Deformation and seismicity associated with continental rift zones propagating toward continental margins

V. Lyakhovsky and A. Segev
Geological Survey of Israel, 30 Malkhei Israel Street, Jerusalem 95501, Israel (vladi@geos.gsi.gov.il)

U. Schattner
Dr. Moses Straus Department of Marine Geosciences, Charney School of Marine Sciences, Faculty of Natural Science, University of Haifa, Haifa 31905, Israel

R. Weinberger
Geological Survey of Israel, 30 Malkhei Israel Street, Jerusalem 95501, Israel

[1] We study the propagation of a continental rift and its interaction with a continental margin utilizing a 3-D lithospheric model with a seismogenic crust governed by a damage rheology. A long-standing problem in rift-mechanics, known as the tectonic force paradox, is that the magnitude of the tectonic forces required for rifting are not large enough in the absence of basaltic magmatism. Our modeling results demonstrate that under moderate rift-driving tectonic forces the rift propagation is feasible even in the absence of magmatism. This is due to gradual weakening and “long-term memory” of fractured rocks that lead to a significantly lower yielding stress than that of the surrounding intact rocks. We show that the style, rate and the associated seismicity pattern of the rift zone formation in the continental lithosphere depend not only on the applied tectonic forces, but also on the rate of healing. Accounting for the memory effect provides a feasible solution for the tectonic force paradox. Our modeling results also demonstrate how the lithosphere structure affects the geometry of the propagating rift system toward a continental margin. Thinning of the crystalline crust leads to a decrease in the propagation rate and possibly to rift termination across the margin. In such a case, a new fault system is created perpendicular to the direction of the rift propagation. These results reveal that the local lithosphere structure is one of the key factors controlling the geometry of the evolving rift system and seismicity pattern.

Components: 12,100 words, 14 figures, 1 table.

Keywords: continental margin; continental rifting; damage rheology; numerical simulations; tectonic force.

Index Terms: 8002 Structural Geology: Continental neotectonics (8107); 8020 Structural Geology: Mechanics, theory, and modeling.

Received 17 October 2011; Revised 1 December 2011; Accepted 2 December 2011; Published 20 January 2012.

1. Introduction

During the last decades considerable efforts were dedicated to modeling of continental and oceanic rifting [e.g., Keen, 1985; Parmentier and Schubert, 1989; Houseman and England, 1986, Sonder and England, 1989; Dunbar and Sawyer, 1989; Melosh, 1990; Negredo et al., 1995; Buck and Poliakov, 1998]. Many authors have discussed how tectonic processes could produce relative tension that initiates and propagates rift zones [e.g., Forsyth and Ueda, 1975; Solomon et al., 1980] and estimated the magnitude of the tectonic rift-driving forces. Spohn and Schubert [1982] and Bott [1991] related the tectonic rift-driving forces to the uplift over a region of abnormally hot mantle and provided the scaling relations between extensional forces, uplift magnitude and the depth of the low density compensation level. Both analytic and semi-analytic models [e.g., McKenzie, 1978; Buck, 1991] as well as numerical simulations [e.g., Braun and Beaumont, 1989; Bassi, 1991; Dunbar and Sawyer, 1989] assume that the tectonic force required to initiate rifting is available. Following these ideas, most recent thermomechanical rifting models combine asthenosphere doming with passive extension of the lithosphere with a given rate of remote plate motion (e.g., Huismans et al. [2001], Ziegler and Cloetingh [2004], Rosenbaum et al. [2010], and others). These models do not consider the available tectonic forces and energy required for the extension.

Lithospheric strength profiles typically include a brittle upper crust overlying a viscous lower crust and a mantle. This viscous part can behave as a ductile layer or change from brittle to ductile behavior with depth [Goetze and Evans, 1979; Brace and Kohlstedt, 1980; Kohlstedt et al., 1995]. The resulting curve is known as the Brace-Goetze strength profile. It provides a general and simple framework for estimating lithosphere strength. This strength is defined as an integrated differential stress over the whole lithosphere under a constant strain rate and given depth-dependent temperature. Based on these considerations, Vink et al. [1984] noted that a continental lithosphere, with its granitic upper crust and deep Moho interface, should be weaker than an oceanic lithosphere. The mechanical behavior of the oceanic lithosphere, with a relatively shallow upper mantle, is dominated by a stronger olivine rheology. Therefore, rift propagating through a continental lithosphere might cease before approaching a transition to an oceanic lithosphere. Buck [2004, 2006, 2009] estimated the minimum tectonic force needed to allow slip on optimally oriented normal faults in a typical continental lithosphere (i.e., 30 km thick crust) with a thermal profile corresponding to surface heat flow below 60 mW/m². He concluded that the available forces are not large enough for rifting in the absence of basaltic magmatism (the “tectonic-force paradox”). It was suggested that continental rifting is always associated with simultaneous dike emplacement and tectonic forces.

The above conclusions are based on the assumption that tectonic stress has to overcome the yielding stress integrated over the whole lithosphere thickness ignoring any weakening mechanism, either in the brittle upper crust or in the olivine-dominated upper mantle. In spite of intensive experimental and theoretical studies, mantle weakening processes are difficult to constrain [e.g., Karaoto, 2010, and references therein]. Among these processes are water weakening, transition from diffusion to dislocation creep in olivine, and perhaps the most important effect is the influence of the change in grain-size. Bercovici et al. [2001] developed a phenomenological two-phase damage theory that has been extended to consider the contribution of grain size reducing damage to the development of strain localization and strength reduction of the lithosphere [Bercovici and Ricard, 2005; Landuyt et al., 2008; Ricard and Bercovici, 2009; Landuyt and Bercovici, 2009]. The two-phase damage theory as well as a new mathematical formulation of continuum damage mechanics [Karrech et al., 2011] are mostly relevant for deep lithospheric layers. However, they are not appropriate for a detailed analysis of brittle failures in the seismogenic zone where elastic deformation and brittle fracturing play the dominant role.

At low temperatures and low pressures, typical for the seismogenic zone, a material is deformed by brittle failure rather than viscous flow. The pervasive damage in the form of micro-cracks and voids develops as part of rock formation, and it usually increases during tectonic loading leading to degradation of strength and rock elasticity. Weakening of the brittle rocks with plastic strain accommodated by evolving fault zones in models of rifting has been considered through the decrease of the cohesion [e.g., Lavie et al., 2000] and the angle of internal friction [e.g., Behn et al., 2002; Huismans and Beaumont, 2007, 2011]. Buck [2009] noted that the reduction in strength of the brittle layer due to cohesion loss is fairly small compared to the strength that remains due to rock friction. Based on rate-dependent frictional strength change observed in laboratory studies, Behn et al. [2002] suggested the model with a degree of strain rate softening...
proportional to the logarithm of the sliding velocity. The adopted velocity weakening of fault zones results in efficient strain localization, but has a minor effect on the overall strength of the brittle layer. Significant strength reduction of the brittle layer was introduced by Huismans and Beaumont [2007, 2011] and others. They assumed an unrealistically low friction angle of 15° (friction coefficient 0.27) for the intact rock, which is not supported by laboratory experiments, and further decreasing its value with accumulated strain up to 1 or 2° (friction coefficient below 0.03). This mathematical model predicts formation of essentially frictionless fault zones.

We study the initiation of a non-magmatic continental rift and the interaction of a propagating rift with a continental margin. We utilize 3-D simulations of the tectonic motion and faulting process in a layered model of the lithosphere. These layers consist of a brittle upper crust governed by a nonlinear continuum damage rheology, overlying a visco-elastic upper mantle of wet olivine. The employed damage rheology constrained by various observations from laboratory friction and fracture experiments provides an adequate tool for studying the evolution of geometrical and material properties of crustal fault zones, along with the evolution of various types of seismicity patterns. Results of the simulations demonstrate that gradual weakening and healing of the brittle rocks during earthquake cycle allows rifting under moderate tectonic stress and provides a feasible solution for the tectonic-force paradox. Results also demonstrate how changes in the Moho depth affect the geometry of a propagating rift toward continental margin, the associated strain field, and its seismicity pattern.

The geology of the Eastern Mediterranean motivated this study providing two examples of continental rifts, the Suez rift [Steckler and ten Brink, 1986] and the Azraq-Sirhan rift [Sharland et al., 2001; Schattner et al., 2006; Segev and Rybakov, 2010, 2011]. Both rifts propagated during the Oligocene toward the northwest and terminated at the Levant continental margin. Some aspects concerning these rift zones and the Andaman Sea back-arc extensional basin will be shortly discussed in the light of the present results.

2. Model Formulation

We study the initiation of a non-magmatic continental rift and the interaction of a propagating rift with a continental margin. We utilize 3-D simulations of the tectonic motion and faulting process in a layered model of the lithosphere. These layers consist of a brittle upper crust governed by a nonlinear continuum damage rheology, overlying a visco-elastic upper mantle of wet olivine. The employed damage rheology constrained by various observations from laboratory friction and fracture experiments provides an adequate tool for studying the evolution of geometrical and material properties of crustal fault zones, along with the evolution of various types of seismicity patterns. Results of the simulations demonstrate that gradual weakening and healing of the brittle rocks during earthquake cycle allows rifting under moderate tectonic stress and provides a feasible solution for the tectonic-force paradox. Results also demonstrate how changes in the Moho depth affect the geometry of a propagating rift toward continental margin, the associated strain field, and its seismicity pattern.

2.1. Mechanics of Rock Deformation

2.1.1. Brittle Deformation

Hey et al. [1980] considered rift propagation analogous to mode-I crack propagation in an elastic material, where the crack propagation is controlled by the stress concentration at its tip. Linear elastic
fracture mechanics postulates that an isolated crack propagates at velocities approaching the speed of sound in a medium, once a critical stress intensity factor is reached [e.g., Freund, 1990]. At lower values of the stress intensity factor the crack remains stable. Therefore, the classical crack theory introduces no limiting physical mechanisms to determine the rate of rift propagation. Several studies [e.g., Phipps Morgan and Parmentier, 1985; Schubert and Hey, 1986; Parmentier and Schubert, 1989] suggested that viscous suction near the tip of a propagating rift segment causes it to resist propagation and could control its rate. Rocks and rock-like materials, subjected to long-term loading, show an accumulation of micro-cracks in the process zone around the tip of a macroscopic crack. The micro-cracks are localized in a narrow weak zone and their coalescence leads to a slow extension of the macroscopic crack at the stress intensity factor values significantly lower than its critical value. This phenomenon, known as subcritical crack growth [e.g., Swanson, 1984; Atkinson and Meredith, 1987; Cox and Scholz, 1988], is another mechanism that controls the rate of rift propagation. A realistic rheological model of rupturing processes should include a subcritical crack growth from very early stages of the loading, material degradation due to increasing micro-crack concentration, macroscopic brittle failure, post failure deformation, and healing. Continuum damage mechanics based on pioneering works by Robinson [1952], Hoff [1953], Kachanov [1958, 1986], Rabotnov [1959, 1988] and further developed in engineering [e.g., Hansen and Schreyer, 1994; Kachanov, 1994; Kračinović, 1996; Lemaitre, 1996; Allix and Hild, 2002] and earth sciences [e.g., Newman and Phoenix, 2001; Shcherbakov and Turcotte, 2003; Turcotte et al., 2003; Bercovici et al., 2001; Bercovici, 2003; Ricard and Bercovici, 2003; Bercovici and Ricard, 2003, Karrech et al., 2011] provide a framework for the rheological model of the faulting and rupturing processes.

[10] In this study we use the damage rheology model of Lyakhovsky et al. [1997] and Hamiel et al. [2004], further developed by Lyakhovsky et al. [2011] and Hamiel et al. [2011]. The model accounts for the following general aspects of brittle rock deformation: (1) nonlinear elasticity that connects the effective elastic moduli to a damage variable and loading conditions; (2) Evolution of the damage state variable as a function of the ongoing deformation and gradual conversion of elastic strain to permanent inelastic deformation during material degradation; (3) Macroscopic brittle instability at a critical level of damage and related rapid conversion of elastic strain to permanent inelastic strain. This approach was adopted by Ben-Zion et al. [1999], Lyakhovsky [2001], Ben-Zion and Lyakhovsky [2002, 2006], Lyakhovsky and Ben-Zion [2009], Ben-Avraham et al. [2010], and Finzi et al. [2009, 2011] to study the evolution of geometrical and material properties of crustal fault zones, along with the evolution of various types of seismicity patterns.

[11] The damage rheology framework allows calculating the simultaneous evolution of damage, \( \alpha \), and its localization into narrow highly damaged zones (faults), earthquakes, and associated deformation fields. Synthetic earthquake catalogs generated during the model run enable analyzing the coupled evolution of faults and seismicity pattern. The elastic moduli of the damage-free rocks (\( \alpha = 0 \)) are estimated from typical values of the rock density and seismic wave velocities (Table 1) widely accepted in the geophysical literature. The condition for the onset of damage accumulation is derived from the basic thermodynamic balance equations [e.g., Lyakhovsky et al., 1997] and is formulated in terms of the strain invariants ratio, \( I_1/\sqrt{I_2} \) (\( I_1 = \varepsilon_{kk} \) and \( I_2 = \varepsilon_{ij} \varepsilon_{ij} \) are two invariants of the elastic strain tensor \( \varepsilon_{ij} \)). The critical value of the strain invariants ratio at the onset of damage accumulation is connected to the internal friction...
angle of the Byerlee’s law [Byerlee, 1978]. For most rocks the coefficient of friction varies between 0.5 and 0.8; a value of 0.6 is a good approximation for the general use, corresponding to the critical strain invariants ratio, $I_1/\sqrt{I_2} \sim -0.8$. The tensile stress (vertical minus horizontal components) required for the onset of damage for the intact rock ($\alpha = 0$; Figure 2) is identical to the Brace-Goetze strength profile for the upper brittle layer. With the onset of damage the rate of its accumulation is controlled by the applied loading and the model kinetic coefficient, $C_d$. Its value is associated with the duration of the sample loading from the onset of acoustic emission till its macroscopic failure in the rock mechanics laboratory experiment. The laboratory-constrained values $C_d = 1–10 \text{ s}^{-1}$ are important for understanding the damage accumulation process leading to failure, but have a minor effect on the geometrical properties of the new-created fault zones and earthquake statistics in regional-scale models [e.g., Finzi et al., 2009; Lyakhovsky and Ben-Zion, 2009]. Damage accumulation leads to significant material weakening toward the seismic event. Lyakhovsky and Ben-Zion [2008] developed a mathematical procedure for the local stress drop that utilizes the Drucker-Prager model, which generalizes the classical Coulomb yield condition for cohesion-less material. They use scaling relations between the rupture area and seismic potency values established in the seismology and typical range of the stress drop (1–10 MPa) during earthquakes to calibrate parameters of their local stress drop procedure and connected them to the dynamic friction ($\sim0.2$) of simpler models with planar faults. A depth-dependent yielding function for the highly damaged zone ($\alpha = 1$) corresponding to the frictional sliding during the simulated seismic event is shown in Figure 2. When the slip associated with the macroscopic brittle failure is arrested, the post-failure material recovery starts. The recovery of elastic moduli and material strengthening is associated with healing of microcracks and is facilitated by a high confining pressure, a low shear stress and high temperatures. Motivated by the observed logarithmic increase of the static coefficient of friction [e.g., Dieterich, 1978, 1979], Lyakhovsky et al. [1997] used an exponential damage-dependent function with coefficients $C_1, C_2$ for the kinetics of the healing. The rate of the damage ($\alpha$), which decreases under a 3-D compaction strain ($\varepsilon$), is

$$\frac{d\alpha}{dt} = -C_1 \cdot \exp \left( \frac{\alpha}{C_2} \right) \cdot \varepsilon^2. \quad (1)$$

Lyakhovsky et al. [2005] showed that the damage model with exponential healing (1) reproduces the main observed features of rate- and state-dependent friction and constrained the coefficients $C_1, C_2$ by comparing model calculations with laboratory frictional data. A field-based constrain of the healing parameters is given by Finzi et al. [2009, 2011], but leaves room for a relatively wide range of possible values. Figure 3 shows the material recovery starting from a total failure ($\alpha = 1$) and lasting up to a thousand years under a constant compaction strain corresponding to 200 MPa confining pressure ($\sim8$ km depth in the earth crust). For every set of

![Figure 2. Strength profile for the damage free upper brittle layer ($\alpha = 0$) and yielding profile for the highly damaged zone ($\alpha = 1$). Colored lines show recovered strength profile hundred years after the seismic event for various kinetic coefficients (see Figure 3 for values and details).](image-url)
the model parameters, the rate of healing is the highest during the initial stage of healing (see Figure 3b for one year zoom in) and becomes slower afterwards, keeping the strength of the damaged rock well below the strength of the intact rock ($\alpha = 0$). We demonstrate the effect of healing on the simulation results with three different sets of the model parameters corresponding to fast and more efficient healing (FH), relatively slow and low healing (LH) and an intermediate case, called here “normal healing” (NH). Colored lines in Figure 2 show the recovered strength profile hundred years after the seismic event for three sets of the healing coefficients adopted in this study.

Brittle failure is apparently irrelevant for the upper mantle accounting for the lack of seismicity below the Moho interface. However, Lindenfeld and Rumpker [2011] recently detected several earthquakes located well within the mantle (53–60 km depth) beneath the East African Rift. Keeping this option in mind, we also allow damage accumulation in the upper mantle using the same conditions for onset of damage accumulation as in the upper brittle crust. We note that the damage healing is much more efficient at the upper mantle than the brittle crust due to a high confining pressure and high temperatures.

2.1.2. Ductile Flow

Ductile deformation can be described by empirical constitutive relations that express the strain rate, $\dot{\varepsilon}$, as a function of stress and temperature for the dominant lithologies comprising the continental and oceanic lithospheres [e.g., Karato, 2010, and references therein]. The common lithologies of the continental lithosphere are quartz-diorite (upper crust), diabase (lower crust) and dunite/olivine (mantle). Most of the experimental constitutive relations are established for minerals, but it is assumed that the weakest among the dominant minerals controls the flow rate of the entire rock [Brace and Kohlstedt, 1980]. Steady state plastic flow of rocks can take place by a variety of mechanisms including dislocation creep, solid state diffusion, and solution-diffusion-

![Figure 3](image-url)
precipitation processes [e.g., Kohlstedt et al., 1995]. Direct observations of ductile shear zones and indirect analyses of seismic anisotropy in mantle rocks demonstrate that dislocation creep dominates at stresses above 10 MPa [e.g., Karato, 2010]. A power law rheology is widely accepted for expressing the strain rate of the dislocation flow [e.g., Weertman, 1978]

\[ \dot{\varepsilon} = A\sigma^n \exp \left( -\frac{Q}{RT} \right) \]  

(2)

The parameters \( A \) and \( n \) are empirical constants, \( Q \) is the activation energy, \( V^* \) is the activation volume, \( P \) is pressure, \( T \) is temperature, and \( R \) is the gas constant. \( A, n \) and \( Q \) values are presented in Table 1. For relatively low pressures corresponding to depths less than 100 km, the \( PV^* \) term in (2) is negligible. For the lower crust we use the material constants appropriated for diabase rocks. Since our study is mostly focused on the brittle processes in the crust, we utilize the common approach of water-weakening of the upper mantle adopting wet olivine rheology below the Moho interface.

[14] The ductile flow pattern in the lower crust and upper mantle is strongly affected by the depth-dependent temperature distribution, corresponding to a regional heat flow. Following Turcotte and Schubert [2002], we assume that the heat production due to radioactive elements decreases exponentially with depth and use the depth-dependent temperature \( T(z) \):

\[ T(z) = T_0 + \frac{q \cdot z + \rho_c \cdot H_r \cdot h_r^2}{k} \left( 1 - e^{-z/h_r} \right) \]  

(3)

where \( T_0 = 300 \) K is the surface temperature; \( k = 3.35 \) Wm\(^{-1}\)K\(^{-1}\) is the coefficient of thermal conductivity; constant heat flux \( q = 50 \) mW/m\(^2\) is used for a relatively cold lithosphere; \( H_r \) is the surface radiogenic heat production and \( h_r \) is the length scale of its exponential decay. We take \( \rho_c \cdot H_r = 3.7 \times 10^{-3} \) mW/m\(^3\) and \( h_r = 10 \) km for the radiogenic component of the continental heat flux. The depth-dependent temperature distribution (Figure 1b) is kept constant during the simulation.

2.2. Rift-Driving Forces and Model Boundary Conditions

[15] The regional stresses in the lithosphere are commonly estimated using a thin plate approximation that treats the velocity and stress distributions in terms of values averaged over the plate thickness [e.g., Elsasser, 1969; England and McKenzie, 1982]. For the problem of a long-term deformation, the England-McKenzie model uses a viscous plate riding over an invicid substrate, assuming a complete decoupling between plate and substrate. This approach lacks an appropriate description for a relatively short-term elastic behavior of the lithosphere and mantle. The Elsasser model uses elastic plate riding over a viscous substrate and therefore, accounts for the drag forces transmitted from the large-scale convection currents in the upper mantle to the base of the plate. Rice [1980], Lehner et al. [1981] and Li and Rice [1987] provided a generalized Elsasser model replacing the viscous rheology of the substrate with a visco-elastic one. Few attempts have been made to calculate crustal deformation in a 3-D model with a depth-dependent rheology [e.g., Ben-Zion et al., 1993; Reches et al., 1994]. The results from those studies show that outside the space-time domain of a large earthquake, the 3-D calculations do not differ significantly from those obtained by the Elsasser and generalized Elsasser models. Reches et al. [1994] presented a direct comparison between a fault-parallel velocity averaged over the thickness of the elastic layer from 3-D calculations and the analytical solution of Li and Rice [1987] for the generalized Elsasser model. After about 50 years into the earthquake cycle the difference between the 2-D and 3-D models is negligible and both models fit the available geodetic data with similar accuracy. Lyakhovsky et al. [2001] proposed some modifications to the generalized Elsasser model following the common seismological assumption [e.g., Kasahara, 1981; Stein et al., 1994] that the instantaneous response of the Earth to sudden stress redistribution is accommodated by an elastic half-space. This assumption is realistic, since the strain relaxation is not instantaneous even for a ductile lower crust. Their modified model provides improvements over the generalized Elsasser model in simulating the elastic component of the geodetic signals, especially close to the master fault.

[16] Following the framework of the original Elsasser model, the horizontal components of a stress tensor, \( \sigma_{km} \), averaged over a plate thickness \( H \) are

\[ \sigma_{km}(x, y) = \frac{1}{H} \int_{-H}^{0} \sigma_{km}(x, y, z) dz. \]  

(4)
In this case, gradual recovery of the local stress field to its regional level, $\tau^{(R)}_k$, calculated using (6), can be approximated by

$$\frac{\partial \tau_k}{\partial t} + \frac{1}{t_{load}} (\tau_k - \tau^{(R)}_k) = 0,$$

(7)

where $t_{load}$ is the characteristic time of the post-seismic relaxation. Substituting (6) for the regional stress into (7) leads to the equation for the drag forces applied at the vertical edges of the simulated volume:

$$\frac{\partial \tau_k}{\partial t} = \frac{1}{t_{load}} \left[ \frac{\eta}{h} \left( V^{(k)}_{\text{plate}} - \frac{\partial u_k}{\partial t} \right) - \tau_k \right].$$

(8)

Equation (8) assumes that the forces driving the rifting process depend on a mismatch between the mantle flow velocity and the rate of extension of the simulated volume. The rate of the stress accumulation is defined by the characteristic time of the post-seismic relaxation, $t_{load}$. In a case of a strong upper layer and lack of a rifting process, the driving forces gradually approach their maximum value

$$\tau_{\text{max}} = \frac{V_{\text{plate}}}{h} \frac{\eta}{h}.$$

(9)

Following this formulation, variable boundary forces are applied to the vertical edges of the simulated volume. Figure 4 shows a schematic plain-view structure of the boundary conditions. A free slip condition is applied to the edges oriented perpendicular to the direction of the propagating rift zone. Virtual springs connected to the edges oriented parallel to the rift trace schematically represent variable tensile boundary forces. Using (8), force values for every boundary node are proportional to the velocity difference between the prescribed constant far-field plate motion and calculated rate of the boundary node motion. These boundary conditions account for the evolution of the elastic properties and accumulated plastic strain in the model region [Lyakhovsky and Ben-Zion, 2009]. A free surface boundary condition is applied to the upper surface, which is the top of the simulated volume. A free slip in the horizontal direction combined with a force balance of the vertical component is applied to the bottom surface. The vertical force component acting on the bottom surface is calculated assuming depth-dependent lithostatic pressure in the upper mantle.

2.3. Numerical Approach and Model Setup

[18] The numerical simulations were done using the Fast Lagrangian Analysis of Continua (FLAC) algorithm [Cundall and Board, 1988; Cundall, 1989;...
This fully explicit numerical method relies on a large-strain explicit Lagrangian formulation originally developed by Cundall [1989] for elasto-plastic rheology and implemented in the well-known FLAC 2D software produced by ITASCA. Poliakov et al. [1993] developed an adaptive time stepping and applied the FLAC algorithm for visco-elasto-plastic rheology. Their adaptive scheme does not require iterating, which makes the numerical model stable even for highly nonlinear damage rheology [Lyakhovsky et al., 1993]. Physical instability is modeled without numerical instability since inertial terms are included in the equilibrium equations. A modified version of this code incorporating heat transport is known as PAROVOZ (locomotive in Russian) and is widely used by many researchers. Ilchev and Lyakhovsky [2001] developed their own 3-D version of the code for visco-elastic damage rheology [see also Hamiel et al., 2004]. This 3-D code was utilized for several studies of a long-term tectonic motion and faulting process in a layered model of the lithosphere (Figure 1); among them recently reported results of Lyakhovsky and Ben-Zion [2009], Ben-Avraham et al. [2010], and Finzi et al. [2009, 2011].

Initially we simulated rifting processes in horizontally layered models (i.e., “flat” structure), corresponding to typical continental or oceanic lithospheres. The simulated volume of 150 \times 150 km and 70 km depth was divided into tetrahedral elements of variable sizes that increase gradually from \sim 1 km in the seismogenic zone to \sim 5 km in the ductile zone. The model consists of three different layers representing a weak sedimentary cover, a crystalline basement (crust) down to the Moho, and an upper mantle. In the case of the continental lithosphere, the thickness of the sedimentary cover is chosen at 3 km and that of the crystalline basement at 30 km (depth to Moho). In the oceanic lithosphere, the thickness of the sedimentary cover is 8 km, and that of the crystalline basement is reduced to 20 km (depth to Moho). These thicknesses mimic the structure of the lithosphere in the Levant area [Segev et al., 2006]. At the next stage, we constructed a model composed of an oceanic lithosphere on the west (50 km width), continental lithosphere on the east (50 km width) and a gradual transition zone (50 km width) between them, representing the continental margin. The simulated rift zone is expected to propagate perpendicular to the continental margin from east to west.

3. Results

3.1. Rifting of Lithosphere With “Flat” Structure

The rate of the mantle flow in all the simulations was \( V_{\text{plate}} = 50 \text{ mm/y} \). Rifting of a continental lithosphere was simulated for low (\( \tau_{\text{max}} = 50 \text{ MPa} \)) and moderate (\( \tau_{\text{max}} = 100 \text{ MPa} \)) tectonic loading, without any additional heating. The simulation starts with an isostatically compensated structure without any tectonic stress except for a depth-dependent lithostatic pressure. Low damage (\( \alpha < 0.2 \)) is randomly distributed through the model region except for a small (two numerical elements) highly damaged (\( \alpha = 1 \)) notch (i.e., pre-defined weak zone) in the middle of the right edge of the simulated volume, which serves as a rift initiation zone. At the initial stage, with the onset of the mantle flow, the tensile stress is accumulated in the upper crust. The total opening rate of the simulated region (Figure 5) steeply decreases from about 100 mm/y, which is close to double the \( V_{\text{plate}} \) value, to 2–3 orders of magnitude lower values corresponding to a
continuous rift opening. A failure in rift propagation (zero opening rate) occurs under a low drag force $\tau_{\text{max}} = 50$ MPa and a normal (violet line in Figure 5) or faster (not shown) rate of healing. The damage starts accumulating in the pre-defined weak zone at the right edge of the simulated volume. This zone serves as a high stress concentrator and enables rift initiation even under low tectonic loading. In the case of the successive rifting, the rift zone propagates into the simulated area and may cross the whole model or terminate inside the region, depending on the tectonic forcing and the rate of healing.

Figures 6, 7, and 8 present results after 300,000 years of simulation. Figures 6a, 7a, and 8a show maps of the tensile strain component ($\varepsilon_{yy}$) in the brittle part of the crystalline basement (8 km depth); and Figures 6b, 7b, and 8b show the deformed grid together with damage distribution along sections crossing the center of the simulated volume in the y-direction. Animations in the auxiliary material present the evolution of strain and damage distribution associated with the rifting process.\(^1\) Presented results include two end-member case studies corresponding to the formation of a well-localized straight rift zone (Figure 6) and the termination of the rift zone (Figure 7). An intermediate case corresponding to a relatively wide rift zone is also shown in Figure 8.

Relatively high forcing ($\tau_{\text{max}} = 100$ MPa) together with low healing initially leads to the formation of a $\sim$50 km wide zone of deformation (Animation S1) and 70–80 km wide damage zone (Animation S2). After $\sim$200,000 years (2/3 of the total simulated time) the damage is localized into a narrow regular zone and most of the deformation is accumulated within a $\sim$20 km wide straight zone crossing the simulated area (Figure 6a). The opening rate of the rift zone is $\sim$1.5 mm/y (Figure 5) leading to thinning of the simulated structure mostly accommodated by a vertical motion of the lower boundary (Animation S3). Maximum uplift of the mantle material below the center of the rift zone is $\sim$1 km (Figure 6b), while the surface subsidence is $\sim$200 m. The ongoing uplift of the mantle material in long-term simulations (not produced in this study) supplies additional heat beneath the rift zone, changing the temperature profile. Several recent studies [Benallal and Bigoni, 2004; Regenauer-Lieb and Yuen, 2004; Kaus and Podladchikov, 2006; Regenauer-Lieb et al., 2006] noted the role of the thermomechanical feedback and demonstrated that spontaneous strain localization in the mantle occurs during the extension of the continental lithosphere [e.g., Weinberg et al., 2007; Rosenbaum et al., 2010]. The thermomechanical coupling provides an additional localization mechanism, accelerating the rift propagation at low stresses.

Reduced forcing and more efficient healing prevent a localization of the rift zone and may even lead to its termination. Such a scenario is obtained for $\tau_{\text{max}} = 50$ MPa and normal healing (Figure 7). Starting from the right edge of the model, the rift propagates $\sim$50 km into the simulated area, and

\(^{1}\)Auxiliary materials are available in the HTML. doi:10.1029/2011GC003927.
then forms a wide distributed strain (Animation S4), associated with a widening and a stagnation of the damage zone (Animation S5). The opening rate of the simulated volume along the section crossing its center gradually decreases and practically stops (<0.1 mm/y) after 150,000 years (Figure 5). This minor stretching is accommodated by a small (up to 100 m, Figure 7b) uplift of the lower boundary (Animation S6), which remains flat, except for some small boundary effect at the model corners. Surface subsidence is negligible. Increasing the healing efficiency (fast healing instead of normal) reduces the ability of the rift zone to develop, while decreasing the efficiency (low healing) leads to the formation of a wide rift zone with a very slow rate of opening ~0.2 mm/y (both cases are not shown here). A higher and more realistic rate of opening (0.6–0.7 mm/y) is obtained for fast healing and $\tau_{\text{max}} = 100$ MPa (green line in Figure 5). Efficient healing prevents strong localization of the strain (Animation S7) and the damage (Animation S8) into a narrow rift zone. After 300,000 years of simulation, the width of the rift zone is ~70 km (Figure 8a), and instead of a single localized deformation zone, opening is distributed between

![Figure 7](image_url)  
**Figure 7.** (a) Tensile strain ($\varepsilon_{yy}$) distribution in the seismogenic zone - map view at 8 km depth and (b) vertical profile across the center of the simulated volume. Note 20 times exaggeration of the grid deformation. The simulation with $\tau_{\text{max}} = 50$ MPa and normal healing.

![Figure 8](image_url)  
**Figure 8.** (a) Tensile strain ($\varepsilon_{yy}$) distribution in the seismogenic zone - map view at 8 km depth and (b) vertical profile across the center of the simulated volume. Note 20 times exaggeration of the grid deformation. The simulation with $\tau_{\text{max}} = 100$ MPa and fast healing.
three parallel segments with a spacing of 10–15 km between them. Thinning of the simulated structure is mostly accommodated by a vertical motion of the mantle material below the rift zone (Animation S9). The uplift of the lower boundary (~700 m) and surface subsidence (~150 m) are distributed within a wide zone with clear undulations corresponding to the different segments (Figure 8b).

Figure 9 shows the distribution of the tensile strain component ($\varepsilon_{yy}$) along the section crossing the center of the simulated volume in the y-direction after 300,000 years for all the analyzed simulations. A comparison between modeling results with different forcing and different rates of healing demonstrates that more regular and localized structures are expected under higher forcing and slow (less efficient) healing; fast healing leads to a widened and more disordered rift zone.

In the oceanic lithosphere (Figure 1a), with a shallow depth to Moho (~20 km), the temperature below the crust is significantly lower than the temperature below the continental crust with the same temperature profile (Figure 1b). Under these conditions, the olivine-rich upper mantle rocks are much stiffer and could hardly be deformed without additional heating, even accounting for their water-weakening. Moreover, the stresses needed for the onset of fracturing (i.e., rifting) of the crystalline rocks increase with depth proportionally to the confining pressure produced by the overburden sediments. Hence, fracture nucleation in the crystalline basement covered by a thick sedimentary layer (~8 km) in the oceanic lithosphere requires higher stresses than that covered by a thin sedimentary layer in the continental lithosphere. This explains why rifting fails to propagate in the oceanic lithosphere even with $\tau_{\text{max}} = 100$ MPa and succeeds only when the force is increased to $\tau_{\text{max}} = 150–200$ MPa, depending on the rate of healing. General features of the rifting process in a thin and flat oceanic lithosphere are similar to those in the continental lithosphere discussed above and are not shown here.

3.2. Propagating Rift Approaching a Continental Margin

A series of modeling with flat lithosphere structure revealed that significantly lower forcing is required for successive rifting of thick continental lithosphere than that of a thin oceanic lithosphere with the same healing parameters and temperature regime. Hence, there are a wide range of conditions leading to a successive rifting of a continental lithosphere, and to ceasing of rift propagation at the continental margin. Similar considerations lead
Vink et al. [1984], Steckler and ten Brink [1986] and others to discuss rift termination at the continental margin. Figure 10 shows one of the possible scenarios obtained in the course of this study, demonstrating the deformation pattern associated with the termination of rifting at the continental margin. The model structure (Figure 10a) includes a 50 km wide transition zone between a typical continental (on the right) and oceanic (on the left) lithosphere. In map view, the distribution of the tensile strain component, $\varepsilon_{yy}$, in the middle part of the seismogenic zone (down to $\sim$10 km depth) is shown for $\tau_{\text{max}} = 100$ MPa and normal healing (Figure 10b). A $\sim$50 km long zone of high strain values (red zone in Figure 10b) spreads from the right edge of the model, where the rift zone was initiated, to the beginning of the continental slope. The rift zone successively propagated only till the onset of the thinning of the crystalline crust (Moho uplift). Similarly, the distribution of the shear strain component, $\varepsilon_{xy}$, shows that the rift zone fails to propagate further along its own plane. Instead, new shear zones, spanning parallel to the continental slope, developed (Figure 10c). The location of these shear zones depends on the applied tectonic load and temperature profile (Figure 1b). An increase of the regional heat flux would weaken the upper mantle and enhance rift propagation across the continental margin. The symmetry of the simultaneously formed left and right-lateral shear zones stems from the idealized structure of the continental margin and the right angle between the propagating rift and the continental slope (Figure 10c). In any realistic structure a certain asymmetry is expected leading to a more pronounced development of one of these zones than the other.

### 3.3. Seismic Activity Associated With Propagating Rift Zone

[27] During the simulation of rifting using damage rheology, the numerical model produces a catalog of synthetic seismic events. The numerical procedure [e.g., Lyakhovsky and Ben-Zion, 2008] allows storing information about every numerical element involved in each event and calculating the magnitude of the synthetic earthquake. The seismic activity starts around the notch at the right edge of the model where the rift is initiated, propagates along the new formed rift zone, and continues with ongoing rift opening or ceases when the rift is terminated. The last scenario happens in the continental lithosphere with flat structure under $\tau_{\text{max}} = 50$ MPa and normal healing (Figure 11a). In this case some moderate activity was generated during the first 60,000 years; this activity lasted up to about 150,000 years, when the rift zone becomes stagnant (see also Figures 5 and 7). The model with $\tau_{\text{max}} = 50$ MPa and low healing generates significantly more seismic events during the first 10,000–20,000 years (Figure 11b), when the rate of opening is above...
1 mm/y (Figure 5). After 50,000–60,000 years, when the rate of opening decreases below 0.5 mm/y, seismic activity decreases, but as opposed to the case with normal healing (Figure 11a), the rifting process is permanently associated with seismic activity. The magnitude range of the simulated seismic events at this stage is relatively narrow around $M/C_{24}$ (orange rectangle in Figure 11b). The frequency-size earthquake statistics significantly deviates from the power-law Gutenberg-Richter distribution and resemble the characteristic one.

Some reduction in the level of seismic activity between the initial stage of the formation of the rift zone and later stage of the ongoing opening was also obtained for higher tectonic forcing ($\tau_{\text{max}} = 100$ MPa), but this change is not so pronounced as in the case with $\tau_{\text{max}} = 50$ MPa. The overall activity in the model with $\tau_{\text{max}} = 100$ MPa is significantly higher especially at the initial stage of rifting (Figure 12). The model continuously generates lots of relatively small events with a magnitude range between $M = 4.5$ (lower model cut-off according to the grid size) and $M = 5.5$. Strong events ($M > 6$) are concentrated in a time with an outstanding main shock and form earthquake swarms (Figure 12a). Five swarms were generated during the very initial stages of rifting from 5,000 to 15,000 years of simulation. During this period the depth of the simulated events (Figure 12b) gradually increases from 3 km (top basement) to ~17 km, corresponding to...

Figure 11. A synthetic record of the seismic events for the simulation with $\tau_{\text{max}} = 50$ MPa and (a) normal or (b) low healing. The orange rectangle marks the stage with characteristic frequency-size statistics.

Figure 12. (a) Magnitude and (b) depth of the synthetic seismic events for the simulation with $\tau_{\text{max}} = 100$ MPa and fast healing. The dash green line shows a deepening of the simulated seismic events at the initial stage and depth of the seismogenic zone.
the typical depth of the seismogenic zone. The seismic activity starts at depths corresponding to the top of the crystalline basement (seismicity in a weak sedimentary layer is not simulated) and then deepens gradually. At the later stage of rifting, swarms are more seldom and generated every 5,000 or even 10,000 years. During the swarms, some events occur as deep as \(20\) km and are located below typical depths of the seismogenic zone where most of the events occur. The overall log number of the simulated seismic events, including swarms, fits well the linear Gutenberg-Richter frequency-size distribution with \(b\)-value \(b = -1\) (green dots in Figure 13). The model run with normal healing and \(\tau_{\text{max}} = 100\) MPa forcing (red makers in Figure 13) leads to some deviation of the statistics from the linear trend with \(b = -0.8\). The large amount of events, generated during the initial stage of rifting, exhibit a wide range of magnitudes and an almost linear distribution (not shown here). At later stages, the rift zone becomes more regular and strong seismic events with magnitudes above \(M = 6.0\) slightly dominate. Therefore, the overall log number of the simulated seismic events shown in Figure 13 with magnitudes \(M = 6.0\) and \(M = 6.5\) are slightly above the values expected for the Gutenberg-Richter statistics. Significant deviation from the linear frequency-size distribution is obtained in model runs with low healing generating the most regular rift zone.

4. Discussion

Previous studies have discussed how tectonic processes produce relative tension to initiate and propagate rift zones and estimated the magnitude of rift-driving forces. Buck [2009] concluded that the rift-driven tectonic force allowing continental rifting due to extension is well above that available for rifting of a thick and strong lithosphere in the absence of basaltic magmatism (the “tectonic force” paradox). This conclusion is based on the assumption that the tectonic stress has to simultaneously overcome the yielding stress of the whole lithosphere thickness, and ignore the gradual rock weakening under long-term loading. Nevertheless, gradual accumulation of rock damage under ongoing, even moderate, tectonic extension, could lead to a considerable weakening of the heavily fractured brittle rocks. These rocks could remain significantly weaker than the surrounding intact rock (long-term memory), which can make rifting feasible. In the case of a flat layered structure of the lithosphere, the formation of a rift system is controlled not only by remote (tectonic) loading, but also by the rate of healing (strengthening) of the rock mass.

The numerical simulations of the plate motion and rifting process adopted water weakening in the upper mantle using dislocation creep parameters for wet olivine and a continuum damage rheology approach for brittle rocks. The damage mechanics employed in this study for brittle failure in the crystalline crust models the effects of distributed cracks in terms of a single scalar damage parameter. Representative elementary volumes with a sufficiently large number of cracks corresponding to given damage values are assumed to be uniform and isotropic. While most of the material parameters controlling ductile flow and damage accumulation...
in a brittle regime are well constrained by field and laboratory observations, values controlling the rate of post-failure material healing are non unique and may vary within a relatively wide range.

[31] Results of the modeling of a flat-layered structure demonstrate a gradual formation of the rift zone in the continental lithosphere. The successive propagation of the rift system, the localization of the rift zone and its final width, and the associated seismicity pattern strongly depend on the applied tectonic force, thickness of the crystalline crust (depth to the top basement and Moho interfaces), and the rate of the crustal rock healing. The presented results for a typical continental lithosphere structure demonstrate various scenarios including a formation of a well-localized straight rift zone (Figure 6), stagnation of the rift zone (Figure 7), and an intermediate case corresponding to a relatively wide rift zone (Figure 8). Passive continental rifting may occur under tectonic forcing as low as $\tau_{\text{max}} = 50$ MPa, but only if the rate of healing is low. Under elevated forcing of $\tau_{\text{max}} = 100$ MPa the rift is successively formed for every rate of healing adopted in this study. With increased rates of healing, the evolving damage zone becomes more disordered and wide. Narrow, localized rift zones form with low rates of healing and remain weak for a very long time. This general tendency for rifting processes is similar to previously studied structural properties and deformational patterns of evolving strike-slip faults [Lyakhovsky et al., 2001; Lyakhovsky and Ben-Zion, 2009; Finzi et al., 2009, 2011].

[32] The present results of modeling demonstrate how the lithosphere structure, in particular the depth to Moho interface, affects the force needed for successive rifting and the geometry of the propagating rift zone. Modeling of a flat-layered structure typical for the oceanic lithosphere with a 20 km depth to Moho and a 8 km top basement depth, show that rifting succeeds only when the force is increased to $\tau_{\text{max}} = 150–200$ MPa, depending on the rate of healing. With the same temperature profile, the olivine-rich upper mantle material is uplifted and has much lower temperatures than in the continental lithosphere. Under these temperatures, the upper mantle rocks are very stiff and could hardly be deformed without additional heating. In addition, fracture nucleation in the crystalline basement covered by a thick sedimentary layer in the oceanic lithosphere requires higher stresses than for that cover by a thin sedimentary layer in the continental lithosphere. This feature explains why rifting fails to propagate in the oceanic lithosphere even with $\tau_{\text{max}} = 100$ MPa and requires significantly larger tectonic forces.

[33] The tectonic evolution of the Middle East region (paleo-geographical reconstruction in Figure 14) shows that the Oligocene Suez [Steckler and ten Brink, 1986; Bosworth et al., 2005] and Azraq-Sirhan rift zones [Schattner et al., 2006; Segev and Rybakov, 2010, 2011] are propagating toward northwest and terminate at the Levant continental margin. The Suez rift is $\sim 70$ km wide and $\sim 500$ km long. It consists of the northwestern branch of the Red Sea rift separating the Arabian and the African (Nubian) plates [McKenzie et al., 1970]. The Azraq-Sirhan rift, about 40 km wide and 700 km long, is located on the northwestern edge of the Arabian continent. The initial stage of rifting occurred before the intense magmatism in the northeastern Afro-Arabia plate at $\sim 30$ Myr [Avni et al., 2011]. Similarly, Whitmarsh et al. [2001] suggested that magmatism was essentially absent during the initial stage of rifting of the Iberian-Newfoundland margins. Another possible example of non-magmatic rifting was discussed by Lie and Husebye [1994] in relation with the north Skagerrak graben of the Oslo rift system.

[34] Under our modeling conditions, a rift propagating through a continental lithosphere may terminate before a transition to an oceanic lithosphere at a continental margin ($\tau_{\text{max}} = 150$ MPa, normal healing, Figure 10). This termination induces the formation of a new fault system perpendicular to the direction of the rift propagation. A similar deformational pattern was discussed by Steckler and ten Brink [1986] who suggested that the northwest continuation of the Red Sea-Suez continental rift system stops propagating toward the Levant Basin oceanic lithosphere, leading to the initiation of the Dead Sea Transform. During the same period, the gradual thinning of the crystalline basement and the lithosphere strengthening of the Levant continental margin in the direction of the propagating Azraq-Sirhan rift system, might explain the rift termination. A further detailed study of the Azraq-Sirhan continental rift evolution will be presented in a forthcoming paper.

[35] The presented results address a relatively short geological time period (up to 300,000 years) of an early stage of rift initiation and propagation. The study is mostly concentrated on the gradual weakening of fractured brittle rocks and ignores effect of regional heating, shear heating and other mechanisms of heat generation. The time scale of the weakening effect is associated with the seismic
cycle ($10^2$ to $10^3$ years) while the thermal effect is essential at time scale of $10^6$ years. Model results demonstrate that rifting of a relatively cold continental lithosphere is possible even under moderate to low tectonic forcing. Temperature increase due to an elevated regional heat flow as well as heat generation in a localized ductile shear zone below Moho provide an additional mechanism of strain localization and acceleration of the rifting process. The role of the thermomechanical feedback at a long geological time scale was discussed in several recent studies [Benallal and Bigoni, 2004; Regenauer-Lieb and Yuen, 2004; Kaus and Podladchikov, 2006; Regenauer-Lieb et al., 2006; Weinberg et al., 2007; Rosenbaum et al., 2010] demonstrating spontaneous strain localization during the extension of a cold continental lithosphere.

[36] An important advantage of the numerical modeling using damage rheology approach is its ability to produce a catalog of synthetic seismic events. This allows us to study simultaneously deformational and seismicity patterns of evolving fault systems and to connect the style of fault zone localization and earthquake frequency-size statistics. Previous studies [e.g., Ben-Zion et al., 1999; Lyakhovsky et al., 2001; Lyakhovsky and Ben-Zion, 2009] show that if new created damage zones remain weak for a very long time, their geometry is regular and frequency-size statistics of the simulated seismic events is characteristic. With increased rates of healing, the evolving damage zone becomes more disordered and wide; frequency-size statistics become more of a Gutenberg-Richter type. A similar tendency is obtained here for the rifting process; event statistics for the model with fast healing fits well a linear Gutenberg-Richter distribution with $b = -1$ (Figure 13). Less efficient healing changes the frequency-size statistics to a characteristic distribution with event magnitude falling within a relatively narrow range. Another remarkable feature of the simulated seismicity is earthquake swarms, i.e., seismic activity concentrated in time without a distinct main shock (Figure 12). The occurrence of earthquake swarms is most common in areas of
active volcanism, but is also abundant in other regions with extensional tectonics, including continental rifting [e.g., Ibs-von Seht et al., 2008]. Several large swarms were observed in the Andaman Sea, a back-arc extensional basin in the Indian Ocean [Mukhopadhyay and Dasgupta, 2008]. The frequency-magnitude distribution of earthquakes during the largest 2005 swarm (651 registered events) is centered around $M = 4.5$. We speculate here that healing in this region is relatively to the extremely high tectonic forcing leading to repeated swarms every several years. Although the presented modeling is far from representing a realistic structure and loading conditions in the Andaman Sea, the general features of the seismic activity are expected to be similar. Frequency-magnitude distribution of earthquake swarms in continental rifts follow the Gutenberg-Richter law with $b$-value between $-0.8$ and $-1$ for the Rio Grande, Kenya, and Eger rift zones [Ibs-von Seht et al., 2008]. This behavior resembles the simulation with fast healing and $\tau_{\text{max}} = 100$ MPa (Figures 12 and 13) generating earthquake swarms and a linear frequency-size distribution.

5. Conclusions

[37] A rifting process under moderate tectonic extension is feasible due to gradual damage accumulation and long-term rock memory. These lead to a considerable rock weakening within the rift zone relatively to the surrounding intact rock and provide a possible solution for the tectonic force paradox. Results of the flat-layered model demonstrate a gradual formation of a rift zone in a continental lithosphere. A successive formation of the rift system and the associated seismicity pattern strongly depends on the applied tectonic force and the rate of healing of the crustal rocks. The geometry of the propagating rift zone, its associated faulting and seismicity patterns are dictated by the local lithosphere structure, and particularly by the depth to Moho interface. Such a rift zone terminates at the continental margins and a new fault system forms perpendicular to the rift propagation.

Acknowledgments

[38] We thank the editor Thorsten Becker, the reviewer Taras Gerya and anonymous reviewers for constructive comments. The study was supported by the Israel Science Foundation (ISF 753/08).

References


Braun, J., and C. Beaumont (1989), A physical explanation of

Byerlee, J. D. (1978), Friction of rocks,

Cox, S. J. D., and C. H. Scholz (1988), Rupture initiation in

Dunbar, J. A., and D. S. Sawyer (1989), How pre-existing


Finzi, Y., V. Lyakhovsky, Y. Ben-Zion, and E. H. Hearn (2009), Structural properties and deformation patterns of evolving strike-slip faults: Numerical simulation incorporating


Ilichev, A., and V. Lyakhovsky (2001), Practical aspects of the hybridization of the boundary integral method with damage rheology modeling for the assimilation of seismic data, in Rockbursts and Seismicity in Mines: Dynamic Rock Mass Response to Mining, edited by G. van Aswegen,


Stein, R. S., G. King, and J. Lin (1994), Stress triggering of the 1994 M-6.7 Northridge, California, earthquake by its predecessors, Science, 265, 1432–1435.


