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The role of frictional plasticity in the evolution of normal fault systems

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Abstract

Tectonic extension applied to the upper crust commonly results in the development of sub-parallel arrays of faults and fault belts. A computational model of an extending elastic-viscous-plastic crust is presented, which describes the evolution of stress, strain and displacement in the vicinity of an active normal fault. Within the upper crust discrete episodes of fault slip accommodate up to half the applied extension rate, with the balance accommodated by bands of frictional plastic shearing representing young proto-faults. Repeated coseismic shear stress reductions in the near-fault vicinity impose a stress-failure shadow where the formation of new faults is prohibited. A near-surface compressional stress regime in the footwall causes the failure shadow to extend disproportionately into this fault block, thereby favoring the formation of new faults in the hanging wall. A continuous shallowing of fault dip is also documented and supports a

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scenario of migration of faulting into the hanging wall of existing faults, consistent with descriptions of the structural evolution in the western Gulf of Corinth, Greece.

Keywords

Normal fault, frictional plasticity, fault rotation, numerical model, extension budget

1. Introduction

Extension of the lithosphere in rifting basins is accommodated through displacement on normal dip-slip faults or by the emplacement of dikes, i.e., intrusive magmatism. Such structures are observed at rifting at both convergent (Taymaz et al., 1991; Wilson et al., 1995) and divergent plate margins (Ebinger, 1989), and intra-plate settings (Illies, 1972; Puskas et al., 2007). Over time the character of normal faulting evolves, with individual fault strands organizing into sub-parallel en-echelon arrays and, with continued linkage, into large fault belts (Cowie et al., 2005). However, fault planes may also become inactive (Jackson, 1999). Extension can cause fault planes to rotate about a horizontal axis, shallowing the dip angle to a point where continued reactivation of the fault becomes unlikely (Sibson, 1985). In addition, stresses generated by an existing fault have been hypothesized (Goldsworthy and Jackson, 2001) to induce a migration of faulting in a basinward direction (Dart et al., 1995; Horton and Schmitt, 1998; Goldsworthy and Jackson, 2001; Sanchez et al., 2010).

During discrete episodes of dip-slip on normal faults, extension is accommodated by translation of one fault block relative to the other. Displacement is primarily parallel to

the fault plane, and therefore horizontal movements are accompanied by vertical displacements. This behavior translates to uplift of the footwall block and a comparatively larger subsidence of the hanging wall (e.g., Barr, 1987). Conceptual models employ lithospheric flexure (Weissel and Karner, 1989; Kusznir et al., 1991), domino rotation (Wernicke and Burchfiel, 1982), or isostatic rebound (Barr, 1987; Westaway, 1992) on either planar or listric fault surfaces to describe this deformation. However, field studies indicate that fault slip aggregated across a rift does not always account for the full geodetic extension rate (e.g., Villamor and Berryman, 2001; Wallace et al., 2004). Additional deformation mechanisms or movements on undiscovered faults must account for the shortfall.

Normal fault systems evolve and grow over many thousands of years in response to the continuous application of far-field tectonic extension (e.g., Villamor and Berryman, 2001; Bell et al., 2009). Such evolution involves the initiation of new faults (Jackson and McKenzie, 1983), which is itself an indication that extension is not completely accommodated by existing structures. Potential causes of this shortfall include: a migration of the locus of extension away from the faulted region; a suspension of activity on pre-existing faults due to fault rotation (Jackson, 1999), restrengthening of the fault (Li et al., 2003; Muhuri et al., 2003) or stress shadowing by other faults (Cowie 1998); distributed frictional deformation (see section 2, below); or a suspension of alternative mechanisms of extension accommodation, e.g., dike emplacement (Rowland et al., 2010).

67 The purpose of this work is to investigate specifically the role that frictional plasticity
68 plays in the evolution of a normal fault system. From a theoretical perspective, frictional
69 plasticity is the non-recoverable deformation of the crust through frictional sliding on
70 hypothetical fracture planes. Such planes are presumed to exist abundantly throughout the
71 crust (Ranalli and Murphy, 1987) at orientations that are optimal for reactivation within
72 the prevailing stress field. Within the crust, these structural discontinuities may represent
73 joints, dilational jogs, bedding planes or extensional shear fractures (Sibson, 1996; 2000)
74 across which frictional shear displacements might be expected to occur. Coalescence of a
75 multitude of these shear offset structures within a well constrained deformation band may
76 be a central process leading to the formation of a new fault (e.g., Pollard and Fletcher,
77 2005). Thus, an investigation of frictional plasticity, treated as a precursor to fault
78 formation, can address where subsequent faults should localize and provide insight into
79 fault zone evolution.

80
81 We build on earlier models of Ellis et al. (2006) and Nüchter and Ellis (2010) that
82 prescribe fault surfaces directly and model frictional sliding between fault blocks over
83 many seismic cycles. We model the evolution of normal fault systems shortly after the
84 onset of rifting. Modeled deformation includes: discrete episodes of slip between fault
85 blocks that displace independently along the plane of contact; frictional plastic yield
86 within the brittle upper crust; and viscous flow in the lower crust. The relative importance
87 of each mechanism in accommodating the regional extension rate indicates the likelihood
88 of development of additional fault structures and the positions of these relative to the
89 primary fault. We observe a gradual shallowing of the dip angle of the fault plane due to

rotation, which is a prerequisite for the development of cross-cutting faults (Profett, 1977; Nur et al., 1986) as older structures rotate beyond the limit for reactivation (Sibson, 1985).

Results from the computational model are interpreted in the context of the western Gulf of Corinth, an asymmetric graben in mainland Greece, comprised of several major north-dipping faults accommodating extension of 12-14 mm yr⁻¹ (e.g., Clarke et al., 1997; Briole et al., 2000; Floyd et al., 2010). Geomorphological and stratigraphic evidence indicate a migration of faulting in a basinward direction (Goldsworthy and Jackson, 2001; Flotté et al., 2005; Rohais et al., 2007) as new faults developed in the hanging walls of older structures (Jackson, 1999). Flotté et al. (2005) propose a tectonic model utilizing a low-angle detachment fault underlying steeper surface faults, although this model is not universally accepted and is not applied by us (e.g., McNeill et al., 2005). In this paper, we propose a model for basinward fault migration based on modeled distributions of stress and frictional plasticity in the vicinity of existing faults.

2. Previous numerical experiments

Computational modeling of fault systems on short timescales has been used to examine rupture propagation along a fault under rate-state friction constraints (Dieterich, 1979; Ruina, 1983), rupture propagation in an asymmetric geometry (Oglesby et al., 2000), and near-fault coseismic deformation of an elastic crust (e.g., Freund and Barnett, 1976; Gupta and Scholz, 1998). Such studies constrain the dynamics of fault rupture and are

useful in modeling ground acceleration and quasi-static coseismic displacements;
however, they do not account for deformation in fault systems over many seismic cycles.

Over longer timescales, numerical models of fault architecture and rift evolution, such as an extension to the propagator matrix method of Rundle (1980) by King et al. (1988) and Stein et al. (1988), have explored topographical development of normal faults over multiple earthquake cycles and in continual depositional and erosional processes. Their model indicates that the degree of lithospheric flexure and the width of the deformation region are controlled by the effective elastic thickness of the crust. Finite element models have expanded on these findings, investigating the effects of finite elastic strength in the upper crust, and tracking patterns of failure in the vicinity of a fault (e.g., Melosh and Williams, 1989; Bott, 1997). Plastic deformation reduces the effective elastic thickness of the crust, thereby controlling graben dimension by migrating the outer flexure hinge closer to the fault scarp (Bott, 1997).

Melosh and Williams (1989) detailed a finite element model that accommodates slip between 65° dipping conjugate elements and which tracks proximity to Mohr-Coulomb failure in the vicinity of a dip-slip fault. Their analysis indicates a maximum propensity to failure in the near surface of the hanging wall, and also at the fault tip where the maximum gradient in dip-slip is encountered. On the basis of strain-energy minimization, they argue that antithetic faults should form in the hanging wall in response to coseismic stress changes caused by rupture of the main fault. This result, which is specific to

systems in which the main fault dips at 65° , does not account for coupling between gravitational potential and strain energy (Dempsey et al., 2012).

Henk (2006) uses a thermo-mechanical model with a viscoelastic-plastic rheology to investigate lithospheric extension and deformation about a reactivated thrust fault, for a variety of strength parameters and strain rates. These models also demonstrate strain localization in the near-surface of the hanging wall, interpreted as a crestal collapse graben (McClay and Ellis, 1987). Others have used 3-dimensional thermo-mechanical models (Braun et al., 2008; Popov and Sobolev, 2008) to model rifting over large spatial and temporal scales (100's of km and millions of years), and have reproduced rift bounding faults through bands of localized strain (Huisman et al., 2005).

3. Numerical model

The finite element method, implemented here using the software suite Abaqus/Standard (http://www.simulia.com/products/abaqus_standard.html), is used to solve deformation and stress equations for an extending crust. Far-field boundary conditions and a cyclical fault-friction regime drive the evolution of stress and deformation in time. This approach was previously used by Ellis et al. (2006) and Nüchter and Ellis (2010) to develop 2-D models of fault systems in contractional and extensional tectonic settings.

We consider deformation parallel to the strike of a dip-slip fault as negligible, which allows us to model a 2-dimensional slice of the crust under plane strain conditions. This slice is oriented perpendicular to the axis of a generic rift system and measures 60 km

horizontally and 25 km vertically. Such dimensions are presumed to be sufficient to represent the relatively shallow seismogenic zone in thermally weakened rifts (approximately 8 km, e.g., Bibby et al. (1995)) and isolate fault behavior from the influence of side and lower boundaries.

Deformation prior to anelastic yield is described by linear, isotropic elasticity, parameterized by Young's Modulus, E , and Poisson's ratio, ν . For a complete summary of mathematical terms and values see Table 1.

When stress in the upper crust exceeds the threshold for frictional plastic failure, subsequent deformation is non-recoverable. This yield criteria can be defined in terms of the Von Mises equivalent stress, q ,

$$q = \sqrt{\frac{3}{2} s_{ij} s_{ji}} \quad (1)$$

where the stress deviator tensor is defined $s_{ij} = \sigma_{ij} - \sigma_{kk} \delta_{ij}/3$. Under plane strain conditions, q reduces to the difference between maximum and minimum principal stresses and is thus equivalent to the differential stress and equal to twice the maximum shear stress, i.e., $q = \sigma_1 - \sigma_3 = 2\tau$. Frictional plastic yield occurs when the shear stress exceeds a Mohr-Coulomb failure criterion

$$\tau > \tau_p = (p - p_f) \cdot \sin(\phi) + C \cdot \cos(\phi) \quad (2)$$

where $p = -\sigma_{kk}/3$ is the mean stress, p_f is the pore fluid pressure, assumed to be hydrostatic, C is the cohesion and ϕ is the internal angle of friction of the crust. As p increases with depth in the crust so too does the shear stress required to achieve frictional yield. Frictional-plastic strain, ε_p , that accumulates at yield is predicted to localize along

conjugate shear bands whose dip depends on the resolution of heterogeneities, the use of dynamic pressure in Eq. (2) (Buiter, 2012) and the friction angle (Kaus, 2010). In these models, we use a modified internal angle of friction, $\sin(\varphi') = \sin(\varphi) \cdot (\rho - \rho_f) / \rho$, where ρ is rock density and ρ_f is the saturating fluid density. This modification reflects weakening of the crust by the hydrostatic pore pressure term in Eq. (2) (Townend and Zoback, 2000). For a modified friction angle of 19° , Kaus (2010) predicts shear bands to localize along the Roscoe angles, which for a non-dilational crust are at 45° to σ_1 . This dip angle is shallower than would be expected for a dilational crust at greater friction angles. The effects of strain softening are not considered in these simulations. Strain softening is not necessary to cause localization of shear deformation into bands (Rudnicki and Rice, 1975), although it does introduce a finite length-scale independent of mesh dimension (Needleman, 1988). Other workers, e.g., Huismans et al. (2005), have incorporated strain softening for modeling the onset of faulting during lithospheric extension.

Anelastic deformation of the lower crust by dislocation creep is modeled using a viscous power-law rheology. The viscous strain rate, $\dot{\epsilon}_v$, depends upon the local stress state via the Von Mises equivalent stress

$$\dot{\epsilon}_v = Aq^n \exp(-Q/RT) \quad (3)$$

where Q is the activation energy, A is a constant containing material and geometrical factors, n is the power-law exponent, R is the gas constant and T is the temperature in Kelvin. In these models, we use viscous parameters from Paterson and Luan (1990) (see Table 1). A linear crustal temperature gradient of 35°C km^{-1} (Figure 1(b)) is used, which is consistent with the elevated heat flux of active rifts, approximately 0.7 W m^{-1}

(Hochstein, 1995). At depths in the crust where viscous deformation is the dominant mechanism accommodating extension, we can rearrange Eq. (3) for the Von Mises stress, q_v , required to accommodate the imposed bulk strain rate

$$q_v = \left(\frac{\dot{\epsilon}_v}{A} \right)^{\frac{1}{n}} \exp(Q / nRT) \quad (4)$$

In this form, it is clear that at shallower depths, where the crust is cooler, a higher Von Mises stress is required to accommodate extension by viscous flow. At any depth in the crust, both q_v and $q_p (=2\tau_p)$ can be calculated as a function of temperature and pressure. If $q_v < q_p$ and $q_v < q$, then viscous flow occurs, and if $q_p < q_v$ and $q_p < q$, then deformation is frictional plastic. When $q_v = q_p$, there is a transition between the two behaviors and the depth at which this occurs is called the brittle-ductile transition (BDT).

Deformation of the crust by tectonic extension occurs due to velocity boundary conditions imposed at the lateral boundaries. GPS investigations of the western Gulf of Corinth indicate that extension is on the order of 12-14 mm yr⁻¹ (e.g., Clarke et al., 1997; Briole et al., 2000; Floyd et al., 2010). In the models detailed here, we apply a horizontal velocity boundary condition of -6 and +6 mm yr⁻¹ to the left and right boundaries, respectively, to account for this forcing (see Figure 1(a)). This differs from previous models (Bott, 1997; Ellis et al., 2006; Nüchter and Ellis, 2010) where one boundary is constrained by a 'no displacement' condition and the full tectonic forcing applied to the other boundary. The lower boundary is fixed at zero vertical displacement and the upper surface is unconstrained. Gravity is applied throughout the model as a body force. A constant density of 2800 kg m⁻³ is applied throughout the crust.

A single normal fault is modeled as a contact surface that intersects the crust, dipping at an angle $\theta=50^\circ$ to a depth of 15 km. This geometry is shallower than previous numerical experiments of faulting (e.g., Melosh and Williams, 1989), but is consistent with normal fault dips used by Bell et al. (2009) to calculate fault slip rates, and by Briole et al. (2000) to model normal fault earthquakes in the Gulf of Corinth. The fault is prescribed as either 'locked', where the surfaces are prevented from sliding against one another, or 'slipping', where frictional sliding according to Amonton's law is permitted on the interface

$$\tau > \mu_d \sigma_n \quad (5)$$

where τ is the shear stress, σ_n is the normal stress and μ_d is the dynamic coefficient of friction on the fault surface, taken here as a slip weakened value of 0.2 to account for heating and fluid effects during dynamic rupture propagation (Scholz and Engelder, 1976). The fault is arbitrarily switched between locked and slipping modes and therefore this model is not suitable for investigating earthquake rupture initiation. Instead we investigate the stress and deformation of a faulted crust during the seismic cycle, i.e., the static response to earthquakes.

The model domain is meshed using ~10 000 plane-strain quadrilateral elements, with smaller elements used near the surface and on the fault contact surfaces. Element dimensions range from a minimum of 250 m on the fault to 1000 m at the base of the model (see Figure 1(a)).

An initial stress state for the model is evolved by applying the velocity boundary conditions over geological time (30 kyr, see Figure 1(d)). This stress 'spin-up' (Hetland

and Hager, 2006; Ellis et al., 2006) phase produces a lithostatic vertical stress, σ_z , and a horizontal stress, σ_x , such that the differential stress is equal to the lower of q_p or q_v , i.e., a critically stressed crust at frictional plastic yield at shallow depths and viscous yield below the BDT (Townend and Zoback, 2000). Deformation that occurs during the stress spin-up period is discarded at the beginning of a new simulation, while the stress state is retained as an initial condition. The initial stress state, characterized by the differential stress (and equivalent to the Von Mises stress under plane strain conditions), is shown as a function of depth in Figure 1(c).

The seismic cycle, in which long periods of fault dormancy are punctuated by short, violent episodes of fault slip and followed by a readjustment period of afterslip and aftershock activity, is approximated in the model by a two-step cycle. The interseismic period refers to the interval between earthquakes in which extension applied at the model boundaries alters the displacement and stress distribution within the model domain. Each interseismic period is followed by a seismic period, in which the friction coefficient on the fault surface is substantially lowered and displacements of the fault blocks are driven by an imbalance between gravitational and elastic forces. The model reaches dynamic equilibrium, where elastic changes accumulated during the interseismic period are offset by those of the seismic period, after 10 seismic cycles. In this paper we analyze results from the 12th cycle (Figure 1(d)). Displacements of the model boundaries caused by the velocity boundary conditions are negligible during the 60 second seismic period. Any slip that occurs during this period is due to stress imbalances accumulated during the

interseismic period. Fault slip that occurs during the seismic period includes all possible afterslip.

4. The seismic cycle

Figure 2 summarizes salient aspects of the modeled seismic cycle including (a) coseismic surface displacements, (b) fault parallel displacement that occurs during the 12th fault rupture, (c) the distribution of the differential stress after 12 seismic cycles, (d) elastic strain accumulated during the 12th fault rupture, and the total accumulated (e) viscous and (f) frictional-plastic strain after 12 seismic cycles.

A sudden decrease in friction on the fault surface during the seismic period permits displacement of the fault blocks relative to one another. As the problem is modeled under plane-strain conditions, along-strike displacement is not considered and the resulting motions are approximately parallel to the dip of the fault. Figure 2(b) plots the fault-parallel displacement field of the crust during the seismic period, in which “up-dip” is taken as the positive direction. Maximum displacements are observed at the fault scarp and decay with distance, consistent with geodetic evidence of large normal fault ruptures at Hegben Lake, Montana and Borah Peak, Idaho (Barrientos et al., 1987). Along the fault plane the footwall exhibits 0.6 m of upward-directed fault-parallel displacement from the surface to 5 km depth (Figure 2(b) inset). Similarly the hanging wall is displaced a constant 0.9 m in a down-dip direction for the same range of depths. The total single-event offset is therefore 1.5 m, which is consistent with the order of magnitude of historical ruptures in the Gulf of Corinth (Koukouvelas and Doutsos, 1996; Roberts, 1996). In

addition to material (Gudmundsson, 2004) and fault geometry (Section 4) considerations, the length of the loading period preceding an earthquake, i.e., the interseismic period, exercises an important control on the size of the subsequent fault rupture. In these simulations we have selected an interseismic period of 220 years that produces fault ruptures of an appropriate size for the region under consideration. Fault slip decays from its maximum value to zero over a depth interval of 5.0-8.5 km, below which viscous deformation maintains a stress state that inhibits slip. The true depth of the modeled fault is therefore determined by material parameters and the applied strain rate and is not a prescribed feature of the geometry.

The distribution of the differential stress (Figure 2(c)) reflects the brittle and viscous rheologies prevailing in the upper and lower crust, respectively (see Figure 1(c) and Eq. (2) and (4)). The maximum differential stress, marking the transition from brittle to ductile behavior, occurs at a depth of 6 km. Perturbations to the profile in Figure 1(c) are observed close to the fault surface and are caused by coseismic movements of the fault blocks and gradients in slip at the fault tip. In the hanging wall, a broad maximum in q is centered on the BDT, while in the adjacent footwall there is a corresponding reduction compared to background levels for that depth.

Viscous deformation is the dominant contribution to deformation at the high temperatures and pressures found in the lower crust (>9 km depth). Frictional plastic deformation or slip on the fault surface are absent from this region of the crust so that, during the interseismic period, 2.64 m ($12 \text{ mm yr}^{-1} \times 220 \text{ yr}$) of tectonic extension must be

accommodated by viscous strain. As the strain rate, $\dot{\epsilon}_v$, is proportional to q^4 (see Table 1 and Eq. (3) with $n = 4$), the additional accumulation of ϵ_v in Figure 2(e) along the fault root at depths >6 km points to an increase in the stress above the background level for this depth. This increase, evident in Figure 2(c), is attributed to the upper crustal displacements of the fault blocks and the accommodation of these motions within the mid-crust. The extra viscous straining directly beneath the fault alters post-seismic deformation at the surface (Ellis et al., 2006) and helps restore the stress state on the fault to its equilibrium state (Nüchter and Ellis, 2010).

Frictional-plastic strain accumulates exclusively in the upper crust at depths shallower than 6 km and localizes in shear bands that dip at approximately 45° from the horizontal, as discussed in section 2. Qualitatively, frictional-plastic strain occurs preferentially in a triangular wedge of the hanging wall block and is conspicuously absent from much of the footwall. Regions of increased plastic strain are also observed at the lateral boundaries of the model; however, these regions are outside the influence of the fault and represent background strain behavior. The presence of multiple shear bands, as opposed to one or two bands of high strain, is consistent with the “distributed” mode of fracturing described by (Schueller et al., 2010). They showed that if the applied extension rate cannot be accommodated by displacements on existing frictional plastic shear zones, which in turn are limited by the slip velocity in both brittle and ductile media, then new shear zones form to accommodate the shortfall. However, with further deformation such that these shear bands experience strain softening due to accumulated plastic strain, we expect that displacement would localize onto fewer structures.

340

341 **5. Extension budget**

342 The extension rate of the modeled rift, which is applied as a velocity boundary condition
343 at the lateral boundaries, must be accommodated within the crust by some combination of
344 the available means of deformation, i.e., by fault slip, or the accumulation of elastic,
345 frictional plastic, or viscous strain. The relative contribution of each to the total extension
346 of the crust during a single seismic cycle comprises the extension budget. For instance, if
347 during a single seismic cycle tectonic extension of 2.64 m is comprised of 0.66 m of
348 horizontal elastic strain, 0.66 m of horizontal frictional-plastic strain, and 1.32 m
349 horizontal component of fault slip, then the relative contribution of each component is
350 0.25, 0.25 and 0.5, respectively. Depending on the means of deformation available, the
351 extension budget will vary with both depth (e.g., viscous strain does not accumulate in
352 the upper crust) and time (e.g., fault slip occurs only during the seismic period). Figure 3
353 shows a representative extension budget with depth for the (a) interseismic and (b)
354 seismic periods, and (c) a full seismic cycle, where each component is normalized against
355 the total extension (2.64 m) during the full cycle.

356

357 Figure 3(a) indicates that viscous strain is the sole contributor to deformation at depths
358 greater than 9 km and, due to its slow accumulation rate, occurs principally during the
359 interseismic period. Small changes in elastic strain are also observed at these depths,
360 although, summed over the interseismic and seismic periods, their net effect is nearly
361 zero, which is characteristic of the dynamic equilibrium of the system.

362

Figure 3(c-d) indicates that the presence of a rupturing fault permits extension of the upper crust to be distributed between the two brittle failure mechanisms of fault slip and frictional plastic deformation. However, the accumulation of each component is mutually exclusive, i.e., fault slip occurs rapidly during the seismic period (Figure 3(b)) whereas frictional-plastic strain accumulates slowly during the interseismic period (Figure 3(a)). Coseismic frictional-plastic strain accumulation is significant for steeper dipping faults (see Section 6). A significant component of transient elastic strain accommodates extension during the interseismic period, but is replaced by fault slip in the following seismic period. It also causes a reduction of the maximum shear stress, $(\sigma_1 - \sigma_3)/2$, throughout the crust wherever the least principal compressive stress is horizontal, i.e., $\sigma_3 = \sigma_{xx}$. Accumulation of frictional-plastic strain ceases for a brief post-seismic period until extension restores the preseismic stress state to frictional yield (Dempsey et al., 2012).

Figure 3(a) and (c) show that the transition from purely brittle deformation, including fault slip and frictional-plastic straining, to purely viscous deformation occurs over a 5.0-9.0 km depth interval. Frictional plastic deformation terminates over a shallower 5.0-6.5 km depth interval, with a small overlap between the brittle and viscous rheologies indicative of the exponential decrease in viscous strain rate with decreasing temperatures (Equation (3)). Fault slip penetrates further into the brittle-ductile transition zone, terminating over depths of 6.5-9.0 km. This bimodal truncation of the two frictional behaviors is due to the relative frictional weakness of the fault ($\mu=0.2$ compared with the intact crust $\mu=\tan(\phi')\approx 0.3$) and indicates a mixed brittle-viscous region between a purely

386 brittle upper crust and purely viscous lower crust. In this region weakened, embrittled
387 faults are able to penetrate the mostly viscous lower crust (Scholz, 1988, 1998), while
388 frictional plasticity is limited to shallower depths.

389
390 Figure 4(a) considers the composition of the full cycle extension budget within the upper
391 crust (< 5.0 km depth) as a function of the loading period that precedes an earthquake.
392 For this analysis each simulation was “spun-up” to the same dynamic stress equilibrium,
393 using a nominal interseismic period of 220 years, and followed by a single seismic cycle
394 at the modified length. Thus, Figure 4(a) considers the effect of seismic cycle length
395 independent of the previous rupture history. It also illustrates extension accommodation
396 in the upper crust at different times during the interseismic period. As in Figure 3,
397 components were normalized against the total extension applied over the full cycle. For
398 longer seismic cycles, distributed frictional-plastic strain accounts for a proportionally
399 larger share, e.g., approaching 50% of the total extension for an interseismic period 220
400 years in length. During the early stages of the interseismic quiescence following fault
401 rupture, tectonic extension is mostly taken up by the accumulation of elastic strain
402 throughout the upper crust, which is later exchanged for fault slip. Extension reduces
403 horizontal stress, σ_{xx} , which causes the differential stress to approach the limit of
404 frictional plastic failure, q_p , and switches large areas of the crust from elastic to frictional
405 plastic behavior. Elastic strain is temporarily saturated in these regions, and subsequent
406 accumulated strain is frictional plastic. Each successive increment of extension during the
407 interseismic period then has two compounding effects: (i) to effect stress changes that
408 switch ever larger regions of the crust from elastic to frictional plastic deformation and

(ii) to accumulate frictional-plastic strain within an increasing volume of yielding crust and elastic strain within a decreasing volume of elastic crust. Horizontal stress is increased in the subsequent earthquake step due to coseismic displacements of the fault blocks.

Fault dip also exerts control on the composition of the upper crust extension budget (see Figure 4(b)). For a system in dynamic equilibrium with a seismic cycle 220 years in length, a shallow fault, $\theta = 40\text{-}50^\circ$, accommodates more than half of the applied extension through fault slip, with the shortfall accommodated by frictional plasticity. However, steeper faults accommodate a decreasing proportion of the total extension through displacement of the fault blocks. This decrease is because the production of a given amount of horizontal extension requires an increasing component of vertical movement, and thus a correspondingly larger earthquake, at steeper angles of fault dip. A reduction in the fault-slip component of the extension budget must be compensated by an increase in other components, which for an upper crust at dynamic equilibrium must be frictional-plastic strain. This result is largely independent of the elastic properties of the crust, and the partitioning of upper crustal extension between the accumulation of frictional-plastic strain and episodic fault slip appears to be controlled principally by the geometry of the system.

6. Distribution of frictional plasticity

By definition, frictional-plastic strain accumulates in those regions where the differential (plane-strain Von Mises) stress exceeds the threshold for failure, i.e., $q > q_p$. The

accumulation of frictional-plastic strain then, to some extent, reflects the stress state of the crust. Figure 5(a-b) plots the proximity of the differential stress to the limit of frictional plastic yield, i.e., $q_p - q$, to a depth of 8 km, prior to and following fault rupture. The frictional plastic yield limit is pressure dependent and must therefore be recalculated following fault rupture to account for changes in mean stress. The sharp transition from brittle to ductile failure is evident at a depth of 6 km, consistent with the thickness of the frictional plastic layer in the extension budget (see section 5). Below 6 km, viscous deformation maintains low stresses and ensures that $q < q_p$. In the brittle upper crust and prior to fault rupture, the majority of the hanging-wall crust is either at or within 0.1 MPa of failure. Coseismic stress changes generally decrease the proximity to failure by between 0.1 and 1 MPa.

Repeated fault slip and a small dynamic friction coefficient maintain the near-fault differential stress at up to 10 MPa below failure, creating a stress shadow. Under plane strain conditions, the Von Mises stress is twice the maximum shear stress, and thus its reduction is due to the repeated coseismic relief of shear stress on the fault plane. Shear stress drops during the modeled episodes of fault slip are approximately 0.5-1 MPa, which indicates that the magnitude of the stress-failure shadow varies by up to 10% over the seismic period. Greater stress drops associated with larger earthquakes and longer recurrence intervals are then permitted for this fault, but the shear stress in the near-fault region cannot increase above q_p . If the differential stress is consistently restored (through a drop in stress) to its postseismic value in Figure 5(b) for any recurrence interval, then it

follows that stress drop, and thus earthquake magnitude (Hanks and Kanamori, 1979), are limited by the frictional strength of the upper crust.

Upward-directed fault slip along the footwall scarp during an earthquake causes an increase in elevation at the surface, but also a component of horizontal displacement away from the fault plane. During the brief seismic slip-events displacements at the side boundaries of the model are approximately zero, and thus coseismic fault displacements result in a horizontal compression of the crust. This elastic deformation lowers the differential stress in the near surface of the footwall and at depths less than 500 m, the compression is so great that the horizontal stress exceeds the vertical, i.e., $|\sigma_{xx}| > |\sigma_{zz}|$, and the stress regime is compressional (see hatched regions in Figure 5(a-b)). The shallow extent of these zones is principally a reflection of the low lithostatic vertical stress at shallow depths, which is exceeded by only a modest horizontal compression.

The presence of a slipping fault perturbs the differential stress in its vicinity, imposing a failure shadow on a crust that is otherwise at frictional yield. This failure shadow broadens near the surface, particularly within the footwall block, and tapers to a narrow width near the BDT. Figure 5(c) shows that frictional-plastic strain does not accumulate within the shadow, but rather the greatest accumulation is outside the shadow in the hanging wall block. In this region, we observe frictional plastic shear-bands dipping at 45° , synthetic to the main fault, and intersecting the surface ($z=0$) outside the edge of the failure shadow. Low levels of frictional plasticity are also observed in the footwall block at depths of 3-7 km where upward-directed fault slip and compression have not

significantly altered the stress state. Except for a small region at the root of the fault, the majority of the fault plane is free from plastic deformation, although we cannot preclude failure from dynamic loading during rupture propagation, which we have not modeled.

Previous numerical experiments by Melosh and Williams (1989) indicated that formation of antithetic faults in the hanging wall of a main fault was energetically favorable over synthetic faults of similar a similar dip. This result was derived assuming immediate secondary fault formation in response to failure stresses induced by slip on the main fault, i.e., coseismic antithetic faults. A similar result is replicated in our model by considering frictional-plastic strain accumulated during the coseismic period in response to slip on a 65° dipping master fault (Figure 6(b)). For the model containing a 50° dipping fault, frictional-plastic strain during the seismic period is negligible as local bending at the surface of the hanging wall is less pronounced than for steeper faults. This result suggests a preference for coseismic antithetic faulting may be limited to steeper dip angles.

In contrast, in Figure 5(c) and Figure 6(b) we demonstrate a preference for synthetic fault formation, caused by a steady accumulation of frictional-plastic strain in shear bands during the comparatively longer interseismic period. It should be emphasized this result is not inconsistent with Melosh and Williams (1989), but rather indicates that the configuration of secondary faults is sensitive to the fault dip relative to optimally oriented fault dip, and to the stage of the seismic cycle during which they form.

7. Fault plane rotation

Coseismic displacement of the fault blocks parallel to the fault plane, which accommodates extension and reduces near-fault shear stresses, is perhaps the most conspicuous aspect of deformation associated with faulting (e.g., Barrientos et al., 1987; Segal, 2010). However, observations of normal faults dipping at shallow angles ($<30^\circ$) that preclude reactivation within an Andersonian stress regime (Wernicke, 1995), or a regime of cross-cutting faults (Profett, 1977; Nur et al., 1986), suggest a secondary process of rotation via a bookshelf or domino-style mechanism. Such motion, resolved on the fault plane, is equivalent to a displacement of the fault block perpendicular to the fault plane. Further, for a rotation as opposed to a rigid body translation, displacement must increase linearly with distance from some pivot, i.e., a point of no displacement.

Figure 7 plots fault perpendicular displacement along the fault and against depth during (a) the interseismic and (b) seismic periods, with a regression line fit to the data in the brittle layer (<8 km depth). During both periods, displacement profiles in the upper crust are approximately linear and have gradients of 59 and 66 mm km⁻¹ (i.e., millimeters of fault perpendicular displacement per kilometer of depth) for the interseismic and seismic periods, respectively.

The magnitude of fault plane rotation can be calculated from the gradients of the regression lines in Figure 7. A linear fault-normal displacement profile, u , is expressed as a function of depth, $u = m(z-z_0)$, where m is the gradient of regression line through the displacement data, and z_0 is the depth of the fixed horizontal axis, or pivot, about which rotation occurs. The depth of the pivot corresponds to the intersection of the regression

line in Figure 7(a-b) with the vertical $u = 0$ axis. During the interseismic period, the pivot ranges in depth from 18 to 18.7 km, with a mean of 18.4 km. For the seismic period, rotation of the fault plane occurs about a pivot at a depth of 4.7 km. A displacement, u , some vertical distance $z-z_0$ above the pivot induces a rotation, $d\theta$, of the fault plane (Figure 7(c)), given by

$$d\theta = \tan^{-1}(u \sin \theta / (z - z_0)) = \tan^{-1}(m \cdot \sin \theta) \quad (6)$$

Using $m = 59$ and 66 mm km^{-1} , we calculate a shallowing of dip of 2.4 and 2.7 millidegrees during the interseismic and seismic periods, respectively. This corresponds to a long term rotation rate of 23.5° every million years (for a seismic cycle that repeats every 220 years), although a reduction in this rate would presumably be observed as the fault rotates out of the range for reactivation.

The total rotation during a seismic cycle is distributed approximately equally between the interseismic and seismic periods. The amount of coseismic rotation is proportional to the length of the interseismic period that precedes it, and is thus also proportional to earthquake magnitude. Further, coseismic rotation about an axis at a depth of 4.7 km, 14 km shallower than that for the interseismic period, suggests that a different forcing operates each rotation. For instance, during the interseismic period, far-field tectonic extension deforms the entire crust, causing vertical contraction in addition to horizontal spreading, effectively flattening the crust and the fault that is hosted within. The effect of this deformation on fault dip is evident in Figure 8(a), which presents the exaggerated change in position of the fault surface and topography during the interseismic period.

In contrast, displacements at the lateral boundaries during the short seismic period are too small to account for the observed rotation. However, large and rapid stress changes in the mid-crust, which cause uplift of the footwall and accommodate collapse of the hanging wall, may be responsible. Footwall uplift is caused by a decrease in stress and rapid decompression of the crust between 6 and 9 km depth of the footwall block. Figure 8(b) shows the decrease in pressure associated with this uplift, superimposed on the deformed topography and fault plane. A drop in pressure corresponds to an increase in volume in that region, which, while primarily directed up-dip along the fault plane, also includes a fault normal component. This expansion applies a force to the hanging wall and a torque about the pivot at 4.7 km depth. This expansion also accounts for a component of the pressure increase in the adjacent hanging wall. The expanding footwall does work on the hanging wall, which, in addition to the compression caused by dip-slip of the fault block above, yields a reduction in volume and an increase in stress. As this rotation occurs due to stress changes at the fault tip, the depth of the pivot about which rotation occurs is primarily controlled by the depth of faulting.

8. Evolution of normal fault systems

Where pre-existing faults do not adequately accommodate the geodetic extension rate, parts of the crust are subjected to a horizontal, positive strain. In regions of the crust where the shear stress is below the Mohr-Coulomb failure limit, extension is accommodated by elastic strain. However, continued elastic stretching generates shear stresses that eventually induce failure on any pre-existing fracture planes and material discontinuities. Shear stresses are maintained at near critical values (Townend and

Zoback, 2000) by incremental frictional failure on these minor structures in response to each subsequent increment of extension. These processes constrain the maximum stress state attainable by the crust and therefore impose an upper limit on the amount of extension that is temporarily accommodated as elastic strain.

However, no such constraint exists for accommodation of extension through frictional-plastic strain, which accumulates following the saturation of elastic strain. In our model, frictional plastic deformation localizes in shear bands that form at 45° to the maximum principal compressional stress, which is vertical in an extensional tectonic regime. That these shear bands represent the development of new faults is speculative, although they do satisfy two criteria associated with fault structures: (i) the shear bands accommodate dip-slip displacement and (ii) are localized, rather than distributed (Schueller et al., 2010), zones of deformation, in this case the width of several elements (approximately 1 km). In these models, shear zone width is sensitive to mesh dimension, as we have not incorporated a description of strain softening behavior (Needleman, 1988). Physically, a localization of frictional plasticity represents a continuous reactivation within a particular region of those small discontinuities, e.g., joints, dilational jogs, fractures, etc., accommodating shear displacements. Providing each subsequent reactivation increases the length of the activated structure, a region of frictional plastic shear may be regarded as a sort of proto-fault. We note that the evolution of a proto-fault-style shear zone into a true fault that perturbs stresses in its vicinity is not modeled here. Instead, we are considering the early onset of proto-fault development.

Basinward migration of normal fault structures has been inferred from stratigraphic studies in Turkey (Dart et al., 1995), the central Basin and Range province, USA (Horton and Schmitt, 1998), the Gulf of Corinth, Greece (Goldsworthy and Jackson, 2001), and the strike-slip Gar Basin in western Tibet (Sanchez et al., 2010). Structurally, new faults develop in the hanging wall block of older faults producing a series of sub-parallel faults spaced every 2-5 km. Goldsworthy and Jackson (2001) proposed several mechanisms for the cessation of rupture on one fault and concurrent development of an adjacent structure including: topographic stresses in a domino-style topography modulated by the strength of the active fault; density contrast between sedimentation of the hanging wall basin and erosion of the uplifted footwall; and rolling-hinge evolution along a migrating lower crustal flow front (Buck, 1988; McKenzie et al., 2000). However, they note that, with the exception of the rolling hinge, which itself requires a migrating magma source, these mechanisms do not account for the preferred basinward direction of migration. Alternatively, inward migration within a rift may result from brittle necking of strong crustal layers, where thinner brittle layers are easier to fault cf. thicker brittle layers outside the rift (Buiter et al., 2006).

The migration of fault activity first requires that new faults form, which occurs when existing faults do not accommodate the total geodetic extension rate. For instance, in Figure 3 we show that for a 220 year interseismic period, a single fault dipping at 50° accounts for approximately half the extension applied to the upper crust, with the balance accommodated by off-fault frictional-plastic strain. As fault slip and frictional-plastic strain accumulate during the seismic and interseismic periods respectively, we consider

the competing accumulation of elastic versus frictional-plastic strain during the interseismic period, with the caveat that the elastic component is replaced by fault slip during an earthquake. Elastic strain accumulates only in those regions of the upper crust maintained below the limit of frictional plastic failure, which is largely within the near-fault stress-failure shadow (Figure 5(a)). Elsewhere, tectonic extension applied for long time periods ensures the crust is close to failure and thus strain is accommodated by frictional plasticity.

For a system containing a single fault, frictional-plastic strain accommodates a significant component of the extension budget (Figure 4(b)). In terms of rift basin evolution, our model describes an area soon after the onset of extension in which only a single mature fault developed (see Figure 9(a)). The full extension rate cannot be accommodated by this structure, and thus frictional-plastic strain accommodates the shortfall. Frictional-plastic strain localizes into shear zones in the hanging wall. In our model these shear zones are synthetic to the main fault, and are therefore consistent with a scenario of migration of fault activity into the hanging walls of existing faults. However, the formation of secondary faults antithetic to the main fault, i.e., a graben geometry, is observed in many locations (e.g., Jackson and McKenzie, 1983) and also in numerical experiments (Melosh and Williams, 1989). Preference of one over the other has been considered from an energy minimization perspective (Melosh and Williams, 1989); however, Figures 5 and 6 indicate that the geometry of the primary fault, and the precise phase of the seismic cycle during which secondary fault formation occurs, may also play a role. Other factors that may control the geometry of secondary structures include the initial localization of shear

bands onto pre-existing planes of weakness or crustal heterogeneities, and the effects of gravitational spreading and basal detachment shear during rifting (e.g., Brun et al., 1994).

Localization of frictional plastic deformation into proto-faults occurs outside the stress-failure shadow of the main fault in the hanging wall block (Figure 9(b)). The extent of the failure shadow therefore exercise a degree of control on the location of these proto-faults and perhaps also the spacing of a sub-parallel array of faults. If the -1 MPa contour is used to estimate the dimension of the failure shadow, Figure 5(a) would suggest a minimum spacing of 10 km for footwall faults and 2 km for hanging wall faults. The latter is slightly smaller than the average fault spacing in the western Gulf of Corinth of 5-8 km (Rohais et al., 2007); however, as the dimensions of the stress-failure shadow depend on the thickness of the brittle upper crust and fault dip, direct comparison between model and field studies is not possible. For example, crustal thickness in the Gulf of Corinth of 10-12 km (Briole et al., 2000) is somewhat thicker than the 6-8 km in our model. For this crust, we would expect larger failure shadows than those modeled and consequently larger minimum fault spacings for both fault blocks. In addition, fault dips in the western Gulf of Corinth are not well constrained, although the dip of our modeled fault planes are consistent with those used by Briole et al. (2000) in modeling normal fault earthquakes in the Corinth Rift.

Coalescence of frictional plastic deformation into a single shear zone, culminating in a new discrete fault structure has important implications for the extension budget. The introduction of a second stress-failure shadow (Figure 9(c)) increases the area available

for interseismic elastic-strain accumulation and reduces that available for frictional-plastic straining. Proto-faults previously developing adjacent to the new fault would become inactive as deformation within its stress-failure shadow is switched to elastic/fault slip (Figure 9). Fault slip will then account for a greater component of the extension budget. Frictional-plastic straining will initiate in the hanging wall of the new fault, although at a reduced rate reflecting the extension accommodated by the second fault.

Rotation of faults to shallower dip angles, either through interseismic stretching or coseismic stress changes in the mid-crust (see section 7), has the opposite effect on the upper crustal extension budget. The eventual removal of a fault that was accommodating extension (Figure 9(d)), and the subsequent recovery of stresses within its failure shadow to Mohr Coulomb yield (Figure 9(e)), increases the extension shortfall, which is then accommodated by frictional plasticity. The formation of proto-faults in the hanging wall of the second fault then accelerates and the cycle continues. Alternatively, the recovery of the stress-failure shadow of the inactive fault would permit the development of new, cross-cutting faults in this region (Nur et al., 1986).

The net effect of the processes discussed above is a preference for the migration of faulting into the hanging walls of existing faults. However, an appropriate timescale to attach to this evolutionary scenario is unclear, largely due to ambiguity in the rate of formation of new faults. In our models, the frictional plastic shear strain rate in the hanging wall where proto-faults form is approximately $7 \times 10^{-16} \text{ s}^{-1}$, which for faulting to

seismogenic depths (6-8 km) represents a slip rate on the order of 0.15 mm yr^{-1} . At this rate, a proto-fault should exhibit several 10's of meters of offset after several hundred kyr. Goldsworthy and Jackson (2001) described uplift of coastal faults in the western Gulf of Corinth and suggested that the morphology might be generated in less than a million years. One might further expect acceleration of the slip rate with increasing rift maturity due to along-strike linkage of fault segments (Cowie and Roberts, 2001; Bell et al., 2009).

9. Conclusion

The onset of extension and subsequent evolution of normal faulting within the brittle crust was investigated via a computational model for an extending elastic-viscous-plastic crust. Extension applied at the boundaries of the model is accommodated in the upper crust through dip-slip on a fault contact surface or the accumulation of frictional-plastic strain, and in the lower crust through viscous deformation. Between these two layers, a mixed brittle-viscous region develops, which exhibits a rapid decrease in frictional-plastic straining, a slower decay in fault slip, and the onset of viscous deformation.

Frictional-plastic strain localizes preferentially within the hanging wall due to the distribution of shear stress near the fault. Coseismic displacements of the fault blocks create a stress-failure shadow that extends disproportionately into the footwall thereby favoring the formation of new faults in the adjacent hanging wall block. The dimension of the failure shadow at the surface is broadly consistent with the observed fault spacing in normal fault arrays of several kms.

707

708 Rotation of the fault plane about a horizontal axis at depth is observed to occur during
709 both phases of the seismic cycle. A long term rotation rate of 23.5° per Myr is divided
710 into almost equal components of far-field tectonic rotation during the interseismic period
711 and rapid rotation induced by mid-crustal stress changes during the seismic period. The
712 subsequent shallowing of the fault plane over many seismic cycles provides one avenue
713 by which faults may become inactive, accelerating the frictional-plastic straining and
714 proto-fault formation accommodates the extension shortfall.

715

716 The juxtaposition of these competing processes within an extending crust is consistent
717 with evolution of subsequent faults into existing hanging walls. However, a basinward
718 migration of faulting cannot occur indefinitely without a corresponding migration of the
719 locus of extension. Where the position of extension is static in time, recovery of the
720 stress-failure shadow around inactive faults may permit the eventual return of faulting to
721 the older footwalls. Cross-cutting of old, shallow structures by younger, steeper faults is
722 permitted, providing that they form outside the stress-failure shadow that extends
723 disproportionally into the footwall block.

724

725 As a final note, this simplified model for normal fault evolution requires careful
726 interpretation in the context of a richly complex crust. Material properties, stratigraphy
727 and tectonic forcings that vary along-strike and with depth also exert important controls
728 on the evolution of deformation and stress around faults, and by extension the
729 development of new structures. The mechanisms and scenarios described here do not

constitute a complete or prescriptive theory, but rather serve as an illustration of the evolutionary processes that may prevail in specific locales, e.g., the western Gulf of Corinth.

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Table 1:

Symbols used in the text. Unless otherwise stated rheological parameters are set at the bracketed values.

Figure 1

(a) A typical model geometry comprising a 60×25 km crust that hosts a single fault dipping 50° at the surface to a depth of 15 km and intersecting the surface 20 km from the left vertical boundary. The lower boundary is constrained to zero displacement in the *z* direction and the left and right boundaries move at a constant -6 and 6 mm yr⁻¹ in the *x* direction, respectively. The upper boundary is unconstrained. Model mesh (~10 000 elements) is shown superimposed on the domain. (b) Linear vertical temperature gradient of 35 °C km⁻¹ imposed through the crust. (c) Vertical differential stress, *q*, profile as prescribed by Eq. (2) and (4). (d) Flow diagram detailing time periods, fault properties and boundary conditions associated with various stages of the model.

Figure 2

Changes to displacement, stress and strain distributions due to fault slip observed during a seismic cycle of loading and slip. (a) Coseismic vertical surface displacements (adapted from Dempsey et al. (2012)) produced by the model (solid line) and at Borah Peak, Idaho (1983, $M=6.9$), and Hegben Lake, Montana (1959, $M=7.3$). Error bars for historical data are smaller than the marker dimension. (b) Change in fault parallel displacement during the 12th seismic period, i.e., fault slip. Note that down-dip displacements are taken to be negative. The inset profiles show fault parallel displacement (m) at the fault plane for the footwall (dashed) and hanging wall block (solid). The vertical scale is the same as for the contour plot. (c) Distribution of the differential stress, q , at the end of the 12th seismic cycle. (d) Change in fault parallel elastic strain during the 12th seismic period. Note that a non-linear contour interval has been used. (e) Total accumulated viscous strain after 12 seismic cycles. (f) Total accumulated frictional-plastic strain after 12 seismic cycles. The position of the fault is marked by the black and white dashed line.

Figure 3

(a) Interseismic and (b) seismic components of the (c) full seismic cycle extension budget. The contribution of fault slip (no marker), elastic (square), frictional plastic (circle) and viscous (triangle) deformation mechanisms at each depth add to give the unit normal extension profile. This extension budget has been constructed for deformation between $10 < x < 40$ km. (d) Extension budget for a crust at a critical stress state in the absence of faulting.

Figure 4

Elastic (grey), frictional plastic (black) and fault slip (black hatched) normalized components of the extension budget averaged over the upper crust ($0 < z < 5$ km). (a) Extension budget components against the length of the seismic cycle. (b) Extension budget components against the dip angle of the fault, for a seismic cycle length of 220 years.

Figure 5

Proximity of the differential stress to Mohr-Coulomb failure, $q_p - q$, (a) prior to and (b) following an earthquake for a 50° dipping fault and a crust at dynamic equilibrium. The BDT is marked by a sharp decrease in proximity to failure and black hatching indicates regions where the crust is in tectonic compression, i.e., $|\sigma_{xx}| > |\sigma_{zz}|$. (c) Distribution of frictional-plastic strain accumulated during the 12th interseismic period (Figure 1(d)). The position of the fault is marked by the black and white dashed line.

Figure 6

Frictional-plastic strain accumulated during (a) the 12th interseismic period, and (b) the 12th seismic period, for a model containing a 65° dipping fault. The position of the fault is indicated by the black and white dashed line.

Figure 7

Fault perpendicular displacement along the fault surface during (a) the interseismic and (b) seismic period. A linear regression profile has been fit to the data above a depth of 8

km. (c) Geometric arguments used to derive fault rotation from fault perpendicular displacement profiles. Note that negative displacements in (a) and (b) correspond to the direction of the vector **u** (positive up, right) in (c). The position of the axis of rotation is indicated in (b) and (c) by a black circle.

Figure 8

Deformation of the entire model domain and fault plane, exaggerated 2000 times, during (a) the interseismic and seismic periods. Boundary extension of 1.3 m at both lateral boundaries, along with a vertical contraction of 1 m, is marked. (b) Coseismic deformation in finer detail for the grey box in (a). Pivots about which rotation occurs are shown by open circles. Contours indicate coseismic mean stress change: light grey, unfilled = -0.25 MPa, light grey, filled = -0.5 MPa, dark grey, unfilled = +0.25 MPa, dark grey, filled = +0.5 MPa.

Figure 9

Conceptual evolution of a normal fault system over time. (a) Shortly after the onset of rifting the crust hosts a single fault that imposes a stress-failure shadow as in Figure 5 (grey shading). (b) Frictional plastic proto-faults, and (c) eventually new faults, develop in the hanging wall. (d) Continual rotation of faults out of the angle for reorientation reduces the component of extension accommodated by fault slip and promotes the development of new proto-faults. (e) Recovery of the failure shadow around inactive faults allows cross-cutting proto-faults to develop.

Symbol	Explanation
<i>Rheological parameters</i>	
ρ	density [2800 kg m ⁻³]
E, ν	Young's Modulus [10 GPa], Poisson's ratio [0.25]
C, ϕ	cohesion [1 MPa], material angle of friction [30°]
μ_d	dynamic fault friction coefficient [0.2]
$\dot{\epsilon}_v$	viscous strain rate [6.3×10^{-15} s ⁻¹]
A, Q, n, R	pre-exponential constant [4×10^{-34} Pa ⁻⁴ s ⁻¹], activation energy [135 kJ mol ⁻¹], power-law exponent [4], gas constant [8.314]
T	temperature
<i>Stress terms, components and conventions</i>	
σ_{ij}, s_{ij}	stress and deviatoric stress tensors
σ_1, σ_3	maximum, minimum principal compressive stress
p, p_f	mean stress, pore fluid pressure
q, q_p, q_v	differential stress (plane strain Von Mises), differential stress at frictional plastic yield, differential stress maintaining viscous strain rate
τ, σ_n	shear, normal stress on fault surface
sign	negative values of stress indicate compression
<i>Displacement and strain terms</i>	
x, z	Cartesian coordinates, z is negative when directed downward
u_x, u_z, u_{para}	horizontal, vertical and fault parallel displacement

Table 1: Symbols used in the text. Unless otherwise stated rheological parameters are set at the bracketed values.

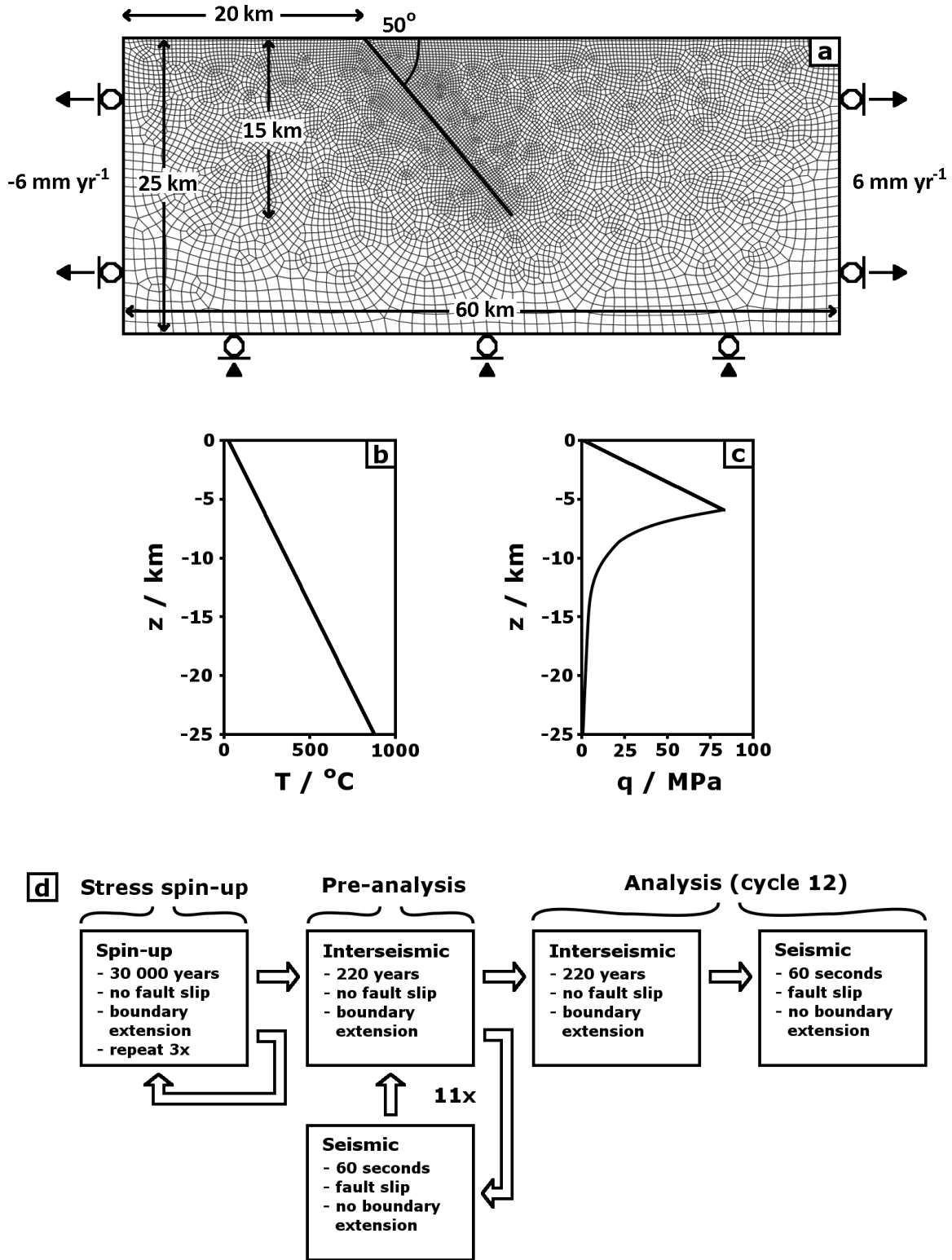


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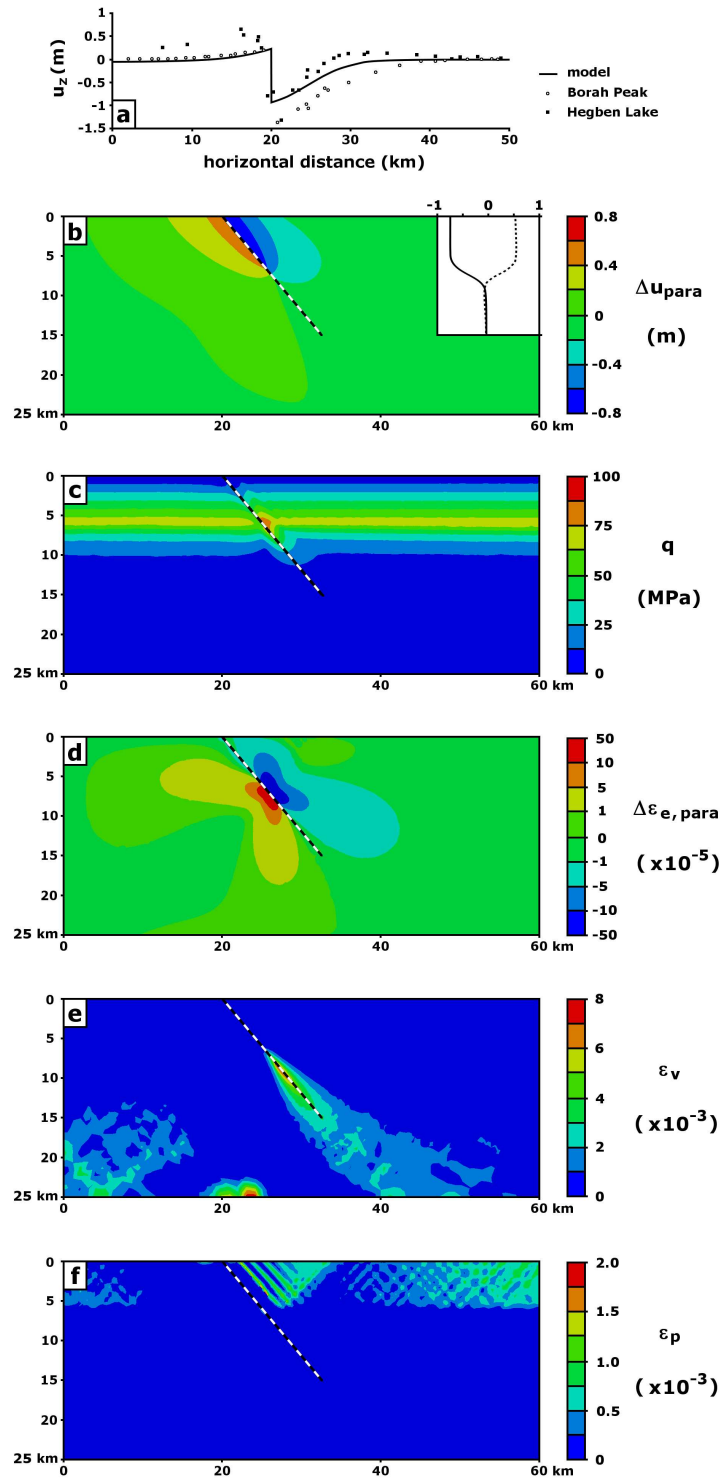


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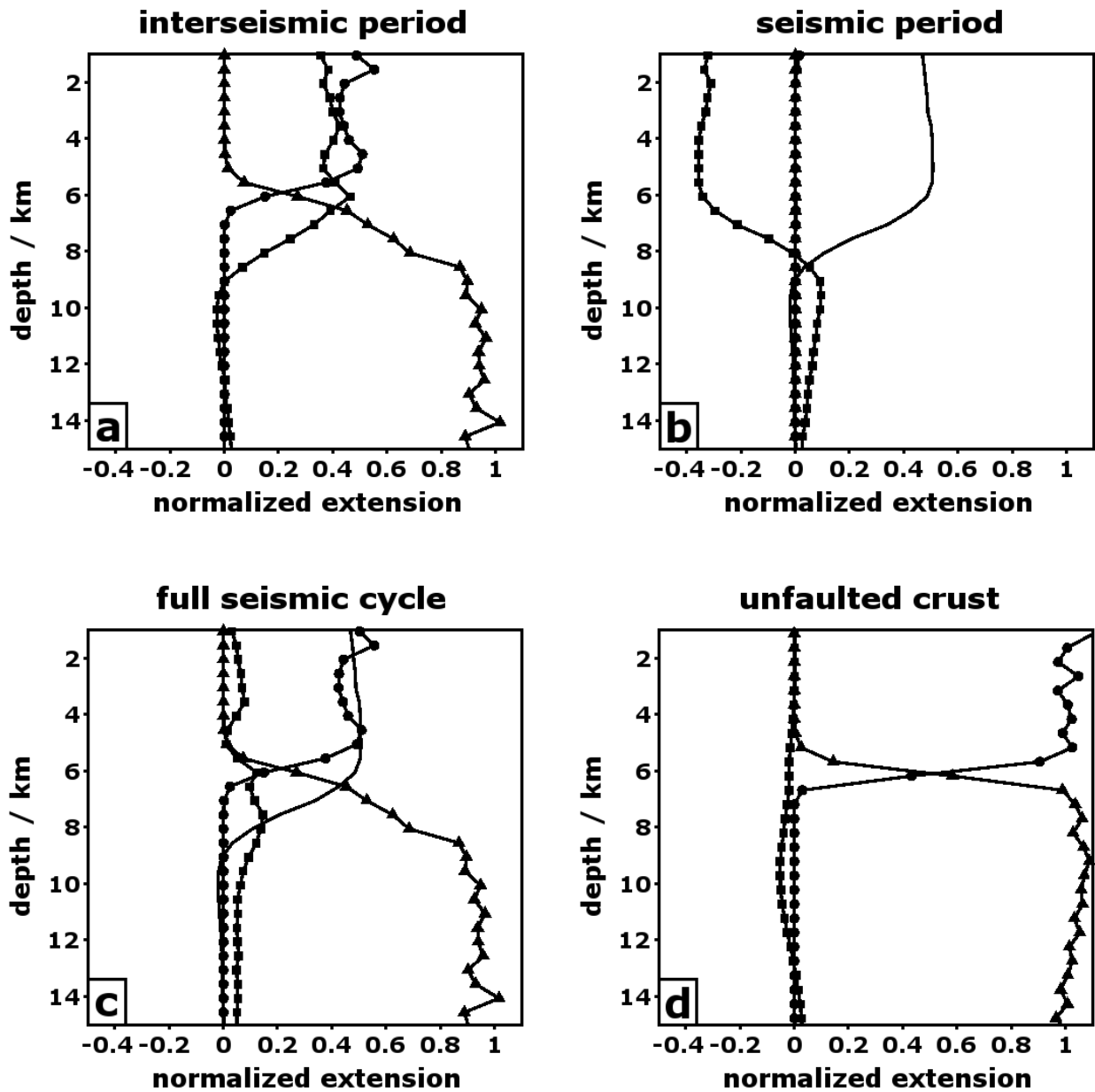


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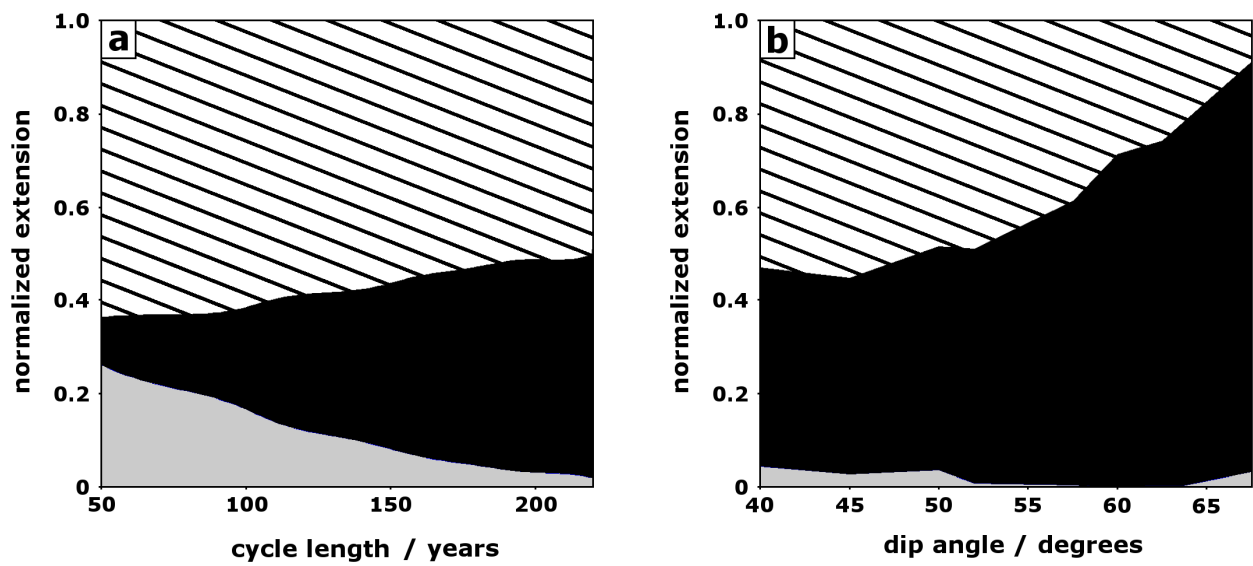


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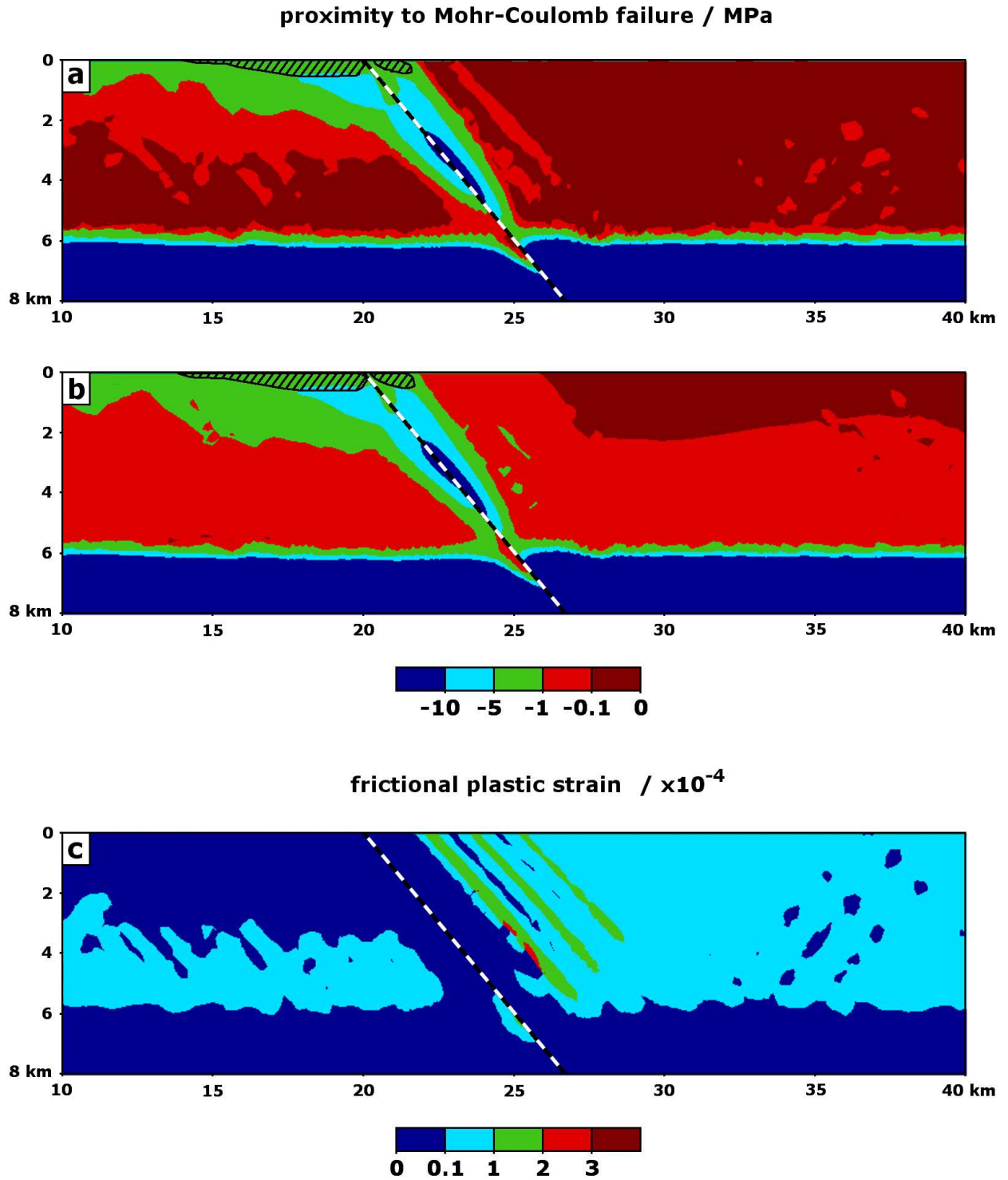


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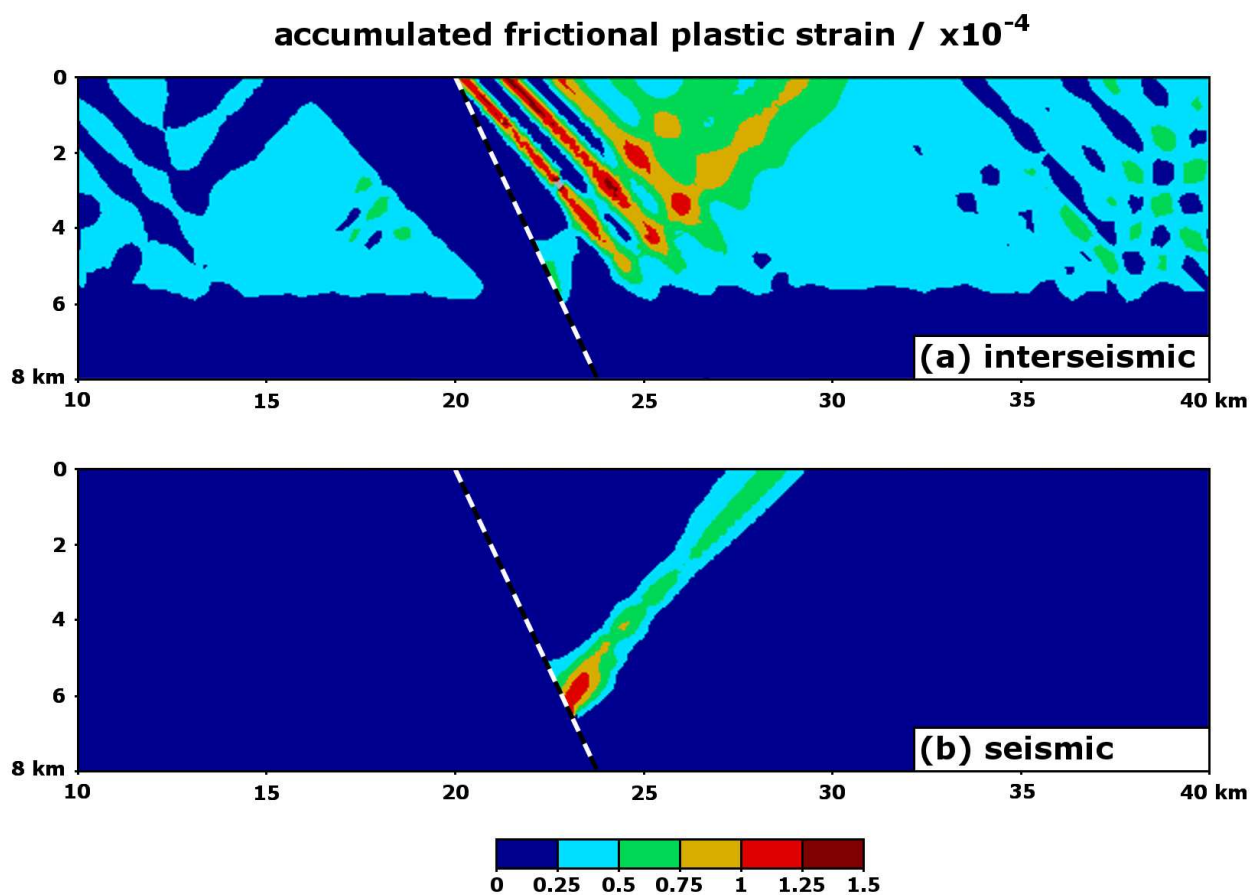


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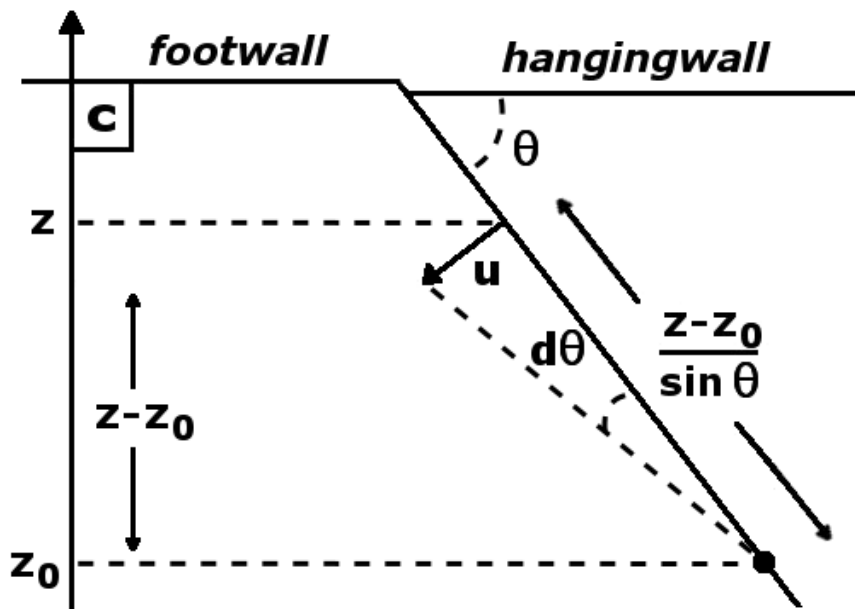
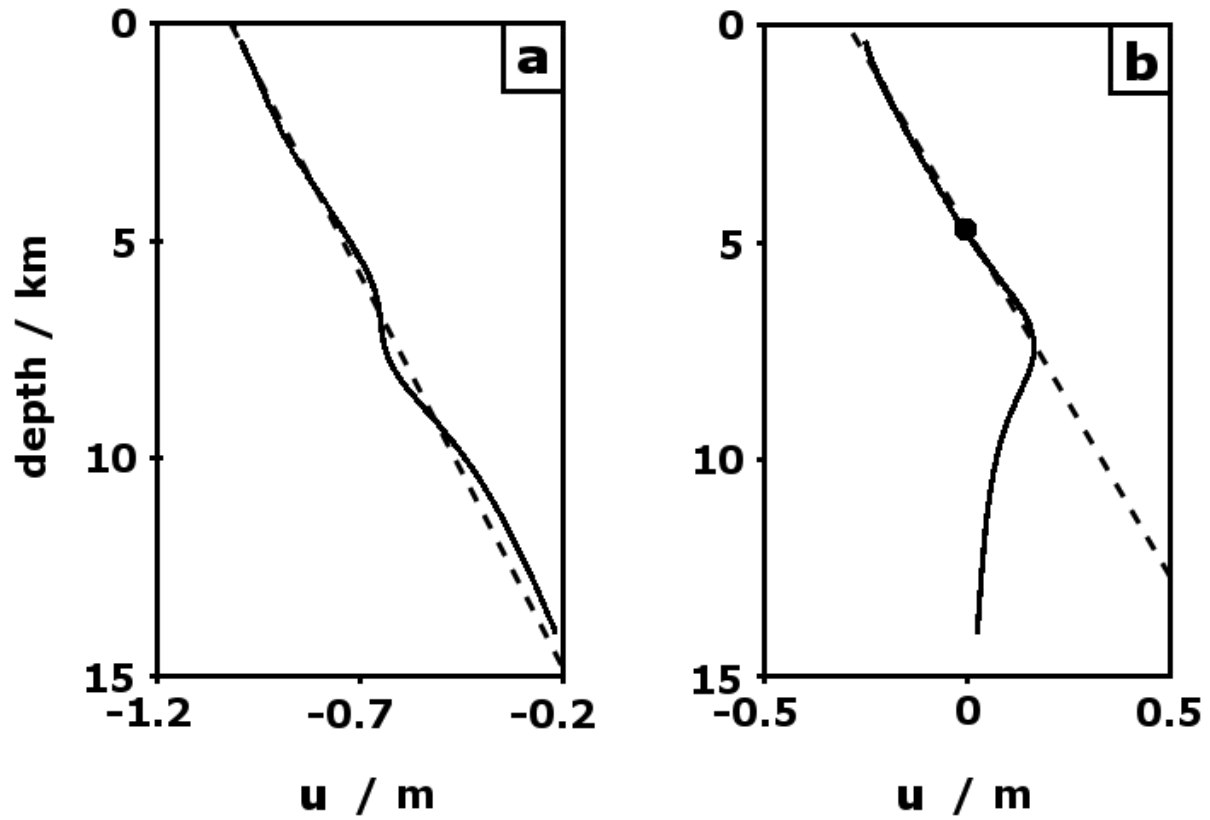


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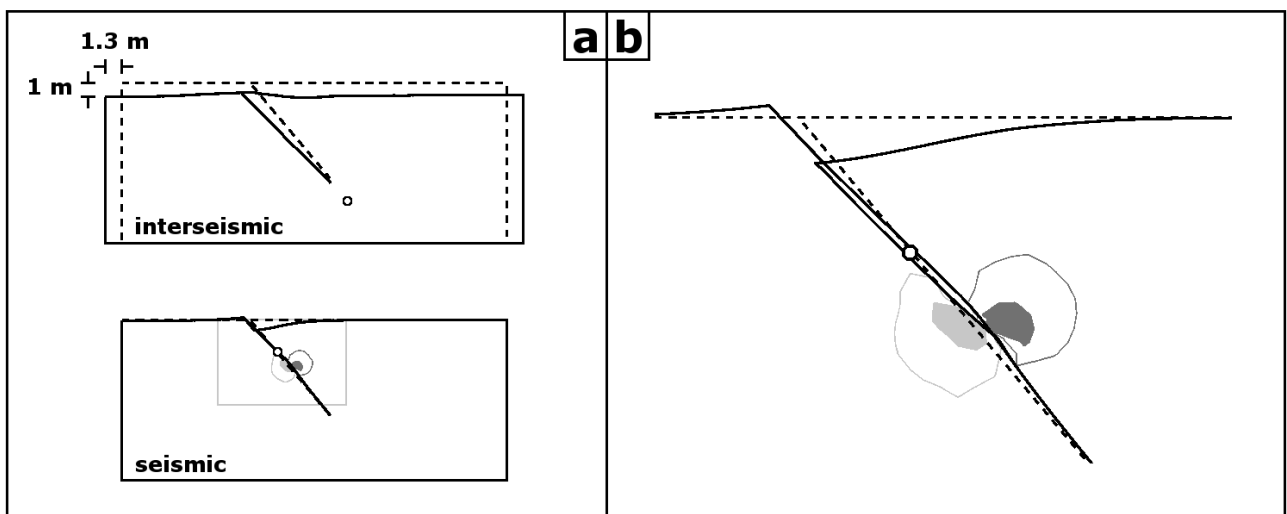


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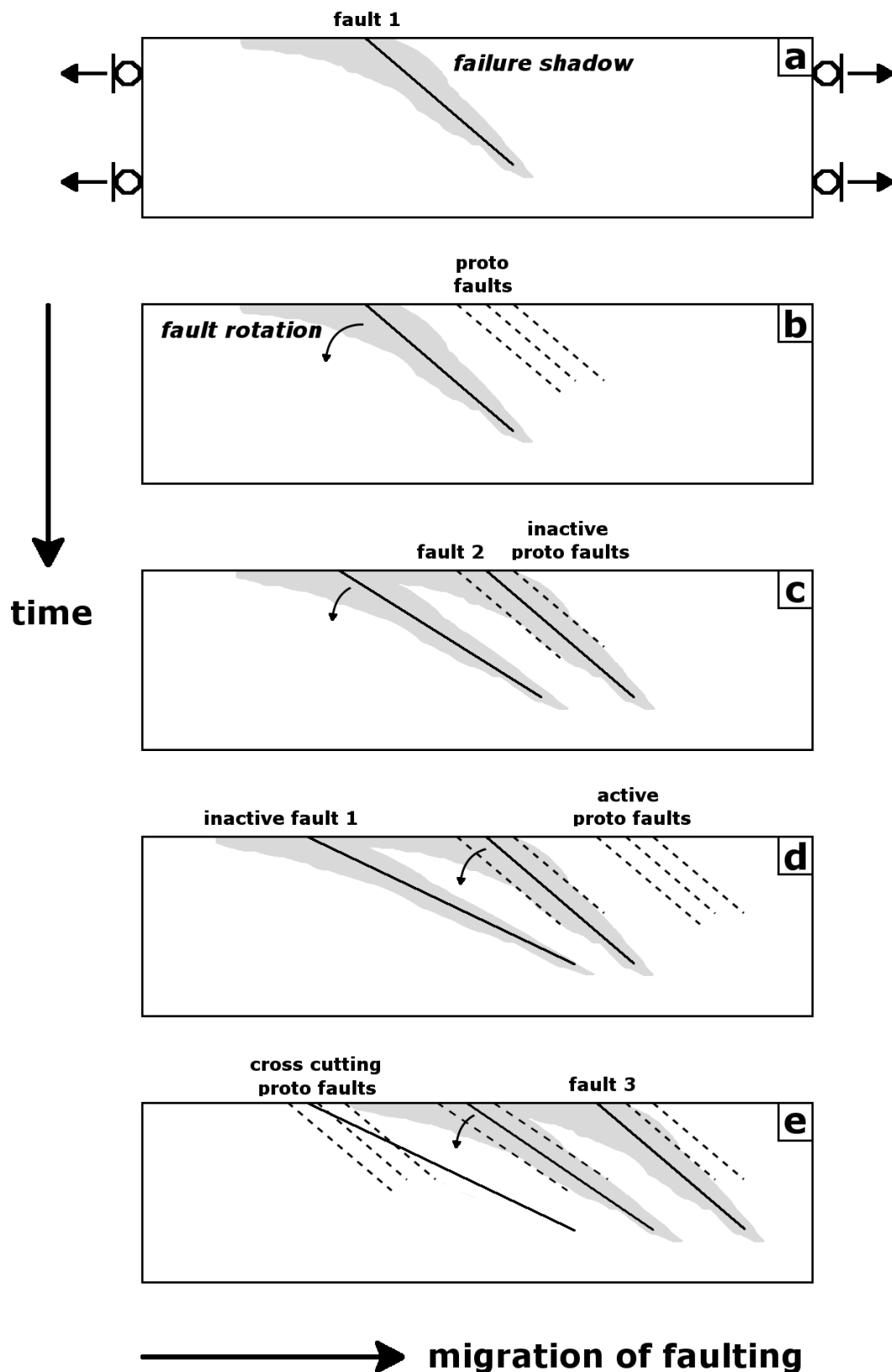


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