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Dynamic Triggering

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(Herrington and Brodsky 2006, Miyazawa et al. 2005). As an example of (1), Rabaul caldera (Papua New Guinea) responded to a M7 earthquake at a distance of 180 km with a pronounced earthquake swarm but produced no detectable activity in response to a second M7 earthquake two month later at a distance of only 60 km (Mori et al. 1989). Recharge times for the Geysers geothermal area appear to be short -- a matter of months or less (Gomberg and Davis 1996). They are apparently much longer (but ill-constrained) for Long Valley caldera where none of the triggered sites are strictly repetitive. Recharge times and their spatial-temporal fluctuations complicate efforts to establish triggering thresholds for dynamic stresses (e.g. Gomberg 1996, Gomberg and Johnson 2005). At the same time, they offer important clues to the processes behind dynamic triggering at any given site.

8.03.5 Peak Dynamic Stresses, Triggered Magnitudes and Durations,

Seismic wave amplitudes responsible for dynamic triggering are variously reported as peak dynamic stress, T_p, or peak dynamic strain, ε_p. In the plane-wave approximation, peak dynamic strain is proportional to peak particle velocity, ů divided by the phase velocity, or $\varepsilon_p \sim \mathring{u}\, v_s^{-1}$ for shear waves, and $T_p \sim G(\mathring{u}\, v_s^{-1})$ where G is the shear modulus (commonly taken as $G \sim 3\text{x}10^4$ MPa) and v_s is the shear wave velocity (e.g. see note 20 in Hill et al. 1993). Although peak dynamic stresses at depth will in general differ from those based on seismograms recorded on the Earth's surface, they can be estimated given a reasonable model for the physical properties of the underlying crust (Gomberg, 1996). In computations of stress amplitudes at depth, for example, the tendency for surface wave displacement amplitudes and strains to decrease with depth will be offset to one degree or another by the tendency of elastic moduli to increase with depth.

Reported peak dynamic stresses associated with remotely triggered seismicity range from 0.01 MPa ($\varepsilon_p \sim 0.3$ microstrain) at the Coso volcanic field 3,660 km from the $M_W = 7.9$ Denali Fault earthquake (Prejean et al. 2004) to ~1 MPa ($\varepsilon_p \sim 3$ microstrain) or more for the Little Skull Mountain earthquake 240 km north of the M=7.4 Landers earthquake (Hill et al. 1993, Anderson et al. 1994). Peak dynamic stresses can easily exceed 4 MPa within the transitional region to the aftershock zone and near field of a large earthquake (Kilb et al. 2002). The large range in peak dynamic stresses (or strains) that have resulted in remote triggering together with variations in intrinsic site characteristics and recharge times indicate that the triggering process does not depend on a simple minimum amplitude threshold for dynamic stresses to be effective. Thus, although most instances of dynamic triggering involve dynamic strains, $\varepsilon_p \geq 1\text{x}10^{-6}$, or dynamic stresses, $T_p \geq 0.03$ MPa (Gomberg and Johnson 2005), this is neither a necessary nor sufficient threshold for dynamic triggering (most areas with $T_p \geq 0.03$ MPa are not triggered and some areas with $T_p < 0.01$ MPa are triggered). The weight of evidence at this point suggests that for a given peak amplitude, dynamic stresses in the periods range 20 to 30 seconds are more effective at inducing a triggered response than those at higher frequencies and that a lower bound on the peak dynamic stress capable of inducing a triggered response may be at the level of tidal stresses, or ~0.001 MPa at periods of 12 to 24 hours (Beeler and Lockner 2003; also see section 8.02.3.4).