Our work evolved from our work in deriving the microphysical basis of rate and state friction [1-3]. As proposed for last two years, we concentrated on damage processes within shallow (upper ~10s m) off-fault environments. We have shown that 3 nonlinear effects are related: (1) The low-amplitude S-wave velocity (measured from repeating earthquakes) in the shallow subsurface decreases following strong shaking. This velocity "heals" with the logarithm of time after the event [e.g., 4-8]. (2) Attenuation of strong seismic waves is nonlinear [9-14]. (3) The dynamic stress from strong seismic waves triggers secondary high-frequency events in the shallow subsurface [15-16]. As proposed, Sleep and Ma [17] obtained conditions where dynamic stress causes an earthquake of a very shallow fault with extreme ground acceleration. Sleep and Hagin [18] constrained the energy balance associated with low-amplitude S-wave velocity changes and showed that it involved significant nonlinear attenuation of the strong seismic wave. This progress lets us propose to continue in a productive manner our work on strong ground motions and to apply it to extreme ground motions and to near-fault damage in clay-poor rocks. We plan to return to micromechanics and to apply rate and state friction to parallel strike-slip faults. The P.I. will maintain his broad interests in seismology and tectonics.

Off-fault damage and nonlinear attenuation. Numerous observations indicate that the low-amplitude S-wave velocity decreases in the aftermath of large nearby events [e.g., 4-8]. Comparison of borehole and surface seismograms indicates that the seismic velocity changes are shallow, <100 m [6,8]. This process shares aspects of damage during frictional sliding predicted by rate and state dependent friction. The S-wave velocity decrease recovers to its pre-quake value with the logarithm of time after the event. The damage from one strong event seems to make the ground more easily damaged by subsequent events [4]. Comparison of the seismograms from weak and intense shaking indicates that attenuation becomes nonlinear at high amplitudes. Recent works include those by *Frankel et al.* [9], *Beresnev* [10], *Hartzell et al.* [11-12], *Bonilla et al.* [13], and *Tsuda et al.* [14]. Comparison of borehole and surface records indicates that the nonlinear response occurs in the shallow subsurface [e.g., 10,13,19,20]. Strong seismic waves trigger small high-frequency events in the shallow subsurface [15-16,21]. *Fischer et al.* [16,21] point out that these events are likely to be both a form of nonlinear attention and a process that damages the shallow subsurface in localized domains.

At the 2008 SCEC meeting, we found that the USC group including Adam Fischer is continuing their work on very shallow triggered earthquakes [21]. John Anderson of UNR has compiled records of extreme ground motions similar to the Parkfield record that we used [17,22] and will search for identifiable P and S arrivals to constrain moment and depth. This work will provide a test of our physical model.

We began our work on the current grant with low-amplitude S-wave velocity changes [18]. An attractive model is that the velocity changes result from dilatancy associated with inelastic shear strain in the shallow subsurface. For purposes of illustration, we model a vertically propagating S-wave with well-known equations. A standing wave represents its reflection from the free surface. The displacement and particle velocity are:

$$u = u_0 \cos(\omega t)\cos(kz); \quad (1) \quad U = -u_0 \omega \sin(\omega t)\cos(kz), \quad (2)$$

where u_0 is the scalar displacement, ω is angular frequency, t is time, k is wave number and z is depth. We do only generic calculations, as seismologists currently measure strong ground motions and repeating earthquakes at different sites. We avoid the use of spectra, as sine waves are not eigenfunctions for nonlinear waves. In fact, blind use of spectral ratios where very shallow triggered events produce high frequencies would yield negative apparent attenuation. Conversely, the raw amplitude and period of the signal on a velocity seismogram provides a stable measure of the energy in a strong seismic wave.

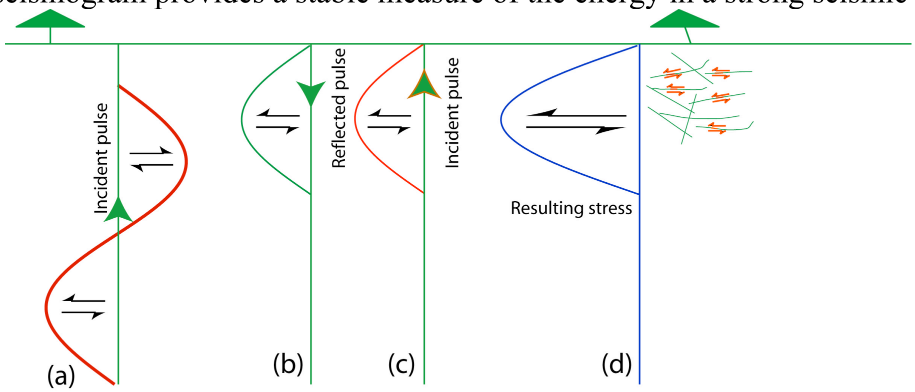

Figure 1: Conceptual model and basic nonlinear physics associated with strong ground motion [18]. Peak dynamic stresses near the quarter-wavelength depth occur when the reflected pulse encounters the second part of the incident pulse. The dynamic stress triggers small earthquakes on pre-stressed fractures. Failure occurs over a range of dynamic stresses below the nominal stress for Coulomb failure in an unstressed medium.

Dynamic stress causes inelastic deformation and attenuation

$$\tau = -u_0 kG\cos(\omega t)\sin(kz);\ (3) \qquad \tau_0 = u_0 k^2 Gz.\ (4)$$

Equation (4) is the Taylor series expression for shallow depths where $kz < 1$. This scalar form yields dimensional approximations for the magnitude of quantities but not their phase. The depth $1/k$ is a natural basis for additional scaling relationships as dynamic stresses are closest to frictional failure criteria $\tau_0 = \mu_0 P = \mu_0 \rho g z$ (where P is confining pressure, μ_0 is the first order coefficient of friction, and g is the acceleration of gravity) in that region [e.g., 12, p. 1614]. That is, the energy of a reflecting wave is kinetic energy near the free surface that does not cause dissipation and shear-strain energy around the quarter wavelength depth $\pi/2k$. Lithostatic stress continues to increase below the quarter wavelength depth while dynamic stresses are bounded by their value at that depth. The combination of significant shear-strain energy and dynamic stresses near confining pressure imply that nonlinear dissipation should occur around the scale depth if the rock is not fully elastic.

Our procedure [18] uses before and after low-amplitude S-wave delays Δt from repeating earthquakes assumes that damage occurs around the scale depth $1/k$ and produces a porosity change Δf over a depth range scaling to $1/k$. This process involves work against lithostatic pressure $\rho g Z$ where ρ is density, g is the acceleration of gravity, and $Z \approx 1/k$ is depth. The work per damaged volume is $W = \lambda \rho g \Delta f / k$ where λ is the ratio of total work over work to open porosity. We use percolation theory to obtain a linearized relationship between the shear modulus change (and hence S-wave velocity change) and the porosity change. The energy (per volume) in a wave as shown in Figure 1 is $E = 0.5\rho U_0^2$ where one in practice obtains velocity amplitude U_0 from a surface strong motion velocity seismogram. With some algebra, the diminution of energy is

$$\frac{\text{energy loss}}{\text{energy in wave}} \approx \frac{D_W W}{2\pi E} \approx \frac{\Delta t_S 2(\gamma - f_{\text{bef}}) g \lambda V_S}{U_0^2 \pi}, \quad (5)$$

where we assume a pulse-like arrival as in Figure 1, damage occurs over at depth range D_W / k, the shear modulus extrapolates to 0 at porosity γ, and the porosity in the rock before the earthquake is f_{bef}. $\Delta t_S \approx 0.007$ s [6]. It is unnecessary to know the depth D_W / k range over which damages occurs in (5). We did need to specify, however, that the damage was concentrated near the scale depth $1/k$ [18].

We obtained generic results of Parkfield at sandstone sites [6,23] where VS30=~300 m s^{-1}. The strong seismic waves had an angular frequency of ~10 s^{-1} and a velocity amplitude of ~ 0.5 m s^{-1}. The scale depth is thus ~30 m, implying that VS30 is an appropriate measure of S-wave velocity. The porosity difference in (5) is constrained, as it needs to yield the observed shear wave velocity in our linearized model. We do not know the value of λ but 2 (equal dissipation in dilation and shear) and ~17 for laboratory gouge are plausible [3, 18]. The former yields the reasonable result for Parkfield of a ~50% diminution of energy; the latter yields excessive computed diminution of energy.

There are practical engineering implications to our results. First we relate are results to engineering practice. Coulomb ratio above the scale depth is

$$\frac{\tau_0 k}{\rho g} = \frac{u_0 G k^2}{\rho g} = \frac{u_0 \rho c^2 k^2}{\rho g} = \frac{u_0 \omega^2}{g} = \frac{A_0}{g}, \quad (6)$$

which is the measured ratio of (sustained) particle acceleration to gravity and independent of material properties. This mechanical definition of strong motion in (6) is equivalent to the conventional definition that sustained accelerations are on the order of that of gravity. This useful relationship is entrenched in work on the nonlinear attenuation of seismic waves. For example, *Beresnev* [10] compiled amplification ratio as a function of peak ground acceleration. Second engineers would like to know the sustained acceleration where nonlinear attenuation becomes significant at a given site. We note that "mild" strong ground motions as at Parkfield are more common that very strong ground motions. One can use S-wave delay changes and small triggered earthquakes to show that a site and similar sites are in the nonlinear regime. Third, our model involves pre-stress [24] that is repeatedly renewed by failure during strong shaking yields increasing attenuation over a range of dynamic stresses similar to the Masing rules [e.g., 12, 25].

Publications. We have published [24] that presents to the energy balance discussed above and examines more sophisticated ways to relate S-wave velocity to porosity and to starting frictional strength. We have found that the *Linker and Dieterich* [26] relationship allows one to extrapolate experimental frictional strength toward low confining pressures (Figure 1). One goal is to calibrate the maximum strength of the shallow subsurface and thus obtain limits on extreme ground motions. Significant nonlinear attention occurs well before the maximum strength. We are preparing paper on the compactional strength of tuff, sandstone, and shallow regolith as near Parkfield. We found that the *Linker and Dieterich* [26] relationship provides a reasonable fit for frictional failure of porous material at low confining pressure. We have also shown that materials like unwelded tuff with pointy real contacts have creep rates that are very strongly dependent on confining pressure. Such materials fail in strong P-waves at stresses moderately higher than their ambient lithostatic stress. Sandstone with broad contacts fails at pressures greatly exceeding the ambient lithostatic stress. We have obtained a unified rate and state and end-cap formalism in terms of stresses at real contacts.

We have paper in press in *BSSA* discussing shallow transient low-amplitude seismic velocity changes asssocied with the Parkfield mainshock. a new simple way to constrain the depth of S-wave velocity changes with existing data. The coda for repeating earthquakes in the Parkfield regions is progressively delayed with increasing coda-S arrival time [6]. The delay of S- and P-coda relative to their primary phases is mostly from circuitous paths at great depths where the seismic velocity does not change. The coda reverberates in the shallow subsurface where that seismic velocity did change. The change in coda-P and coda-S is comparable to the change in S-P as observed. Conversely modeling the coda a scattered Rayleigh or body where most of the primary to coda delay occurs in the shallow subsurface predicts excessive changes in coda-primary delays. The noise seismograms [27] are mostly surface waves (~4 s period, ~1.7 km/s group velocity), which give little depth resolution. The "vertical" S-wave delay ~0.007 s by Rayleigh's principle [e.g., 28, p. 304] gives relative Rayleigh wave velocity change ~0.05% if we assume the perturbations are shallow using [29]. The two data sets are thus compatible with most of the damage being shallow. However the fractional Rayleigh wave velocity change is sensitive to its group velocity and period. However, we autocorrelated borehole seismograms and obtained P-wave (but not S-wave delays) of the free surface reflection. The P-wave delay changes are comparable to the S-P delay changes as expected.

We are preparing paper on nonlinear attenuation of reverberative waves in basins like Palm Springs and Los Angeles. An empirical site-response treatment as used for body waves is inadequate as energy passes through the near surface several times. Modest nonlinear attenuation each time the energy passes through the near surface will add up. Our tasks are to see whether these onerous calculations are necessary for reverberative waves and to check a potential shortcut. We begin with a generic approach: elastic energy that resides at Coulomb stress ratios above 0.1 suffers modest nonlinear attenuation. We find that a modest fraction of elastic energy resides at Coulomb stress where failure is expected in a prestressed material. A synthetic record with a maximum surface particle velocity of ~1.8 m s^{-1} from Petashake with local seismic velocity provided guidance.

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