A hypothesis for delayed dynamic earthquake triggering

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[1] It’s uncertain whether more near-field earthquakes are triggered by static or dynamic stress changes. This ratio matters because static earthquake interactions are increasingly incorporated into probabilistic forecasts. Recent studies were unable to demonstrate all predictions from the static-stress-change hypothesis, particularly seismic rate reductions. However, current dynamic stress changes do not explain delayed earthquake triggering and Omori’s law. Here I show numerically that if seismic waves can alter frictional contacts in neighboring fault zones, then dynamic triggering might cause delayed triggering and an Omori-law response. The hypothesis depends on faults following a rate/state friction hypothesis, submitted to Geophys. Res. Lett., 32, L04302, doi:10.1029/2004GL021811. Citation: Parsons, T. (2005), A hypothesis for delayed dynamic earthquake triggering, Geophys. Res. Lett., 32, L04302, doi:10.1029/2004GL021811.

1. Introduction

[2] Increased regional seismicity accompanies most large earthquakes, peaking immediately after, and then fading away as a function of time [Omori, 1894]. The phenomenon of aftershocks is well recognized, but not well explained. Even if aftershocks are assumed to result from mainshock-caused stress increases, there is no consensus whether dynamic stresses induced by the passage of seismic waves [e.g., Cotton and Coutant, 1997; Belardinelli et al., 1999; Kilb et al., 2000; Gomberg et al., 2003], or static stresses induced by fault offset [e.g., Yamashina, 1978; Das and Scholz, 1981; Stein and Lisowski, 1983; King et al., 1994] are more important for near-field earthquake triggering. The issue is significant because static stress triggering is an increasingly common component of earthquake probability forecasts [e.g., Working Group on California Earthquake Probabilities, 2003]; if dynamic triggering is shown to be more important, then some probabilistic methods might require revision.

[3] There are theoretical diagnostic differences between dynamic and static earthquake triggering. Static stress changes are calculated to increase and decrease failure stress in the region near the mainshock [e.g., King et al., 1994], whereas dynamic stress changes are expected only to increase stress, although the increase might be asymmetric [Gomberg et al., 2003]. In addition, modeling studies of dynamic stress changes conclude that earthquakes are triggered as seismic waves pass through the crust, or very shortly thereafter [Gomberg et al., 1998; Belardinelli et al., 2003], implying that dynamic waves are not responsible for delayed, Omori-law aftershock sequences. Static stress changes are permanent and could thus trigger earthquakes according to Omori’s law [e.g., Dieterich, 1994].

[4] Seismic rate increases correlate with both the static and dynamic models, thus diagnosis of static stress triggering comes through looking for ‘stress shadows’, post-mainshock seismicity rate reductions associated with calculated stress decreases [Harris and Simpson, 1998]. Seismic rate increases can be many-fold and are obvious, whereas rate reductions are difficult to identify because pre-mainshock rates are often low to begin with. Generalized searches for stress shadows have questioned their universal property (K. Felzer and E. Brodsky, Testing the stress shadow hypothesis, submitted to Journal of Geophysical Research, 2004). However, the dynamic stress-transfer hypothesis lacks an explanation for Omori-law, delayed earthquake triggering. In this paper I present a simple idea for how the passage of dynamic waves might cause delayed triggering and an Omori-law response.

2. Earthquake Triggering by Contact-Area Change

[5] Much of the explanation for aftershocks and earthquake triggering is derived from laboratory fault analogs and rate- and state-dependent friction [Dieterich, 1979]. Dynamic triggering models show that earthquakes can be triggered if a stress pulse emitted by a mainshock increases another fault’s slip speed (‘rate’ of rate-state) above a threshold value [Gomberg et al., 1998; Belardinelli et al., 2003]. A dynamic stress pulse might also enhance triggering by reducing the ‘state’ of a fault (frictional increase with contact age). This was suggested to occur through slip-induced renewal of contacts, which could be time-delayed [Gomberg et al., 1997]. However, slip-induced state reduction is calculated to have minimal influence on failure time [Gomberg et al., 1998; Belardinelli et al., 2003]. Here I also call upon rate/state friction, but suggest that fault state is changed by minor ‘damage’ to fault contacts induced by passing seismic waves.

[6] This paper investigates the ramifications of a fundamental assumption: that shaking at fault zones can change the status of some frictional contacts. In a later section I investigate some potential mechanisms for how this might happen. Here it is demonstrated that Omori-law delayed triggering could result without a static stress change if, for the purpose of illustrating the concept, it is assumed: (1) fault contacts can be physically altered by a dynamic stress pulse, (2) rate/state friction is applicable, (3) contacts suffer about the same magnitude of change regardless of initial area of contact, and (4) all other material properties remain constant.

[7] Under rate- and state-dependent friction, an earthquake occurs when fault slip speed increases to an unstable
Parameter the rate/state equations [Dieterich, 1979] numerically [Gomberg et al., 1998]; each point represents a separate calculation rather than repeated stick-slip episodes.

Dependence of calculated time-to-instability on the mean critical slip distance \( D_c \). The time to failure is reduced if \( D_c \) is reduced. Calculations were made solving the rate/state equations [Dieterich, 1979] numerically [Gomberg et al., 1998]; each point represents a separate calculation rather than repeated stick-slip episodes.

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\[
\mu(t) = \mu_0 + a \ln[V(t)/V_0] + b \ln[\theta(t)V_0/D_c] \tag{1}
\]

the slip speed can be expressed as

\[
V(t) = \left[\left(\mu(t) - \mu_0 + b \ln[\theta(t)V_0/D_c]\right)/a\right] \tag{2}
\]

where \( V(t) \) is slip velocity, \( \mu(t) \) is friction, \( \mu_0 \) is reference friction (set to 0.7), \( a \) and \( b \) are dimensionless constants (set to 0.005 and 0.010 respectively), \( \theta(t) \) is the state, \( V_0 \) is a normalizing constant (set to \( 10^{-16} \) m s\(^{-1}\)) and is equal to the initial loading velocity, and \( D_c \) is the critical slip distance (varies in this example from 0.03 to 0.05 m). The constants were chosen to create stick-slip behavior [Gomberg et al., 1998] and were fixed throughout the calculations. The state evolves according to the slowness law (equation (3)) [e.g., Ruina, 1983]. I solved for the slip velocity numerically by using a 4th-order Runge-Kutta algorithm to solve the differential equations

\[
\frac{d\theta(t)}{dt} = 1 - \frac{\theta(t)V(t)}{D_c} \tag{3}
\]

\[
\frac{d\mu(t)}{dt} = k(V_b - V(t)) \tag{4}
\]

for \( \theta(t) \) and \( \mu(t) \) which were iteratively substituted into the expression for \( V(t) \) (equation (2)). The parameter \( k \) is the system stiffness (0.025 m\(^{-1}\)), selected so that the system is unstable \( (k < \text{critical stiffness} k_c = (b - a)/D_c) \). Failure is defined as initiation of failure, when \( V(t) \) reaches high values just prior to the onset of dynamic motion [Gomberg et al., 1998].

Calculations show that time to failure increases with increasing average \( D_c \) (Figure 1). This result is intuitive. Since unstable slip occurs after \( D_c \) has been traversed, it follows that failure would take longer for a larger mean critical distance. If the critical slip distance is reduced somehow, then time to failure is calculated to be reduced (Figure 2). The time-advance from the unperturbed evolution, and the time delay of failure after the perturbation depend on when in the earthquake cycle the change in \( D_c \) occurs (Figure 2). Perturbations very late in the cycle have almost no effect because slip is rapidly accelerating towards instability. However, before that latest stage, a change in \( D_c \) is calculated to cause nearly constant delayed earthquake triggering because there is still a period of slip evolution required before unstable slip occurs. Thus if passing seismic waves can reduce \( D_c \), then delayed dynamic triggering can happen. It also follows that if \( D_c \) were increased, then

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![Figure 1](image1.png)

**Figure 1.** Dependence of calculated time-to-instability on the mean critical slip distance \( D_c \). The time to failure is reduced if \( D_c \) is reduced. Calculations were made solving the rate/state equations [Dieterich, 1979] numerically [Gomberg et al., 1998]; each point represents a separate calculation rather than repeated stick-slip episodes.

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![Figure 2](image2.png)

**Figure 2.** Slip speed evolution vs. time. A perturbation of \( D_c \) is introduced approximately 100 years into the earthquake cycle, which has the effect of advancing the time to instability ahead of when it would have happened. Triggering does not happen at the time of the perturbation because some slip speed evolution must still be completed. Thus if passing seismic waves could affect \( D_c \), then delayed dynamic triggering might be possible.
earthquakes would be suppressed for a period of time after the passage of seismic waves.

3. How Dynamically Triggered Aftershocks Could Have Omori-Law Rates

[10] If for now it is accepted that a dynamic stress pulse could affect the average critical slip distance of a nucleation patch, then there is a simple means of replicating Omori-law decay in the triggered earthquake rate. Experimental direct imaging of a variety of materials shows that contact populations are distributed according to a power law relating the number of contacts \( N \) and their areas \( a \) as

\[
\frac{dN}{da} = C a^{-b},
\]

where \( C \) and \( b \) are constants [Dieterich and Kilgore, 1996]. This fractal relationship implies that smaller contacts greatly outnumber larger ones, depending on the exponent \( b \).

[11] If the critical slip distance \( D_c \) is taken as roughly the diameter of the mean contact area, then a fault (or group of faults) with a power-law distribution of contact areas would also have a power-law distribution of critical slip distances. The last major assumption in this hypothesis is that the process that changes \( D_c \) does so roughly equally everywhere. That is, \( D_c \) is reduced by the same amount regardless of mean contact area.

[12] A simulation was made using nucleation zones with \( D_c \) values populated in a power-law array according to equation (5), and that were scaled for short cycle times (\( \sim 10 \) years) appropriate for small earthquakes. Cycle initiation times were staggered using a random number generator so that their expected unperturbed failure rate would be approximately steady over a 5-year period (Figure 3a). Next, \( D_c \) values were reduced by a small amount (equivalent to a 10% reduction in area of the smallest contacts in the array). Resulting changes in time-to-failure were calculated numerically for each \( D_c \) value in the distribution. The power-law \( D_c \) distribution caused more large advances relative to cycle time than smaller, changing the seismicity rate (Figure 3b). Since there were more large advances than small, the seismicity rate change is highest near the time of perturbation and falls off with time (Figure 3c). Instantaneous triggering cannot happen under this model because perturbations during self-acceleration have no effect; however, small delays are possible if earthquake cycle times are short (Figure 3c). If this process were to occur in the Earth, it is expected that the instantaneous peak of the Omori-law sequence would be enhanced by dynamically triggered events that occur through slip caused directly by passing seismic waves [Gomberg et al., 1998; Belardinelli et al., 2003], and by static stress changes.

4. Possible Mechanisms of Seismically Induced Fault Contact Change

[13] The hypothesis presented here depends on the ability of passing seismic waves to damage, or alter some fault zone contacts. The magnitude of the stress pulse imparted by seismic waves is significantly less than lithostatic at most depths, and the yield strength of fault zone rocks would not be exceeded. However, laboratory experiments on faults with synthetic gouge that applied 1-Hz cyclical variations of normal stress observed changes in the critical slip distance \( D_c \), probably as a result of compaction [Richardson and Marone, 1999; Sleep et al., 2000]. In addition, the presence of highly pressured fluids within fault zones that partly offset lithostatic stress might be a component to the model. Observations of groundwater pressure changes induced by seismic waves have long been noted after distant earthquakes [e.g., Brodsky et al., 2003, and references contained therein]. It’s more difficult in the near field to separate the effects of dynamic and static stress changes on pore fluid pressure. Study of fault zones in situ, and laboratory analogs reveal very complex interactions among varying thicknesses of gouge, clay minerals, granulated and fractured rock, discrete slip planes that evolve into damage zones, porosity, permeability and fluid pressure [e.g., Marone and Kilgore, 1993; Caine et al., 1996; Chester and Chester, 1998; Schulz and Evans, 2000; Faulkner, 2004]; it may be possible for compaction or fluids moving through this arrangement to change the status of some fault contacts.

[14] If periodic strains from passing seismic waves cause compaction and fault-zone fluids to migrate, then pore spaces might collapse or expand, and/or granular particles within the fault gouge might be eroded or moved. These processes could alter the size and scale of fault contacts, potentially changing the mean critical slip distance. Fault damage zones may be 100’s of m wide, yet individual slip events localize in zones \( \sim 10^{-1} – 10^{-2} \) m wide [Sibson, 2003], suggesting the possibility of migrating slip planes.

Figure 3. (a) A synthetic catalog of expected earthquakes with \( \sim 10\)-year cycle times. (b) Distribution of calculated time-advances resulting from uniform reduction in critical slip distance of the distribution. (c) The synthetic catalog is perturbed with the advances in (b) and shows a rough Omori-law decay. The difference in seismicity rate between (a) and (c) is plotted in the inset with a 1/t curve for comparison.
Thus in some cases an entirely new slip plane with a different characteristic $D_c$ might be adopted.

[15] There is no obvious reason to only consider seismically induced reductions of the mean critical slip distance. Instead, reductions and increases in $D_c$ might occur in the same fault zone. The signal (seismicity rate reduction) from places where $D_c$ increased could be masked by the signal from decreased $D_c$ (seismicity rate increase), particularly if these effects were mixed on the same fault planes. To summarize, observations suggest that fault zones operate under a delicate balance between frictional state, pore fluid pressure, and a variety of chemical reactions and metamorphoses that respond to stress, strain, and temperature changes. The idea posited here is that a dynamic stress pulse passing through a fault zone might upset that balance, changing its evolution toward failure.

5. Conclusions

[16] If seismic waves passing through a fault zone could subtly change the scale of some frictional contacts, then many aspects of aftershock sequences can be replicated numerically. By changing the critical slip distance $D_c$ in rate-state slip speed calculations, it is shown that the time-to-failure can be changed. Triggering is not expected to be instantaneous with the perturbation, but would be delayed until the evolution to failure is complete. A long-term Omori-law response to dynamic triggering can be replicated if fault contacts are distributed according to a power law with smaller contacts occurring in exponentially greater numbers as laboratory experiments show. An approximately uniform reduction of $D_c$ across the power-law distribution is necessary to produce an Omori-law sequence. Perhaps the most likely way for seismic waves to affect the critical slip distance is by strain-induced compaction and/or fluid migration, which might alter gouge thickness, pore spaces, and/or move granular fault materials.

References


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