

2013 Annual SCEC report

Numerical models of stress transfer from plate motion to mature transform faults

Publications resulted from this project:

Takeuchi, C. and Y. Fialko, On the effects of thermally weakened ductile shear zones on postseismic deformation, J. Geophys. Res., 118, doi:10.1002/2013JB010215, 2013.

Summary of results:

We have extended 2-D anti-plane strain models of *Takeuchi and Fialko* (2012) to three dimensions to investigate effects of viscous shear zones that result from thermomechanical coupling on the geodetically observable deformation transients following an earthquake on a vertical strike-slip fault. We also explored potential kinematic similarities between viscoelastic models incorporating shear zones, and elastic models incorporating rate-strengthening friction on a deep aseismic fault root. We find that the thermally-activated shear zones have little effect on postseismic relaxation. In particular, the presence of shear zones does not change the polarity of vertical displacements in cases of rheologies that are able to generate robust postseismic transients. Stronger rheologies can give rise to an opposite polarity of vertical displacements, but the amplitude of the predicted transient deformation is generally negligible. We conclude that additional (to thermomechanical coupling) mechanisms of strain localization are required for a viscoelastic model to produce a vertical deformation pattern similar to that due to afterslip on a deep extension of a fault. We also investigated the discriminating power of models incorporating Burgers and power law rheology. These rheologies were proposed to explain postseismic transients following large ($M7$) earthquakes in the Mojave desert, Eastern California (e.g., *Freed and Bürgmann*, 2004; *Pollitz*, 2003). Numerical simulations indicate that it may be difficult to distinguish between these rheologies even with high-quality geodetic observations for observation periods less than a decade. Longer observations, however, may potentially allow discrimination between the competing models, as illustrated by the model comparisons with available GPS and InSAR data.

Technical report:

Large earthquakes are often followed by spatially- and temporally- varying deformation as the Earth’s crust and mantle respond to stress perturbations produced by coseismic slip. Imaging of this transient deformation has dramatically improved in recent years as the density of GPS and InSAR observations has increased in both time and space around seismogenic faults. The primary mechanisms invoked to explain postseismic deformation include viscoelastic relaxation (*Elsasser, 1969; Nur and Mavko, 1974; Savage and Prescott, 1978*), aseismic fault creep (*Ruina, 1983; Tse and Rice, 1986*), and poroelastic rebound (*Booker, 1974; Jonsson et al., 2003*). Poroelastic relaxation and shallow afterslip are upper crustal processes that have predominantly near-field effects, while deeper afterslip and viscoelastic relaxation occur mainly in the lower crust and upper mantle and thus produce broad-ranging surface displacements.

The individual contributions of these mechanisms to postseismic relaxation following a given event may be difficult to identify, largely due to the non-uniqueness of inverse models. For instance, for an infinitely long strike-slip fault undergoing uniform coseismic displacement, viscoelastic relaxation predicts surface deformation that is indistinguishable from that due to an appropriately configured elastic dislocation model (*Savage, 1990*). It was proposed that in the case of three-dimensional (3-D) deformation due to finite ruptures, the afterslip and viscoelastic relaxation mechanisms may in principle be distinguished as these mechanisms predict vertical postseismic velocity patterns that are opposite in polarity (e.g., *Pollitz et al., 2001*). In addition, viscoelastic relaxation models typically predict relatively large fault-normal velocities up- and down-strike from a finite strike-slip rupture, while such velocities are small or absent in afterslip models (*Hearn, 2003*). However, if multiple mechanisms contribute to postseismic relaxation, even high-quality observations may not allow for robust discrimination between the candidate mechanisms. For example, postseismic deformation in the Eastern California Shear Zone (ECSZ) following the 1992 M_w 7.3 Landers and 1999 M_w 7.1 Hector Mine earthquakes has been ascribed to viscoelastic relaxation (*Pollitz, 2003*), (*Freed and Bürgmann, 2004*), a combination of poroelastic relaxation and afterslip (*Fialko, 2004a; Peltzer et al., 1998*), poroelastic and viscoelastic relaxation (*Masterlark and Wang, 2002*), as well as other mechanisms (*Jacobs et al., 2002; Massonnet et al., 1996*).

Afterslip on the deep extension of the rupture plane may be considered to be kinematically analogous to a narrow ductile shear zone, with shear deformation governed by the constitutive equation for stress-driven frictional afterslip rather than that for viscoelastic flow (e.g., *Barbot et al., 2009*). Several previous investigations of postseismic relaxation have allowed for strain localization in the viscoelastic medium by incorporating a tabular region of reduced effective viscosity along the fault root (e.g., *Freed et al., 2007; Hearn et al., 2009; Kenner and Segall, 2003*). However, the introduction of such shear zones into numerical models is usually *ad hoc* and neglects details of how such zones formed in the first place. In a previous study, we demonstrated that shear heating and thermomechanical coupling in the ductile substrate give rise to long-lived localized shear zones beneath mature strike-slip faults (*Takeuchi and Fialko, 2012*). These shear zones participate both in loading faults interseismically and relaxing coseismic stress perturbations. Figure 1 shows the structure of a thermal shear zone in terms of the effective viscosity $\eta_{eff} = \sigma_d / \dot{\epsilon}_d$, where σ_d and $\dot{\epsilon}_d$ are the deviatoric stress and strain rate, respectively, and relaxation time $t_r = \eta_{eff} / G$, where G is the assumed shear modulus of the substrate (32 GPa). This structure was developed in the result of 20 Ma of steady slip on a fault at a rate of 4 cm/a.

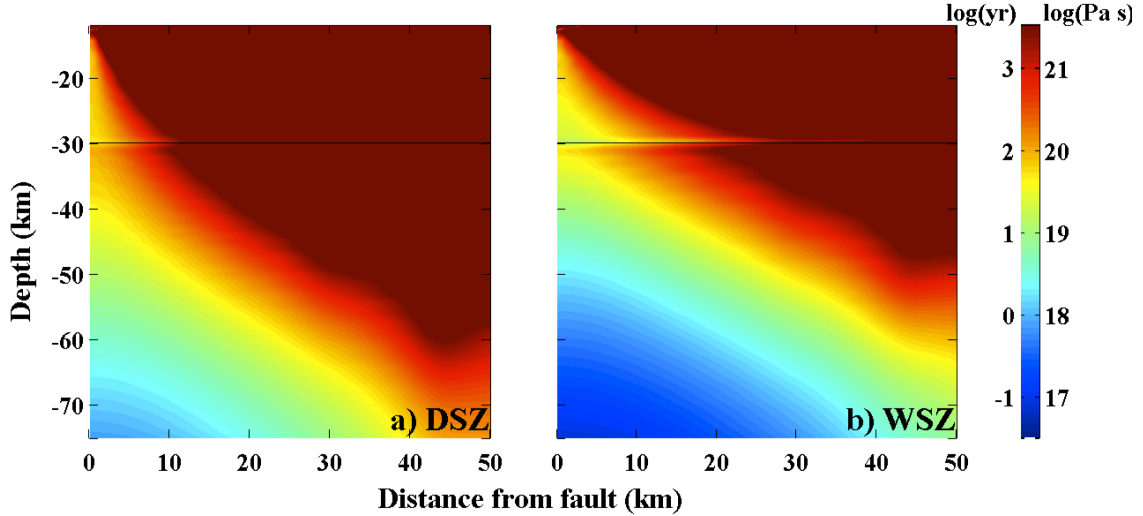


Figure 1: Effective viscosity and relaxation times for models of a mature thermal shear zone assuming (a) dry and (b) wet rheology. Relaxation times are calculated as a ratio of the effective viscosity over the shear modulus.

We extended our 2-D results to 3-D finite-rupture scenarios. We used our models to test the hypothesis that, under assumptions of laboratory-derived rheologies and far-field loading, a model incorporating highly localized ductile shear zone produced by shear heating may produce a postseismic deformation field similar to that predicted by a frictional afterslip model. In this case, different patterns of postseismic deformation might be expected depending on the effective fault age and slip rate, with afterslip-like localized viscoelastic shear dominating postseismic relaxation in the case of mature faults, and diffuse viscoelastic relaxation dominating in the case of immature faults. We find that, contrary to the hypothesis, thermally-induced shear zones have little effect on postseismic relaxation. It follows that the degree of shear localization necessary for a viscoelastic relaxation model to mimic postseismic surface deformation due to frictional afterslip model requires additional localization mechanisms.

Most of numerical simulations were carried out using the finite element software Abaqus/Simulia (www.simulia.com/products/abaqus.fea.html). The model domain is a 600 km (fault-normal, x coordinate) \times 600 km (along-strike, y coordinate) \times 75 km (vertical, z coordinate) rectangular block. The domain is composed of three horizontal rheological layers: a 12 km-thick elastic upper crust overlying an 18 km-thick viscoelastic lower crust and a 45 km-thick viscoelastic upper mantle. A 600 km-long ($y = -300$ to 300 km) vertical planar fault is introduced within the domain at $x = 0$. The fault penetrates through the entire upper crust and roots in the lower crust at a depth of 17 km.

The model domain is discretized into 612,000 elements, with 34 element layers in the fault-normal direction, 240 layers in the strike direction, and 75 vertical layers. Fault-normal node spacing decreases towards the fault, from 93.58 km in the far-field to 0.5 km on the fault. Along-strike node spacing varies from 19.57 km in the far-field to 0.5 km for $|y| > 35$ km; nodes are spaced by 0.5 km within $|y| \leq 35$ km. Nodes are spaced vertically by 1 km.

Rheology

We explored four candidate rheologies of the viscoelastic lower crust and upper mantle. Two

Table 1: Laboratory-derived material properties of rocks.

Rock Type	A (MPa ⁻ⁿ s ⁻¹)	n	Q (kJ mol ⁻¹)	ρ (kg m ⁻³)	Ref.
Dry diabase	8.0	4.7	485	2850	1
Wet diabase	2.2×10^{-4}	3.4	260	2850	2
Dry olivine	1.1×10^4	3.5	535	3320	3
Wet olivine	3.6×10^5	3.5	480	3320	3

¹*Mackwell et al. (1998)*; ²*Shelton and Tullis (1981)*; ³*Hirth and Kohlstedt (2004)*

models incorporate linear rheologies for these layers. The first model incorporates a Maxwell rheology for the entire ductile substrate. The mantle has a viscosity η_m of 1.6×10^{17} Pa s, corresponding to the transient rheology of *Pollitz (2003)*, and a shear modulus μ_m of 70 GPa. The lower crust in this model has a viscosity η_c of 3.2×10^{19} Pa s and a shear modulus μ_c of 38 GPa. Poisson’s ratio is 0.25 in all model layers. We refer to this model as MS.

The second model incorporating linear rheology assumes a biviscous Burgers body rheology, in which Maxwell and Kelvin viscoelastic elements are in series. We select Burgers body parameters $\eta_1 = 4.6 \times 10^{18}$ Pa s, $\eta_2 = 1.6 \times 10^{17}$ Pa s, $\mu_1 = 70$ GPa, and $\mu_2 = 70$ GPa, which were the best-fitting values of *Pollitz (2003)* for the rheology of the mantle beneath the Mojave Desert, California. We also use the best-fitting model of *Pollitz (2003)* for the lower crust, which incorporates a Maxwell rheology with a viscosity $\eta_c = 3.2 \times 10^{19}$ Pa s and a shear modulus $\mu_c = 38$ GPa (this value represents an average of the depth-varying shear modulus of *Pollitz (2003)* over our lower crustal depth range of 12-30 km). The lower crust for the biviscous-mantle model and the Maxwell-mantle model are thus identical. Poisson’s ratio is 0.25 for all materials in the biviscous-mantle model. We refer to this model as BL.

The remaining two models incorporate non-linear, temperature-dependent rheology. We assume mafic (diabase) and ultramafic (olivine) composition for the lower crust and upper mantle, respectively (*Rudnick and Fountain, 1995*); (*Karato and Wu, 1993*). To account for variability in ductile strength, we consider end-member models of hydrated (weak) and dry (strong) mineral compositions. Laboratory-determined material parameters for these models are listed in Table 1. In all power law models, the elastic behavior of both the upper crust and the ductile substrate is governed by the linear isotropic Hooke’s Law, with a shear modulus and Poisson’s ratio of 32 GPa and 0.25, respectively.

Earthquake Simulations

In this study we employed kinematically-driven earthquake cycles, i.e. coseismic fault slip is prescribed as a boundary condition. Such models generate unphysical stresses in the lithosphere; in particular, repeated kinematically-prescribed earthquakes produce negative (i.e. having sense opposite to that of fault slip) stresses in the seismogenic layer (*Takeuchi and Fialko, 2012*). However, surface velocities are relatively insensitive to the type of boundary condition on a fault surface (*Takeuchi and Fialko, 2012*).

The models are “spun up” by prescribing creep on a fault in the upper crust and at the far sides of the model for 10 Ma, followed by ten earthquake cycles on the entire 600 km-long fault. In these cycles, we prescribe an instantaneous coseismic slip of 8 m on the upper 12 km of the

fault. Fault slip is again cosine-tapered to zero at 17 km depth. We then lock the fault for an interseismic period of 200 years. The total relative plate velocity (40 mm/yr) is maintained on the far-field boundaries of the model domain for the duration of the earthquake cycle sequence. Because deformation is anti-plane strain, only fault-parallel motion is allowed.

Following the final interseismic period of the last system-wide earthquake, we simulate a finite rupture on a 70 km-long segment in the middle of the domain ($|y| \leq 35$ km). Slip amplitude is the same as in previous system-wide events. The respective moment is 1.08×10^{20} N m, similar to the moment estimates for the 1992 M_w 7.3 Landers rupture (Fialko, 2004b; Kanamori *et al.*, 1992; Sieh *et al.*, 1993), and slightly higher than those for the 1999 M_w 7.1 Hector Mine rupture (Dreger and Kaverina, 2000; Simons *et al.*, 2002).

We also used observations of postseismic transients due to the 1992 Mw7.3 Landers and 1999 Mw7.1 Hector Mine earthquakes in the Mojave Desert (eastern California) to investigate whether longer time series can help discriminate between the previously proposed deformation mechanisms. Figure 2 shows the observed and predicted horizontal displacements, referenced to site OPRD.

The biviscous model predicts quasi-constant horizontal velocities after robust transient deformation within the first ~ 3 months. A comparison to the assumed secular velocity (relative to station OPRD, as denoted by the dotted lines in Figure 2) shows that the coseismic stress changes remain essentially unrelaxed after 13 years of postseismic deformation. In contrast, the power law model indicates that horizontal velocities nearly returned to secular values by the end of the 13 year period. While neither model provides a satisfactory fit to all the data (possibly due to model simplifications), it is worth noting that the biviscous model tends to overpredict the observed horizontal velocities, while the power law model generally underpredicts the data. Because the coseismic source is identical between the models, these differences are due to the assumed rheological properties alone. Note that the biviscous model has systematically higher misfits than the power law model for east, north, and vertical displacement components, regardless of which station is used as a reference site.

Comparisons of model predictions to observations of postseismic deformation in the Eastern California Shear Zone suggest that available GPS and InSAR data do not allow one to distinguish viscoelastic relaxation models incorporating Burgers and power law rheologies on short (< 5 years) observation intervals. However, differences in the decay rate of the transient deformation predicted by the two models indicate that the models may be distinguished given continued GPS and InSAR monitoring on decadal timescales.

Outreach and broader impacts:

This project provided training and support for one graduate student. The PI (Fialko) used results of this study in a graduate class taught at SIO.

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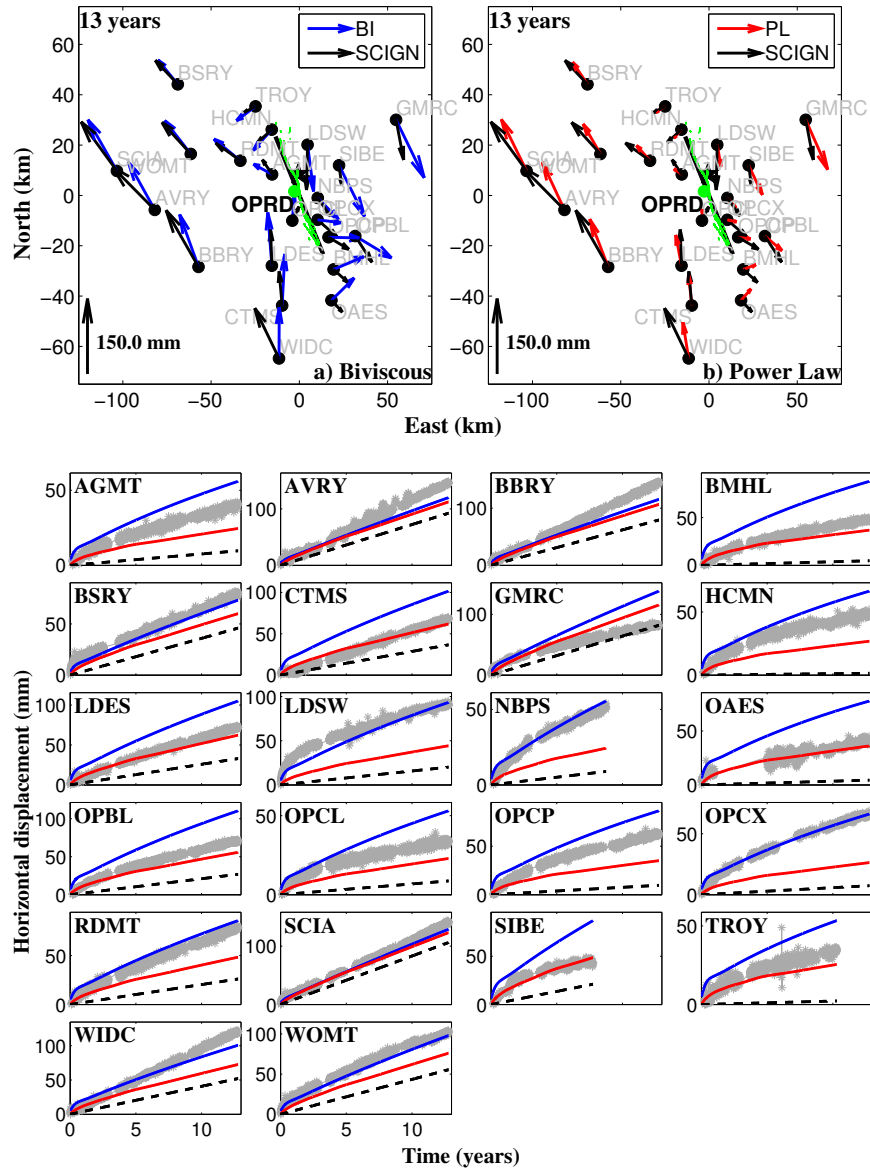


Figure 2: (a, b) Thirteen years of horizontal displacement, in millimeters, due to the Hector Mine earthquake. Displacements referred to the continuous GPS station OPRD (green dot) and shown by black (observed), blue (biviscous model), and red (power law model) arrows. Black circles denote GPS station locations. Black star denotes the epicenter of the Hector Mine earthquake, heavy black line denotes the simulated Hector Mine rupture, and green line denotes the geologically mapped fault trace. Other panels show time series of observed (grey stars) and modeled (solid blue line = biviscous model, solid red line = power law model) horizontal displacements, referred to station OPRD, for all near-field continuous GPS sites. Dashed black line denotes displacement due to modeled station secular velocity referred to simulated station OPRD.

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