A hypothesis for delayed dynamic earthquake triggering

Tom Parsons

U.S. Geological Survey, Menlo Park, California, USA

[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 seismicity rate reductions. However, current dynamic stress change hypotheses do not explain delayed earthquake triggering and Omori's law. Here I show numerically that if seismic waves can alter some 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 law, and on seismic waves changing the mean critical slip distance (D_c) at nucleation zones.   **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]   Seismicity 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 static stress decreases [*Harris and Simpson*, 1998]. Seismicity 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 universality (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
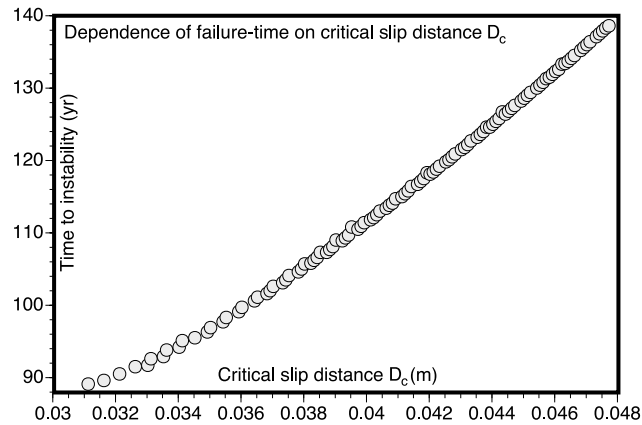
Figure 1. Dependence of calculated time-to-instability on the mean critical slip distance D_c. The time to failure is reduced if D_c is reduced. Calculations were made solving the rate/state equations [*Dieterich*, 1979] numerically [*Gomberg et al.*, 1998]; each point represents a separate calculation rather than repeated stick-slip episodes.

level [*Dieterich*, 1979]. This occurs after fault contacts move a critical slip distance D_c ($\sim 10^{-2}$ m [*Scholz*, 1988]) proportional to the displacement required to renew a population of fault contacts [*Dieterich*, 1979]. The critical slip distance is a function of fault state, and can be thought of as the average diameter of the contact areas. The critical slip distance can also be associated with the thickness of localized shear strain and the fault gouge layer [*Marone and Kilgore*, 1993]. If some process related to the passage of seismic waves through a fault, or part of a fault, can change D_c, then evolution towards failure can be changed. Not every contact need be changed to affect slip evolution, just enough such that the mean critical distance is altered.

[8] To illustrate the dependence of failure time on the parameter D_c, I adopted the spring-slider formulation of *Gomberg et al.* [1998] to make numerical calculations. This method is used to simplify a very complex process. Numerical results for single rupture patches are thus extrapolated in concept to more complex, multiple-patch ruptures in nature. By rearranging the rate/state equation

$$\mu(t) = \mu_0 + a\ln[V(t)/V_0] + b\ln[\theta(t)V_0/D_c] \qquad (1)$$

the slip speed can be expressed as

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

where $V(t)$ is slip velocity, $\mu(t)$ is friction, μ_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^{-10} m^{-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 Runga-Kutta algorithm to solve the differential equations

$$\frac{d\theta(t)}{dt} = 1 - \frac{\theta(t)V(t)}{D_c} \qquad (3)$$

$$\frac{d\mu(t)}{dt} = k(V_b - V(t)) \qquad (4)$$

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 < 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].

[9] 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
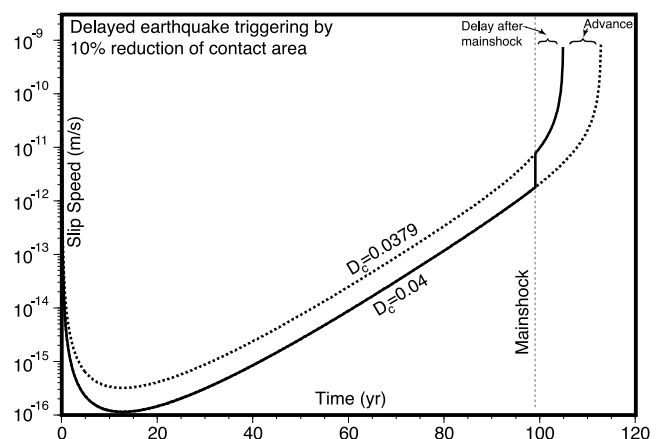

Figure 2. Slip speed evolution vs. time. A perturbation of D_c is introduced approximately 100 years into the earthquake cycle, which has the effect of advancing the time to instability ahead of when it would have happened. Triggering does not happen at the time of the perturbation because some slip 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

$$\frac{dN}{da} = Ca^{-\beta}, \qquad (5)$$

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

[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 (~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
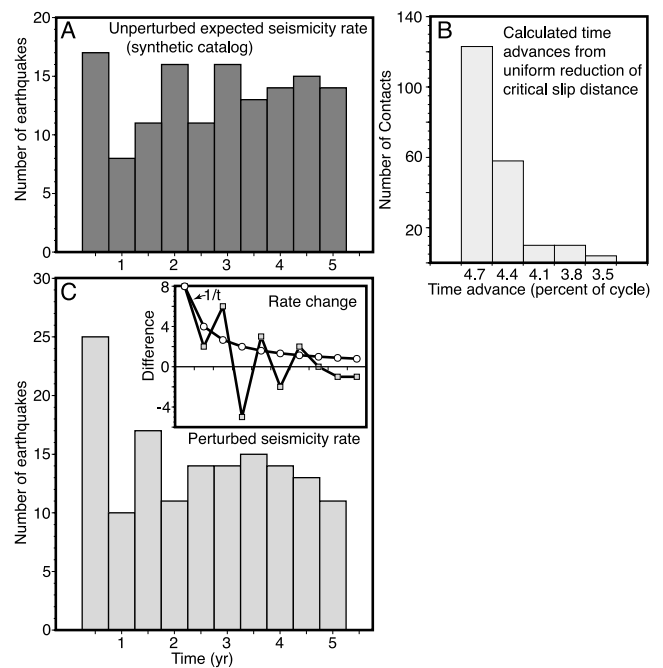

Figure 3. (a) A synthetic catalog of expected earthquakes with ~10-year cycle times. (b) Distribution of calculated time-advances resulting from uniform reduction in critical slip distance of the distribution. (c) The synthetic catalog is perturbed with the advances in (b) and shows a rough Omori-law decay. The difference in seismicity rate between (a) and (c) is plotted in the inset with a 1/t curve for comparison.

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.

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

Belardinelli, M. E., M. Cocco, O. Coutant, and F. Cotton (1999), Redistribution of dynamic stress during coseismic ruptures; evidence for fault interaction and earthquake triggering, *J. Geophys. Res.*, *104*, 14,925–14,945.

Belardinelli, M. E., A. Bizzarri, and M. Cocco (2003), Earthquake triggering by static and dynamic stress changes, *J. Geophys. Res.*, *108*(B3), 2135, doi:10.1029/2002JB001779.

Brodsky, E. E., E. Roeloffs, D. Woodcock, I. Gall, and M. Manga (2003), A mechanism for sustained groundwater pressure changes induced by distant earthquakes, *J. Geophys. Res.*, *108*(B8), 2390, doi:10.1029/2002JB002321.

Caine, J. S., J. P. Evans, and C. B. Forster (1996), Fault zone architecture and permeability structure, *Geology*, *24*, 1025–1028.

Chester, F. M., and J. S. Chester (1998), Ultracataclasite structure and friction processes of the San Andreas Fault, *Tectonophysics*, *295*, 199–221.

Cotton, F., and O. Coutant (1997), Dynamic stress variations due to shear faults in a plane-layered medium, *Geophys. J. Int.*, *128*, 676–688.

Das, S., and C. Scholz (1981), Off-fault aftershock clusters caused by shear stress increase?, *Bull. Seismol. Soc. Am.*, *71*, 1669–1675.

Dieterich, J. H. (1979), Modeling of rock friction: 1. Experimental results and constitutive equations, *J. Geophys. Res.*, *84*, 2161–2168.

Dieterich, J. H. (1994), A constitutive law for rate of earthquake production and its application to earthquake clustering, *J. Geophys. Res.*, *99*, 2601–2618.

Dieterich, J. H., and B. D. Kilgore (1996), Imaging surface contacts: Power law contact distributions and contact stresses in quartz, calcite, glass, and acrylic plastic, *Tectonophysics*, *256*, 219–239.

Faulkner, D. R. (2004), A model for the variation in permeability of clay-bearing fault gouge with depth in the brittle crust, *Geophys. Res. Lett.*, *31*, L19611, doi:10.1029/2004GL020736.

Gomberg, J., M. L. Blanpied, and N. M. Beeler (1997), Transient triggering of near and distant earthquakes, *Bull. Seismol. Soc. Am.*, *87*, 294–309.

Gomberg, J., N. M. Beeler, M. L. Blanpied, and P. Bodin (1998), Earthquake triggering by transient and static deformation, *J. Geophys. Res.*, *103*, 24,411–24,426.

Gomberg, J., P. Bodin, and P. A. Reasenberg (2003), Observing earthquakes triggered in the near field by dynamic deformations, *Bull. Seismol. Soc. Am.*, *93*, 118–138.

Harris, R. A., and R. W. Simpson (1998), Suppression of large earthquakes by stress shadows: A comparison of Coulomb and rate-and-state failure, *J. Geophys. Res.*, *103*, 24,439–24,452.

Kilb, D., J. S. Gomberg, and P. Bodin (2000), Triggering of earthquake aftershocks by dynamic stresses, *Nature*, *408*, 570–574.

King, G. C. P., R. S. Stein, and J. Lin (1994), Static stress changes and the triggering of earthquakes, *Bull. Seismol. Soc. Am.*, *84*, 935–953.

Marone, C., and B. Kilgore (1993), Scaling of the critical slip distance for seismic faulting with shear strain in fault zones, *Nature*, *362*, 618–621.

Omori, F. (1894), On aftershocks, *Rep. Imp. Earthquake Invest. Comm.*, *2*, 103–109.

Richardson, E., and C. Marone (1999), Effects of normal stress vibrations on frictional healing, *J. Geophys. Res.*, *104*, 28,859–28,878.

Ruina, A. (1983), Slip instability and state variable friction laws, *J. Geophys. Res.*, *88*, 10,359–10,370.

Scholz, C. H. (1988), The critical slip distance for seismic faulting, *Nature*, *336*, 761–763.

Schulz, S. E., and J. P. Evans (2000), Spatial variability in microscopic deformation and composition of the Punchbowl fault, southern California: Implications for mechanisms, fluid-rock interactions, and fault morphology, *Tectonophysics*, *295*, 223–244.

Sibson, R. H. (2003), Thickness of the seismic slip zone, *Bull. Seismol. Soc. Am.*, *93*, 1169–1178.

Sleep, N. H., E. Richardson, and C. Marone (2000), Physics of friction and strain rate localization in synthetic fault gouge, *J. Geophys. Res.*, *105*, 25,875–25,890.

Stein, R. S., and M. Lisowski (1983), The 1979 Homestead Valley earthquake sequence, California: Control of aftershocks and postseismic deformation, *J. Geophys. Res.*, *88*, 6477–6490.

Working Group on California Earthquake Probabilities (2003), Earthquake probabilities in the San Francisco Bay region: 2002 to 2031, *U.S. Geol. Surv. Open File Rep., 03-214*.

Yamashina, K. (1978), Induced earthquakes in the Izu Peninsula by the Izu-Hanto-Oki earthquake of 1974, Japan, *Tectonophysics*, *51*, 139–154.

————————————

T. Parsons, U.S. Geological Survey, Mail Stop 999, 345 Middlefield Road, Menlo Park, CA 94025, USA. (tparsons@usgs.gov)