Spatiotemporal evolution of a fault shear stress patch due to viscoelastic interseismic fault zone rheology

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Abstract

We conducted numerical studies to explore how shear stress anomalies on fault planes (shear stress patches) evolve spatiotemporally during the interseismic period under the influence of viscoelastic rheology assigned to fault zones of finite thickness. 2-D viscoelastic models consisting of a fault zone and host rock were sheared to simulate shear stress accumulation along fault zones due to tectonic loading. No fault slip along a distinct fault planes is implied in the model, thus all fault shear motion is accommodated by distributed deformation in the viscoelastic fault zone. Results show that magnitudes of shear stress patches evolve not only temporally, but also spatially, especially when the stress anomaly is created by a geometrical irregularity (asperity) along the interface of an

elastic host rock and viscoelastic fault zone. Such shear stress anomalies diffuse spatially so that the spatial dimension of the shear stress patch appears to grow over time. Models with varying fault zone viscoelastic properties and varying fault zone viscosity both show that such spatial diffusion of shear stress is enhanced by increasing the contribution of the viscous behavior. The absolute rate at which shear stress patches grow spatially is generally not influenced by the size of the shear stress patch. Therefore shear stress patches with smaller dimensions will appear to grow quicker, in the relative sense, compared to larger stress patches. These results suggest that the minimum dimensions of shear stress patches that can exist along a fault could be governed by the effective viscosity of the fault zone. Therefore patterns of accumulated shear stress could vary along faults when viscous properties are heterogeneous, for instance due to depth or material heterogeneity, which has implications on how earthquake rupture behavior could vary along faults.

Keywords

fault shear stress; stress heterogeneity; viscoelasticity; interseismic tectonics; fault zone rheology

1. Introduction

Earthquake fault slips are spontaneous phenomenon driven by shear stresses acting on fault planes. Therefore mechanical understandings of earthquake ruptures require the knowledge of (1) the stress condition along faults prior to an earthquake, and (2) conditions leading to fault frictional instability. In the past decades, there have been many studies addressing the latter aspect, which have investigated the cause of fault instability (Brace and Byerlee, 1966) and revealed various fault weakening mechanisms such as rate and state friction (Dietrich, 1979), thermal pressurization (Lachenbruch, 1980), and rock dynamic weakening at high velocities (Tsutsumi and Shimamoto, 1997; Mizoguchi et al., 2006). On the other hand, relatively fewer studies have dealt with the former aspect of characterizing

the origin and distribution of fault-resolved shear stress which accumulates between earthquakes.

The importance of understanding fault shear stresses is evident from geophysical observations and theoretical studies. Geodetic studies from subduction zones reveal heterogeneous distribution of fault locking degree which indicates that slip deficits relative to the steady plate motion accumulates unevenly along faults (Burgmann et al., 2005; Chlieh et al., 2008), creating a heterogeneous fault shear stress distribution. Such patterns of locking degree have been shown to exhibit close spatial correlation with the distribution of coseismic fault slip inverted from seismic waveforms of megathrust earthquakes (Moreno et al., 2010; Loveless and Meade, 2011). Recent dynamic rupture models incorporating fault roughness and initial stress heterogeneities (Ampuero et al., 2006; Ripperger et al., 2007; Dunham et al., 2011c; Fang and Dunham, 2013) also show that the final slip distributions of modeled earthquake ruptures are heavily influenced by the initial distribution of stress along modeled faults.

It is also suggested that the degree of the stress heterogeneity varies along a fault based on frequency contents of radiated waves. Recent observations from megathrust earthquakes shows that sources of high-frequency radiation revealed from back-projection methods tend to locate at the deeper regions of the seismogenic zone, although fault slip is generally greater at shallower depths (e.g., Wang and Mori, 2011; Yao et al., 2011, 2013; Kiser and Ishii, 2012; Lay et al., 2012). Results by Yao et al. (2013) specifically suggest that there may be a continuous deepening trend of radiation source locations, from the trench to the down-dip limit of the seismogenic zone, as the frequency of the band-passed waveforms increase. If high-frequency waves are to be emitted from the acceleration and deceleration of rupture fronts, these observations suggest that the degree of heterogeneity of fault shear stress and/or frictional properties increases with depth. Along-strike contrast in frequency content of radiated waves was also seen from strong motion data during the Chi-Chi earthquake (Ma et al., 2001). Whether this case should also be interpreted as a consequence

of difference in stress or fracture energy heterogeneity is unknown, but suggests that fault stress heterogeneity could have bearing on ground motions caused by earthquakes.

The development of fault shear stress is typically considered to result from steady elastic loading on a fault interface, which is either fully or partially locked obeying some fault friction law. As mentioned above, geodetic observations allow us to invert for the locking degree distribution along a fault, but the limited resolution only yield first order patterns of fault locking. Alternatively, interseismic fault behavior can be forward modeled in numerical models capturing fault slip behavior spanning multiple earthquake cycles (Lapusta et al., 2000; Hetland, 2010). These studies provide one complete picture of how interseismic and coseismic processes influence each other, but spatial scales of the heterogeneities produced in these models are often restricted to the length scale of the heterogeneity provided a priori in the model (e.g. the distribution of frictional properties, segmentation between seismic and aseismic regions). We also argue, as discussed below, that interseismic fault behavior may not be best represented by friction laws that govern localized deformation along zero-width fault planes. We still have limited knowledge about how fault stress heterogeneity develops at various scales.

In this study, we focus on one fundamental aspect of interseismic fault shear stress development. We investigate numerically how a single stress anomaly on a fault plane evolves over time when viscoelastic rheology is introduced for materials constituting fault zones. The aim is to gain a basic understanding of a source process, namely the behavior of a single stress patch on a fault, which shall lead to our comprehension of fault stress heterogeneities and their development in general.

There are several reasons why we focus on the influence of viscoelastic rheology in contrast to common fault models which consider frictional interfaces embedded in elastic host rocks. First of all, rocks do exhibit viscoelastic properties even at depths shallower than the brittle-ductile transition without occurrence of crystal plastic creep of silicate minerals. The process of fault stress loading takes place over long time scales on the order of decades to centuries which suggests that slow bulk deformational processes such as

volumetric creep assisted by compaction (Dudley et al., 1998; Hagin and Zoback, 2004; Chang and Zoback, 2009; Sone and Zoback, 2013) or pressure solution (Weyl, 1959; Durney, 1972; Gratier and Guiguet, 1986) can also become an important process that influences the stresses loaded on faults during the interseismic period. Sone and Zoback (2014) studied the time-dependent deformational properties of shale gas reservoir rocks and showed that rocks with Young's modulus as high as 20-60 GPa can exhibit enough time-dependent behavior to relax significant fractions of differential stresses applied to the rock over engineering time scales (i.e. less than decades).

These time-dependent rock behaviors caused by the presence of some porosity are usually considered to be effective at shallower conditions compared to seismogenic depths (few tens of km). However, if we acknowledge the facts that (1) rock materials within and around faults are highly damaged (Chester et al., 1993), (2) many tectonic and petrophysical evidences point to pore pressure ratios (pore pressure divided by lithostatic stress) close to 1 along subduction plate boundaries (Audet et al., 2009; Saffer and Tobin, 2011), (3) temperature along plate boundaries are significantly lower than the surroundings (Hyndman and Wang, 1993; Hyndman et al., 1995; Wang et al., 1995), it is not unreasonable to expect that rock rheological behaviors typically considered at engineering depths ($< \text{few km}$) also apply to seismogenic depths along subduction zones. Studies of postseismic crustal deformation have also indicated that either time-dependent afterslips or off-fault viscous deformation is needed to explain transient crustal deformations observed after large earthquakes (Thatcher and Rundle, 1984; Savage, 1995). Thus it is plausible to believe that many rock materials at seismogenic depths can also exhibit significant viscous behavior over geological time scales. If fault rocks and the surrounding host rock exhibit time-dependent behavior, tectonic loading of fault stresses may not be as simple as described by frictional surfaces embedded in elastic mediums. Time-dependent viscous deformation relaxes stress simultaneously as it is also being loaded steadily over time.

In this paper, we argue that viscous deformation around asperities can alter the spatial pattern of stress anomalies along faults in addition to its effect on stress magnitudes. We

also study how differences in fault rheological properties (e.g. viscosity, type of constitutive law) affect the spatial evolution of fault stress anomalies during interseismic loading. This is motivated by the possibility that change in fault rock properties may help explain the difference in fault stress heterogeneity with depth, and possibly along-strike, suggested by seismological observations.

2. Finite Element Model Setup

2.1 General Description

We conducted 2-D time-dependent quasi-static calculations using the finite element modeling software PyLith 2.0.3 for Windows (Aagaard et al., 2013; Aagaard et al., 2015) to investigate how a patch of shear stress anomaly on a locked fault plane evolves over time when viscoelastic rheology is introduced to the materials within and around a fault zone. The models consist of a 2-D (plane strain) rectangular block of 200 m by 30 m dimension where one of the long edges is considered to be the fault plane and the other side considered to be some boundary within the host rock (Figure 1). For convenience, we call this boundary on the opposite side of the fault the “far-field” boundary, although they are not particularly far away (30 m) from the fault. The scale of the model is chosen to roughly represent fault structure at the field scale and above. Note that the model does not contain any fault plane within the rectangular domain. All fault-parallel shear motion is thus accommodated by distributed deformation in the continuum and no fault slip is implied in the model. Due to the model geometry, we are assuming in this study that the problem is symmetric about the fault plane although this may not be true in nature. Two kinds of models were considered which employ meshes with different internal geometries and assign viscoelastic rheology in different domains of the models as explained in the following sections.

We impose displacement boundary conditions on the fault plane boundary and far-field boundary of the model to induce fault shear stress or simulate interseismic tectonic loading

on the fault. The two short edges of the model are kept as free surfaces in all calculations. PyLith solves a Lagrangian description of the conservation of momentum using the Galerkin's method to formulate equations of the finite element model, and use PETSc solvers to obtain solutions to the problem. See Aagaard et al. (2013b) for details on the specific equations and computational methods used in PyLith. All models in this study were meshed using triangular elements whose sizes were 0.08 m where local fault shear stress concentration appeared, and mesh sizes gradually increased away from the fault plane. We confirmed that the chosen mesh sizes were small enough for the solutions to converge. Uniform time step sizes were chosen between 0.04-1.0 years and ensured to be smaller than one-fifth of the relaxation time arising from the shear modulus and viscosity values. This was confirmed to assure convergence of the solution.

2.2 Viscoelastic Host Rock Model Setup

In the first model, the entire model is just a simple block of viscoelastic medium of 200 m length and 30 m width (Figure 1a). This represents a case where the host rock behaves as a viscoelastic medium and the fault zone is simply the bottom boundary of the model with zero thickness. The viscoelastic constitutive model used here is a Maxwell model, consisting of a linear spring and dashpot connected in series, and the time-dependence is incorporated only in the shear component of the deformation. Thus, volumetric deformation is all due to elastic deformation but not affected by viscous deformation. The material constants were chosen to reflect typical values considered in seismology and geodynamics (Table 1). In order to introduce a fault shear stress anomaly, 45 MPa of initial shear traction was applied to a 10-m long section in the middle of the fault plane with the displacement fixed at the far-field boundary. This produces an approximately elliptical (fault-parallel) displacement distribution along the fault plane (Eshelby, 1957). After the application of this initial shear traction, all displacement along the fault plane and the far-field boundary are fixed to observe how the stress perturbation introduced in the model evolves over time due to the Maxwell rheology. We may consider this model to be analogous to a stress perturbation caused by a constant-stress drop fault slip and its successive evolution due to

viscoelastic deformation of the host rock. We refer to this model as the viscoelastic host rock model.

2.3 Viscoelastic Fault Zone Model Setup

The second model considers the fault zone to be of finite thickness. A 2-m thick sliver along the long edge of the model represents the fault zone and this fault zone is assigned some viscoelastic constitutive properties. The rest of the model is elastic, thus the model represents a viscoelastic fault zone embedded within an elastic host rock (Figure 1b). In this model, the displacement is fixed at the fault boundary throughout the calculation while the far-field boundary is displaced parallel to the fault zone to cause shear stress and deformation in the model. In order to introduce a shear stress perturbation along the fault zone, a geometrical irregularity is introduced along the interface between the fault zone and the host rock as shown in Figure 1b. This irregularity is a circular projection on the elastic host rock wall intruding into the viscoelastic fault zone, which we refer to as the “asperity”. The asperity height is 1 m and the radius of curvature (ρ) is 2 m, which results in an asperity width in the fault-parallel direction (w) of about 3.5 m. In addition, we also assign a contrast in the instantaneous elastic properties of the fault zone and the host rock such that the fault zone is more compliant compared to the host rock. Due to the geometrical irregularity and contrast in elastic properties, a stress concentration develops around the asperity when a uniform displacement along the far-field boundary is applied.

The correspondence of this model to actual geological structures found in nature can be considered in different ways. For instance, the fault zone in our model can represent the highly sheared fault gouge/breccia that constitutes the fault core region embedded in the host rock. Alternatively, we may also view the fault zone to represent larger scale structures such as the fractured damage zone around a fault caused by dynamic rupture propagation (Dunham et al., 2011a, 2011b; Johri et al., 2014). In the former case, the elastic/viscoelastic boundary is the boundary between the fault core and the damage zone, whereas in the latter case it is the boundary between the damaged and intact host rock. We refer to this model as the viscoelastic fault zone model.

The material constants used in the viscoelastic fault zone model are also shown in Table 1. The contrast in the short-term properties of the fault zone and host rock is set to be about 16 %. This was chosen to be consistent with studies of fault zone trapped waves which estimated that the velocity contrast between the fault zone and host rock can be as high as 40 % (Li et al., 1990, 1994, 2004). For this model setup, we also varied the viscosity of the linear dashpot or assigned a standard linear solid rheology to the fault zone to explore how results may change with varying viscoelastic properties. A standard linear solid represents a “long-term solid” behavior as opposed to the “long-term fluid” behavior of the Maxwell model. Viscosity values were chosen to be close to values recovered from geodetic and geophysical studies (Hearn et al., 2004; Hu et al., 2004; Muto et al., 2013) or slightly lower due to our proposition that damaged fault rocks will appear less viscous.

3. Impulse Responses

We first consider in both models a situation where some initial shear stress perturbation is applied at $t = 0$ years. Then the displacements at the two boundaries, fault plane and far-field, are fixed throughout the rest of the calculation while the elements inside the model are allowed to deform based on the specified constitutive properties. This allows us to first explore the basic time-dependent impulse responses of the system upon application of a sudden fault shear stress perturbation.

In the case of the viscoelastic host rock model, first, the node displacements along the fault plane caused by a 45 MPa shear traction in the central segment (see Figure 1) is calculated while the far-field boundary is fixed. This fault plane node displacement is then treated as the initial displacement boundary condition along the fault plane which introduces the shear stress perturbation in the model. For the viscoelastic fault zone model, 3 cm of fault-parallel displacement is applied at the far-field boundary to generate shear deformation in the entire model and also cause a stress concentration around the geometrical asperity. These examples would be relevant, for instance, in considering how a sudden stress change caused by some fault slip may evolve over time.

3.1 Impulse Response in Viscoelastic Host Rock Model

Figure 2 shows several snapshots of the shear stress (σ_{xy}) distribution at different times obtained from the viscoelastic host rock model. There is a region of high shear stress around the stress patch at the center of the fault which decays in amplitude over time. Figure 3 shows the shear traction profiles along the fault plane drawn every 10 years after the application of the initial traction (Figure 3a). At $t = 0$ years, there is 45 MPa of shear traction anomaly at the central region of the fault which decays in magnitude over time. It appears that the shape of the shear traction profile does not change over time, and thus the fault-parallel width of the region with enhanced shear stress stays the same. In fact, when all shear traction profiles are normalized against the maximum values, all profiles from different time steps collapse into an identical normalized traction profile (Figure 3b). We also confirmed that changing the patch size of the initial shear traction (1 – 30 m) does not affect this observation and results in the same conclusion that all profiles collapse into an identical normalized profile. This indicates that the time-dependence due to the viscoelastic rheology is affecting the stress magnitudes of the solution over time, but not the spatial distribution.

These observations are in fact a manifestation of the elastic-viscoelastic correspondence principle in linear viscoelastic theory (Findley et al., 1976; Lakes, 2009). According to the correspondence principle, viscoelastic problems can be solved as an equivalent elastic problem in the Laplace domain, and then the time-dependence from viscoelasticity can be brought back through inverse Laplace transform without distorting the spatial aspects of the equivalent elastic solution; however, with the restriction that *the boundary conditions remain constant over time and the time-dependence of the material is spatially homogeneous* (Paulino and Jin, 2001). In the case of our viscoelastic host rock model, the displacements at the fault and far-field boundaries are fixed and the other two boundaries are free surfaces throughout the problem. Also the viscoelastic constitutive properties are uniform throughout the model. Therefore stresses at all points in the model obey the same time-dependence and the resulting stress distributions at different times were similar to

each other (Figure 3b). In summary, our viscoelastic host rock model with minimal geometrical features shows that shear stress simply decays over time in the model preserving the spatial features due to viscoelastic stress relaxation.

3.2 Impulse Response in Viscoelastic Fault Zone Model

The shear stress (σ_{xy}) distribution resulting from the viscoelastic fault zone model is shown in Figure 4 for several time frames after the application of the far-field displacement. As intended from the model configuration, a stress concentration is developed around the asperity where the shear stress is higher relative to the surrounding region. As time elapses, the overall shear stress level is diminished due to the viscous shearing that takes place in the fault zone. When the stress distribution inside the fault zone around the asperity is examined in detail (Figure 4, $t = 30$ years), we see that the shear stress is generally enhanced around the asperity but the distribution is not necessarily uniform in the fault-normal direction. There exists a chain of high shear stress region that emanates diagonally from the asperity towards the plane of symmetry. While the detailed understanding of such structure and its consequence on earthquake rupture can be important to discuss, from here on, we only focus on the along-fault variation and take the fault-normal average between $y = 0 - 1$ m in order to obtain a representative shear stress profile within the fault zone.

Figure 5a shows the shear stress profiles within the fault zone plotted every 10 years. As expected from Figure 4, the overall magnitude of the shear stress decreases over time due to viscoelastic relaxation. When we focus on the shear stress anomaly around the asperity, we notice that the magnitude and shape of the anomaly changes significantly over time. To examine this in detail, we subtract the background shear stress level, defined by the average shear stress between $x = 30 - 40$ m (Figure 5a), from the original shear stress profiles to obtain a *relative* shear stress profile as shown in Figure 5b. The relative shear stress profiles show that the evolution of the shear stress anomaly occurs roughly in two stages.

As a precursory stage, a rapid increase in the shear stress anomaly is first observed. This is caused by the fact that the background shear stress decays faster than the shear stress

concentration below the asperity during the first decade. (see appendix for explanations about this precursory stage) Once the shear stress anomaly builds up, the peak relative shear stress at $x = 0$ m starts to decay over time. At the same time we also observe that the spatial extent of the region with enhanced relative shear stress also increases. For instance at around $x = 5$ m, the relative shear stress is nearly 0 MPa at $t = 0$ years but rises to about 2 MPa after 40 years (Figure 5b). Therefore the stress anomaly not only decreases in magnitude due to viscous relaxation, but also appears to diffuse spatially over time. To describe this observation quantitatively, we extracted the spatial dimension of the shear stress anomaly (e.g. shear stress patch size/width) by reading off the along-fault width of the region with relative shear stress exceeding some threshold value, τ_{thres} . Figure 6 shows how the shear stress patch size changes as a function of time according to several different threshold values. In some cases with higher threshold values, the patch sizes vanish as the entire relative stress profile becomes smaller than the threshold. But it is clear that patch sizes generally increases over time for any choice of threshold values. In all cases the patch size at least doubles within the first decade and quadruples by 50 years.

This apparent growth in stress patch size is a markedly different behavior from what was observed in the previous model with a viscoelastic host rock. In the previous model, the spatial extent of the shear stress anomaly did not change over time. The difference can be understood as the inapplicability of the correspondence principle to the viscoelastic fault zone model. In the viscoelastic fault zone model, the interface geometry between the fault zone and the host rock is constantly deforming over time, which violates the time-independent boundary condition necessary to apply the correspondence principle. Alternatively, we can also explain the inapplicability by viewing the model as a composite of two domains with different time-dependent properties, which also does not permit the application of the correspondence principle. If the correspondence principle does not hold, the spatial stress distribution within the viscoelastic fault zone will not simply scale with time as in the viscoelastic host rock results, and the spatial feature of the mechanical solution will evolve over time.

4. Steady Loading on Viscoelastic Fault Zone

4.1 Fault Zone with Maxwell Rheology

We now consider a case where the load is applied steadily over time in the viscoelastic fault zone model, analogous to interseismic loading on a fault due to plate motion. The far-field displacement is applied at a constant rate of 0.3 cm/year in the positive x -direction starting at $t = 0$ year. Figure 7a shows the fault shear stress profile drawn every 10 years according to the same method as the previous section (i.e., average within 1 m of the symmetry plane). Because load is constantly added to the system, the overall shear stress continues to increase with time. But after about 60 years, the shear stress magnitude changes only a small amount and settles to a seemingly steady state level below 50 MPa.

To examine the shear stress anomaly around the asperity, we obtain the relative shear stress profile by subtracting the background shear stress level (Figure 7b). The peak value of relative shear stress increases in magnitude as expected from the continuous loading, but also approaches a steady-state value at around 35 MPa. The widening of the spatial dimension of the shear stress anomaly is also more evident compared to the impulse response shown in the previous section. Figure 7c shows the change in width of the shear stress patch extracted with a relative shear stress threshold of 3 MPa. After 10 years, the stress patch size is about 3.5 m, comparable to the geometrical width of the host rock asperity, but reaches about 18 m after 100 years. Interestingly, when the magnitude of the peak relative shear stress below the asperity is plotted against time (Figure 7c), we see that it stabilizes after about 60-70 years although the width continues to grow throughout the calculation. This suggests that the apparent spatial growth of the stress patch is not simply caused by the increase in magnitude of the shear stress anomaly.

4.2 Comparison with Elastic and Standard Linear Solid (SLS) Rheology

To demonstrate that this growth in patch size is indeed due to the viscoelastic behavior, we also modeled the same problem with different rheological properties assigned to the fault zone. First, an elastic fault zone with the same instantaneous elastic properties as the

viscoelastic fault zone is considered. The resulting absolute and relative shear stress profiles from the elastic fault zone are shown in Figure 8a, b. Because the entire model is now linear elastic, it is obvious that the magnitude of the solution scales linearly with time. Thus the stress profile increases overall at equal vertical spacing (Figure 8a) and the stress perturbation around the asperity also simply increases in magnitude with time (Figure 8b). If we extract the shear stress patch size with a 3 MPa threshold, we observe a slight increase in patch size but the increase is limited to about 5 m and does not grow significantly beyond the geometrical width of the asperity (Figure 9). Therefore it is clear that without viscous deformation, growth in shear stress patch size is suppressed. In contrast to the Maxwell fault zone model, relative shear stress under the asperity continues to increase but the width stays relatively constant.

We further tested a case with a standard linear solid (SLS) rheology assigned to the fault zone. This is to test a rheology with a “long-term solid” behavior as opposed to the “long-term fluid” behavior possessed by the Maxwell model. Also as described below, we can vary properties of the SLS model to study a spectrum of behavior spanning between completely elastic and “long-term fluid” behavior. Similar to the Maxwell rheology we used above, the time-dependence of the SLS model is incorporated only in the shear component of the deformation in the model.

A schematic of the SLS model is shown in Figure 10. The sum of the two spring stiffness, $G_1 + G_2$, governs the instantaneous elastic property which is chosen to match the Maxwell and elastic fault zone models used above. The dashpot viscosity is also set as 10^{18} Pa s to match the Maxwell model above. In the SLS model, the ratio of the first shear modulus to the total shear modulus, $R = G_1 / (G_1 + G_2)$ or the *shear modulus ratio*, governs how much of the instantaneous shear modulus remains in the long-term after the dashpot has fully relaxed. If $R = 0$, the model possesses no (shear) loading capacity in the long-term, thus equivalent to a Maxwell rheology. If $R = 1$, the first spring carries the entire shear load, thus equivalent to an elastic medium. Thus, Maxwell and elastic behaviors are extreme cases of an SLS behavior and R is a convenient parameter to vary in order to study the

influence of varying degree of viscous behavior. Here we show two results from SLS fault zone models with shear modulus ratio of $R = 0.5$ and 0.1 together with the previous results of Maxwell ($R = 0$) and elastic ($R = 1$) fault zone properties.

Figure 8 compares the absolute and relative shear stress profiles of the SLS fault zone models for shear modulus ratio values between 0 and 1. Because both $R = 0.5$ and $R = 0.1$ yields long-term solid fault zones, the overall shear stress level increases at nearly constant spacing with time for these cases (Figure 8c, e). However with decreasing shear modulus ratio, the shear stress level attained in the model decreases, and spatial diffusion of the shear stress anomaly at the asperity become more evident.

The evolution of shear stress patch size is shown in Figure 9 for the two SLS cases compared with the extreme cases. It is observed that decrease in R from 1 (elastic) to 0 (Maxwell) increases the stress patch growth rate. Thus increase in dominance of the viscous effect promotes stress patch growth. In general, we also find that a shear stress ratio of $R = 0.5$ yields results similar to an elastic fault zone and it requires a shear stress ratio as low as $R = 0.1$ to achieve about half of the stress patch growth rate as seen in the Maxwell fault zone model (Figure 9).

5. Effect of Viscosity and Asperity Size

We further investigate how the growth of shear stress patch size is influenced by other parameters in the model. First, the viscosity (η) of the dashpot element in the Maxwell model is varied. Then the width of the asperity (w) is varied by changing the size of the circular projection.

5.1 Effect of Dashpot Viscosity

The dashpot viscosity was fixed at 10^{18} Pa s in the previous sections, but here we perform calculations using viscosities ranging from 10^{17} to 10^{19} Pa s. There are two major consequences of changing viscosity. First, the overall shear stress level is heavily influenced by the viscosity because the flow shear strength in Maxwell rheology is

proportional to the viscosity when the fault zone thickness is fixed (2 m) and the far-field displacement rate is constant (0.3 cm/year) as shown in Figure 11. For viscosities higher than 10^{19} Pa s, the stress profile will likely converge into the elastic solution, which is also shown in Figure 11 for reference. Secondly, the viscosity variation changes the characteristic relaxation time ($\tau = \eta/G$) in Maxwell rheology, which influences the time scale required for the viscous deformation to take effect. Given findings from the previous section, it is the viscous deformation that causes diffusion of the stress anomaly, thus the change in viscosity is expected to control the spatial diffusion of the shear stress.

To compare the spatial diffusion effects from stress profiles with vastly different magnitudes in Figure 11, we normalized the shear stress profiles by the background shear stress level to homogenize the magnitude differences. We refer to these profiles as the *normalized* shear stress profiles. Note that from here on, the background shear stress level is defined by the average between $x = 50 - 60$ m. Figure 12a displays normalized profiles drawn every 50 years for each dashpot viscosity, each group of normalized profiles offset by 1. We recognize that the profiles are wider for lower viscosity and narrower for higher viscosity. At higher viscosities, the normalized profiles grow vertically in magnitude with time (lighter to darker lines). This is analogous to the elastic solutions (Figure 7d) and can be understood by recognizing that high viscosities effectively produce elastic behavior since only negligible time-dependence is permitted for a given amount of time. At lower viscosities, the normalized profiles grow horizontally but not vertically, because the characteristic relaxation time is on the order of months and thus equilibrium at the asperity is reached quickly at the beginning of the model. Thus lower viscosity and shorter relaxation time are allowing more spatial diffusion of shear stress to occur under the given loading condition.

Figure 13a shows the evolution of the shear stress patch size defined by a threshold level of 0.3 MPa. Note that this threshold is lowered by an order of magnitude compared to previous sections because the peak stress anomaly in the relative stress profile for the lowest viscosity (10^{17} Pa s) was only about 3 MPa. The patch size continues to increase

monotonically at a decelerating rate in all cases, but the contrast caused by the difference in viscosity is clear. After 300 years, patch size only increases to about 15 m for the highest viscosity, 10^{19} Pa s, but reaches 65 m for the lowest viscosity, 10^{17} Pa s. Another way of comparison is to say that it takes only a decade for the patch size to grow to 20 m for viscosity of 10^{17} Pa s, but nearly 300 years if the viscosity is 5×10^{18} Pa s. Thus, viscosity is influencing the longevity of a shear stress patch of a certain size.

5.2 Effect of Asperity Size

Next we explore the effect of the geometrical asperity size. We achieved this by performing the same set of calculations as the previous section using finite element meshes with different asperity geometries. Specifically, the asperity geometry was varied by changing the radius of curvature (ρ) from 2 m to 20 m or 200 m, but keeping constant the fault zone thickness (2m) and asperity height ($h = 1$ m). As a result, the width of the geometrical projection (w) varied from about $w = [3.5 \ 12.5 \ 40]$ m when $\rho = [2 \ 20 \ 200]$ m, respectively.

Figure 12b and 12c shows the normalized shear stress profiles obtained with wider asperity sizes $w = 12.5$ m and 40 m, respectively. As an obvious consequence of widening the asperity, the shear stress anomaly becomes wider in the fault parallel direction and less sharp as w increases. We find that the main time-dependent features found from the previous section also apply to the wider asperities. First the growth in magnitude of the stress anomaly over time occurs more in the higher viscosity results. On the other hand, the lateral widening of the stress anomaly is more pronounced for lower viscosity results.

Figure 13b and 13c shows the stress patch size data against time for $w = 12.5$ m and 40 m (or $\rho = 20$ m and 200 m), respectively. In both cases, a rapid increase in patch size is observed at the beginning and then a more steady growth for the rest of the calculation for all viscosities. Since the initial rapid increase in stress patch size is comparable to the asperity width, we understand that it mostly reflects the instantaneous elastic stress perturbation caused by the asperity. The following steady increase in patch size is rather

similar between results from different w values. The increases in patch size from 10 years to 300 years for $w = 3.5$ m and 40 m are about 10-50 m and 11-35 m, respectively, depending on the viscosity. Therefore the spread between the lowest and highest viscosity results is slightly decreased for the larger asperity but the growth rates are basically similar for different asperity size.

6. Discussion

6.1 Effect of Viscosity on the Minimum Stress Patch Dimensions

The results demonstrate that a viscoelastic fault zone embedded in an elastic host rock can produce complicated time-dependent mechanical responses. Viscoelastic responses not only relax the magnitude of shear stress generated in the model, but also cause spatial diffusion of stress concentrations when combined with internal geometrical irregularities. We demonstrated by varying the contribution of the viscous component, either by controlling the shear modulus ratio in the Standard Linear Solid model or by varying the viscosity in the Maxwell model, that the spatial diffusion effect is indeed caused by the contribution from the viscous deformation. It appears that the additional dimension in the fault normal direction given to the fault zone, in contrast to zero-thickness frictional interfaces, is a key factor in allowing viscous flow and spatial diffusion of stress.

Viscosity affects the degree to which spatial diffusion of shear stress occurs after a given time because it controls the characteristic time scale of the viscoelastic model. In this regard, changing the length scale of the asperity did not affect the spatial diffusion behavior as seen by the minor differences in the growth rate of stress patch widths in Figure 13. So the rate at which a stress patch grows is dependent on viscosity but not on asperity size. We argue that the consequence of such viscosity-dependence is that small asperities become mechanically invisible over time when the viscosity is low.

Figure 14 shows the change in stress patch size normalized by the width of the geometrical asperities (*normalized patch width*). Because the widening of stress patches is

relatively insignificant compared to the dimension of the asperity for larger asperities, normalized patch widths stay between 1-2 for all viscosities when $w = 40\text{m}$. However as the asperity sizes become smaller, the variation in the growth trend with changing viscosity becomes more significant. After 300 years, the normalized patch width varies between 4.5 and 18.8 depending on the viscosity when $w = 3.5\text{m}$.

This means that when the fault zone viscosity is relatively low, small stress patches cannot retain its dimension due to rapid spatial diffusion. However when the fault zone viscosity is higher, the sizes of stress patches will remain close to the geometrical asperity size, preserving the geometrical heterogeneity for a longer time. Comparing the normalized stress profiles for the highest viscosity in Figure 12 (top row), we see that the final profiles after 300 years appear substantially different from each other because the stress concentration due to the asperity is not unfocused due to diffusion. However for the lowest viscosities (Figure 12 bottom row), the final profiles look similar to each other, especially towards the base of the stress anomaly, despite the significant differences in the dimension of the asperities causing the stress anomalies. Therefore fault zone viscosities may control the minimum dimension of stress patches that will remain mechanically visible during interseismic loading.

6.2 Implication for Variations in Fault Stress Heterogeneity

If small stress patches tend to disappear in low viscosity fault zones, then we can alternatively suggest that fault zone viscosity controls the minimum dimension of the stress patches that can exist after some interseismic loading. We speculate that variations in fault zone viscosity exist in nature, either dependent on depth or due to variation in the mineral compositions and porosity of fault-related rocks. For instance, as the magnitude of the in-situ stresses increase with depth, rocks should become more compact (less porosity) and exhibit less compaction-assisted ductile deformation (Dudley et al., 1998; Hagin and Zoback, 2004; Chang and Zoback, 2009; Sone and Zoback, 2013), thus higher apparent viscosity. Such a trend would lead to the presence of a smooth fault shear stress distribution

at shallow depth which gradually increases its spatial heterogeneity at greater depth due to the small stress patches that persist due to the increasing viscosity.

Although no quantitative evaluation can be provided at the moment, this may be qualitatively consistent with seismological observations from mega earthquakes. Sources of high-frequency radiation tend to project back to the down-dip limit of the seismogenic zone (Lay et al., 2012) and it has also been shown that there is a gradual deepening of radiation sources from the trench to seismogenic depths as the radiation frequency content increases. Small earthquakes along the 2011 Tohoku-oki earthquake rupture plane are also found to exhibit a depth dependency of high-frequency radiation intensity inferred from corner frequency estimations (Uchide et al., 2014). We also acknowledge, though, that a lower limit will exist in the applicability of the above notion, since high temperature at depth will start to promote ductile deformation of the fault zone rocks by crystal plastic deformation (i.e. mylonites), thereby reducing the viscosity. However, temperature effects on compaction-assisted creep mechanisms are not quantified as well as crystal plastic mechanisms (e.g. dislocation creep, diffusion creep) and it is still difficult to evaluate when temperature-enhanced ductility dominates the compaction hardening effect.

Along-strike variation in the frequency characteristics of radiated seismic waves are also reported from the 1999 Chi-Chi earthquake (Ma et al., 2001). Some authors have argued that hydrodynamic lubrication occurred (Brodsky and Kanamori, 2001) during the rupture of the northern part of the Chelungpu fault, which may have suppressed the radiation of high-frequency waves, but our results may provide additional insights into how we might explain the difference in seismic radiation emitted from different portions of the fault. Differences in frictional properties of fault rocks between the northern and the southern portion of the Chelungpu fault is reported (Tanikawa and Shimamoto, 2009) such that the fault may be velocity-strengthening in the north but velocity weakening in the south. If such contrast in mechanical properties causes differences in the stress distribution pattern, characteristics of seismic radiation may also obey such contrast between the northern and southern portion of the fault. For instance, if velocity-strengthening materials are more

ductile compared to velocity-weakening materials, this may be consistent with the depleted high-frequency radiation from the northern portion of the fault during the Chi-Chi earthquake.

6.3 Magnitude Differences between Profiles of Different Viscosity

In section 5, we recognized large differences in stress magnitudes that arise from the different fault zone viscosities, but left out the discussion to focus on the spatial spread of the stress anomaly. Because stress magnitudes are limited by rock strengths at a given depth, it may be unrealistic to invoke and compare results with such large stress magnitude contrasts. However, we believe that these factors do not affect our conclusions. For example, if a high fault zone viscosity model results in shear stress magnitudes well over the suggested rock strength, it implies that we are running the calculation too long since the fault zone would have slipped before the end of our calculation. If the fault slips before the end of the calculation, there will be less time for interseismic loading and also less time for spatial diffusion of shear stress, further reducing the effect of stress patch growth in higher viscosity fault zones. Thus the general conclusion that higher viscosity limits stress patch growth is preserved even when limits to shear stress magnitudes are imposed. The above arguments also imply that lower viscosity involves longer recurrence times between earthquakes. Then there will be longer time for spatial diffusion of shear stress patches before the next slip event, possibly exaggerating the role of viscosity to widen stress patches for lower viscosity fault zones. It may also be interesting to note that this also implies that the stress accumulation can be much slower than typically expected from the plate convergence rate in shallow portion of megathrust earthquakes, such as the resultant large slip area of the 2011 Tohoku-oki earthquake.

We also note that the stress magnitudes of the resulting profiles are easily influenced by the far-field displacement rate. In fact we confirmed that the stress results from the viscoelastic fault zone model scales linearly with the far-field displacement rate. If the displacement rate is doubled, the stress profile simply doubles in magnitude preserving all spatial features, a consequence of the linearity of the viscoelastic model we employed.

These are all important aspects of the model that requires careful discussion to gain further insights about how viscous fault zone deformation influences the interseismic accumulation of fault shear stress. Because of the sensitivity of our model to the far-field displacement rate, we will also need to address what the appropriate far-field displacement rate is to impose in these models, which was chosen in this study without rigorous justification.

6.4 Uncertainty in Fault Zone Rheology and Geometry

We presented here numerical results that emphasize the effect of viscous deformation on the temporal and spatial evolution of fault shear stresses. The degree of ductility was varied by varying some constitutive parameters to demonstrate the effect of viscous rheology. However, it is still difficult to directly compare the numerical models in this paper quantitatively with natural fault zone process, since parameters and formulations in the model will likely vary beyond those considered here in natural settings.

First, the linear viscoelastic constitutive relations we used were chosen for convenience and may not represent true deformational behavior of fault zone materials. Viscosity was handled as a representative parameter defining the viscous nature of the fault zone, but it is merely a proxy of ductility as there is no guarantee that linear viscoelastic models can adequately describe earth material behavior. For instance, volumetric creep of unconsolidated sediment have been described as either an exponential or power law function of time (Dudley et al., 1998; Hagin and Zoback, 2004), and triaxial creep of organic-rich shales described as a power law function of time (Sone and Zoback, 2014). Compaction-assisted creep deformation of porous and/or fractured material, which we invoke as a major physical mechanisms for viscous fault zone rheology in this study (see Introduction), involve time-dependent volumetric deformation with complicated plastic effects. Sheared materials exhibit anisotropic behavior, thus may require more involved tensorial constitutive formulations rather than scalar flow laws typically used for the asthenosphere with negligible porosity. Much study is needed to constrain realistic constitutive formulations that describe fault zone materials.

The choice of fault zone viscosity values between 10^{17} - 10^{19} Pa s in this study are lower than those typically assumed for the lithosphere although apparent asthenosphere viscosities on the order of 10^{19} Pa s are not uncommon values recovered from inversion of crustal deformation due to afterslips and postseismic rebound (e.g. Hearn et al., 2004; Hu et al., 2004; Suito and Freymueller, 2009; Ohzono et al., 2012). Laboratory creep experiments using low porosity ($< 10\%$) shales (Sone and Zoback, 2014) also imply apparent viscosity values in the range of $10^{18} - 10^{20}$ Pa s for time scales of 1 to 100 years. As argued in section 1, if fault zone materials undergo brittle deformation once every seismic cycle, it is plausible that fault rocks exhibit even lower viscosities than the intact shales samples mentioned above. Mechanical properties of fault rocks are almost exclusively argued in terms of their frictional properties and their bulk deformational characteristics are not understood. It may also be important to study the bulk time-dependent deformational behavior of fault rocks affected by brittle damage.

Finally, geometrical parameters such as fault zone thickness and asperity height are certainly not constant along fault zones. Thus a thorough parameter study needs to be conducted to better understand the relevance of the stress diffusion mechanism at a wider range of spatial scales. Potential similarities and differences with other asperity geometries (e.g. host-rock fragments within viscous matrix of fault zone material, continuous distribution of multiple asperities, asymmetric structure, 3-D model) would also be important to demonstrate in future studies to understand the general applicability of our results to more complicated geometries observed in nature.

6.5 Comparison with other interseismic models

It is important to note that numerous past studies have already addressed postseismic and interseismic fault mechanical behavior that may occur due to time-dependent rheology. These studies are motivated from the transient crustal deformation observed after earthquakes and have revealed the importance of considering off-fault viscoelastic rheology (Wang, 1995; Freed and Bürgmann, 2004; Hearn et al., 2009), fault afterslip behavior (Marone et al., 1991; Miyazaki et al., 2004; Montés, 2004; Johnson et al., 2006), and

624 poroelastic deformation (Jónsson et al., 2003; Fialko, 2004) in explaining these geodetic
625 observations.

626 Hetland et al. (2010) investigated aseismic fault slip behaviors along a strike slip fault
627 that occur in response to steady far-field loading and coseismic fault slips imposed
628 periodically in the asperity region of the model. Note that asperity in their study is defined
629 by the fault region which slips co-seismically but otherwise locked, different from our
630 geometrical definition. Hetland et al. (2010) made a comprehensive comparison between
631 different fault zone rheologies including linear/non-linear viscous rheologies, rate-
632 dependent friction, and rate-state friction, in which all results show some extent of spatial
633 growth of the region of afterslip. Since afterslip in their models occurs as a result of shear
634 stress concentration that develops at the edge of the asperity, their model also demonstrates
635 how the spatial extent of enhanced shear traction grows over time due to time-dependent
636 fault zone rheology (i.e. analogous to our growing shear stress patch).

637 The difference in our viscoelastic fault zone model is that we generate spatial stress
638 heterogeneity interseismically by introducing a geometrical heterogeneity rather than by
639 imposing a discontinuity in coseismic fault slip. Thus we address the behavior of fault
640 stress heterogeneities that would naturally occur solely as a result of the fault architecture,
641 without any external perturbation except for the far-field tectonic movement. Clearly,
642 studies considering the response to coseismic slips are more complete in that they discuss
643 the interaction between different fault domains during the entire seismic cycle. However,
644 these models only capture the largest fault slip that defines the seismic cycle and do not
645 capture any smaller scale mechanical heterogeneities that may exist in nature. Such
646 heterogeneity is evident from natural earthquakes with a wide range of magnitudes
647 occurring at all scales. When only the largest events are considered in the model, stress
648 concentrations only occur at the edge of the coseismic slip. Therefore these class of models
649 may not be suitable to compare and discuss stress diffusion phenomenon that may occur at
650 various length scales.

Our model, although rather simplistic and only for the interseismic period, attempts to address the difference in spatiotemporal behavior of fault shear stress anomalies at multiple length scales, thus provides additional insights into how fault zone rheology may influence the fault stress distribution, and consequently the characteristics of earthquakes rupture, occurrence, and statistics along a fault.

Conclusion

We have demonstrated through numerical analyses that shear stress acting on faults could evolve both temporally and spatially if viscoelastic rheology is assumed in fault zones of finite thickness. Shear stress anomaly produced by geometrical irregularities of the viscoelastic fault zone will tend to diffuse spatially and grow in spatial dimension during steady tectonic loading. This is a marked difference from pure elastic behavior where stress anomalies would simply increase in magnitude, but not in spatial dimension. We also showed that varying the viscoelastic properties of the fault zone, by changing the viscosity in the Maxwell model or the shear modulus ratio in the SLS model, influences the degree to which spatial diffusion of shear stress occurs. In general, increasing the contribution of viscous deformation enhances the spatial diffusion of shear stress. On the other hand, the size of the geometrical irregularity does not influence the spatial rate of stress diffusion. Therefore we conclude that the apparent viscosity of fault zones may determine the minimum dimension of shear stress patches that could exist along faults. If apparent viscosity varies along a fault, patterns of shear stress distribution will also vary accordingly influencing the rupture characteristics during earthquakes along the fault.

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Appendix: Precursory Growth in Peak Relative Shear Stress

In the impulse response of the viscoelastic fault zone model (section 3.2), the stress anomaly in the relative shear stress profile increases at the beginning (Figure 5b) because, initially, the peak stress decays slower compared to the surrounding region. After the far-field displacement is applied at the beginning, the shear stress below the asperity (figure 4 point A) is higher compared to the surroundings (figure 4 point B) by about 8 % in the current case (Figure 5a), which results in a higher fault zone shear strain rate below the asperity. However because the fault zone thickness below the asperity is only half compared to the surroundings, the fault-parallel velocity of the host rock, product of the strain rate and the thickness of the fault zone, at the asperity (point A) will be lower than

the rest of the host rock (point B) by about $108 \% \times 0.5 = 54 \%$, which results in the asperity being dragged behind compared to the surroundings (Figure A1). This results in further development of the shear stress concentration until enough shear stress difference is built up for the asperity to accelerate its fault parallel movement and move together with the surrounding host rock. This happens at around $t = 14$ years for the current case (Figure A1) when the velocity of the asperity overtakes the host rock velocity, at which point excess shear stress will start decaying together with the background level. Note that the exact timing of when the asperity velocity catches up with the background (or whether this precursory phenomenon exist or not) will depend on the contrast of elastic properties between the fault zone and host rock, as well as the fault zone thickness ratio between the asperity (A in Figure 4) and background (B in Figure 4) regions.

Table 1

	Viscoelastic Host Rock Model	Viscoelastic Fault Zone Model		Standard Linear Solid Fault Zone Model	
		Host Rock	Fault Zone	Host Rock	Fault Zone
Density [kg/m^3]	2500	2500	2250	2500	2250
P-wave Velocity [m/s]	6000	6000	5000	6000	5000
S-wave Velocity [m/s]	3000	3000	2500	3000	2500
Dashpot Viscosity [Pa s]	10^{20}	-	10^{17} - 10^{19}	-	10^{18}
Shear Modulus Ratio [-]	-	-	-	-	[0.1 0.5 1]

Figure/Table Captions

Table 1: List of material constants used in the finite element model.

Figure 1: Schematic diagrams of the 2-D finite element model used in this study. (a) Viscoelastic Host Rock Model (b) Viscoelastic Fault Zone Model.

Figure 2: Shear stress (σ_{xy}) distribution results from the viscoelastic host rock model with instantaneous initial loading. Snapshots from $t = 0, 50, 100$ years.

893 Figure 3: (a) Shear traction profiles along the fault plane in the viscoelastic host rock model
894 with instantaneous initial loading. Profiles are drawn every 10 years from light to dark
895 colors. (b) Shear traction profiles from (a) normalized with maximum absolute value. All
896 normalized profiles are collapsed into a unique profile.

897 Figure 4: Shear stress distribution results from the viscoelastic fault zone model with
898 instantaneous initial loading. Snapshots from $t = 0, 30, 60$ years. Insert in second panel is a
899 close-up view of the shear stress distribution around the asperity. Points A and B denotes
900 location where fault-parallel velocities are plotted in Figure A1.

901 Figure 5: (a) Shear stress profiles along the fault zone in the viscoelastic fault zone model
902 with instantaneous initial loading. Profiles are drawn every 10 years. (b) Relative shear
903 stress profiles obtained by subtracting background shear stress ($x = 30-40$ m) from stress
904 profiles in (a).

905 Figure 6: Shear stress patch widths plotted against time in the viscoelastic fault zone model
906 with instantaneous initial loading. Stress patch widths were determined using various
907 values for shear stress thresholds.

908 Figure 7: (a) Shear stress profiles along the fault zone in the viscoelastic fault zone model
909 with constant loading of 0.03 cm/year at the far-field boundary. Profiles are drawn every 10
910 years. (b) Relative shear stress profiles obtained by subtracting background shear stress
911 from stress profiles in (a). (c) Shear stress patch width and peak relative shear stress plotted
912 against time. Patch widths determined with a 3 MPa shear stress threshold.

913 Figure 8: Absolute and relative shear stress profiles in the viscoelastic fault zone model
914 with elastic, standard linear solid (SLS) model, and Maxwell model assigned to the fault
915 zone. Profiles are drawn every 10 years. (a, b) Elastic fault zone ($R=1$). (c, d) SLS fault
916 zone with $R=0.5$. (e, f) SLS fault zone with $R=0.1$. (g, h) Maxwell fault zone ($R=0$).

917 Figure 9: Shear stress patch width plotted against time for the relative shear stress profiles
918 in Figure 8. Patch widths determined with a 3 MPa shear stress threshold.

919 Figure 10: Schematic diagram of a standard linear solid model.

920 Figure 11: Absolute shear stress profiles in the viscoelastic fault zone model with dashpot
921 viscosities between 10^{17} - 10^{20} Pa s. Profiles after 100 years of far-field displacement at 0.3
922 cm/year. The elastic solution, corresponding to infinite viscosity, is also shown for
923 reference.

924 Figure 12: Comparison of normalized shear stress profiles in the viscoelastic fault zone
925 model with various dashpot viscosities and asperity width. Profiles are drawn every 50
926 years. Results from different dashpot viscosities are offset by 1 for display purpose. (a)
927 Profiles from asperity width of 3.5 m. (b) Profiles from asperity width of 12.5 m. (c)
928 Profiles from asperity width of 40 m.

929 Figure 13: Shear stress patch widths plotted against time. Each panel compares results from
930 different dashpot viscosities ($1, 2, 5 \times 10^{17}$, $1, 2, 5 \times 10^{18}$, and 1×10^{19} Pa s). (a) $w = 3.5$
931 m (b) $w = 12.5$ m (c) $w = 40$ m.

932 Figure 14: Normalized shear stress patch widths plotted against time. Each panel compares
933 results from different dashpot viscosities ($1, 2, 5 \times 10^{17}$, $1, 2, 5 \times 10^{18}$, and 1×10^{19} Pa s).
934 (a) $w = 3.5$ m (b) $w = 12.5$ m (c) $w = 40$ m.

935 Figure A1: Fault-parallel velocity of 2 points, A and B, on the host rock wall. Point A is
936 located at the asperity. Point B is located 10 m away from the asperity.

Figure 1

Figure 1: Schematic diagrams of the two dimensional finite element model used in this study. (a) Viscoelastic Host Rock Model (b) Viscoelastic Fault Zone Model.

Figure 2

Figure 2: Shear stress (σ_{xy}) distribution results from the viscoelastic host rock model with instantaneous initial loading. Snapshots from $t = 0, 50, 100$ years.

Figure 3

Figure 3: (a) Shear traction profiles along the fault plane in the viscoelastic host rock model with instantaneous initial loading. Profiles are drawn every 10 years from light to dark colors. (b) Shear traction profiles from (a) normalized with maximum absolute value. All normalized profiles are collapsed into a unique profile.

Figure 4

Figure 4: Shear stress distribution results from the viscoelastic fault zone model with instantaneous initial loading. Snapshots from $t = 0, 30, 60$ years. Insert in second panel is a close-up view of the shear stress distribution around the asperity. Points A and B denotes location where fault-parallel velocities are plotted in Figure 5b.

Figure 5

Figure 5: (a) Shear stress profiles along the fault zone in the viscoelastic fault zone model with instantaneous initial loading. Profiles are drawn every 10 years. (b) Relative shear stress profiles obtained by subtracting background shear stress ($x = 30-40$ m) from stress profiles in (a).

Figure 6

Figure 6: Shear stress patch widths plotted against time in the viscoelastic fault zone model with instantaneous initial loading. Stress patch widths were determined using various values for shear stress thresholds.

Figure 7

Figure 7: (a) Shear stress profiles along the fault zone in the viscoelastic fault zone model with constant loading of 0.03 cm/year at the far-field boundary. Profiles are drawn every 10 years. (b) Relative shear stress profiles obtained by subtracting background shear stress from stress profiles in (a). (c) Shear stress patch width and peak relative shear stress plotted against time. Patch widths determined with a 3 MPa shear stress threshold.

Figure 8

Figure 8: Absolute and relative shear stress profiles in the viscoelastic fault zone model with elastic, standard linear solid (SLS) model, and Maxwell model assigned to the fault zone. Profiles are drawn every 10 years. (a, b) Elastic fault zone ($R=1$). (c, d) SLS fault zone with $R=0.5$. (e, f) SLS fault zone with $R=0.1$. (g, h) Maxwell fault zone ($R=0$).

Figure 9

Figure 9: Shear stress patch width plotted against time for the relative shear stress profiles in Figure 8. Patch widths determined with a 3 MPa shear stress threshold.

Figure 10

Standard Linear Solid

Figure 10: Schematic diagram of a standard linear solid model.

Figure 11

Figure 11: Absolute shear stress profiles in the viscoelastic fault zone model with dashpot viscosities between 10^{17} - 10^{20} Pa s. Profiles after 100 years of far-field displacement at 0.3 cm/year. The elastic solution, corresponding to infinite viscosity, is also shown for reference.

Figure 12

Figure 12: Comparison of normalized shear stress profiles in the viscoelastic fault zone model with various dashpot viscosities and asperity width. Profiles are drawn every 50 years. Results from different dashpot viscosities are offset by 1 for display purpose. (a) Profiles from asperity width of 3.5 m. (b) Profiles from asperity width of 12.5 m. (c) Profiles from asperity width of 40 m.

Figure 13

Figure 13: Shear stress patch widths plotted against time. Each panel compares results from different dashpot viscosities (1, 2, 5 $\times 10^{17}$, 1, 2, 5 $\times 10^{18}$, and 1 $\times 10^{19}$ Pa s). (a) $w = 3.5 \text{ m}$ (b) $w = 12.5 \text{ m}$ (c) $w = 40 \text{ m}$.

Figure 14

Figure 14: Normalized shear stress patch widths plotted against time. Each panel compares results from different dashpot viscosities ($1, 2, 5 \times 10^{17}$, $1, 2, 5 \times 10^{18}$, and 1×10^{19} Pa s). (a) $w = 3.5$ m (b) $w = 12.5$ m (c) $w = 40$ m.

Figure A1

Figure A1: Fault-parallel velocity of 2 points, A and B, on the host rock wall. Point A is located at the asperity. Point B is located 10 m away from the asperity.