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Modeling dynamic triggering of tectonic tremor using a brittle-ductile friction model

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[1] We study the physics of dynamically triggered tectonic tremor by applying a brittle-ductile friction model in which we conceptualize the tremor source as a rigid block subject to driving and frictional forces. To simulate dynamic triggering of tremor, we apply a stress perturbation that mimics the surface waves of remote earthquakes. The tectonic and wave perturbation stresses define a phase space that demonstrates that both the timing and amplitude of the dynamic perturbations control the fundamental characteristics of triggered tremor. Tremor can be triggered instantaneously or with a delayed onset if the dynamic perturbation significantly alters the frictional state of the tremor source. **Citation:** Trugman, D. T., E. G. Daub, R. A. Guyer, and P. A. Johnson (2013), Modeling dynamic triggering of tectonic tremor using a brittle-ductile friction model, *Geophys. Res. Lett.*, *40*, doi:10.1002/grl.50981.

1. Introduction

[2] Earthquakes are caused by local stress conditions in the Earth's crust. However, a growing body of evidence indicates that earthquakes may in fact interact at remote distances [e.g., *Velasco et al.*, 2008; *Pollitz et al.*, 2012], with earthquake occurrences at one location influencing seismicity on distant faults. In particular, the transient stress changes due to passing seismic waves can dynamically trigger seismicity [e.g., *Stein*, 1999; *Freed*, 2005]. While numerous studies have focused on triggering of earthquakes by remote seismic waves [e.g., *Gomberg et al.*, 1998; *Hill and Prejean*, 2007], earthquakes occur relatively infrequently, making it difficult to assess the physics of triggering.

[3] Here we focus instead on triggering of tectonic tremor, tiny earthquakes that occur much more frequently in the deep part of the crust and that are particularly sensitive to stress perturbations [*Rubinstein et al.*, 2007; *Thomas et al.*, 2009]. First observed by *Obara* [2002], tremor consists of weak but long-duration seismic activity in the brittle-ductile transition zone beneath active fault zones [e.g., *Rogers and Dragert*, 2003; *Shelly et al.*, 2006]. These tremor episodes are hypothesized to be comprised of a sequence of individual low-frequency earthquake (LFE) events, the result of shear slip on the deep extension of the fault [e.g., *Shelly et al.*, 2007; *Beroza and Ide*, 2011].

[4] Dynamic triggering of tremor has been observed both in strike-slip regimes like Parkfield, California [e.g., *Gomberg et al.*, 2008; *Peng et al.*, 2009; *Ghosh et al.*, 2009], and in subduction zones [e.g., *Miyazawa and Mori*, 2005; *Rubinstein et al.*, 2007; *Gomberg*, 2010]. The properties of triggered tremor are, to date, indistinguishable from nontriggered (i.e., spontaneous) tremor [*Shelly et al.*, 2011], with the triggering perturbation simply modifying the timing of spontaneous tremor episodes. Triggered tremor can occur in two ways: instantaneously triggered tremor (i.e., tremor concurrent with the passage of the remote seismic waves [*Rubinstein et al.*, 2007]) and delayed-onset-triggered tremor (i.e., tremor that occurs well after the stress perturbation but earlier than in the absence of a perturbation [*Shelly et al.*, 2011]). Because of the rich suite of observations of triggered tremor and its sensitivity to small stresses, we use the brittle-ductile friction model of tremor described below to gain insight into the general mechanics of triggered seismicity.

2. Brittle-Ductile Friction Model for Spontaneous Tremor

[5] We employ the brittle-ductile friction (BDF) model of *Daub et al.* [2011] to study the physics of triggered tremor. We model the tremor source region as a rigid block subject to both driving and frictional forces. The block is driven along by a spring that represents tectonic loading of the tremor source region (Figure 1a). Both brittle and ductile friction restricts the block's slip motion. Brittle friction is conceptualized as the sum of discrete frictional contacts that represent individual asperities and fail in a brittle manner. Each brittle contact imparts a shear stress proportional to the block's displacement. When the displacement becomes greater than the characteristic failure distance for a particular contact, that contact breaks, and a new contact is formed at the block's current position. Stronger contacts have longer failure distances, and these failure distances a_i are drawn from a power law distribution $p(a) \sim a^{-2}$, consistent with laboratory experiments of sliding roughness in rocks [*Dieterich and Kilgore*, 1996]. Ductile friction is rate strengthening, also consistent with these experiments.

[6] The dynamics of the BDF model are described by

$$\frac{M}{A}\frac{dV}{dt} = K(V_0 t - x) - \sum_{i=1}^{N_c} \gamma(x - x_{0i}) - \sigma_d \log\left(\frac{V + V_0}{V_0}\right) + F_{perturb}(t); \quad (1)$$

$$\frac{dx}{dt} = V. \quad (2)$$

Equation (1) is Newton's second law divided by fault area A for a block of mass M and position $x(t)$. The four terms on the

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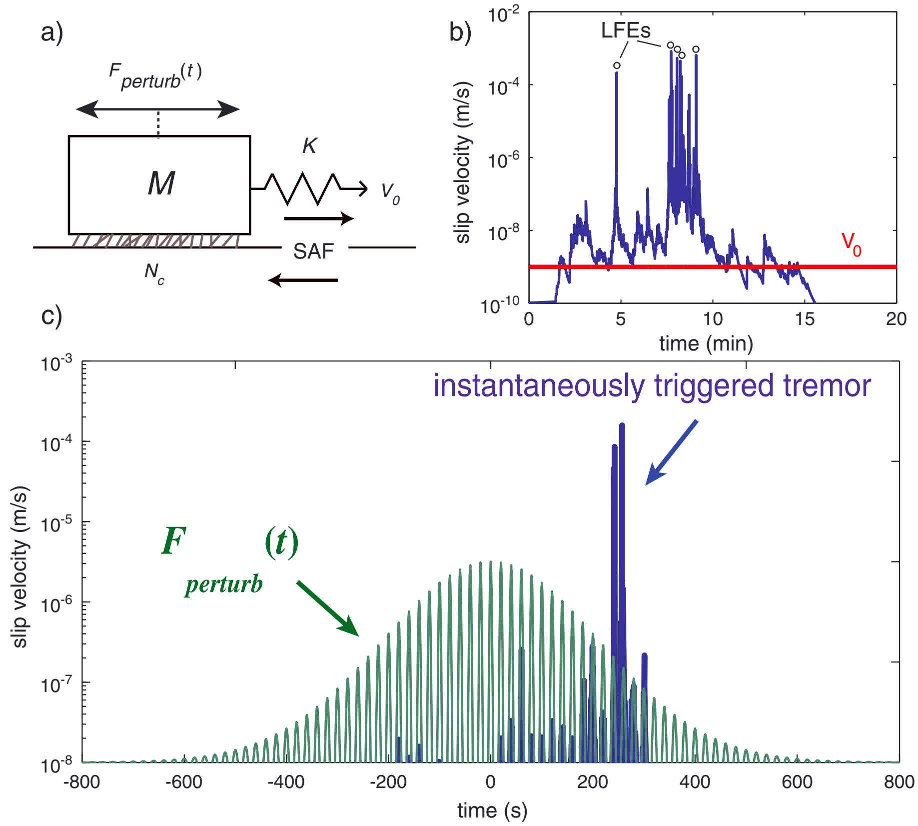

Figure 1. Brittle-ductile friction (BDF) model for triggered tremor. (a) Map view of tremor faulting model for the San Andreas Fault (SAF). Tremor at a single source location is modeled as a rigid block (mass M) pulled across a rough surface (with N_c brittle contacts) at constant velocity V_0 by a spring of stiffness per unit area K. A dynamic shear perturbation, $F_{perturb}(t)$, is applied to disturb the system. (b) Slip velocity as a function of time for a typical tremor episode. Each tremor episode consists of a cascade of distinct failures of the brittle contacts, simulating a sequence of distinct LFEs (denoted by circles). (c) Typical example of instantaneously triggered tremor. A dynamic perturbation $F_{perturb}(t)$ with peak stress amplitude $F_0 = 18$ kPa (maximal at $t = 0$) and 20 s period initiates a tremor episode, which manifests as spikes in slip velocity. Each individual spike represents an LFE. $F_{perturb}(t)$ is an oscillatory Gaussian pulse and is plotted for reference (in nondimensionalized form, with only positive stress amplitudes shown for clarity). A cascade of brittle contact failures is initiated just before $t = 0$, triggering tremor during the latter half of the perturbation.

right-hand side of equation (1) are, respectively, the shear stress exerted by the spring of stiffness per unit area K, the friction due to brittle contacts, the friction due to ductile contacts, and the dynamic perturbation, $F_{perturb}(t)$. Brittle friction is parameterized in terms of brittle stiffness γ and $x - x_{0i}$, the relative displacements of N_c brittle contacts. Ductile friction is determined by the ductile damping strength σ_d and the block's velocity V relative to the tectonic driving velocity V_0. We solve for slip velocity V numerically using a linearly implicit trapezoidal method [e.g., *Press*, 2007] to mitigate the stiffness inherent to equations (1) and (2). Additional model details and parameter choices are described by *Daub et al.* [2011] and the auxiliary material therein.

[7] Each tremor episode consists of a sequence of failures of individual contacts (Figure 1b), analogous to individual LFE events. Tremor in the BDF model is thus characterized by a "cascade" of brittle contact failures, in which the rupture of an initial (usually strong) contact generates a large drop in friction and a corresponding spike in slip velocity V (equation (1)). If the block slips far enough, more contacts will rupture and the block will slip further, rupturing more contacts, eventually causing a self-sustaining cascade of contact ruptures and rapid slip of the block. The ratio of brittle to ductile friction controls the observable characteristics of tremor, such as the recurrence time between tremor episodes [*Daub et al.*, 2011].

3. Modeling Dynamic Triggering of Tremor

[8] To study the effects of dynamic triggering on the BDF tremor model, we augment the equations of motion [*Daub et al.*, 2011] with a forcing function $F_{perturb}(t)$ that simulates the shear stress (resolved along the fault interface) applied by a passing seismic wave. Within the Earth, tremor sources are subject to repeated dynamic perturbations from earthquakes at different locations and times. Within the model, however, we restrict our attention to the application of a single dynamic perturbation: $F_{perturb}(t)$ is representative of the seismic waves radiated from a single earthquake. We model $F_{perturb}(t)$ as an oscillatory Gaussian wave packet with a 20 s period and a 600 s pulse duration, typical of the long-period Love waves known to efficiently trigger tremor beneath the San Andreas Fault [*Rubinstein et al.*, 2007; *Peng et al.*, 2009]. We also fix the level of brittle friction to 67% of total friction, which produces tremor episodes with a characteristic recurrence time ~5 days, consistent with observations of tremor near Parkfield, California [*Shelly*, 2010; *Daub et al.*, 2011].

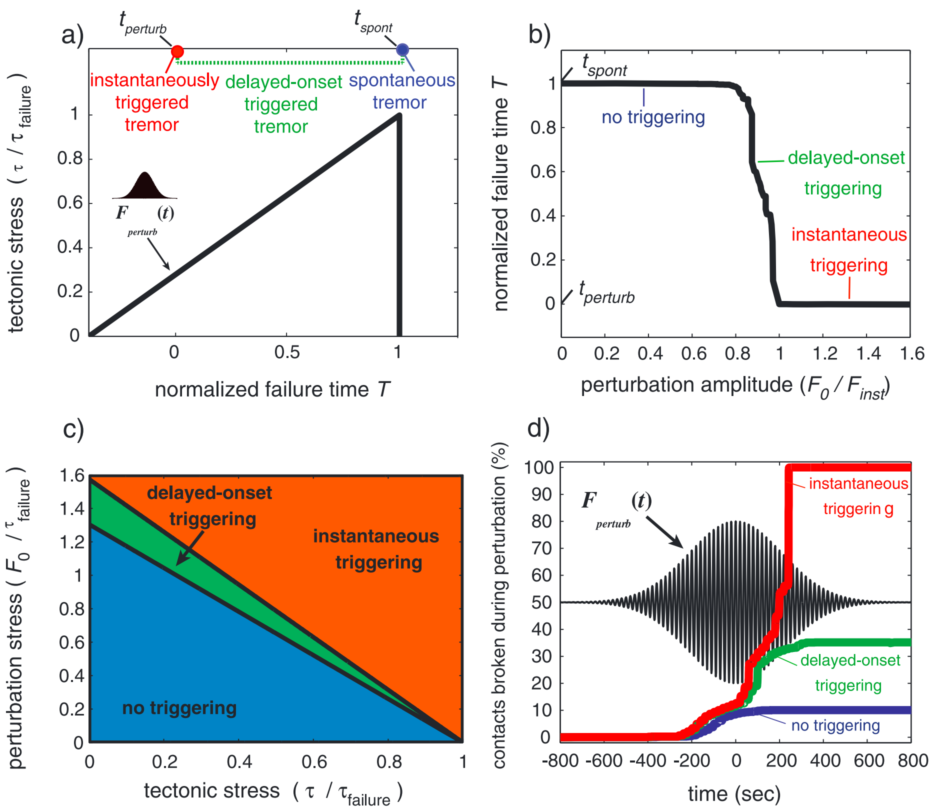

Figure 2. Characteristics of triggered tremor. (a) Tectonic shear stress τ normalized by failure stress $\tau_{failure}$ for a representative tremor episode. In the absence of perturbation, spontaneous tremor occurs at a failure time $t_f = t_{spont}$. For a perturbation applied at $t_{perturb}$, we define the normalized failure time for a tremor episode as $T = (t_f - t_{perturb}) / (t_{spont} - t_{perturb})$. Tremor that occurs with $T = 0$ is instantaneously triggered, while tremor that occurs in the range $0 < T < 1$ is triggered with delayed onset. (b) Normalized timing T of tremor episodes, plotted as a function of normalized perturbation stress amplitude F_0 / F_{inst} (where F_{inst} is the minimum amplitude perturbation required to instantaneously trigger tremor). For low F_0, tremor is not triggered ($T = 1$), while for $F_0 > F_{inst}$, tremor is triggered instantaneously ($T = 0$). Perturbations with F_0 just below F_{inst} trigger tremor with delayed onset ($0 < T < 1$). (c) Phase space for triggered tremor. Tectonic stress τ (normalized by failure stress $\tau_{failure}$) is plotted against perturbation stress F_0 (also normalized by failure stress $\tau_{failure}$). No triggering is observed for low values of τ and F_0, while instantaneous triggering is observed for high values of τ and F_0. Delayed-onset triggering is observed for F_0 just below the instantaneous triggering threshold. (d) Representative example of the cumulative number of distinct brittle contacts ruptured as a function of time during $F_{perturb}(t)$. All the brittle contacts rupture during instantaneously triggered tremor. A smaller fraction (~30%) rupture for tremor that is triggered with delayed onset—enough to substantially modify the frictional state of the source but not enough to induce an immediate tremor episode. Low-amplitude perturbations result in no triggering.

[9] In the absence of any perturbation, a spontaneous tremor episode will occur at a time t_{spont} at which the background tectonic shear stress exerted by the spring $\tau = K(V_0 t - x)$ reaches a failure threshold, $\tau_{failure}$, that depends on the state of the brittle contacts. To explore how the timing and amplitude of an applied perturbation affects the occurrence of tremor, we first simulate a tremor episode without any perturbation to determine the time t_{spont} at which a spontaneous tremor episode occurs. We then repeat the calculation with the same initial set of contacts and apply a perturbation, varying the perturbation time $t_{perturb}$ and amplitude F_0 to examine the effect of amplitude and timing on the triggering process. Note that within the Earth, the nonperturbed failure time is not known, making identification of triggering inherently more difficult.

[10] The timing of the tremor episode following perturbation can exhibit three different outcomes: instantaneous triggering, delayed-onset triggering, and no triggering. If the failure time t_f occurs at $t_f \approx t_{perturb}$, then the tremor episode is instantaneously triggered (Figure 1c). In contrast, if the perturbation has very little effect on the timing of the tremor the tremor episode will not be triggered. The event will still occur when the tectonic stress τ exceeds the failure stress $\tau_{failure}$, and hence, $t_f \approx t_{spont}$ (Figure 2a). Delayed-onset-triggered tremor can occur at any time between $t_{perturb}$ and t_{spont}. Quantitatively, we can characterize these three regimes by examining a normalized failure time T, which is the failure time relative to the perturbation time and normalized by the known time to spontaneous failure, or $T = (t_f - t_{perturb}) / (t_{spont} - t_{perturb})$. If $T = 0$, then the episode is instantaneously triggered, while if $T = 1$, the episode is not triggered. Delayed-onset triggering has occurred if $0 < T < 1$. The plot in Figure 2b illustrates how T varies with perturbation amplitude, revealing that a small change in triggering amplitude can have a large effect on the failure time.

[11] In Figure 2c, we show the phase space that characterizes triggered tremor in the BDF model. When low-amplitude dynamic perturbations are applied to systems with low tectonic stress ($F_0 + \tau \lesssim \tau_{failure}$), no triggering is observed. Conversely, when high-amplitude perturbations are applied to systems with high tectonic stress ($F_0 + \tau \gtrsim \tau_{failure}$), tremor

episodes are instantaneously triggered. Within an instantaneously triggered tremor episode, spikes in block slip velocity (i.e., LFE events) tend to coincide with the times when $F_{perturb}(t)$ is positive (Figure 1c), a robust observation of triggered tremor in the Earth [e.g., *Ghosh et al*., 2009; *Hill*, 2010]. Instantaneous triggering in the BDF model is thus analogous to simple Coulomb failure, with the stress amplitude of $F_{perturb}(t)$, F_0, augmenting the background tectonic stress τ to induce immediate failure and rapid slip of the tremor source region. Note, however, that because $F_{perturb}(t)$ is a transient perturbation with *maximum* stress amplitude F_0, the *average* impulse delivered to the tremor source by one positive stress lobe of $F_{perturb}(t)$ is less than that of an analogous, quasistatic impulse of stress amplitude F_0. Thus, F_0 may need to exceed $\tau_{failure}$ (which is a quasistatic stress due to tectonic loading) in order to instantaneously trigger tremor. Delayed-onset triggering occurs in a limited band of perturbation amplitudes just below the threshold amplitude for instantaneous triggering.

[12] Dynamic stresses imparted by $F_{perturb}(t)$ alter the frictional dynamics of the BDF system (Figure 2d). For low perturbation amplitudes, only a small percentage of the brittle contacts (less than~10%) rupture, and no triggering is observed. Because the strengths of the individual brittle contacts follow a power law distribution, most of the brittle contacts are weak. Thus, for low perturbation amplitudes, failure of only the weakest brittle contacts occurs, which results in negligible block motion and does not substantially modify the frictional state of the system. As perturbation amplitude increases, however, more and more contacts are ruptured. For intermediate levels of contact ruptures (~10–35%), some of the stronger contacts rupture, altering the frictional state of the system enough to cause delayed-onset triggering of tremor.

[13] At a given level of tectonic stress, there exists a critical perturbation amplitude, above which any dynamic perturbation will break the strongest brittle contacts, initiating a cascade of brittle contact failures that ruptures 100% of the contacts and causes instantaneously triggered tremor. As shown in Figure 2d, delayed-onset and instantaneously triggered tremor episodes exhibit similar time evolution of the state of the brittle contacts until the dynamic stress amplitude of $F_{perturb}(t)$ peaks, triggering tremor instantaneously only for sufficiently large F_0. The system therefore reacts nonlinearly to stress perturbations, as minute changes in perturbation stresses can result in sizable changes in frictional dynamics (e.g., Figure 2b), a feature that is not present in strictly linear-elastic failure mechanisms [*Johnson et al*., 2012].

4. Discussion

[14] The dynamic stress threshold required to trigger tremor in BDF model simulations is *not* absolute—it depends on the background tectonic stress τ relative to $\tau_{failure}$ (see Figure 2c)—providing one explanation for the wide-ranging estimates of such triggering amplitude thresholds within the Earth [e.g., *Gomberg*, 2010; *Chao et al*., 2013]. Furthermore, because triggering potential depends on both the perturbation amplitude and the background tectonic stress, the presence or absence of dynamically triggered tremor events can be used to gauge the in situ state of stress within the Earth (as suggested by *van der Elst et al*. [2013]), as well as to better constrain the fundamental characteristics of tremor.

[15] For example, consider a scenario in which instantaneously triggering is *not* observed at a known tremor source region during the passage of the seismic waves (e.g., the lower left portion of the phase space in Figure 2c). Since $F_0+\tau \lesssim \tau_{failure}$ in this portion of the phase space, the stress amplitude F_0 of the seismic wave perturbation (resolved at the tremor source) can then provide a lower bound on the average stress drop, $\overline{\Delta\tau}$, of tremor episodes at that source location. For a source of characteristic source dimension L and shear modulus G, the average slip, $\overline{\Delta u}$, associated with these tremor episodes can then be approximated as $\overline{\Delta u} \sim L\ \frac{\overline{\Delta\tau}}{G} \gtrsim L\ \frac{F_0}{G}$.

[16] Various physical mechanisms have been proposed for delayed-onset dynamic triggering of seismicity, many of which are based upon subsurface fluid flow [*Pollitz et al*., 1998; *Hill*, 2012] or secondary triggering through slow slip phenomena [*Shelly et al*., 2011; *Zigone et al*., 2012], both of which occur over a time scale much longer than the triggering perturbation. The BDF model, however, demonstrates that nonlinearity in the governing friction law alone can provide a credible explanation for delayed-onset triggering. In BDF model simulations, delayed-onset triggering occurs as a result of the failure of a subset of brittle contacts during the perturbation $F_{perturb}(t)$ (Figure 2d). This subset of brittle contacts must both be large enough to substantially modify the frictional state of the system and thus allow for delayed-onset triggering yet not so large as to induce immediate failure and instantaneously trigger an event. This phenomenon can also be interpreted in terms of a rate-and-state friction law [e.g., *Dieterich*, 1979; *Scholz*, 1988], where the dynamic perturbation reduces the critical slip distance through modification of the frictional contacts, as suggested by *Parsons* [2005].

[17] While delayed-onset triggering comprises a relatively small part of the phase space of triggered tremor in the BDF model (Figure 2c), delayed-onset triggering within the Earth should still be common. Low-amplitude perturbations (which are more frequent, given a Gutenberg-Richter distribution of remote seismicity) are more likely to give rise to delayed-onset-triggered tremor and hence reduce the number of tremor sources susceptible to instantaneous triggering. Moreover, delayed-onset triggering within the Earth may be caused not only by such alterations to local frictional dynamics but also by the aforementioned large-scale processes related to movement of crustal fluids and induced fault creep. We therefore anticipate that delayed-onset triggering is a pervasive natural occurrence, a better understanding of which may significantly improve our ability to assess seismic hazard in real time.

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