Stress transfer, dynamic triggering, and stress correlations: How earthquake occurrence effects the timing and slip of subsequent earthquakes

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Abstract

Earthquakes produce physical changes in the region surrounding their occurrence. A number of these changes influence the environment in which subsequent earthquakes occur, potentially effecting their timing and slip distribution. The process of stress transfer, where coseismic slip and postseismic processes modify the stress field surrounding an earthquake, can influence both the timing and slip distribution of subsequent events. The dynamic strains induced by the passage of large amplitude seismic waves can trigger earthquake failure at large distances from the original event. When laboratory derived frictional laws are added to these processes, inherent time-dependent effects arise. Finally, it has been suggested that the action of many smaller earthquakes can smooth a regional stress field and prepare the way for larger events. To be realistic, simulations of the earthquake process must account for these effects.

Introduction

It has been known for a long time that earthquake occurrence is non-random in space and time. Foreshock-mainshock-aftershock sequences and swarms are common in the earthquake record; indeed, in some cases aftershocks may make up as much as half an earthquake catalog (Reasenberg, 1985[18]). Any simulation of the earthquake process must be able to reproduce these features before it can be considered a viable model. Knowledge of the physical processes underlying such clustering is therefore essential to produce accurate earthquake simulations.

There have been numerous studies looking into the physical processes by which one earthquake effects the timing and slip distribution of subsequent events. In the following sections I will review some of the recent work in this area, pointing out what I believe are the main processes at work. This review will be by no means exhaustive; there are numerous other examples I could have used to illustrate my point. Following the brief review I will discuss the implications of these results for earthquake simulation models, and what I believe are the important observations these simulations must reproduce to be viable models of the earthquake process.

How one earthquake effects the timing of subsequent earthquakes

Over the past several years there have been many studies of earthquake triggering, where the occurrence of one earthquake has triggered others. Typically, there seem to be two major ways in which one earthquake can trigger others: 1) stress transfer - the deformation of the earth's crust due to fault slip changes the surrounding stress field and increases or decreases stress on surrounding faults. This can potentially include coseismic slip, postseismic slip, and relaxation of the lower crust and mantle, and 2) dynamic triggering - the dynamic stress changes during the passage of seismic waves triggers earthquakes. There are now several well-constrained observations of both processes acting to trigger earthquakes. I will briefly review each in turn.

Stress transfer

Since earthquakes are a result of failure of a fault surface under shear and normal stress, it is not surprising that earthquakes modify the stress field on and in the region around the slipped area. Such stress transfer can occur both coseismically and through several postseismic processes such as postseismic creep (both on the slipped and the down-dip portion of the slipped fault) and postseismic viscous relaxation of the lower crust and mantle. The change in the stress tensor for all of these processes can be calculated for the region surrounding an earthquake, and subsequent seismicity compared to the stress changes.

Stress changes are commonly quantified as a change in a Coulomb Failure Function (CFF), defined as (Oppenheimer et al., 1988[15]):

$$\Delta CFF = \Delta\tau - \mu(\Delta\sigma - \Delta p) \tag{1}$$

where τ is the shear stress, σ is the normal stress, p is the pore fluid pressure, μ is the coefficient of friction, and Δ represents a change in the following quantity. $\Delta\tau$ and $\Delta\sigma$ can be calculated from the change in the stress tensor, but Δp cannot. Simpson and Reasenberg (1994[21]), among others, have explored different ways of taking into account the potential effect of Δp on ΔCFF. A fault orientation must be assumed in to resolve the change in the stress tensor into $\Delta\tau$ and $\Delta\sigma$. Usually, either an *a priori* set of fault orientations are assumed (e.g., Jaumé and Sykes, 1996[12]), or a background stress state is assumed and ΔCFF calculated on optimally oriented planes (e.g., King et al., 1994[14]).

Numerous studies have compared the ΔCFF due to coseismic slip to the spatial distribution of activity subsequent to an earthquake. Aftershock locations correlate very strongly with regions of positive ΔCFF (e.g., King et al., 1994[14]). Harris et al. (1995[10]) found that, within 1.5 years following a $M \geq 5.0$ earthquake in southern California, most subsequent $M \geq 5.0$ earthquakes fall in regions of increased ΔCFF from a previous event. Great earthquakes also create large regions of negative ΔCFF; moderate magnitude earthquake activity is observed to be strongly depressed in these regions for time periods of up to decades following the great event (Jaumé and Sykes, 1996[12]; Harris and Simpson, 1998[11]). Other studies have examined stress transfer by other means (e.g., postseismic relaxation; Pollitz and Sacks, 1997[16]) and also find correlations between these stress changes and subsequent activity.

A more recent development in this area is the combination of stress changes with laboratory based frictional relationships (Harris and Simpson, 1998[11]; Gomberg et al., 1998[8]). This introduces an inherent *time dependence* into the effect of stress changes on subsequent earthquake activity. In these models the temporal aspects are controlled not only by the friction-law parameters and the magnitude and sign of the stress changes, but also by the stress loading rate and how close the fault was to failure prior to the stress change. Some predictions of these models are: a) a step increase in shear stress produces a seismicity increase that decays with the same functional form as Omori's Law, and that the temporal length of an aftershock sequence is inversely proportional to the stressing rate (Dieterich, 1994[5]), and b) stress changes induced earlier in the fault loading cycle have a greater effect on failure time than those applied later (Gomberg et al., 1998[8]).

Dynamic triggering

Another means by which one earthquake can effect the timing of subsequent earthquakes is triggering by stress induced during the passage of large amplitude seismic waves. This "dynamic triggering" was most clearly observed following the 1992 Landers earthquake in the western USA (e.g., Anderson et al., 1994[1]). Since then a number of studies have looked into the mechanics of how this triggering occurs (e.g., Gomberg and Bodin, 1994[6]).

One important study in this area is that by Gomberg and Davis (1996[7]) who examined the triggering of micro-earthquakes at The Geysers, a geothermal area in northern California. Micro-earthquakes at The Geysers are triggered by a number of processes, including dynamic triggering during the passage of large amplitude surface waves, stress transfer from other earthquakes, and fluid pressure changes caused by geothermal power production. Gomberg and Davis (1996[7]) propose a model where the triggering stress is frequency dependent, and that larger triggering stresses are needed at low frequencies.

Gomberg et al. (1998[8]) also examined dynamic triggering in the context of the rate-and-state-friction models. They found that triggering by transient loads (i.e., seismic waves) only produces a clock advance and that a larger stress amplitude is needed for the same clock advance as compared to static triggering. Transients applied later in the cycle produce a greater clock advance than those applied earlier. Because of the time dependency of fault strength built into the rate-and-state-friction formulation, triggering can be delayed for some time after the passage of the seismic waves.

How one earthquake effects the slip distribution of subsequent earthquakes

Besides effecting the timing of subsequent earthquakes, there is also evidence that stress transfer can effect their slip distribution. There are several cases where slip in an earthquake either stops or is greatly reduced at the boundary between regions where stress has increased and decreased. It has also been noted that the stress transfer from one or more smaller earthquakes has apparently effected the slip distribution of succeeding larger events. And, more interestingly, a number of recent studies suggest that the global effect of the many small earthquakes in a region is to smooth the long wavelength features of the regional stress field, and thus prepare the way for progressively larger events.

Two recent studies have pointed out that the boundary between regions of coseismic stress increase and decrease have apparently limited rupture in an earthquake occurring a short time later. King et al. (1994[14]) point out that the northeast extent of slip in the Big Bear aftershock of the 1992 Landers earthquake terminates where the stress changes from Landers become negative. Caskey and Wesnousky (1997[4]) show that the southward extent of rupture in the 12 December 1954 Dixie Valley earthquake may have been controlled by the transition from positive to negative stress transfer from the Fairview Peak earthquake that occurred 4 minutes earlier. Although not conclusive, these studies suggest that stress transfer from a large earthquake can influence not only the timing but also the dynamics of slip of succeeding earthquakes.

It also appears that smaller earthquakes can have effects upon the slip distribution of succeeding larger events. Again, King et al. (1994[14]) have one of the better examples. They show how stress transfer from moderate events in the 20 years preceding the Landers earthquake acted to increase ΔCFF on the southern 2/3's of the subsequent rupture surface. More recently, Perfettini et al. (1998[17]) suggest that the 1988 and 1989 Lake Elsman earthquakes effected the slip distribution of the succeeding Loma Prieta earthquake. They note that the greatest slip during the 1989 event was located where the Lake Elsman earthquake substantially reduced the normal stress on the Loma Prieta fault plane.

An important recent development in the study of interacting fault systems is the suggestion that stress transfer from many small earthquakes may act to prepare the way for the occurrence of much larger events. This hypothesis was presaged by Hanks (1992[9]), who pointed out that small earthquakes are just as important as larger ones in redistributing driving forces along active faults. Several simulation studies of the earthquake process, using very different models, have suggested that small events act to smooth the stress field at long wavelengths, and that these long-wavelength correlations in the stress field are a necessary pre-condition for the occurrence of a large earthquake (e.g., Schmittbuhl et al., 1996[20]; Ben-Zion, 1996[2]). This idea is perhaps best developed in the so-called "Critical Point" hypothesis, which considers a large earthquake as being analogous to a phase transition (e.g. Saleur et al., 1996[19]; Sornette and Sammis, 1995[22]). As reviewed by Bowman et al. (1998[3]), one aspect of this model is that smaller earthquakes smooth the stress field at long wavelengths while roughening it at shorter wavelengths. This increases the correlation length of the stress field and allows earthquakes that initiate at later times to overcome stress barriers and grow to larger sizes. The occurrence of an very large earthquake, an event than spans a significant part of the fault system, decorrelates the stress field at long wavelengths and the process starts again.

Discussion

The work reviewed above suggests the occurrence of any earthquake produces physical changes in the surrounding environment that can effect the timing and rupture characteristics of subsequent earthquakes. The degree to which this occurs is clearly dependent upon the size of the event in question; a single great earthquake can have an observable influence on regional seismicity over decades but a micro-earthquake may need to co-operate with thousands of others to prepare a region for the occurrence of a large event.

With this in mind, I have to question the viability of models of regional seismicity that include only nearest-neighbor interactions. Such models do clearly show complex behavior and some observable aspects of seismicity (e.g., power-law size statistics and in some cases aftershock sequences). But observational evidence suggests that earthquakes interact over distances greater than even a single fault depth. This interaction can also be very three-dimensional; coseismic slip both increases and decreases stress on nearby faults, depending upon their orientation, sense of slip, and location relative to the initial event, and the amplitude of dynamic strains are very dependent upon the orientation and rupture directivity during an earthquake.

How would such effects change the model dynamics if included in earthquake simulations? A clue may be in simulations of seismicity in southern California performed by Ward (1996[24]). Performed on a map-like set of faults and including the effects of stress transfer, the model exhibits clustering of small events and quasi-periodicity in the larger events. Earthquakes on individual fault segments exhibit varying amounts of characteristic and power-law size distributions, and increased rates of smaller earthquakes precede increased rates of larger events. Similar variations in the nature of earthquake size distributions dependent upon total fault slip have been reported by Stirling et al. (1996[23]) and increased rates of moderate events are seen before numerous large and great earthquakes (Jaumé and Sykes, 1999[13]).

What does this mean for the future of simulations of the earthquake process? Most importantly, I think, is that there is more to natural seismicity than frequency-magnitude statistics. Earthquakes exhibit considerable space-time clustering and other patterns, and a number of physical mechanisms have been proposed to explain this occurrence. Numerical simulations can include these effects, and the results compared to various aspects of natural seismicity to see if they reproduce these space-time patterns.

Conclusions

Earthquakes interact with each other by a number of physical processes that can modify the timing and even the size and slip distribution of succeeding events. These interactions clearly modify the space-time patterns of seismicity, and should be included in simulations of the earthquake process. A simulation that includes the effects of stress transfer reproduces more of the richness of real seismicity observations than many simpler models. To be a realistic representation of the earthquake process, a simulation model should reproduce the observed space-time patterns of natural seismicity.

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