Remotely Triggered Earthquakes Following Moderate Mainshocks (or, Why California Is Not Falling into the Ocean)

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INTRODUCTION

On several occasions in recent memory California has experienced apparent clusters of earthquake activity that are too far apart to be considered related according to a classic taxonomy that includes foreshocks, mainshocks, and aftershocks. During a week-long period in July 1986, California experienced the M 6.0 North Palm Springs earthquake, the M 5.5 Oceanside earthquake, and a swarm of smaller events beneath San Diego Bay. The recent M 6.0 Parkfield earthquake was followed approximately 30 hours later by the M 5.0 Arvin event, which was located well outside the traditional aftershock zone for a M 6.0 mainshock. These periods of apparently heightened activity lead to understandable consternation among California residents, who wonder if activity will build further. The recent, memorably dramatic television mini-series, *10.5*, was based on what might be considered an end-member doomsday scenario, culminating in a large part of California literally falling into the ocean. While the public did seem to recognize the gross liberties that were taken with science in this movie, old myths die hard, and seismicity maps showing activity in different parts the state are not reassuring. Neither is what used to be conventional wisdom on the part of the experts, that far-flung earthquakes are not related (even though this might remain a possibility).

Since 1992, however, scientists have come to understand that earthquakes can be related over greater distance and time scales than previously recognized. In addition to developing theories of static stress transfer (*e.g.*, Das and Scholz, 1981; King *et al.*, 1994; Toda and Stein, 2003), remotely triggered earthquakes were first identified in 1992 (Hill *et al.*, 1993) and have subsequently been observed following large (generally M > 7) earthquakes in California as well as in other regions. Investigations of "earthquake interactions" remain at the forefront of earthquake science, with many key questions still unanswered: What is the mechanism by which remote triggering occurs? What role is played by dynamic versus static stress, at small as well as large distances? Does remote triggering occur only in volcanic and geothermal regions? Does remote triggering occur following mainshocks smaller than M 7.0?

Notwithstanding these questions, our emerging understanding of earthquake interactions provides a new context for discussions with the public. In this paper I discuss seismicity following the recent Parkfield earthquake within our developing paradigm of earthquake interactions.

REMOTELY TRIGGERED EARTHQUAKES

What Do We Know?

As noted, remotely triggered earthquakes occur predominantly in active geothermal/volcanic regions, leading to theories that the earthquakes are triggered when passing seismic waves cause disruptions in magmatic or other fluid systems.

Following recent large earthquakes, triggered seismicity was observed to occur preferentially (although not exclusively; *e.g.*, Bodin and Gomberg, 1994) in regions such as Long Valley Caldera, The Geysers, and the Salton Sea region (*e.g.*, Gomberg and Davis, 1996; Stark and Davis, 1996; Prejean *et al.,* 2005). Triggering has also been observed at geothermal and volcanic sites elsewhere around the world (*e.g.*, Power *et al.,* 2001), leading some to conclude that triggered earthquakes do not occur in other seismotectonic settings (Scholz, 2003).

A number of previous studies have presented compelling evidence that remotely triggered earthquakes are caused by the dynamic stress changes associated with transient seismic waves, typically the high-amplitude *S* and/or surface-wave arrivals (*e.g.*, Gomberg and Davis, 1996; Kilb *et al.*, 2000; Gomberg *et al.,* 2004). The association of triggered earthquakes with dynamic stress changes is in contrast to aftershocks, which have been assumed to be caused primarily by local, static stress changes associated with fault movement (*e.g.*, Das and Scholz, 1981; King *et al.,* 1994; Toda and Stein, 2003). (According to convention, aftershocks are generally, albeit vaguely, assumed to be events within one to two fault lengths of a mainshock.) Recent studies (*e.g.*, Felzer *et al.*, 2003) suggest that dynamic stress changes might also play an important role in controlling the distribution of aftershocks. While both types of stress change may play roles in aftershock generation, investigations of remotely triggered earthquakes have focused only on dynamic stress changes.

Because almost all of the initial examples of remotely triggered earthquakes were in regions with active volcanic processes or shallow hydrothermal activity—both of which are associated with abundant heat and fluids at shallow depths in the Earth's crust—scientists were led initially toward triggering mechanisms that require, or are greatly facilitated by, crustal fluids such as magma and ground water. A number of theoretical investigations have proposed triggering mechanisms that involve the effects of seismic waves on bubbles within fluid systems, such as advective overpressure (Linde *et al.,* 1994) and rectified diffusion (Sturtevant *et al.*, 1996; Brodsky *et al.,* 1998). Advective overpressure describes the phenomenon whereby stress can be raised in a fluid system by rising gas bubbles; rectified diffusion describes the process whereby stress increases when a gas bubble is caused to expand and contract. More recently, Brodsky and Prejean (2004) proposed a barrier-clearing model whereby long-period waves generate fluid flow and pore pressure changes within fault zones.

In addition to the above studies, other studies describe remotely triggered earthquakes in a wide range of tectonic settings. Hough (2001) and Hough *et al.* (2003) presented evidence that remotely triggered earthquakes occurred during both the 1811–1812 New Madrid earthquake sequence and the 1886 Charleston, South Carolina earthquake. One of the supposed New Madrid mainshocks may in fact have been a triggered earthquake in the Wabash Valley (Mueller *et al.,* 2004). The results of Seeber and Armbruster (1987) also provided evidence for intraplate triggering during the 1886 sequence. Although this study talks about the "aftershocks" of the 1886 Charleston, South Carolina earthquake, the inferred locations for the events are distributed throughout the state of South Carolina, at distances well outside those of classic aftershocks given the size of the mainshock. Additionally, Hough *et al.* (2004) presented both macroseismic and early instrumental evidence that the 1905 Kangra, India earthquake was followed by a substantial (M 7+) earthquake approximately 6–7 minutes later, at a distance of approximately 150 km. The above studies suggest that triggering occurs commonly—or at least occasionally—following large (M 7+) earthquakes in intraplate settings.

TRIGGERING FOLLOWING MODERATE MAINSHOCKS?

My own recent investigations further suggest that triggering occurs commonly, albeit at low levels, following even moderate (M 5.5–7) earthquakes in both intraplate and interplate settings (Hough, 2005). In this report I highlight results from analysis of 14 recent moderate earthquakes in central and southern California, for which good earthquake catalogs are available (Table 1). Using catalogs from one month (30 days) before and after each event, I investigated seismicity changes using a standard beta-statistic approach (Matthews and Reasenberg, 1988; Reasenberg and Simpson, 1992). The beta statistic, β, is defined as

$$\beta = \frac{N_a - N_e}{\sqrt{v}} \tag{1}$$

where N_a is the number of events occurring following an event, N_e is the expected number given the premainshock seismicity rates (assuming seismicity is stationary), and v is the variance of N_e. The value of β will be large and positive in regions where seismicity increases.

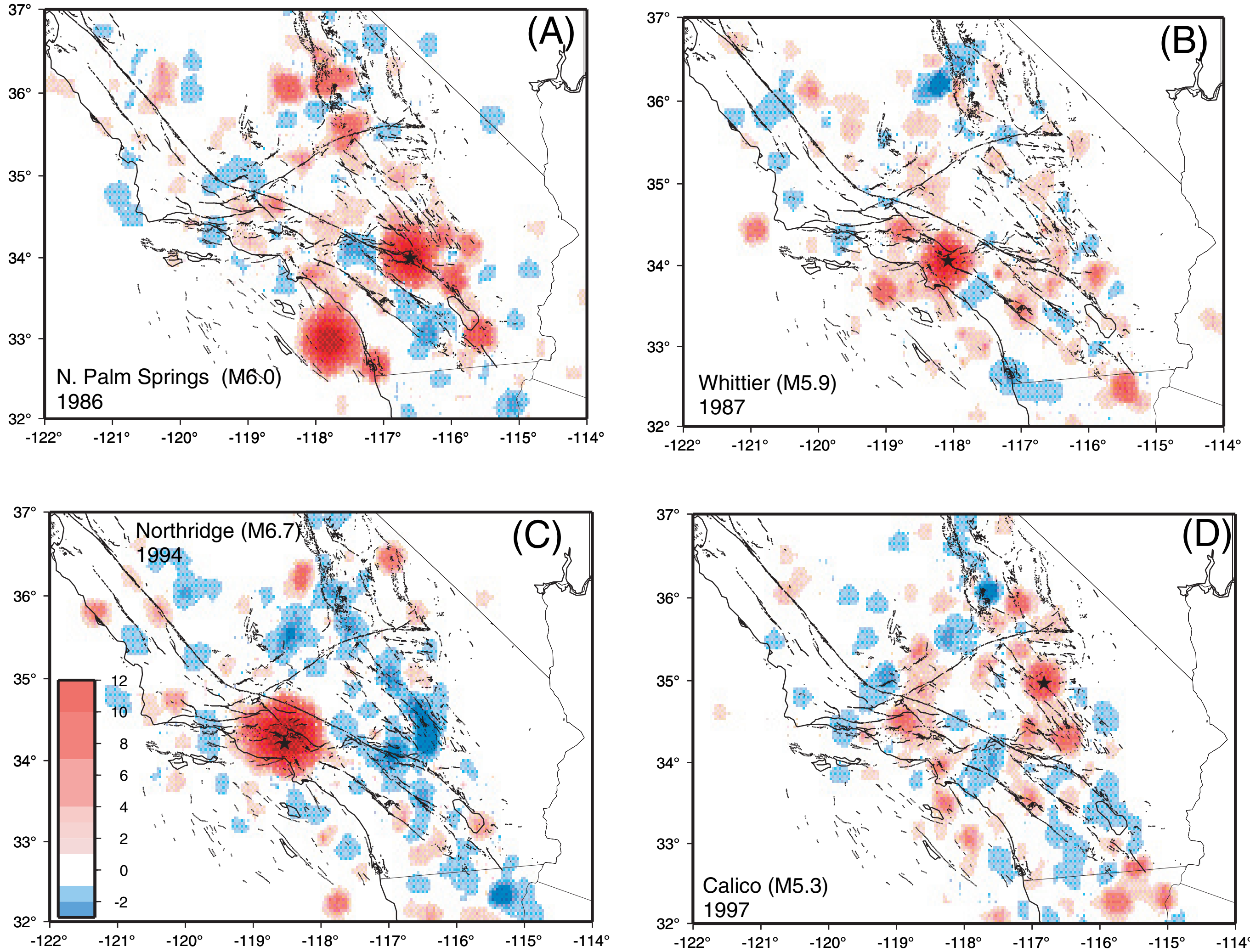
As defined by Matthews and Reasenberg (1988), β will not be equal to zero for the case that there are no earthquakes both before and after the mainshock within a given radius. Considering the expected rate, N_e, as a probability density function, a value of 4 (for example) is equivalent to a rate between 3.5 and 4.49, with uniform distribution between these limits. If there are 0 events in a pre-event window, the equivalent range of the probability density function is 0.0 to 0.49, and so N_e is set equal to 0.25. If N_a is also 0 (hereinafter referred to as the "null case"), β will be slightly negative: approximately –0.7 for the parameters used in this study.

An additional note regarding the beta statistic is that, because seismicity levels commonly fluctuate significantly, even a high value of β does not prove that a seismicity increase was caused by a preceding mainshock. Other evidence, such as a close temporal correspondence between the mainshock timing and the initiation of subsequent events, is needed to establish a causal relationship. The beta-statistic maps do not reveal evidence for widespread triggering following moderate mainshocks (Figure 1). Many of the maps do suggest a similar feature, however: an apparent seismicity increase at approximately 100 km epicentral distance, well beyond the presumed after-

TABLE 1
Recent Moderate Earthquakes in Southern and Central California Analyzed in This Study

Date	Event	M_W	Lat. (°)	Long. (°)
4/26/1981	Westmoreland	5.8	33.096	–115.625
5/2/1983	Coalinga	6.1	36.228	–120.318
7/8/1986	North Palm Springs	6.0	33.999	–116.608
7/13/1986	Oceanside	5.5	32.971	–117.874
10/1/1987	Whittier	5.9	34.061	–118.079
11/24/1987	Superstition Hills	6.2	33.090	–115.792
6/28/1991	Sierra Madre	5.8	34.270	–117.993
4/23/1992	Joshua Tree	6.1	33.960	–116.317
1/17/1994	Northridge	6.7	34.213	–118.537
8/17/1995	Ridgecrest	5.4	35.776	–117.662
3/18/1997	Calico	5.3	34.971	–116.819
10/16/1999	Hector Mine	7.1	35.702	–121.108
2/22/2003	Big Bear	5.4	34.319	–116.848
12/22/2003	San Simeon	6.4	35.647	–121.034
9/28/2004	Parkfield	6.0	35.819	–120.364

Locations and magnitudes are from SCSN/NCSN/CISN online catalogs.

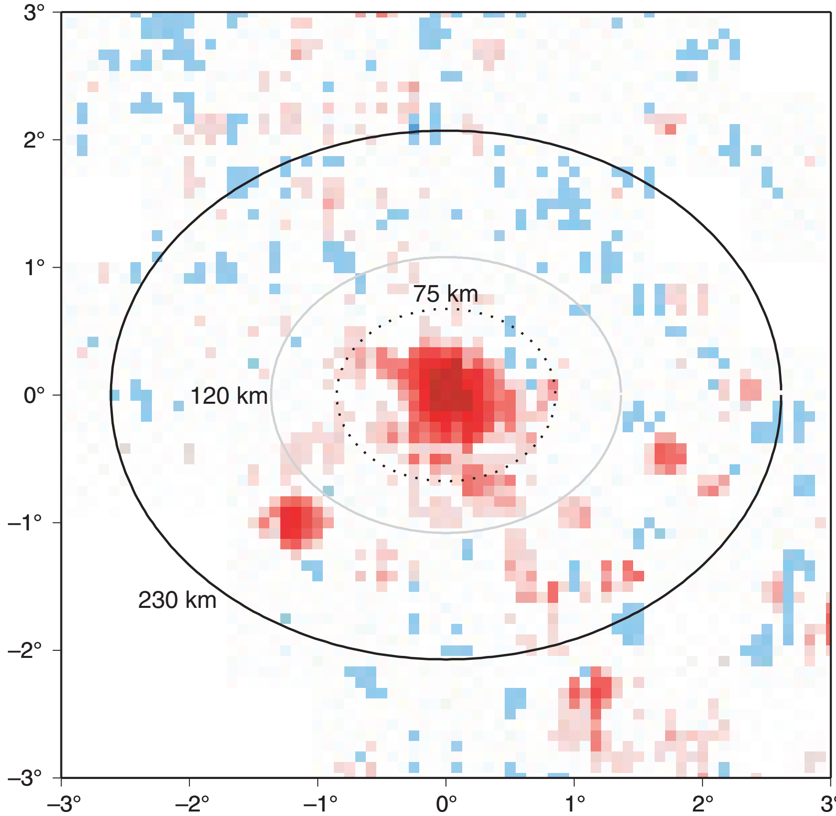


▲ **Figure 1.** Beta statistic calculated from seismicity during the 30 days following four recent earthquakes in central/southern California: (A) 1987 Whittier earthquake, (B) 1993 Coalinga earthquake, (C) 1994 Northridge earthquake, and (D) a M 5.3 earthquake near Calico in 1997. Mainshock epicenters are indicated (black stars). Scale bar shown on panel (C) indicates shading of β values between –3 and 12. Within immediate aftershock zones, β values are much higher. (Same scale is used for all four panels.)

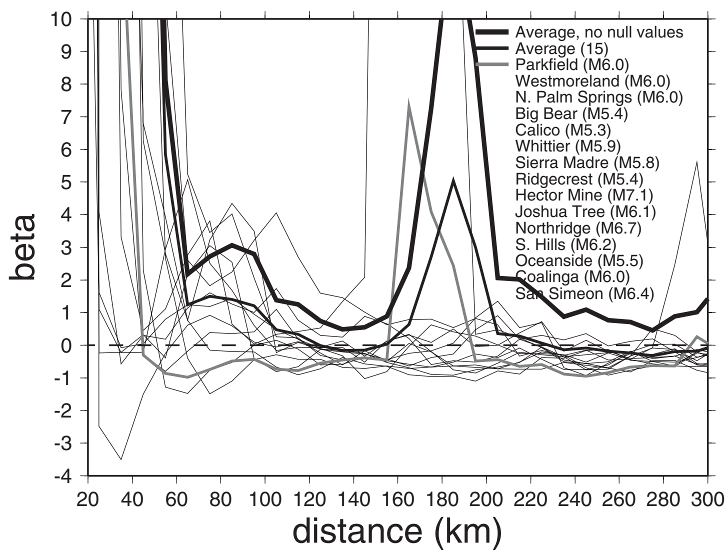
shock zone for M 5–6 earthquakes. To investigate this result further I shifted the epicenter of the beta-statistic maps from all 14 earthquakes, as well as the recent Parkfield mainshock, to 0 latitude/longitude, then contoured the combined result. The resulting image is clearly more red than blue (*i.e.*, seismicity is observed to increase) to a distance of 120 km; on average, $\beta(r)$ is generally positive to a distance of 230 km as well (Figure 2). For most of the moderate earthquakes in central/southern California, as well as the Hector Mine earthquake, β decreases outside of the immediate aftershock zone but increases slightly at a distance of 70–110 km (Figure 3). The peak of this increase is at a distance of 75 km. The very large peak at 170–180 km corresponds to the 1986 North Palm Springs earthquake; this earthquake was followed approximately five days later by the M 5.5 Oceanside earthquake, which had an energetic aftershock sequence of its own.

The inferred seismicity increase at 120–230 km is weak. The results shown in Figure 3 are influenced strongly by the null case, however, which contributes more than 70% of the overall β values. That is, in the overall results, fewer than 30% of the β values are from regions that experienced at least one earthquake during the time periods considered (either before or after the mainshock). Recalculating the average $\beta(r)$ curve using only this smaller set of results (heavy line in Figure 3), the above conclusions are more strongly supported. Interestingly, when the null case is excluded, the results reveal not only an elevated overall β level but also more pronounced peaks at both ~80 km and ~180 km.

As discussed by Hough (2005), the most straightforward explanation for the peak near 75 km is that the events are triggered by postcritical Moho reflections (*SmS* arrivals), which are known to increase ground motions significantly at a distance range of approximately 70–120 km in California. Somerville and Yoshimura (1990) showed that postcritical Moho reflections, or *SmS* arrivals, contributed to damage in the San Francisco Bay area during the 1989 Loma Prieta,



▲ **Figure 2.** Average seismicity fluctuations following 15 recent M 5.3–7.1 earthquakes in central and southern California. To generate this (Mercator-projection) map, 15 beta-statistic maps such as those shown in Figure 1 were shifted to zero origin and combined to reveal the average spatial pattern of seismicity fluctuations. The ovals correspond to three radii: (1) 75 km, the distance at which a persistent β(r) peak is centered; (2) 120 km, the distance over which seismicity clearly increases on average; and (3) 230 km, the distance over which average seismicity rates increase weakly. (Same color scale as shown in Figure 1.)



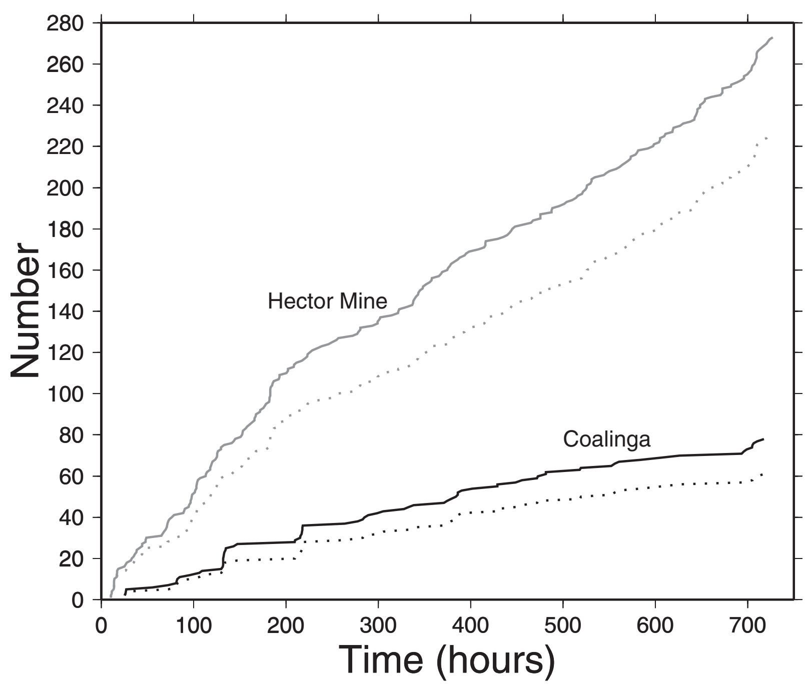
▲ **Figure 3.** Averaged seismicity fluctuation, expressed in terms of the beta statistic, as a function of epicentral distance following 15 recent M 5.3–7.1 earthquakes in central and southern California. Gray line indicates results from recent Parkfield earthquake; medium black line indicates average of 15 individual curves; heavy black line indicates average of 15 individual curves with the “null values” omitted (see text). Dashed line indicates β value of zero.

California earthquake. Somerville and Yoshimura (1990) showed that *SmS* arrivals were larger than the direct *S* arrivals at distances of 50–100 km. Mori and Helmberger (1996) showed that for some ray paths in Southern California, *SmS* arrivals are two to five times larger than the direct *S* phase. The range at which *SmS* waves appear depends, of course, on Moho depth. In Southern California, *SmS* arrivals first appear at a distance of approximately 70 km and can be larger than or comparable to the direct *S* amplitude at distances of 70–170 km (Mori and Helmberger, 1996). Although not always larger than the direct *S* wave, *SmS* arrivals are typically of high enough amplitude to increase shaking and damage during large earthquakes (Somerville and Yoshimura, 1990; Hough *et al.*, 2004). Because they are body waves, *SmS* triggering would be expected to occur anywhere along the ray path where the wave is of substantial amplitude.

Results such as those shown in Figures 2 and 3 are not as statistically compelling as the triggering that was observed following the Landers mainshock (Hill *et al.,* 1993). First, as mentioned, the increases in seismicity are small; none is individually significant and, even if one were, would not necessarily be linked to the inferred triggering event. Second, given the definition of the beta statistic, an increase in β can result from a particularly low local standard error of the background rate. For the events considered in this study, however, the beta statistic increases occur in areas where the seismicity rate increases. The inference of triggering is based on two observations: the average increase of seismicity out to a distance of approximately 230 km, and the persistence of the seismicity increases at a narrow range of distances.

The inferred seismicity increases occur within a month of the respective mainshocks. The choice of a one-month time period follows previous remotely triggered earthquake studies (*e.g.*, Reasenberg and Simpson, 1992; Gomberg *et al.,* 2004). This period represents a compromise between the need to have sufficient data to resolve seismicity fluctuations and the desire to focus on events that might be associated with a given mainshock. The simplest explanation for delayed triggering is that transient stress changes cause very early triggered events, either large or small, and these initial triggered events cause local disturbances that generate subsequent local sequences (Hough and Kanamori, 2002; Hough *et al.*, 2003). This hypothesis implies that immediate triggering occurs in locations where delayed triggering occurs. Immediate triggering will be very difficult to detect, however, unless the events are especially large. Especially in the absence of very local broadband data, it is impossible to know if triggered earthquakes occurred in these locations in the immediate aftermath of their respective mainshocks.

To explore the temporal behavior of the inferred triggered earthquakes, I considered the two earthquakes that have the largest (inferred) *SmS* signals: the 1983 Coalinga and 1999 Hector Mine earthquakes. Considering only the rates of earthquakes that occurred between 70 and 110 km of each mainshock, I found that the rates of these events do decrease with time following their respective mainshocks (Figure 4). The time decay of the (inferred) triggered events does not change substantially if one considers events between 80–110 km. (The limit of 70–80 km is meant to distinguish conventional aftershocks from inferred triggered events. Both



▲ **Figure 4.** Cumulative number of (inferred) triggered earthquakes as a function of time following the Coalinga earthquake (black lines) and Hector Mine earthquake (gray lines). The solid lines indicate temporal characteristics of earthquakes at 70–110 km distance from the respective epicenters; the dotted lines indicate earthquakes at 80–110 km distance.

distances are considered because, for the Hector Mine earthquake in particular, the choice of a value is somewhat subjective.) In effect, Figure 4 suggests that the events at 70–110 km distance "look like aftershocks" of their respective mainshocks in terms of their sequence statistics.

TRIGGERED EARTHQUAKES FOLLOWING PARKFIELD

As previous studies have pointed out (*e.g.*, Reasenberg and Simpson, 1992), it is difficult to assess the statistical significance of any beta-statistic result. Although one can infer strict confidence levels for different β values, β can increase or decrease substantially because of usual, random seismicity fluctuations that may be completely unrelated to the phenomenon under investigation. For example, the substantial increase in β at 175 km following the 1986 North Palm Springs earthquake is clearly a statistically significant fluctuation, but the β statistic alone cannot establish a causal link between the mainshock and subsequent fluctuation. Hough (2004) used a Monte Carlo approach to show that a persistent $\beta(r)$ increase at a narrow distance (*i.e.,* the inferred *SmS* signal) is very unlikely to result as an artifact.

Such analysis illustrates a conceptually obvious point: While random seismicity fluctuations can generate $\beta(r)$ increases comparable to those shown in Figure 3, such fluctuations will occur at random distances for any given event. Still, it is clear that seismicity is triggered only weakly following moderate mainshocks; the inferred *SmS* triggering is also a subtle effect.

With these caveats in mind, I now address the question of whether remotely triggered earthquakes occurred following the 2004 Parkfield mainshock. The occurrence of the M 5.0 Arvin event, which followed the Parkfield mainshock approximately 30 hours later (29 September, 22:54 GMT; 35°23.4N, 118°37.4W) at a distance of ~170 km, appeared to put the answer beyond dispute. Yet here again the question of statistical significance must be considered.

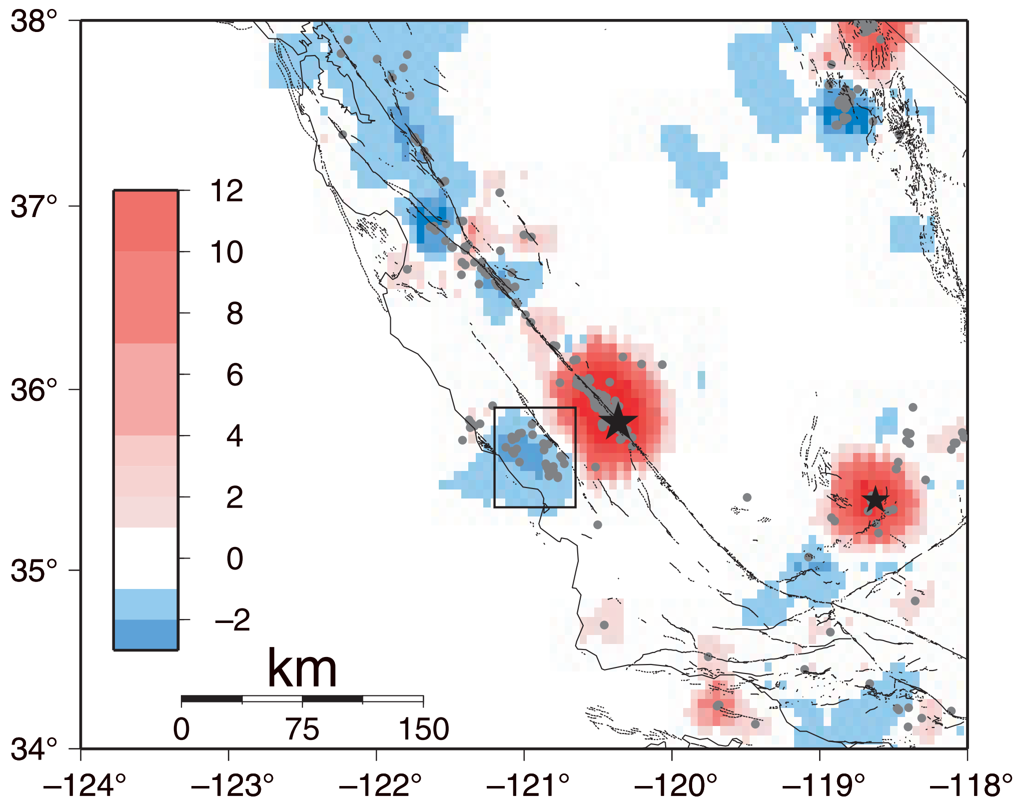
Over the past 20 years, 80 M 5 or greater earthquakes (including aftershocks) have occurred in Southern California, for a rate of 4 per year. The odds of a M 5 or larger event in a given 48-hour window are thus on the order of 2%. Excluding obvious aftershocks, the number of M 5 or greater earthquakes in Southern California is about 50, a rate of 2.5 per year. The odds of seeing an independent M 5 or larger event in a given 48-hour window is then only about 1.3%. This result is not in itself compelling, as the mainshock was chosen for analysis because it was followed by a substantial event outside of the classic aftershock zone. (I note that neither the 1934 nor the 1966 Parkfield earthquakes were followed by any events of similar size outside of their expected aftershock zones.)

Considering overall seismicity fluctuations over a two-week window before and after the Parkfield mainshock, one finds both positive and negative fluctuations (Figure 5). Interestingly, the seismicity rate decreased in the epicentral region of the 2003 San Simeon earthquake; small events did occur in this region following the Parkfield mainshock, but at a lower rate than prior to the mainshock. In fact, isolating the immediate San Simeon aftershock zone defined by (somewhat arbitrary) spatial limits 35.35°–35.9°N and 121.2°–120.65°W, one finds an apparently abrupt decrease in seismicity at the time of the Parkfield mainshock (Figure 6). The result shown in Figure 6 is reminiscent of the "toggling" of seismicity observed by Toda and Stein (2003) and, more fundamentally, the stress shadow hypothesis (*e.g.*, Harris and Simpson, 1992; Jaume and Sykes, 1996). Such an effect would presumably be related to the static stress change caused by the Parkfield mainshock in the San Simeon region, an issue well beyond the scope of this paper. (The existence of stress shadows has been debated in recent years; *e.g.*, Felzer *et al.*, 2004.) It is moreover possible that the apparent decrease in seismicity is caused by catalog incompleteness due either to difficulty locating San Simeon events during the ongoing Parkfield sequence or a backlog in data processing.

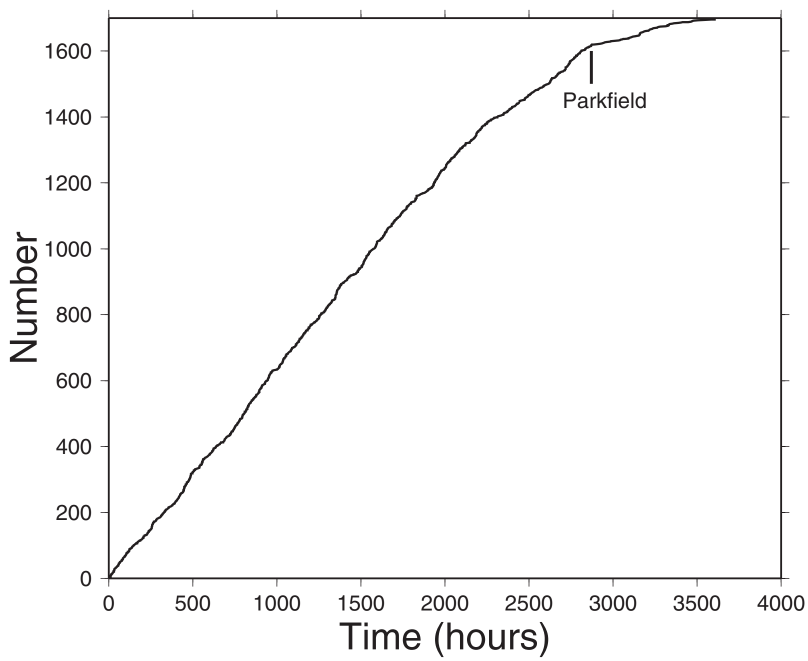
The seismicity decrease in the San Simeon region contributes negative β values at distances of approximately 40–70 km; this is shorter than the distance range of interest in this study. Considering the results at larger distances, the triggering hypothesis would still fail using the beta-statistic criteria by which remotely triggered earthquakes have been identified in the past following individual mainshocks.

As noted, perhaps the most diagnostic test of triggering is whether or not a seismicity increase commences immediately after the mainshock *S*/surface wave arrivals at a given site. No immediate triggered seismicity is apparent in filtered broadband recordings of the Parkfield mainshock at stations ISA and BAK; early small events would be difficult to detect within the coda of a large regional mainshock, however. (Stations ISA and BAK are, respectively, 33 and 44 km away from the epicenter of the Arvin earthquake.) Notwithstanding these limitations, two results support the conclusion that the 2004 Parkfield earthquake was followed by remotely triggered earthquakes beyond the immediate aftershock zone: (1) the low probability of a M 5 event occurring by random chance in a given two-day window, and (2) the fact that the Arvin earthquake occurred at a distance at which average seismicity increased following 14 previous moderate earthquakes in central/southern California.

On average, beta-statistic analysis of the 15 events discussed in this paper suggest that triggering occurs preferentially at 70–120 km, the distance range at which *SmS* arrivals are known to increase amplitudes significantly. In general, several additional factors will obviously be important for triggering: (1) overall wave amplitudes, which can be significantly dependent on directivity effects as well as mainshock magnitude, and (2) the presence of faults that are susceptible to triggering. The (inferred) *SmS* signal is identifiable only because it occurs at a predictable and narrow distance range. The results discussed in this paper further suggest that weak triggering occurs out to a distance of approximately 230 km. Two of the earthquakes analyzed in this study reveal pronounced seismicity increases at a distance of 170–180 km: the 1986 North



▲ **Figure 5.** Beta statistic calculated from seismicity during the two weeks following the 2004 Parkfield earthquake. Large black star indicates location of Parkfield mainshock epicenter; small black star indicates the epicenter of the M 5.0 Arvin earthquake; small gray circles indicate all events in the two-week window following the Parkfield mainshock. Box indicates approximate San Simeon aftershock zone (see Figure 6).



▲ **Figure 6.** Number of events in the San Simeon region (box shown in Figure 5) as a function of time (hours) following 1 June 2004. (The start date is arbitrary, chosen to provide a reliable indication of the pre-Parkfield event rate.) The time of the 28 September 2004 Parkfield mainshock is indicated.

Palm Springs earthquake and the 2004 Parkfield earthquake. Although the corresponding beta-statistic peak is high, the signal at this distance is far less persistent from event to event than the increase at 70–120 km. I thus conclude that, while weak triggering does occur to a distance of approximately 230 km, and *SmS* triggering might be expected to persist to a distance of 170–180 km (*e.g.*, Mori and Helmberger, 1996), the observations do not suggest preferential triggering *at* a distance of 170–180 km.

DISCUSSION

Investigations of "earthquake interactions" remain very much within the realm of emerging science. Any number of important issues remain open to debate, including the importance of dynamic stress changes at short (100 km or less) distances, whether triggering occurs away from geothermal/volcanic regions, and whether triggering does occur follow mainshocks smaller than 7.0. The inference of triggering following the Parkfield earthquake may also be open to debate.

The Parkfield sequence is, however, scarcely without precedent in California. In 1986 the M 5.5 Oceanside earthquake struck just five days after the M 5.7 North Palm Springs earthquake; the two events were, as noted, separated by approximately 175 km. The Oceanside earthquake was in turn followed by a smaller sequence near San Diego Bay, at a distance of about 70 km (Hauksson and Jones, 1988). In 1987 the M 6.6 Superstition Hills earthquake struck 54 days after the M 5.9 Whittier earthquake, a distance separation of about 240 km. (The M 6.2 Elmore Ranch earthquake, which occurred just 12 hours before the Superstition Hills event, can be understood within the framework of Coulomb stress change theory [Hudnut *et al.*, 1989]).

When far-flung earthquakes appear to be clustered in time there are only three possible explanations: (1) The apparent clustering is only a fluke; (2) the events are linked by a mechanism we do not understand; or (3) the events are linked by a mechanism that we do understand. Even when (3) must be followed by the caveat that we might not be able to prove the mechanism (or the link), the answer is useful in the context of public discussions. When recent seismicity maps convey the impression that California is about to fall into the ocean, it is, in the author's experience, a useful and welcome message that recent events are consistent with our emerging understanding of earthquake interactions.

Results from recent earthquakes provide the basis for additional statements to the public, albeit with qualification. Figure 2 suggests that, while earthquake sequences can "cascade" beyond the classic aftershock zone, triggering is unlikely beyond a distance of about 250 km. Also, while the statistics of remotely triggered earthquakes have not been established, recent experience certainly suggests that triggered earthquakes, as with aftershocks, are likely to be smaller than the mainshock.

Figure 2 does raise a further question: What does one call events that follow a moderate mainshock at a distance of 70–200 km? The original, somewhat ad-hoc, definition of "aftershock" specifies events within one to two mainshock rupture lengths—at most a radius of about 20 km for most of the earthquakes analyzed in this study. It is possible that this definition is too narrow. The inference of *SmS* triggering suggests, however, that the more distant events are caused by dynamic stress changes associated with transient seismic waves. It thus seems most appropriate to classify such events with the remotely triggered earthquakes that occur at greater distances.

ACKNOWLEDGMENTS

I thank Stephanie Prejean, Karen Felzer, and Ross Stein for careful and constructive reviews of the earlier, longer, version of this paper; and Ned Field, Allan Lindh, Jeanne Hardebeck, and Associate Editor Susan Bilek for their constructive (and prompt!) reviews of this manuscript. Finally, I thank Janice Sellers for patience above and beyond the call of duty.

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