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Triggered tremor, phase-locking, and the global clustering of great earthquakes

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ABSTRACT

A system of non-linear coupled relaxation oscillators is used to show how the communication between large earthquakes on a global scale can align their seismic cycles to produce a worldwide clustering of large events. Our model builds on recent observations that the seismic waves from a large earthquake can trigger an episode of non-volcanic tremor at the base of a distant fault. We assume that tremor is indicative of creep on the ductile extension of the fault zone that loads its overlying seismogenic layer thus advancing the fault toward failure. If this advance is larger toward the end of the seismic cycle, we show that two or more interacting faults will align their cycles, even if their recurrence intervals are not identical.

1. Introduction

A plot of great earthquakes (magnitude Mw > 8.0) since 1900 (Fig. 1) shows periods of enhanced activity: roughly between 1900 and 1910, 1950 and 1965, and from 2005 to the present (Ammon et al., 2010; Bufe, and Perkins, 2005). The significance of these temporal clusters has been questioned for two reasons: 1) the periods of increased activity could be Poisson clusters in a random distribution, and 2) no plausible physical mechanism has been proposed to explain worldwide temporal clustering (Michael, 2011; Shearer and Stark, 2012).

The reason that a viable physical mechanism has been so elusive is that the direct transfer of stress between faults, which explains the communication between earthquakes on a local scale, does not work on a global scale. On a local scale, the changes in stress caused by an earthquake give a good description of the spatial and temporal clustering of its aftershocks. These stress changes activate areas of less-than-average slip on the fault plane as well as the myriad of favorably oriented smaller surrounding faults. The only remaining controversy is an identification of the mechanism that delays the response and produces Omori's law for the observed decrease in activity with time.

Direct stress transfer models have also been proposed to explain the temporal clustering of large events on a given section of a plate boundary that is observed in many paleoseismicity studies (Dolan et al., 2007; Plamen et al., 2010; Rockwell et al., 2000). Unlike the aftershock models that involve direct (although delayed) triggering, the role of stress transfer in regional clustering is to align the seismic cycles of nearby large events (Sammis et al., 2003; Scholz, 2010). The idea is that stress transfer from each large earthquake advances the seismic cycles on other large faults in the area. If this advance is larger toward the end of the seismic cycle, and if the large events in the region have about the same recurrence interval, then the individual cycles of these large regional earthquakes will tend to lock into phase and produce the observed temporal clustering.

There is no reason, in principle, why this mechanism should not work on a global scale to align the cycles of great earthquakes, thereby producing the worldwide clustering in time as shown in Fig. 1. Note that the three clusters in Fig. 1 probably represent independent synchronizations since each contains a different collections of events and they are spaced only about 50 years apart, which is much shorter than the several hundred year recurrence interval of great earthquakes. The central question is: how does a large earthquake on one fault advance the seismic cycles of other faults at global distances where static Coulomb stress changes are insignificant?

The answer may lie in the mounting evidence that dynamic stress changes in the surface waves from a large earthquake can trigger episodic tremor and episodes of slow slip at the base of large faults at global distances. The phenomenon has been observed in Japan, Sumatra, Cascadia, Mexico and Northern California (Hill et al., in press; Peng et al., 2010; Rubinstein, et al., 2010; Zigone et al., in press).

Episodic tremor is clearly an important but not well-understood expression of fault activity. Many questions remain about the physical processes involved (Ben-Zion, 2012). For our purposes however, we do not require specific knowledge about the nature of tremor, only that its presence in or below a fault zone is associated with slow slip and that the dynamic stresses present in the waves from large earthquakes at great distances can trigger its occurrence. There has also

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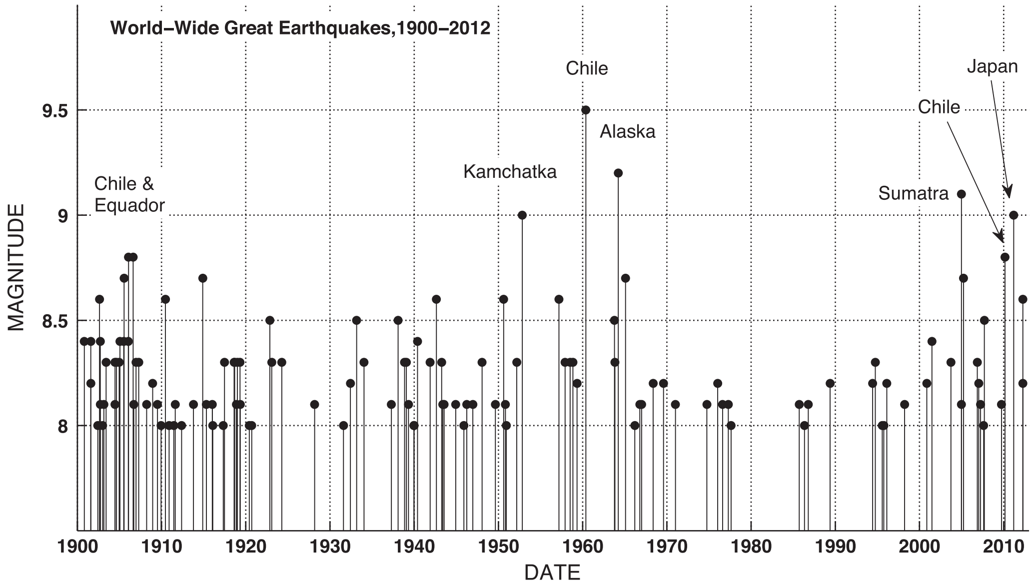

Fig. 1. Worldwide earthquakes of M 8 or greater between 1900 and 2012. Magnitude M is the larger of Mw or Ms. If neither is available, then the magnitude reported in the NGDC/WDC Significant Earthquake Database (2012) was used. Note three clusters separated by relative lulls in the late 1920s and early 1980s.

been some debate about the typical scale of triggered tremor — whether it is just very isolated asperities that are triggered, or whether it is indicative of a larger scale episode of slow slip. Recent papers by Johnson et al. (2012), Shelly et al. (2011), and Zigone et al. (in press) support a link between slow slip and tremor, and we assume in the following that triggered episodes of tremor correspond to episodes of significant slow slip, or that episodes of slow slip without tremor can also be triggered by seismic waves from great earthquakes.

Bursts of tremor are closely associated with geodetically detected slow slip events in both subduction and transform fault environments (Shelly, et al., 2009). Although there was no geodetically detected acceleration of creep prior to the 2004 Parkfield earthquake, tremor beneath the San Andreas in the 3 months prior to that event was attributed to accelerating creep beneath the eventual earthquake epicenter (Shelly, 2009). Even without making specific assumptions about the physical process producing tremor, it appears that tremor may be a sensitive tool to infer regions of deep accelerating creep where surface deformation may be below the detection threshold for geodetic measurement. If we accept the hypothesis that tremor is a proxy for a creep event on the deep extension of a large fault, then tremor episode will load the overlying seismogenic portion of the fault thereby advancing its seismic cycle and making phase-locking possible for large faults over global distances. Although it has been observed that great earthquakes can trigger seismicity at great distances (Gomberg et al., 2001; Hill, et al., 1993), we emphasize that the mechanism proposed here does not involve the direct triggering of distant earthquakes, but rather a more subtle coupling through deep creep that modifies the seismic cycles to produce an emergent alignment of their cycles and, thereby, the observed temporal clustering.

2. A coupled relaxation oscillator model for global seismicity

It is well known that an ensemble of globally coupled relaxation oscillators can phase-lock under a broad range of conditions to produce synchronizations observed in a variety of biological, electrical, and mechanical systems (Bottani, 1995). In this section we develop the simple model for interacting faults originally presented by Sammis et al. (2003).

To first order, the seismic cycle of a large characteristic earthquake can be modeled as a relaxation oscillator that is loaded by a uniform tectonic strain rate $\dot{\varepsilon}$ and fails at a critical strain ε_f. We assume that the stress on the shallow seismogenic portion of a fault plane is either linearly related to the strain in which case

$$\sigma = \sigma_f \frac{\varepsilon}{\varepsilon_f} \qquad 0 \le \varepsilon \le \varepsilon_f \tag{1}$$

or has the convex non-linear form

$$\sigma = \sigma_f \sin\left(\frac{\pi}{2}\frac{\varepsilon}{\varepsilon_f}\right) \qquad 0 \le \varepsilon \le \varepsilon_f. \tag{2}$$

Both are plotted in Fig. 2a where it is evident that Eq. (2) has the negative curvature typical of the stress–strain curves commonly observed in triaxial laboratory experiments on solid rock (Lockner, 1998) or in a frictional slip on fractured surfaces (Goebel et al., 2012). Our results are not sensitive to the exact non-linear form chosen, as long as it is convex.

When an event on one fault increases the stress on a second, the advance in strain can be found by differentiating Eqs. (1) or (2). For the linear case, this is just

$$d\varepsilon = \frac{\varepsilon_f}{\sigma_f} d\sigma. \tag{3}$$

The advance in strain is independent of where the fault is in its seismic cycle. For the convex case, this increment is given by:

$$d\varepsilon = \frac{2\varepsilon_f}{\pi\sigma_f}\left[\cos\left(\frac{\pi}{2}\frac{\varepsilon}{\varepsilon_f}\right)\right]^{-1} d\sigma. \tag{4}$$

In this case the strain increment $d\varepsilon$ produced by a stress increment $d\sigma$ increases as $\varepsilon \to \varepsilon_f$.

To see why a convex stress–strain curve leads to a phase lock, consider two faults #1 and #2 that produce large earthquakes having about the same recurrence intervals T. Let T_{12} be the waiting time between an event on fault 1 and the next event on fault 2, and assume