Accelerating seismicity and stress accumulation before large earthquakes

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Abstract. The stress field that existed before a large earthquake can be calculated based on the known source parameters of the event. This stress field can be used to define a region that shows greater seismic moment rate changes prior to the event than arbitrarily shaped regions, allowing us to link two previously unrelated subjects: Coulomb stress interactions and accelerating seismicity before large earthquakes. As an example, we have examined all $M{\geq}6.5$ earthquakes in California since 1950. While we illustrate the model using seismicity in California, the technique is general and can be applied to any tectonically active region. We show that where sufficient knowledge of the regional tectonics exists, this method can be used to augment current techniques for seismic hazard estimation.

1. Introduction

In recent years, observational and theoretical descriptions of spatio-temporal patterns of seismicity have focused on two fundamental (and controversial) observations: static stress (Coulomb) interactions between earthquakes [e.g. *Harris*, 1998; *King and Cocco*, 2001] and accelerating seismic moment release before large earthquakes. While there have been several documented examples of static stress changes influencing the space-time pattern of seismicity *following* great earthquakes (main shocks and aftershocks), there have been few attempts to link this method to the evolution of seismicity *before* great earthquakes (precursory seismicity and foreshocks).

In this paper, we describe a simple physical model that links static stress modeling to accelerating moment release before a large event. By combining these two techniques, we are able to provide a physically reasonable framework for understanding the evolution of seismicity associated with great earthquakes. Many modern models for the evolution of seismicity during the seismic cycle rely on concepts borrowed from statistical mechanics [e.g. *Rundle*, 1989; *Sammis and Smith*, 1999; *Lyakhovsky et al.*, 2001]. While these models are quite sophisticated, the underlying physical means of interaction between events is often poorly constrained. Coulomb stress modelling provides a physical basis for calculating not only the stress transmitted between events, but also the spatial extent of these interactions. Likewise, the combination of a statistical mechanics framework with the Coulomb approach

facilitates the analysis of the complex temporal evolution of the system described by the theoretical models.

2. Regional Seismicity Before Large Earthquakes.

A number of publications have suggested that seismicity over a wide region appears to be modified before a major event [see *Bowman et al.*, 1998; *Jaumé and Sykes*, 1999 and references therein]. *Bufe and Varnes* [1993] suggested that a simple power-law time-to-failure equation derived from damage mechanics could be used to model the observed seismicity. They showed that the seismicity before the 1989 $M{=}7.0$ Loma Prieta earthquake can be fitted by a relation of the form

$$\varepsilon(t){=}A{+}B(t_c{-}t)^m \qquad (1)$$

where t_c is the time of the large event, B is negative and m is usually approximately 0.3. A is the value of $\varepsilon(t)$ when $t = t_c$, i.e., the final Benioff strain up to and including the largest event. The cumulative Benioff strain at time t is defined as

$$\varepsilon(t) = \sum_{i=1}^{N(t)} E_i(t)^{\frac{1}{2}} \qquad (2)$$

where E_i is the energy of the ith event and $N(t)$ is the number of events at time t.

A power-law time-to-failure function is also predicted by models where the crust is in a state of "intermittent criticality" [*Sammis and Smith*, 1999; *Bowman and Sammis*, 2001]. This hypothesis is an outgrowth of efforts to characterise large earthquakes as a critical phenomenon [e.g. *Rundle*, 1989; *Sornette and Sammis*, 1995], and suggests that a large earthquake is the end result of a process in which the stress field becomes correlated over increasingly long scale-lengths. The scale over which the stress field is correlated sets the size of the largest earthquake that can be expected at that time. The largest event possible in a given fault network can not occur until regional criticality has been achieved. This large event then destroys criticality on its associated network, creating a period of relative quiescence after which the process repeats by rebuilding correlation lengths towards criticality and the next large event.

3. Pre-Earthquake Distribution of Stress

Many studies have demonstrated that the history of seismicity in a region strongly influences the location, and presumably the timing, of subsequent earthquakes. Such works include the observation of Coulomb stress interactions [see *King and Cocco*, 2001 and references therein] and stress shadows following great earthquakes [*Simpson and Reasenberg*, 1994; *Jaumé and Sykes*, 1996; *Harris and Simpson*, 1996]. From these studies we can draw the conclusion that stress must accumulate not only on the fault itself, but also in

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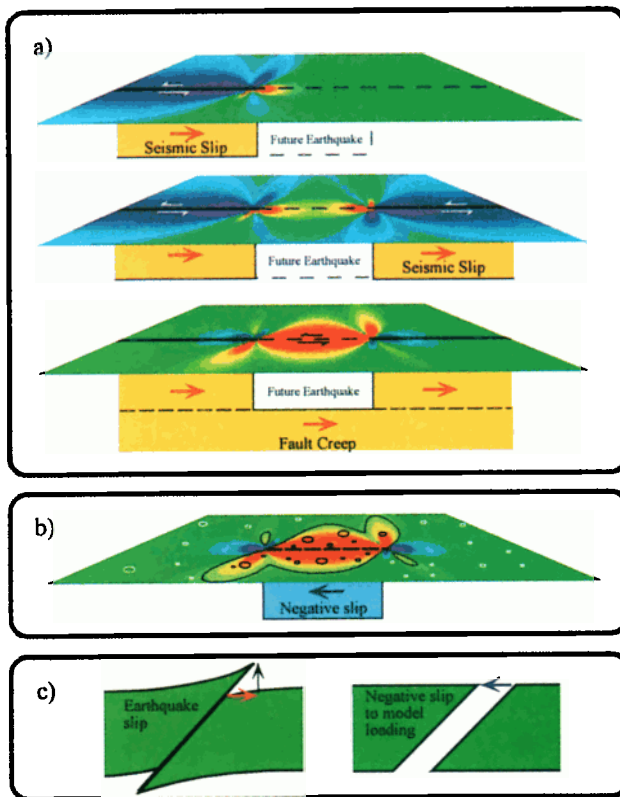
Figure 1. Evolution of Coulomb stresses prior to an earthquake. The Coulomb stress is calculated using the relations in *Ogata* [1992]. Friction is assumed to be 0.4. A regional stress oriented at 45° to the fault favors strike-slip faulting in the same sense as the main event (or antithetic to it). Red colors indicate positive static stress change, while blue colors indicate negative static stress change. (a) Perspective view of the static stress changes calculated at the Earth's surface. The fault slip at depth is indicated below the calculated stress change field. The three figures depict the cumulative stress changes due to slip on adjacent segments of the fault (top two panels) and loading by aseismic slip on the fault below seismogenic depths (bottom panel). (b) For a strike-slip fault this stress can be modeled by reversing the sense of slip observed to occur in the earthquake. Other parameters are as above. Regions where stress must have been high prior to the main event are examined for accelerated moment release. The choice of the contour bounding the region is discussed in the text. (c) Dip-slip events are associated with permanent deformation of the crust and not simple elastic rebound. In this paper, we are only concerned with the stresses from tectonic loading, which generally lie in a horizontal plane. Since only horizontal slip will release horizontal displacement boundary conditions, we can neglect vertical motion on the fault. Thus the negative of the horizontal component of slip can be modeled as an opening or closing dyke. The left side of (c) illustrates the slip and associated inelastic deformation for a thrust earthquake. The right side shows the elastic portion of this deformation with the sense of slip reversed.

a large region surrounding the fault prior to an earthquake. It is this stressed volume that we seek to define. One approach to solving this problem is to combine the historical and geological record of seismicity with the long-term loading to define regions that have been stressed by prior earthquakes [e.g. *Deng and Sykes*, 1997; *Stein et al.*, 1997; *Nalbant et al.*, 1998; *Hubert-Ferrari et al.*, 2000], as summarised in Figure

1a. However, in practice neither the regional tectonic loading nor the history and slip distributions of previous events are well-determined. An alternative approach is to calculate the stress field required to move a fault with the orientation, displacement, and rake observed for the mainshock.

This approach rests on the concept of elastic rebound, which is that the stress released in an earthquake existed as a pre-stress prior to the event (Figure 1b). In general, the stress existing before rupture on a fault of finite length can be determined by calculating the stress that results from slipping the fault backwards by the amount that it moved in the earthquake [*Savage*, 1983; *Matsuura et al.*, 1986]. The Coulomb stress field from this "negative earthquake" can then be calculated using established techniques [see *King and Cocco*, 2001]. This produces the same field that would result from an accurate knowledge of the tectonic history and loading mechanism in the region (Figure 1a). For dip-slip earthquakes, only the horizontal part of the slip is required to model the pre-stress since the motion is not purely elastic rebound (Figure 1c).

4. Methods

We now use the distribution of static (Coulomb) stress before an earthquake as calculated above to determine the regions expected to show accelerating moment release before the event. Figure 1b shows an idealised example of our methodology. The Coulomb pre-stress field is calculated based on a simple slip model of the earthquake under consideration. The pre-event seismicity is then iteratively tested for accelerating activity within a range of calculated Coulomb stress values. At each stress value, the cumulative Benioff strain of earthquakes within the stress contour (Figure 1b) is fitted to a power-law time-to-failure function and to a straight line. We define a curvature parameter as the ratio of the residuals to the power-law fit to the residuals of the linear fit. Thus, accelerating sequences will have a curvature parameter less than 1, linear sequences will have a curvature parameter of 1, and decelerating sequences will be greater than 1. The critically stressed region is then defined as the Coulomb stress contour that minimises the curvature parameter and is thus a meaningful determination of the critical region [see *Bowman et al.*, 1998 and *Brehm and Braile*, 1998]. The method for defining the critical region in this study is similar to the methodology established by *Bowman et al.* [1998], with the important difference that the region is not defined by varying the size of an arbitrary volume, but rather by searching over the range of coulomb stress values determined by the pre-stress of the impending earthquake.

We apply this method to all of the $M{\geq}6.5$ earthquakes in central and southern California since 1950. Detailed source models are not necessary when calculating the far-field stress distributions required [*King and Cocco*, 2001], thus uniform average slip distributions are adequate for all the Coulomb calculations. For each of the events in our study, a Coulomb stress field contour is found that defines the region of precursory accelerating activity (Figure 2). The parameters of the region optimisation (including the chosen contour) are listed in Table 1. Also listed is the asymptote of the power-law time-to-failure function (Eq. 1). As pointed out by *Bufe and Varnes* [1993] and *Sornette and Sammis* [1995], the asymptote time, t_c, should correspond to the time of the actual event. Examination of Table 1 shows that this is the case. Note that
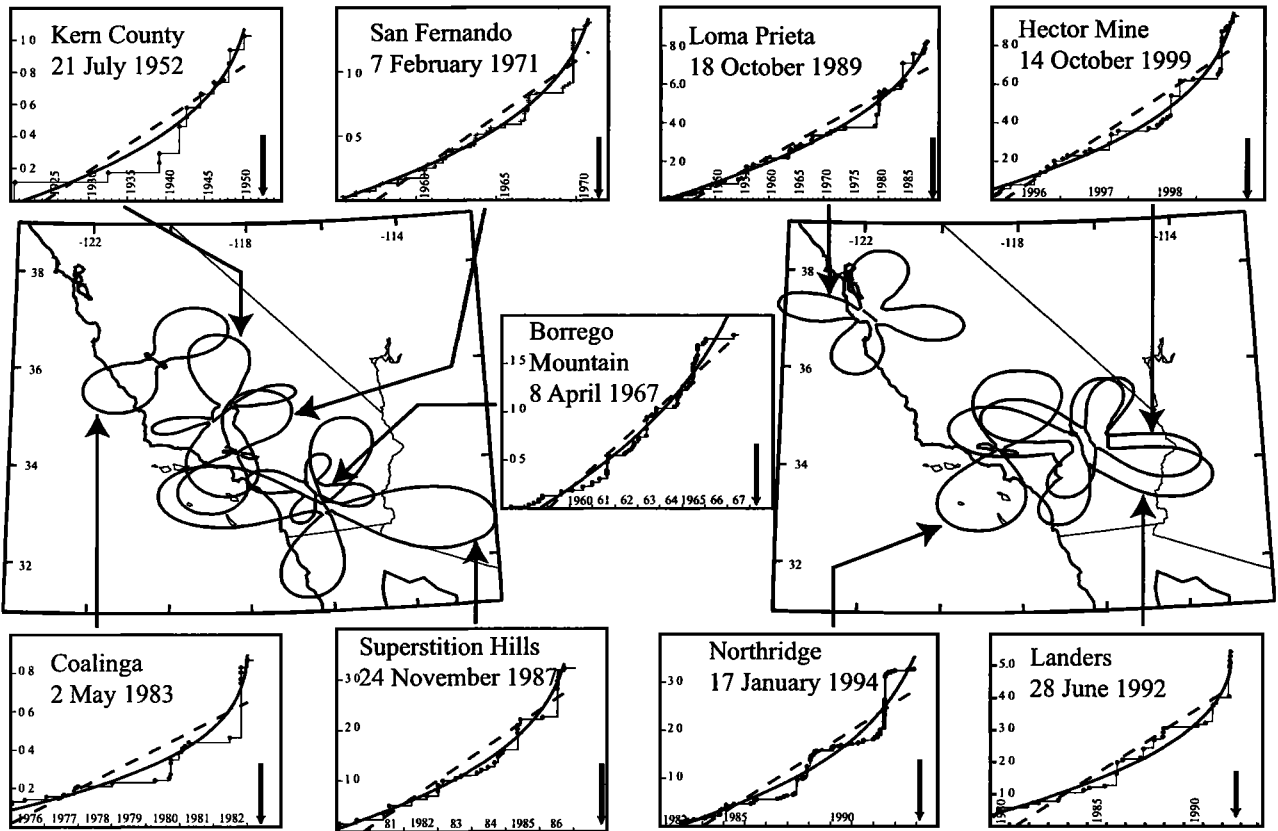
Figure 2. Critical regions and cumulative Benioff strain for all events M≥6.5 since 1950 in central and southern California. The solid curves in the cumulative Benioff strain plots are the fit of the data to the power-law time-to-failure equation, while the dashed lines are the fit of the data to a straight line (see text for details), with the time of the main event in each sequence indicated by a vertical arrow. The maps show the Coulomb stress contour used to define the critical region for each of the corresponding seismicity curves. Table 1 lists the value of the stress contour and the parameters of the curve fit for each event.

the stress change contours found using our optimisation algorithm agree well with the magnitude of static stress changes found by *Anderson and Johnson* [1999] to trigger post-earthquake seismicity.

5. Discussion

The examples that we offer suggest that the region of increased activity before a large earthquake can be identified with the area that must have been subject to an increase of stress prior to the event; the larger the event, the larger the region. Previous work that has searched for circular critical regions has found that the radius of the region scales as the magnitude of the great earthquake being "predicted" by the acceleration [*Bowman et al.* 1998; *Brehm and Braile*, 1998; *Jaumé and Sykes*, 1999]. Unlike circular regions, areas of increased Coulomb stress loading provide a natural explanation for both the widely distributed increase of seismicity before

Table 1. Summary of parameters used to model the accelerating moment release. Stress cutoff is the contour value defining the region where events are fitted to the power-law time-to-failure function. Minimum magnitude is the lowest magnitude in the catalog used in the curve fitting. The catalog is known to be complete at or below this magnitude in all cases except the Kern County event. The quality of the earthquake catalogs before 1952 is questionable.

Event	Date	Magnitude	Stress Cutoff (bars)	Minimum Magnitude	T_f	Curvature parameter
Kern County	July 21, 1952	7.5	0.05	4.6	1952.2	0.56
Borrego Mountain	April 8, 1968	6.5	0.1	3.5	1968.8	0.71
San Fernando	Feb. 9, 1971	6.6	0.01	3.5	1971.4	0.49
Coalinga	May 2, 1983	6.7	0.01	3.5	1983.0	0.49
Superstition Hills	Nov. 24, 1987	6.6	0.01	3.5	1987.0	0.39
Loma Prieta	Oct 18, 1989	7.0	0.1	4.5	1990.8	0.49
Landers	June 28, 1992	7.3	0.1	4.5	1992.4	0.36
Northridge	Jan. 17, 1994	6.7	0.005	3.5	1995.7	0.58
Hector Mine	Oct. 14, 1999	7.1	0.1	3.5	1999.6	0.47

large events and the observed scaling of the region size. Because the spatial extent of the stress perturbation caused by an earthquake scales with the moment of the event, or in this case the impending event, the mechanism described above provides a natural explanation for the observed scaling.

Finally we emphasise that the use of this technique as method of earthquake prediction should be approached with caution, since the method requires forehand knowledge of the fault or fault segments that will rupture in a future event. As recent earthquakes such as Landers, Northridge, Kobe, Hector Mine, and Bhuj have illustrated, this frequently is not the case. Using our procedure, the region of increased stress is clearly easier to identify *after* the event than *before* it. Prediction is much harder when it concerns the future. Nevertheless, our approach has allowed us to elucidate an important feature of the underlying physics of the earthquake cycle that should be incorporated into models of regional seismicity and seismic hazard.

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