one location are compensated for by increased strain in other, across-strike locations (Movie S7). Rift-wide strain profiles (i.e., the summation of deformation across the analyzed model in Figure 5) result in constant rates directly proportional to the extensional velocity applied to the crust. The increased stability of strain rates at greater spatial scales supports the ideas of Nicol et al. (2006), who suggest that each fault is a component of a kinematically coherent system, in which all faults interact and have interdependent earthquake histories. Our results suggest that a kinematically coherent system is established from the onset of extension, giving rise to the rapid establishment of the fault pattern observed in the models (e.g., Figures 4 and 5).

The accommodation of small-scale strain that leads to the accumulation of overall uniform strain may highlight sampling bias limitations. A temporal bias may occur, given that measured slip rates are more likely to be obtained from episodic fault activity akin to the “localized” endmember shown from the model results, rather than interseismic periods of aseismic slip, diffuse deformation, and relative quiescence (i.e., the distributed endmember in Figures 8d–8f), which are less likely to be recorded. Alternatively, a spatial bias may occur as it may be more likely that only the most prominent fault scarps are mapped (e.g., Nicol et al., 2009) and only the largest faults interpreted in seismic data (e.g., Pan et al., 2022), as discussed by Cowie et al. (2012). Small-scale, distributed deformation in the form of near-fault drag folding can account for up to 30% of the geodetically derived slip rates (Oskin et al., 2007). We suspect measured rates of slip in our models could be higher if the spatial scale of observation is able to recover all distributed deformation, particularly at higher extension rates (10 mm yr⁻¹) where distributed deformation is relatively widespread throughout the crust (Figure 5).

Our results suggest that conventional D-L profiles measured for static fault geometries may provide only a limited understanding of fault growth, given they do not capture stress- and time dependent stress interactions crucial to revealing the short- to intermediate-timescale variations in faulting that control earthquake magnitude and location. Our work suggests that a better understanding of fault interaction and stress feedback mechanisms are crucial to understanding and potentially forecasting patterns of fault growth and slip accumulation. Further observational data that capture strain migration (most likely in active rift settings) may therefore better constrain

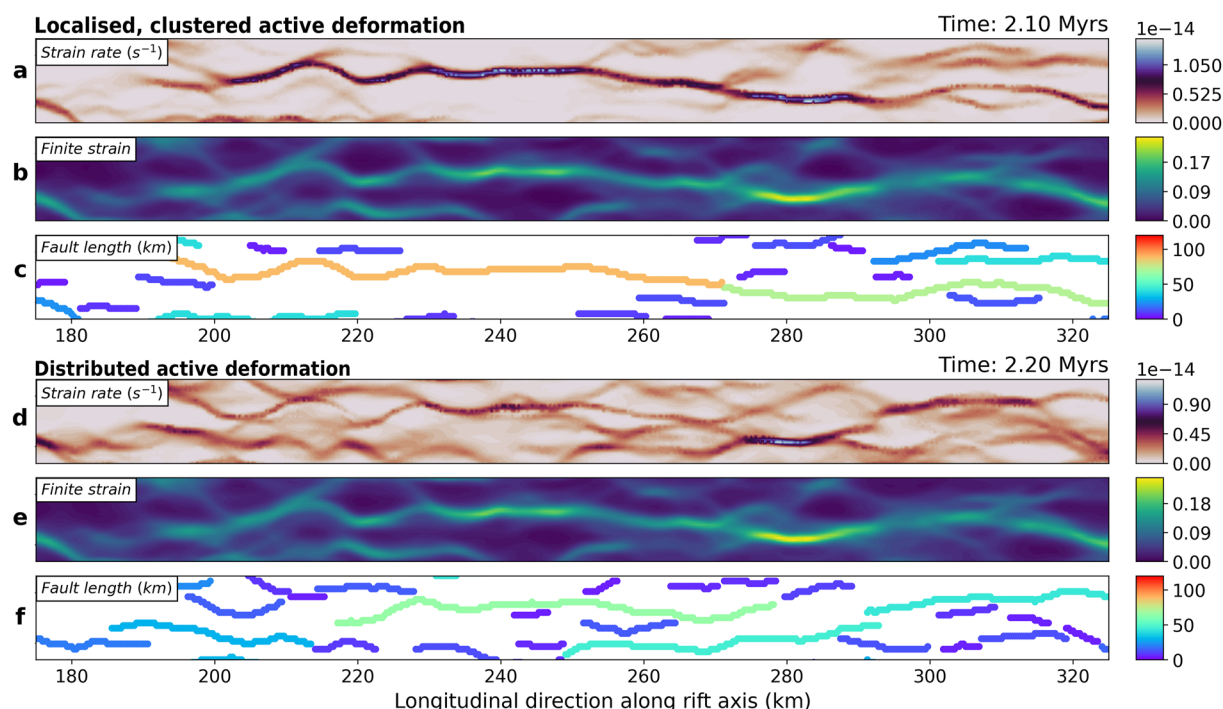


Figure 8. End-member behavior of transient deformation. Along-strike maps from a subset of the model that experienced 10 mm yr^{-1} extension. The strain rate second invariant (a), finite strain (plastic strain invariant) (b), and extracted faults (c) at 2.1 Myrs reveal localized, continuous behavior. The strain rate second invariant (d), finite strain (e) and extracted faults (f) at 2.2 Myrs reveal distributed, segmented behavior.

the mechanisms of stress feedback and subsequent geohazards. Numerical modeling coupled to high-resolution geomorphology and surface processes may further enable our practical understanding of slip distribution and recurrence. Additionally, higher resolution modeling and investigation of initial strain distribution parameters will improve our understanding of early fault growth processes.

6. Conclusions

Numerical models of continental extension allow for a better understanding of normal fault growth via incremental slip accumulation. Modeled fault patterns are rapidly formed from the onset of extension (<100 kyrs from initiation) due to competing stress interactions from an initially randomised distribution of plastic strain. Subsequent strain accumulation is highly transient and active deformation oscillates between being localized and distributed throughout the crust. Our findings demonstrate that fault network evolution is more complex than currently inferred from observing finite displacement patterns on now-inactive structures (i.e., finite strain in Figures 8b and 8e appear nearly identical), which provide only a time-averaged picture of fault kinematics. Short-term slip rates subsequently capture a transient snapshot of the long-term average. We suggest that short-term slip rates coupled to a better understanding of fault interaction and strain migration may greatly benefit our overall understanding of fault network evolution and their underpinning seismic hazards.

Data Availability Statement

The parameter file and additional inputs required to reproduce the models results are provided in the supplementary material. Models were run with ASPECT 2.2.0-pre with deal. II 9.1.1 on 720 processors (15 nodes). This version of ASPECT can be obtained with git checkout ab5eead39 from the main branch after cloning ASPECT from GitHub (see <https://doi.org/10.5281/zenodo.5131909>). All modelled timesteps are stored in interpolated numpy format and can be accessed from: <https://doi.org/10.6084/m9.figshare.19750408.v1>. The Python codes used for the automated fault extraction, data analysis and figure generation are included as supplementary files.

The python script used to generate the initial composition field and geothermal profile are included as supplementary files. All supplementary files are stored and accessible in <https://doi.org/10.6084/m9.figshare.19750408.v1>.

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