Quantifying the Remote Triggering Capabilities of Large Earthquakes Using Data From The ANZA Seismic Network Catalog (Southern California)

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ABSTRACT

Various studies have examined remote earthquake triggering in geothermal areas, but few studies have investigated triggering in non-geothermal areas. We search the ANZA (southern California) network catalog for evidence of remote triggering. Using three statistical tests (Binomial, Kolmogorov-Smirnov and Wilcoxon Rank-sum) we determine the significance of the rates and timing of earthquakes in southern California following large teleseismic events. To validate our statistical tests, we identify 20 local mainshocks ($M_L \geq 3.1$) with obvious aftershock sequences and 22 local mainshocks ($M_L \geq 3.0$) that lack obvious aftershock sequences. Our statistical tests quantify the ability of these local mainshocks to trigger aftershocks. Assuming that the same triggering characteristic (i.e., a particular seismic wave amplitude, perhaps in a specific frequency band) is evident for both local and remote mainshocks, we apply the same tests to 60 remote mainshocks ($M_b \geq 7.0$) and assess the ability of these events to trigger seismicity in southern California. We find no obvious signature of remote triggering. We find minimal differences between the spectral amplitudes and maximum ground velocities of the local triggering and non-triggering earthquakes. Similar analysis of a select few of our remote earthquakes shows that the related ground motion regularly exceeds that of local earthquakes both at low frequencies and in maximum velocity. This evidence weakly suggests that triggering requires larger amplitudes at high frequencies and that a maximum ground velocity alone is not the primary factor in remote triggering. Our results are complex, suggesting that a triggering threshold, if it exists, may depend on several factors.

1. Introduction

The observation that aftershocks can occur at remote distances, up to ~17 mainshock fault lengths from the causative mainshock earthquake, has led to a significant advance in our understanding of mainshock/aftershock sequences (e.g., Hill *et al.,* 1993; Anderson *et al.,* 1994). This observation changed not only where researchers look for aftershocks, but also altered theories about the triggering mechanisms of earthquakes. Prior to the Hill *et al.* (1993) study, it was believed that static stress changes (changes associated with permanent fault offset from the mainshock rupture) were among the primary agents of aftershock generation. Theoretical models of static stress changes produced by a mainshock show regions of increased static stresses that correlate with the spatial distribution of local aftershocks (e.g., Das and Scholz, 1981; Reasenberg and Simpson, 1992; King *et al.,* 1994). In the Hill *et al.* (1993) study, however, it was noted that the remote aftershocks were far too distant for static stresses to be sufficiently large to be the primary triggering agent, leading them to propose triggering by means of dynamic deformation. Additional studies provided evidence that dynamic stress changes generated by the passage of seismic waves might trigger distant aftershocks. Dynamic stress changes, although transient, can be an order of magnitude larger than static stress changes and are thought to be a more likely triggering agent of remote aftershocks (e.g., Cotton and Coutant, 1997; Voisin *et al.*, 2000; Moran *et al.*, 2004; Pankow *et al.*, 2004; Prejean *et al.*, 2004; Brodsky and Prejean, 2005; Felzer and Brodsky, 2006).

Evidence of remotely triggered aftershocks in geothermal areas has been found following the 1999 M_W7.0 Karluk Lake, Alaska earthquake at the Katmai Volcanoes in Alaska (Power *et al.*, 2001); following the 2002 M_W 7.9 Denali, Alaska earthquake at locations such as Mount Rainier, Coso, the Geysers, and Long Valley caldera (Prejean, *et al.*, 2004); and following the great 2004 M_W9.0 Sumatra earthquake at many Alaskan volcanoes (West *et al.*, 2005). Only a few studies have found evidence of remote triggering in non-geothermal settings, such as seismicity noted along the Wasatch Fault Zone in Utah following the 2002 Denali, Alaska earthquake (Pankow *et al.,* 2004). The prevalence of triggering in geothermal areas led some to suggest that passing seismic waves generated by large earthquakes promote the movement of magma within the geothermal structure, and that this redistribution of stresses within the system subsequently triggers the aftershocks (Hill *et al.,* 2002). This theory has since been modified to include all fluids, allowing triggering in non-geothermal areas too (Brodsky *et al*., 2003; Brodsky and Prejean, 2005; Manga and Brodsky, 2006).

Although it is clear that earthquakes can be triggered at remote distances from a mainshock (and at close distances as well), there is little consensus as to how dynamic triggering (i.e., that initiated by the passage of seismic waves) takes place (Stein, 1999; Voisin *et al*., 2004; Main, 2006). Proposed models include triggering by frictional failure (either linear or non-linear), sub-critical crack growth, Coulomb failure resulting from dynamic stress changes, excitation of crustal fluids, fluid transport, and pore pressure changes including magmatic fluids (see summary in Hill and Prejean, 2007). Some of these models suggest that the amplitude of the seismic waves is most important in triggering earthquakes whereas others suggest the frequency content of the passing dynamic seismic waves also may be important (Gomberg, 1996; Gomberg, 2001; Pankow *et al*., 2004; Prejean *et al*., 2004; Pollitz and Johnston, 2006). Alternatively, it could be a more complicated mechanism that depends on the past history of stress in the region (Steacy *et al*., 2005).

In this study, we search for evidence of remote earthquake triggering in the ANZA network data collected near the San Jacinto fault zone in southern California. We focus on remote triggering because the causative deformations can more clearly be associated with seismic waves alone since the static stresses are very much smaller at remote distances. We choose the southern California data because of the robust catalog for the region (data return rate of 99% for three-component broadband data), ease of accessibility to the data, and non-geothermal environment. We use three independent statistical tests to quantitatively evaluate the triggering capabilities of remote and regional large earthquakes. We validate our tests by applying them to known local mainshock/aftershock sequences, with the assumption that the triggering mechanism attributed to local mainshock earthquakes is the same as that for remote mainshocks.

We divide our local mainshock data into two end-member models: one consisting of mainshocks that trigger observable aftershocks (hereafter referred to as 'triggering mainshocks') and the other consisting of mainshocks for which we do not observe aftershocks (hereafter referred to as 'non-triggering mainshocks'). The 'non-triggering' designation merely defines the case in which the catalog shows no significant post-mainshock changes in seismicity rate based on our analysis; this 'non-triggering' designation could be in error as it is much more difficult to prove non-existence than existence. We examine the spectral content of the triggering and non-triggering events, with the intent of identifying the waveform characteristics of earthquakes that strongly trigger aftershocks in southern California. We compare these characteristics to those of the non-triggering events to test the hypothesis that a frequency/amplitude threshold exists for earthquake triggering, and if it does, to constrain its characteristics.

2. Definitions and Data Selection

We use data recorded by the ANZA seismic network in southern California (Figures 1a-b). The ANZA network is a permanent array of three-component broadband stations that surrounds the San Jacinto fault zone and encompasses the Anza seismic gap (Berger *et al.,* 1984; Vernon, 1989). The seismic instruments in the ANZA network were upgraded to STS-2 seismometers in 1992 and early 1993 with 24-bit dataloggers installed in 1994, resulting in improved data quality. Additional instruments were installed after 1994, for a total of twenty-one stations returning data during the full period. In this study we focus on data recorded after the upgrade to STS-2 seismometers.

The ANZA catalog is created by first using automatic detection and association to determine the initial P-phase arrivals, event times, and locations with a detection threshold of M_L~1.0. An analyst reviews these picks and determines the subsequent S-phase arrivals, in addition to searching for missed events. The SCSN and NEIC catalogs are used for comparison to identify additional events. Calculated local magnitudes are based on the Richter magnitude definition. This formula may not be the most accurate for small magnitude events that have small source/station distances (Bakun and Joyner, 1984). From yearly histograms of earthquake magnitudes from 1993-2004 (i.e., examining where the magnitude bins appear to break from the Gutenberg-Richter relationship), we estimate the overall ANZA catalog completeness level is ~M_L1.5, with better coverage near the San Jacinto fault and the ANZA network.

Our local study region of interest extends from 33.0° < latitude < 34.0° and -117.0° < longitude < -116.0°, where over 22,000 earthquakes from 1993 to 2004 have been cataloged. Of these earthquakes, only 151 are above magnitude 3 and only 18 are over magnitude 4.

We categorize data in the ANZA catalog into three groups based on the epicentral distance to the ANZA network centroid (33.5°, -116.6°). These include: (1) local events that are within 0.5° (≤ ~50 km); (2) regional events that are between 0.5° and 5° (~50-550 km); and (3) remote events that are at distances greater than 5° (≥550 km). Within the local, regional and remote groups we consider 'mainshocks' greater than magnitude 3.0, 5.0 and 7.0, respectively. We define an 'aftershock' to be any local earthquake in our defined study region that occurs within a short period of time following a local, regional or remote mainshock. Though our 'local' area is larger than that assumed in most studies (i.e., can be greater than 20 fault lengths for an M3 earthquake, as compared to the more commonly used 2-3 fault lengths), we use the full region for simplicity.

3. Methods

3.1 Statistical Tests

We hypothesize that local, regional and remote mainshocks are capable of triggering seismicity. To quantitatively test this hypothesis we search for evidence of a change in the number of earthquakes within our study region following a mainshock, or a change in how the earthquakes are distributed in time. We assume that these local potential aftershocks are not caused by other factors (i.e., known prior local mainshocks or mining activity, both of which are cause for exclusion from the dataset) and that the mainshock earthquakes are otherwise random and independent. We assume that triggering depends on local deformation no matter how a primary mainshock generates such deformation (i.e., static or dynamic stress changes). Our null hypothesis is that mainshocks produce no change in the temporal distribution of local seismicity.

To quantify the ability of a mainshock earthquake to trigger a local aftershock, we apply three statistical tests to analyze changes in local seismicity: (1) the Binomial test; (2) the Kolmogorov-Smirnov test; and (3) the Wilcoxon Rank-sum test. The Kolmogorov-Smirnov test and the Wilcoxon Rank-sum test are non-parametric tests that analyze changes in the temporal distribution of seismicity prior to and following a mainshock. The Binomial test analyzes only changes in the number of earthquakes prior to and following the mainshock. (See Appendix A for a more detailed discussion of these statistical tests.)

We apply our statistical tests to catalog data that span one day on each side of the mainshock earthquake (two days total). Each test produces a p-value ranging from 0 to 1, where a low value indicates a statistically unlikely change between the local seismicity prior to and following the mainshock (i.e., triggering), and a higher value indicates no statistically significant change (i.e., no triggering). Specifically, the p-value gives the probability that the changes observed deviate from what is expected simply by chance, assuming the null hypothesis is true. We refer to the p-values as P_{Bin}, P_{KS} and P_{RS} for the Binomial, Kolmogorov-Smirnov and Rank-sum tests, respectively. For all three tests, we assume p-values less than 0.10 (the 90% confidence level) suggest earthquake triggering, and that a low p-value in any two of the three tests is sufficient evidence that earthquake triggering may have occurred. A low p-value, indicating a change in the seismicity pattern, could be due to an increase or a decrease in the number of events or seismicity rate. In this study, we focus on an increase in seismicity rate, and use our qualitative tests to confirm the statistical results we seek.

3.2 Quantifying Remote Triggering in Southern California

We test the hypothesis that regional and remote mainshock earthquakes trigger aftershocks in southern California by searching for evidence of seismicity rate changes in the local ANZA catalog. Under the null hypothesis, seismicity should not be noticeably different one day before versus one day after regional and remote mainshocks. To avoid complications from local mainshock/aftershock sequences, we do not include earthquakes in our local aftershock catalog that are caused by known factors such as other local mainshocks (M_L>~3) or quarry mining. We do not consider any mainshocks (remote, regional, or local) that occur within 24 hours following a previous mainshock, which removes 37 mainshocks. Removing events attributed to mining activities (shallow events that occur during the daylight hours in known mining areas) reduces the catalog by ~4%.

For large local earthquakes ($M_L \geq 5$) that produce higher rates of seismicity for much longer periods of time, we remove what we believe are local aftershocks from our catalog, based on when the post-mainshock seismicity rate returns to the pre-mainshock rate. Using this criterion, we remove 30 days of data following the 31 October 2001 M_L5.1 Anza earthquake (4328 events) and 30 days of data following the 26 July 1997 M_L5.0 earthquake (145 events) from our aftershock catalog. The substantial difference in number of cataloged aftershocks can be attributed to greater network coverage and more careful aftershock analysis following the 2001 event. These were the only two events over magnitude 5 that occurred in our study region during our observation time period (1993-2004).

For mainshocks of interest, we compute the power spectral density (PSD) of each seismogram using the Thomson multitaper method (e.g., Vernon, 1989) as implemented in the MATLAB© routine PMTM. We also enforce a normalization factor (not automatically included in the PMTM routine) in the multitaper result that is consistent with Parseval's theorem (e.g., Press *et al.*, 1986). We do not consider clipped waveforms in our analyses. Uniformity in data recording and our spectral computations allows us to compare the spectra of one event with those of another and test for triggering thresholds in both amplitude and frequency. We also catalog the peak ground velocity of the mainshock by identifying the maximum absolute amplitude of the seismic velocity waveform at each station; peak ground velocity is proportional to strain under a first-order approximation and this allows us to constrain changes in the local dynamic strain (Gomberg and Agnew, 1996). The peak velocity is determined after applying a 1 Hz low-pass filter to remove microseismic noise and detrending the data to remove any large DC offsets. Other studies have examined velocity measurements and found a correlation between high peak velocity and triggering (Gomberg and Johnson, 2005).

Local mainshocks will have slightly different waveform characteristics at the different ANZA stations due to propagation distance and radiation pattern variations, in addition to local site effects. We find the PSDs calculated at four ANZA stations (PFO, KNW, FRD, and RDM) do not differ much in relative values (the local triggering mainshocks have higher values than the local non-triggering mainshocks). We present spectral results for the ANZA station at Pinon Flats Observatory (PFO) alone, and median values of maximum ground velocity results as measured at the four ANZA stations.

4. Results

4.1 Using Local Earthquakes to Validate our Statistical Tests & Identify Significance Levels

We first search for qualitative evidence of local mainshock (M_L>3) triggering by looking for obvious increases in the seismicity rate following these earthquakes (local mainshocks). Our results are somewhat dependent on the spatial extent of our study area, but for simplicity we assume that we can obtain adequate results by limiting the temporal extent of the data to ±1 day with respect to the mainshock event and using our full local catalog. We categorize each M_L>3 earthquake as having one of the following: an obvious aftershock sequence, lack of an obvious aftershock sequence, or triggering indeterminate (Figures 2a-b). We define two end-member cases from the two more definitive categories (obvious triggering and lack thereof). For our local data, this yields 20 obvious triggering mainshocks and 22 non-triggering mainshocks (Tables 1a-b). The triggering (and non-triggering) events show no systematic temporal or spatial pattern.

The triggering sequences typically have a gradual increase in the seismicity rate immediately following the mainshock that then decays to the background rate within a few hours or days. The rate change curves for some triggering mainshocks more closely resemble a step function because of extra diligence taken in cataloging very small magnitude events (M_L<0.5) following these mainshocks (Figure 2a). Unfortunately, most small mainshocks are not subject to such rigorous aftershock detection. If we limit the aftershock data to only earthquakes above magnitude 1.5 (our completeness level), we see a gradual change in rate similar to the other sequences. For our purposes, we are only interested in the relative increase in local seismicity and we do not consider the shape of the seismicity rate curves.

More quantitatively, we apply our three statistical tests to our 42 (20 triggering, 22 non-triggering) $M_L \geq 3$ local mainshocks. We use all events within our local study region that have local magnitudes greater than 0.0 in this analysis to maximize the number of data used in the statistical tests. For all three tests we determine p-values from the distribution of event times within one day before and one day after each mainshock. Using the criteria that triggering has occurred if two out of three tests have sufficiently low p-values (≤ 0.10), these statistical tests conservatively confirm 11 of the 20 triggering cases and 21 of the 22 non-triggering cases (see Tables 1a-b). The verification of the non-triggering cases is encouraging. Though several of the mainshock/aftershock sequences in our triggering dataset exhibit an increase in the seismicity rate for only a few hours after the mainshock, our statistical tests extend for a full day before and a full day after the mainshock. The extended time frame of the statistical tests (one day versus a few hours) may account for the more conservative triggering result in our quantitative tests. Restricting the time period to ± 0.5 day for these local triggering events produces more positive results for triggering (an additional 6 events in the local triggering category, for a total of 17 of the 20 events tested). However, because the total number of earthquakes in these more limited tests is significantly smaller, we maintain the ± 1 day windows for all our tests. This window is more reasonable for larger magnitude mainshocks, which are typically followed by a longer period of aftershocks. We maintain a single length one-day before and one-day after time window in all cases for uniformity, and assume that for our needs, the variability in sequences for mainshocks ranging from M3 to M5 is negligible. We conclude that our quantitative statistical tests are more stringent than the qualitative analysis, and that for our purposes they adequately identify local mainshock triggering and local mainshock non-triggering earthquakes.

We next extend our quantitative tests to systematically determine statistical p-values for all local earthquakes in our catalog over magnitude 2 (over 700 events return statistically viable results). We find that many of the largest events ($M_L \geq 4$), which we expect to produce observable local aftershocks, indeed have p-values less than 0.10 (the statistical 90% confidence level) for all three tests (Figure 3). Our analysis of local $M_L \geq 2$ mainshocks suggests that the statistical tests are sensitive to the changes we hope to observe. Since the three statistical tests examine different attributes of the local seismicity, we assume that positive results (p-values less than 0.10) for at least two out of three of the tests is indicative of possible triggering, whereas a positive result for only one test could be a false positive. In the former case, for example, sometimes the local seismicity prior to and following a mainshock exhibits a significant temporal change (expressed by low p-values for the Kolmogorov-Smirnov test and the Wilcoxon Rank-sum test) without much change in number of events (expressed by Binomial test p-values which fall above 0.10).

4.2 Testing our Remote Triggering Hypothesis

We assume aftershock triggering occurs when a local threshold condition (i.e., seismic wave amplitude, perhaps at a particular frequency) is exceeded, and that the occurrence of triggering is impartial to how far away the mainshock is, such that local and remote mainshocks trigger local aftershocks via the same mechanisms. Consequently, a remote mainshock that produces triggering conditions equivalent to those of a local mainshock should result in the same local effect. We therefore use the same methods to evaluate local triggering and remote triggering.

We apply our statistical tests to a catalog of regional and remote mainshocks (72 events total, of which 60 are in the area defined as remote and 12 are regional), and look for evidence of triggering in southern California. We find that very few events (seven remote events and no regional events) produce statistical values indicative of triggering (Table 2). This suggests that either remote triggering is not that common in the Anza region of southern California, or that our analysis is unable to effectively detect remotely triggered seismicity in the ANZA catalog. Examining the empirical cumulative distribution functions of the calculated p-values for all datasets (local triggering, local non-triggering, regional and remote), we find that the distribution functions show a distinct difference between the local triggering and non-triggering groups (Figure 4). The local triggering distribution has a more sloped line which displays the higher prevalence of low p-values as compared to the local non-triggering distribution. The regional and remote distribution functions are more similar to the local non-triggering, suggesting that triggering at great distances is either not occurring, is rare, or we are unable to detect it.

The M_W 6.7 Northridge, California (17 January 1994) earthquake is one of our regional events. The large magnitude and relatively close epicentral distance between this mainshock and the ANZA seismic network (~1.75° or ~200 km) makes this event a hopeful candidate for triggering near Anza. Our statistical analysis indicates that no triggering occurred in the southern California region following the Northridge earthquake (P_{Bin} = 0.9375; P_{KS} = 0.0793; P_{RS} = 0.1905). On first glance, qualitative analysis of the seismicity rate changes suggests triggering, but closer inspection shows that a gap in data collection a few days prior to the mainshock greatly influences the slope of the rate change curve and that seismicity following the mainshock matches the longer term overall trend (Figure 2c). Qualitatively examining the seismicity within one day before and after the mainshock confirms a lack of triggering in the Anza region following the 1994 Northridge mainshock. Two regional events (28 June 1992 Landers, CA earthquake and the 16 October 1999 Hector Mine, CA earthquake) were shown to trigger remote seismicity at other locations (Gomberg *et al.*, 2001) though we find no evidence of remote triggering at Anza (Landers: P_{Bin} = 0.0037; P_{KS} = 0.3149; P_{RS} = 0.6857 and Hector Mine: P_{Bin} = 0.5806; P_{KS} = 0.6840; P_{RS} = 0.5444). We do not consider these regional events further since their ground motions exceeded the range of the seismic instruments, resulting in clipped waveforms that cannot be used to test for a triggering threshold.

In the remote earthquake catalog, we focus on four specific mainshocks: M_W8.4 Peru (23 June 2001), M_W7.9 Denali (03 November 2002), M_W9.0 Sumatra (26 December 2004), and M_W7.3 Mariana Islands (12 October 2001) earthquakes (Figure 1a). The Sumatra and Peru earthquakes are among the largest magnitude events in the ANZA catalog and were chosen for that reason. Remotely triggered aftershocks were documented in several locations following the Denali earthquake and we test that scenario here (Husker and Brodsky, 2004; Moran *et al.,* 2004; Pankow *et al.,* 2004; Prejean *et al.,* 2004). We focus on the Mariana Islands earthquake because of favorable results from our statistical analysis; it is the only remote earthquake in our catalog with $p \leq 0.10$ for all three tests.

The 2001 M_W8.4 Peru earthquake epicenter is more than 7000 km from the ANZA network. Seismic activity recorded by ANZA displays no noticeable change in the ANZA network's local rate (Figure 2c). Statistical analysis confirms this lack of remote triggering: all p-values are well above the 0.1 value we assume indicates triggering ($P_{Bin} = 0.7483$; $P_{KS} = 0.7686$; $P_{RS} = 0.9394$).

In agreement with a study by Gomberg *et al.* (2004) we find no evidence of remote triggering in our southern California study region following the 2002 M_W7.9 Denali earthquake. Again, our statistical tests do not support triggering ($P_{Bin} = 0.6682$; $P_{KS} = 0.0911$; $P_{RS} = 0.5495$), and qualitatively the seismicity rate does not appear to change (Figure 2c). Though triggering is not apparent in our southern California data near Anza, the Denali earthquake produced noticeable triggering in much of North America (Gomberg *et al.,* 2004). It is thought that the rupture directivity to the southeast played a role in triggering these remote aftershocks, and southern California may be too far to the southwest of the main directivity swath to have been strongly affected.

Quantitative analysis of data from the 2004 M_W9.0 Sumatra earthquake (the largest magnitude event in our catalog) suggests that triggering occurred in the Anza region. Two out of three of the statistical tests detect a temporal change in local seismic activity ($P_{Bin} = 0.7483$; $P_{KS} = 0.0005$; $P_{RS} = 0.0011$), and the seismicity rate appears to show a jump in activity near the local arrival time of the Sumatra mainshock (Figure 2c). However, closer examination shows that this increase in seismicity occurred prior to the arrival of mainshock energy and can likely be attributed to a much smaller local mainshock in the Anza vicinity. No change from the long-term seismicity rate is apparent. This case illustrates that our quantitative analysis identifies a change in seismic activity, but such a change might not be what we are seeking to detect.

Our statistical tests suggest that the Mariana Islands earthquake is the most likely remote mainshock to have triggered aftershocks in the southern California Anza region. This 12 October 2001 M_W7.3 earthquake was south of the Mariana Islands (epicentral distance from the centroid of our network is ~90° or ~10,000 km). Our statistical tests suggest this event may have triggered seismicity in the Anza region. P-values for all three tests are low (P_{Bin} = 0.0000; P_{KS} = 0.0283; P_{RS} = 0.0489) and indicate a significant change in both quantity and rate of local seismicity (Figure 2d). This mainshock is not as large as the others we have considered, and is not known to have triggered seismicity in other locations. It is interesting to note that a local earthquake (M3.1) occurred within an hour following the remote P-wave arrival, and the magnitude 5.1 Anza (31 October 2001) earthquake occurred only nineteen days after the Mariana Islands mainshock.

4.3 Spectral Frequency and Amplitude Analysis: Determining Triggering Thresholds

Our next goal is to compare the waveform spectra and amplitudes from earthquakes that we have deemed 'triggering' with those from earthquakes that we have deemed 'non-triggering' to determine if we can identify any obvious triggering threshold limits. We start with our local earthquake data. Comparing the spectra of the 42 local mainshocks, we find that the spectral shapes are very similar for the triggering and non-triggering categories. In addition, the median spectral amplitudes (Figure 5) and the median maximum obtained ground velocities (Figure 6) for each of the two categories display few differences. The similarity of the median spectral amplitudes extends over the full range of frequencies examined. The mean of the triggering group is skewed upwards due to the higher spectral amplitudes of three larger local earthquakes (M_L>4.5); since these events have no equivalent magnitude representation in the non-triggering category, the median is a more reliable measure. From the local data, we conclude that we are unable to distinguish a threshold for triggering with these data.

We next look at the regional data. As expected, the Northridge waveforms display high spectral amplitudes (comparable to the largest local mainshocks of M_L5+) below 2 Hz that then fall off fairly rapidly at higher frequencies. The maximum amplitude is also among the highest values of all events considered. Quantitative (statistical) and qualitative analysis of these data show no signs of triggering in the local ANZA data, suggesting that if triggering is manifested only by a mechanism apparent in amplitude of spectral observations it is likely a mechanism that has effects at frequencies above 2 Hz.

Spectral analysis of our four key remote events (Denali, Peru, Mariana Islands, and Sumatra) shows high spectral amplitudes (comparable to local M_L5+ triggering mainshocks) at low frequencies up to 0.1 Hz followed by a rapid fall-off towards background noise levels by 0.5 Hz. The maximum amplitudes are higher than those for most local non-triggering mainshocks and are comparable to or higher than the medians of both local triggering and non-triggering mainshocks (Figures 5 and 6). Our quantitative and qualitative results indicate that the Denali, Peru, and Sumatra mainshocks did not induce remote triggering in southern California near Anza. This confirms our conclusions based on the Northridge mainshock: if the triggering mechanism is evident in the spectral domain, it likely is a higher frequency (above 2 Hz) effect. The Mariana Islands mainshock is the most likely mainshock to have triggered seismicity near Anza, according to our statistical tests. Supporting the triggering nature of this mainshock, a local earthquake of M3.1 occurred within one hour following the P-wave arrival. However, spectral shape and amplitude, as well as maximum velocity measured for the Mariana Islands mainshock, do not differ significantly from the remote mainshocks that our analysis classified as non-triggering.

Our results weakly suggest that frequency content might play a role in triggering. The events that do not show any evidence of triggering near Anza (Northridge, Denali, Peru, and Sumatra mainshocks) have, in general, high maximum amplitudes and high spectral amplitudes at low frequencies. The Mariana Islands mainshock, which is the most likely triggering candidate, shows similar spectral results. However, the Mariana Islands event has lower spectral amplitudes than the Northridge mainshock and lower maximum amplitude than the other remote (non-triggering) mainshocks. In addition, the spectral amplitudes of local earthquakes (both triggering and non-triggering) exceed those of the remote earthquakes at higher frequencies (above 1 Hz). Alternatively, if observable triggering requires a specific local condition (i.e., a state of stress dependent upon the tectonic loading history, amplitude, and direction), perhaps the Mariana Islands mainshock occurred at a time more suited for triggering whereas the Northridge mainshock occurred at a time that did not.

5. Discussion

To date, the exact cause of dynamic triggering is still being debated (Harris, 1998; Stein, 1999; Steacy *et al.,* 2005). Some studies suggest large amplitude seismic waves, such as those resulting from rupture directivity effects, are the most important factor in the triggering process (e.g., Gomberg and Johnson, 2005; Felzer and Brodsky, 2006) whereas others suggest triggering has a strong dependence on seismic waves within a certain frequency band (e.g., Brodsky *et al*., 2003). Many triggering theories rely on some aspect of changes in pore pressures caused by the passage of seismic waves (Brodsky *et al.*, 2003; Brodsky and Prejean, 2005), but there is no consensus on whether a triggering threshold exists, or to what extent such thresholds might be temporally or spatially variable.

It is possible that remote aftershock triggering is present in the southern California region, but that it is difficult for us to identify in this study because the effect is not easily detected. The completeness level of the ANZA catalog, although already low at magnitude ~1.5, may not be low enough. Borehole data might be beneficial to include in future studies, because the at-depth recordings reduce the level of seismic noise that makes detection of small magnitude earthquakes challenging. Most borehole instruments in the Anza region have only recently been installed, however, and the relatively short catalog durations initially limit robustness and repeatability tests.

It is often assumed that the passage of large amplitude seismic waves causes fluid or magma movement at locations far from the mainshock event, and that the movement of these fluids (i.e., from dike openings or the growth of bubbles) might help initiate aftershock triggering (Linde, 1998; Hill *et al.*, 2002; Manga and Brodsky, 2006). If this is the case, fluid availability plays a role in the susceptibility of a region to remote earthquake triggering (Brodsky *et al.,* 2003; Brodsky and Prejean, 2005). It is possible that there are not sufficient amounts of fluids in the Anza region for this type of triggering to occur, and/or that the fault zone connectivity in the Anza region is not optimal. Since local aftershock triggering does occur in the Anza region, this would suggest that the importance of fluid applies differently to remote and local triggering. This would invalidate our assumption that remote and local triggering mechanisms are equivalent.

Our study suggests that some earthquakes in the Anza region produce local aftershock sequences but that remote and regional large earthquakes very rarely produce aftershocks in the Anza region. Perhaps this is because more complex conditions for triggering are required than those that we have considered – a possibility strengthened by the fact that there appears to be no obvious triggering threshold even for the local earthquakes. For example, triggering might be more pronounced when the stress changes (generated either locally or remotely) act to enhance the already established regional stress field (Stein, 2004; Tanaka *et al.,* 2004). The depth of the remote mainshock may also play a role in the triggering capabilities, although we are unable to test this here because the remote mainshocks we consider all occurred at similar depths (~25 km or less).

Previous deformation of the region could play a role in dynamic triggering capabilities, as ongoing deformation might produce a temporally changing triggering threshold. There might also be a time-to-failure component of remote triggering, such that the Mariana Islands earthquake may have occurred during the requisite triggering conditions whereas the Denali, Peru, and Sumatra earthquakes did not. However, to be consistent with our data, both of these scenarios would require the highly coincidental conditions of favoring triggering for many local mainshocks while inhibiting triggering for most remote mainshocks. We conclude that a strong dependence on a temporal triggering effect is possible, but that it seems improbable based on this study.

6. Conclusions

Selecting earthquakes from a large dataset (>22,000 local earthquakes, >2,500,000 waveforms and >800,000 cataloged seismic wave arrival times) recorded from 1993 to 2004 by the ANZA seismic network, we test the hypothesis that large regional and remote earthquakes can trigger aftershocks in the Anza area of southern California. We use three statistical tests (Binomial test, Kolmogorov-Smirnov test and Wilcoxon Rank-sum test) to determine if distant large earthquakes can alter local seismicity rates and distributions. We validate our tests by demonstrating that they conservatively confirm qualitative estimates of observed mainshock/aftershock triggering (non-triggering) within our local study region.

Of the 72 (12 regional and 60 remote) events that we test, only one, the remote M_W 7.3 2001 Mariana Islands mainshock, shows strong evidence (all three statistical tests have p-values of ~0.05 or less) of remote triggering in southern California. Because this is only one positive example of remote triggering, out of many tested, we conclude that remote triggering in this region either does not occur or that it is not currently detectable with our existing data sets and methods. We suggest that there might be other factors we have not considered. We find minimal evidence that our southern California study region is routinely affected by remote (distances > 5°) aftershock triggering.

Appendix A

We use three statistical tests to determine the likelihood that one earthquake triggered another. Each of these tests relies on comparing seismicity prior to and following the arrival of the mainshock energy, and the tests can be applied uniformly to both local and remote triggering. Let $X_1,\ldots,X_M$ represent local earthquake origin times prior to the mainshock arrival time in the region of interest, and $Y_1,\ldots,Y_N$ represent local earthquake origin times in the region of interest following the mainshock arrival. The data are bivariate containing a coordinate for time and a coordinate for magnitude, and the indices M and N may be random variables themselves. Under the presumption that the mainshock in no manner influenced later seismicity – the null hypothesis – it should follow that distribution $Y_1,\ldots,Y_N$ be the same as $X_1,\ldots,X_M$ as well as the distribution of N being equal to the distribution of M. Any evidence that disqualifies the null hypothesis suggests that the mainshock had some influence on the subsequent seismicity. This conclusion is referred to as the alternative hypothesis. We explore three statistical tests to determine if enough evidence exists to reject the null hypothesis or if a lack of evidence persists in which case the alternative hypothesis cannot be accepted.

Our tests use only two attributes: time of earthquake occurrence and number of earthquakes within a specific time frame. The magnitudes of local earthquakes (aftershocks) are assumed to have no influence on our analysis; future more complex studies might want to incorporate magnitude information.

Test #1: The Binomial Test

This test involves only the number attribute – it only considers the discrepancy between M and N. Let n denote the number of local earthquakes within ±1 day from a particular mainshock. Under the null hypothesis, we would expect the number of local seismic events in the one day after (or equivalently prior to) the mainshock to have a binomial distribution with probability p=0.5. To test whether this is the case, the number s of earthquakes in the one-day window after the mainshock is compared with n. If indeed the null hypothesis were true, then s should be about $n/2$, but if in fact s is significantly different from this value, then the null hypothesis is rejected.

$$P_{Bin} = 2\sum_{k=0}^{s}\binom{n}{k}p^k(1-p)^{n-k} \qquad \text{(A1)}$$

Test #2: The Wilcoxon Rank-Sum Test (aka Mann-Whitney Test)

The Wilcoxon Rank-sum test examines only the time attribute: this test determines if the relative median times of local earthquake origin times occurring prior to and following the mainshock differ. It is a non-parametric statistical test that determines how significantly observations from two samples overlap.

Let x_i represent the relative time between a prior earthquake X_i and the mainshock, and similarly let y_i represent the relative time between the mainshock and a following earthquake Y_i. Then a Wilcoxon test is performed on the x and y data to determine if there is a shift in relative time occurrence of the pre- and post-mainshock seismicity. If increased seismicity follows the mainshock, then the relative times of earthquakes following the mainshock are expected to be more concentrated around time zero (mainshock arrival time) whereas the relative times of pre-mainshock earthquakes are expected to be more uniformly distributed in time.

To perform the Wilcoxon test, we first rank the times of the local earthquakes relative to the mainshock from lowest to highest time (i.e., an earthquake occurring 4 hours after the mainshock would have a lower rank than an earthquake occurring 6 hours prior to the mainshock), The statistic, U, is calculated by

$$U = n_1 n_2 + \frac{n_1(n_1+1)}{2} - R_1 \tag{A2}$$

where n_1 and n_2 represent sample sizes before and after the mainshock, respectively, and R_1 represents the sum of the ranks in sample 1. The choice of sample 1 vs. sample 2 does not affect the result because the two values of U will be complementary. The statistic U corresponds to a p-value (P_{RS}) which can be looked up in a Wilcoxon Rank-sum table, or by using a built-in test in a statistical computation package such as MATLAB. We use the MATLAB function RANKSUM.

Test #3: The Kolmogorov-Smirnov Test

The Kolmogorov-Smirnov test determines if the data (in this case, the origin times of local earthquakes relative to the mainshock arrival) prior to and following the mainshock come from different probability distributions. The null hypothesis requires that the data are from the same distribution, and the alternative hypothesis suggests differing distributions. The reported statistic gives a quantitative comparison between the two distributions.

The Kolmogorov-Smirnov statistic, *D*, is calculated by examining the cumulative distribution functions for two samples ($\Phi_1(t)$ and $\Phi_2(t)$). *D* is the maximum deviation between the two cumulative distribution functions, given by:

$$D = \max\left|\Phi_1(t) - \Phi_2(t)\right| \tag{A3}$$

The p-value is subsequently calculated using a discrete sum over an exponential until convergence:

$$P_{KS}(D > observed) = Q_{KS}\left(\left[\sqrt{N_e} + 0.12 + 0.11/\sqrt{N_e}\right]D\right) \tag{A4}$$

where

$$Q_{KS}(\lambda) = 2\sum_{j=1}^{\infty}(-1)^{j-1}e^{-2j^2\lambda^2} \tag{A5}$$

$$N_e = \frac{N_1 N_2}{N_1 + N_2}. \tag{A6}$$

λ is the argument of Q_{KS} in equation (A4), and N_1 and N_2 are the number of data in sample 1 and sample 2, respectively.

The Kolmogorov-Smirnov test is least sensitive near the endpoints because both cumulative distribution functions must start at zero and end at one. Because of this, the Kolmogorov-Smirnov test will be more effective at giving statistical significance to changes in seismicity farther in time from the mainshock origin time. The test is more sensitive to a time shift in the occurrence of aftershocks, and less sensitive to a grouping of aftershocks close to the mainshock. We calculate the p-value for the Kolmogorov-Smirnov test using the MATLAB function KSTEST2.

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TABLES

Table 1a: Local mainshocks (M_L>3) in our study area that show qualitative evidence of local aftershock triggering. The distance refers to the epicentral distance between the ANZA network centroid and the earthquake origin. The final three columns indicate if the resulting p-value for each test is significant under our cutoff of p=0.10.

Table 1b: As in Table 1a, but for mainshocks that do not show evidence of local aftershock triggering.

Table 2: Regional and remote mainshocks that we analyze for evidence of triggering in the southern California ANZA region. Not all of the mainshocks shown in Figure 1a are present here; we have removed mainshocks that return no statistical results due to a lack of sufficient data. Columns are the same as in Table 1a.

FIGURES

Figure 1a. Map of $M_W \geq 7$ global earthquakes (small white symbols, 77 events) recorded by the ANZA network from 1993-2004. Of these, we focus on 5 earthquakes (dark spheres) that include: M_W8.4 Peru (23 June 2001), M_W7.9 Denali (03 November 2002), M_W9.0 Sumatra (26 December 2004), M_W6.7 Northridge (17 January 1994), and M_S7.3 Mariana Islands (12 October 2001).

Figure 1b. Local study region in southern California (see inset map). This map includes >22,000 earthquakes (points) recorded after 1993 by ANZA network stations (triangles; stations codes listed) and primary faults in the region (lines). Shading represents topography of the region, where light colors indicate higher elevations. Larger magnitude earthquakes ($M_L \geq 3$) are depicted with stars, where gray stars represent mainshocks that have an obvious aftershock sequence (20 events) and white stars represent mainshocks that do not have an obvious aftershock sequence (22 events).

Figure 2. Temporal behavior of select earthquakes. The upper half of each figure shows only local earthquakes (stems with 'o' symbols) and the lower half of each figure shows only non-local earthquakes (stems with 'x' symbols). The stems provide information about the time of occurrence versus the magnitude (left axis) and the monotonically increasing lines display the cumulative number of earthquakes with time (right axis). The total number of earthquakes is also noted (N). Times are shifted such that the mainshock event of interest is at time t=0. The date and magnitude of each mainshock titles each subplot; also included in the plot headings are p-values calculated for ±1-day windows around the mainshock for each of the 3 statistical tests (Binomial, Kolmogorov-Smirnov, Wilcoxon Rank-sum). (a) Examples of obvious local triggering in our southern California study region. (b) Examples of local non-triggering in our southern California study region. (c) Examples from our remote earthquake data that do not show triggering: M_W8.4 Peru (23 June 2001), M_W7.9 Denali (03 November 2002), M_W9.0 Sumatra (26

December 2004) and M_W6.7 Northridge (17 January 1994; note that there is a gap in data collection prior to the mainshock). (d) Example of a possible triggering remote mainshock: M_S7.3 Mariana Islands (12 October 2001).

Figure 3. Bar graph showing the percentage of local earthquakes ($M_L \geq 2.0$) with p-values less than or equal to 0.10 for each test. Results are binned into half-magnitude bins, with the last bin showing all earthquakes $M_L \geq 4.5$. P-values are calculated using ±1 day of data on each side of the earthquake. Values above each bar group give the number of earthquakes in the bin, for a total of 773 earthquakes.

Figure 4. Empirical cumulative distribution functions (CDF) for each group of events (local triggering, local non-triggering, regional, and remote) analyzed in this study. The distributions of p-values from (a) the Binomial test; (b) the Kolmogorov-Smirnov test; and (c) the Wilcoxon Rank-sum test are shown. There is a noticeable difference between the local triggering and local non-triggering distributions, with the triggering curve showing a much higher fraction of results with p-value below our assumed threshold limit of 0.10 (shown by the vertical grid line at $p = 0.10$). The regional and remote results fall close to the non-triggering distribution, suggesting that triggering from mainshocks at greater distances is not a prevalent feature of the ANZA network.

Figure 5. Power spectral densities (computed using the Thomson multi-taper method) of seismograms recorded at ANZA station PFO. Data from the vertical component is shown. The medians of the local triggering and non-triggering spectra are plotted as a thicker line in the corresponding color. Background seismic noise is shown in yellow; noise spectra were calculated using at least 5 seconds of data ending 5 seconds before the P-wave arrival (longer noise windows were used for the remote mainshocks).

Figure 6. Median peak ground velocity as a function of median source/station distance for each mainshock, as measured on the vertical component at ANZA stations PFO, KNW, RDM, and FRD. Error bars show the extrema of the measurements. The local triggering events generally reach higher maximum ground velocities than the local non-triggering events, though they are not significantly different. The horizontal lines extending over the local region show the median peak velocities for all local triggering and all local non-triggering mainshocks.