Is the Aftershock Zone Area a Good Proxy for the Mainshock Rupture Area?

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Abstract The locations of aftershocks are often observed to be on the same fault plane as the mainshock and used as proxies for its rupture area. Recent developments in earthquake relocation techniques have led to great improvements in the accuracy of earthquake locations, offering an unprecedented opportunity to quantify both the aftershock distribution and mainshock rupture area. In this study, we design a consistent approach to calculate the area enclosed by aftershocks of 12 Mw≥5.4 mainshocks in California, normalized by the mainshock rupture area derived from slip contours. We also investigate the Coulomb stress change from mainshock slip and compare it to the aftershock zone. We find that overall, the ratios of aftershock zone area to mainshock rupture area, hereinafter referred to as "aftershock ratio", lie within a range of 0.5 to 5.4, with most values larger than 1. Using different slip inversion models for the same mainshock can have a large impact on the results, but the ratios estimated from both the relocated catalogs and Advanced National Seismic System (ANSS) catalog have similar patterns. The aftershock ratios based on relocated catalogs of Southern California fall between 0.5 and 4.3, while they exhibit a wider range from 1 to 5.4 for Northern California. Aftershock ratios for the early aftershock window (within 1 day) show a similar range but smaller values than using the entire aftershock duration, and we propose that continuing afterslip could contribute to the expanding aftershock zone area following several mainshocks. Our results show that areas with positive Coulomb stress change scale with aftershock zone areas, and spatial distribution of aftershocks represents stress release from mainshock rupture and continuing postseismic slip.

Introduction

Beginning in the 1930s, scientists believed that aftershock zone area corresponds to the area where strain is accumulated and released during an earthquake sequence (Utsu, 1969). However, large uncertainty in earthquake location and the lack of slip models in early studies prevented scientists from verifying this hypothesis (Das and Henry, 2003). With the advent of modern slip models and large increase in the amount of seismic data, studies have qualitatively examined the aftershock distribution of large earthquakes with respect to mainshock slip areas. They found that early aftershocks (within the first 24 hours) tend to occur on the periphery of the eventual aftershock zone, and the region in the center of the zone corresponds to the extent of the coseismic rupture area (C. Mendoza and Hartzell, 1988; Dietz and Ellsworth, 1990; Oppenheimer, 1990; Das and Henry, 2003). However, in spite of the wealth of literature on aftershock distribution, there has been a dearth of studies that quantify the relationship between the aftershock zone area and the mainshock rupture area, both immediately after the mainshock and over time. Recent developments in seismological techniques have led to great improvements in the accuracy of earthquake locations and finite fault solutions, and provided an unprecedented opportunity to characterize the aftershock zone and the mainshock rupture using the latest earthquake catalogs and slip models.

A better characterization of aftershock distribution is also the foundation for understanding the underlying mechanisms of aftershock generation. Different mechanisms have been used to explain the differences in aftershock distribution: static stress change, postseismic deformation, and transient dynamic stress change (Freed, 2005). Transient dynamic stress change can be used to account for earthquakes up to thousands of kilometers away (Belardinelli et al., 2003; Van Der Elst and Brodsky, 2010; Parsons and Geist, 2014; Fan and Shearer, 2016), but due to the large distances involved it is not relevant to this study. Static stress change is the stress change in the earth’s crust surrounding the fault planes due to slip on the faults (King et al., 1994, etc.). In particular, Coulomb stress change became popular in the past few decades with numerous studies attempting to correlate static Coulomb stress change with aftershocks (King et al., 1994; Stein et al., 1997; Hardebeck et al., 1998; Toda et al., 1998, 2011; Kilb et al., 2002; Marsan and Lengliné, 2010, etc). Many of these studies found that the distribution of the aftershocks appears to be co-located with regions of positive Coulomb stress change.

However, not all studies agree that static stress change is the only predictor of aftershock distribution, especially the temporal evolution of aftershocks (Cattania et al., 2015). By investigating over two hundred slip inversions, Meade et al. (2017) found that other stress change components such as max shear stress and stress invariants, or postseismic deformation such as afterslip/postseismic relaxation may account for the spatial distribution of aftershocks better. Both afterslip and viscoelastic relaxation have been shown to be able to explain aftershock distribution (Pollitz et al., 1998; Diao et al., 2014). For example, Perfettini and Avouac, (2004, 2007) found that the aftershocks of the 1999 Chi-Chi earthquake and the 1992 Landers earthquake correlate well with afterslip in both space and time. Savage et al., (2007) concluded that fault creep alone is insufficient to explain the postseismic deformation and aftershocks of those earthquakes, thus viscoelastic relaxation has to be considered.

In this study, we design a consistent approach to calculate the aftershock zone area and estimate its relationship with the mainshock rupture area for both early aftershocks and the whole aftershock duration. This approach applies the beta statistic (Matthews and Reasenberg, 1988) to provide a more universal criterion to estimate the aftershock duration and aftershock boundary, which are key factors in the calculation of the aftershock zone area and were previously determined upon certain empirical assumptions. We select recent moderate to large ($M_w \geq 5.4$) mainshocks in California that have both relocated aftershocks and resolved slip inversion models (Figure 1). To further understand the aftershock generation mechanism, we also compute the static Coulomb stress change from mainshock slip and compare it with the aftershock zone area. Meanwhile, we compile published works on afterslip models and discuss their roles in modulating the aftershock distribution.

Data and Method

We choose 12 moderate-to-large (Mw≥5.4) earthquakes in California as candidate mainshocks (Figure 1). The slip inversion models are obtained from the Finite-Source Rupture Model Database (SRCMOD) (see Data and Resources Section). We use both relocated catalog (either double-difference or waveform relocated) and Advanced National Seismic System (ANSS) catalog for following analysis. The relocated catalogs generally have better relative locations. However, a certain percentage of earthquakes might be dropped during the relocation process and potentially affect seismicity rate estimation. The ANSS catalog includes all archived earthquakes but endures relatively larger location error. We include both catalogs to evaluate the consistency. More specifically, the double-difference catalogs are acquired from the Northern California Earthquake Data Center (NCEDC) (Waldhauser and Schaff, 2008; Waldhauser, 2009). Waveform relocated catalogs are obtained from the Southern California Earthquake Data Center (SCEDC) (Hauksson et al., 2012). We download earthquakes that occurred up to 1 year before and after each mainshock within the surrounding area. We use an area that is deliberately much larger than needed to avoid creating an artificial upper limit when calculating the aftershock zone area. A grid of ±1-degree latitude and longitude relative to the mainshock epicenter is used to download earthquakes from NCEDC/SCEDC, while a circle with a radius of five times the source dimension is used to download earthquakes from the ANSS catalog.

Earthquake Selection

To choose earthquakes associated with mainshock faults, only earthquakes with off-fault distances less than 2km to the fault plane from the slip inversion are kept for further analysis. We use 2km because earthquake epicenter location uncertainties typically fall within 2km. We have tried different off-fault distances from 1 to 20km and find that off-fault distances below 5km do not show a large difference. As hypocenter locations given by the slip inversion data and the catalogs are slightly different, we shift earthquake locations in the catalog using the hypocenter in the slip inversion as a reference for some mainshocks. This ensures that the selection of earthquakes by off-fault distance is accurate and does not affect the calculation of the aftershock zone area. The fault planes are extrapolated past each end, and the earthquakes are then projected onto the nearest fault plane (i.e., smallest fault-normal distance).

Magnitude of Completeness

To remove bias in calculating the change in seismicity rate, we need to ensure that the catalog is complete for both the periods before and after the mainshock, i.e. there are no missing earthquakes for the magnitude range we use. Hence, we calculate the magnitude of completeness (Mc) for both time periods, and only earthquakes with magnitudes above the larger Mc are used. In a few cases, Mc cannot be calculated for either before or after the mainshock, due to the sparsity of data or the shape of the magnitude-frequency distribution (MFD). Hence, we use Mc of the time period that can be calculated instead. The most straightforward way of calculating Mc is the maximum curvature method, which often underestimates Mc for gradually curved bulk MFDs. The Mc95 and Mc90 methods, which calculate the lowest Mc value that gives a best fit of 95% and 90%, provide a closer estimate, but sometimes Mc cannot be calculated when the MFD curve never reaches a 90% fit. Hence, we use the best combination method, whereby an initial estimate is calculated using the max curve, and then the algorithm searches for the Mc95 value and Mc90 value in a fixed range around the estimate. These methods are described thoroughly in Mignan and Woessner (2012), and we use the open-source MATLAB code written by Schorlemmer and Woessner (2004) to calculate Mc (see Data and Resources Section). We set the magnitude bin size to be 0.1 and do not apply any correction.

Beta Statistic: A measure of change in seismicity rate

The β-statistic quantifies seismicity rate change based on the difference between the observed and expected number of events occurring in a time period, normalized by the standard deviation of the expected value (Kilb et al., 2000; Aron and Hardebeck, 2009). The standard deviation is calculated by assuming a binomial distribution where earthquakes occur either inside or outside the time period Ta (Matthews and Reasenberg, 1988). A β value of 2, which means 95% significance of increase in seismicity when the β value is normalized by its standard deviation, is used as the threshold to determine if there is a significant increase in seismicity. The equation to calculate the β value is shown below (Equation 1).

$$\beta = \frac{N_a - N * T_a/T}{\sqrt{N(T_a/T)(1 - T_a/T)}} \qquad (1)$$

Where Na is the number of events in the time period of interest, N is the number of events in the entire time period, Ta is the duration of the time period of interest and T is the duration of the entire time period (background window duration Tb plus above defined Ta).

Defining Aftershock Zone

Determining aftershock zone area is difficult as aftershocks can occur over a large and continuous area especially in places with high background seismicity such as Parkfield, and deciding which earthquakes constitute aftershocks can be quite challenging. Methods used by previous studies include fitting ellipses (Utsu, 1969), drawing energy contours (Tajima and Kanamori, 1985), terminating the aftershock zone based on gaps between the earthquakes (Meng and Peng, 2016), or drawing a simple boundary around the aftershocks (Sykes, 1971). We use the β-statistic to calculate the aftershock zone area by defining the aftershock zone as the region with significant increase in seismicity rate after the mainshock. To find the aftershock zone, we create a grid for the fault plane and the surrounding regions and calculate the β value for each grid cell. A convex boundary is then drawn around those areas with significant change in seismicity using the MATLAB function “boundary” with a ‘shrink factor’ of 0, which is consistent with results using Delaunay triangulation to denote the boundary. The area enclosed by the boundary is then taken to be the aftershock zone area. Another possible method of calculating the aftershock area is to add up area of cells with significant seismicity rate increase. However, we choose not to use this method mainly because the aftershock zone area increases with cell size, which could be subjective to provide a consistent way to estimate aftershock zone area spanning different magnitudes (Figure S1). In comparison, drawing a boundary around the aftershocks is a robust way to define the aftershock zone area that is largely unaffected by cell size (Table S1). Figure 2 illustrates the calculation of the aftershock zone area. Previous calculations of β values have used different cell sizes such as 2 km (Aron and Hardebeck, 2009) and 6 km (Kilb et al., 2000). As earthquakes are represented as points in the grid, the choice of cell size has an impact on the β values. Using different cell sizes that range from 1 to 4 km, we find that as long as the cell size is large enough such that each earthquake is not isolated, the pattern of β values remains similar. However, a larger cell size like 6 by 6 km (Kilb et al., 2000) is not appropriate as it is close to the rupture length of the mainshock, which ranges from 9 to over 100 km in our analysis. Hence, we use a cell size of 2 by 2 km. We locate the areas where the β value is larger than 2 and terminate the aftershock zone area when there is a gap larger than 15 km (Meng and Peng, 2016). We test a range of gap sizes from 5 to 20 km and find that for small off-fault distances (≤5 km), and the gap size does not affect the results.

The choice of time periods T and Ta (as in Equation 1) can greatly affect the calculation of the aftershock zone area by controlling the number of earthquakes that constitute change in seismicity rate. To estimate the background seismicity rate, we adopt a long-term averaged rate before the mainshock. Previous studies reported obvious increasing foreshocks before some large earthquakes (Dodge et al., 1995; Hauksson et al., 2002, etc). However, the short-term foreshock activity should not significantly impact our calculation since we use a much longer window before the mainshock. To test this, we use background window lengths of one year and two years and find that the ratios are generally consistent except for the Whittier Narrows and North Palm Springs earthquake (Figure S2). The range of ratios also remain the same using both pre-shock windows. Since using a pre-shock duration of one year generates more consistent results between the relocated and non-relocated catalogs, we use a pre-shock duration of one year to calculate β values and aftershock zone areas.

Defining Aftershock Duration

Scientists have previously pointed out that there was no formal agreement on a consistent space-time windowing algorithm to select aftershocks (Knopoff et al., 1982), which is still true to this date. The choice of aftershock duration is complicated by the fact that the aftershock zone area could expand with time (Tajima and Kanamori, 1985), and different mechanisms could tangle together with longer durations. Different aftershock durations ranging from one day (Kanamori, 1977), weeks (Wetzler et al., 2018) to years (Parsons, 2002; Perfettini and Avouac, 2007) have been used, depending on the need of each study. Some studies suggested that earthquakes may still have aftershocks decades or centuries later (Bouchon et al., 2013; Ebel and Chambers, 2016). Here we define aftershock duration as the time period when there is still a significantly elevated rate of seismicity in the region. We then calculate the sliding-window β value for the entire faulting system (fault plane and the extended regions) using the aftershocks within an off-fault distance of 2km, with N in equation (1) equal to all the earthquakes that occurred in the region and Ta equal to 10 days after the mainshock. We then slide the time window with a time interval of 5 days and track the evolution of the β value through time. The entire aftershock duration is given by the first time-window when the β value drops below the threshold value of 2. The aftershock duration gives Ta, the time period of interest used in the calculation of the β value in each grid cell. The aftershock duration can vary between a few weeks and over a year (Figure 3).

Aftershock Ratio

The coseismic rupture area is defined as the area enclosed by a contour of 0.15 of the maximum slip (Wetzler et al., 2018). A slip contour is used because areas with very low slip may not be well resolved and depend greatly on the smoothing method used in the kinematic source inversion.

Hence, we also estimate the uncertainty from the slip models by calculating rupture areas with slip contours of 0.1 and 0.2 of maximum slip. We then calculate the ratio of the aftershock zone area to the coseismic rupture area to investigate how well the aftershock zone area approximates the rupture area. Since each earthquake model is unique, some of them require special processing procedures as listed in Table S2.

Coulomb Stress Change Area

In order to examine how the mainshock slip impacts the aftershock zone area, we utilize the Coulomb 3 software to calculate the resulting Coulomb stress change of each earthquake (King et al., 1994). We use the entire slip model and the orientation of the main fault plane as the receiver fault to find the Coulomb stress change of the region. Assuming that earthquakes below a certain off-fault distance lie on the same fault plane as the mainshock, we use the orientation of the main fault plane as the receiver fault to find the Coulomb stress change of the region. We also use a friction coefficient of 0.6, although faults have a large range of plausible values between 0 to 0.75 (King et al., 1994). The cross-section of the fault and its surrounding region are calculated with a cell size of 1 by 1 km. We test thresholds of 0.1 and 1 MPa and find that both will result in a similar trend, but the area enclosed by the 1 MPa cells are more similar to the aftershock zone area observed from the β-statistic. Hence, we sum the area of the cells that have a positive Coulomb stress change of 1 MPa or more to compare with the aftershock zone area (Figure 4). The results of our Coulomb stress change calculations are listed in Table S3.

Results

We analyze a total number of 12 Mw≥5.4 California mainshocks (Table 1), with 3 events from the NCEDC double-difference catalog and the rest from the SCEDC waveform relocated catalog. Most of them are strike-slip events, except for the 1994 Northridge earthquake with a thrust mechanism and the 1989 Loma Prieta earthquake with an oblique mechanism. We calculate the aftershock ratio of each mainshock, often for multiple slip inversion models (Figure 5). The parameters that we used are summarized in Table S4. We also list the data types used by each slip model in Figure 5. Strong ground motion data are predominantly used for Northern California (NC) earthquakes, while various data types are used for Southern California (SC) earthquakes.

We find that aftershock zone areas are within a range of 0.5 to 5.4 times of the mainshock rupture area (Figure 6). Some earthquakes such as the 1989 Loma Prieta earthquake have consistently higher ratios, while others such as the 1999 Hector Mine earthquake and the 1994 Landers earthquake have consistently lower ratios. We explore their potential causes further in the Discussion section. Ratios of the same earthquake estimated from different slip inversion models can vary widely. For example, the ratio for the Gallovič (2016) model of the 2014 South Napa earthquake is more than 3 times of the ratio for the Wei et al. (2015) model. This is partially because the Wei et al. (2015) model has a slip area that is twice as large as the Gallovič (2016) model. The two slip models also assume significantly different fault planes. Since only earthquakes within 2km of the fault planes are included as potential aftershocks, the aftershock zone area estimated for Wei's model is smaller than that for Gallovič's model. Our results also show a similar pattern between the ratios estimated from the ANSS and relocated earthquake catalogs. Table 2 shows that both types of catalogs have almost identical average ratios, but the ANSS catalog has larger variance. We also note that the ratios for 2012 Brawley and 1987 Elmore Ranch earthquakes differ by a factor of 2 across the relocated and ANSS catalogs. The similar ratios estimated from different catalogs demonstrate that aftershock zone area is a macroscopic source feature that is not sensitive to the differences of earthquake locations in catalogs. We do not observe a clear correlation between moment magnitudes and aftershock ratios either (Figure S3).

Early Aftershock Zone

Above results are based on the entire aftershock duration. Since different aftershock generation mechanisms could affect the long-term aftershock evolution, we also measure ratios using only early aftershocks to exclude postseismic deformation if existed. The early aftershock window is set as 1-day after the mainshock (Kanamori, 1977) and the results are shown in Figure 7. Since the Ta/T term is close to 0 in our study, every single earthquake in each cell after the mainshock would be significant in the output beta value. Hence, only the off-fault distance, gap size and Mc are used to determine which aftershocks to include in the analysis. Generally, the ratios for the 1-day duration are smaller than or equal to those for the entire aftershock duration. But the range of ratios (0.5-3.5) is comparable to the range for the entire aftershock duration (0.5-5.4). The early aftershock ratios are listed in Table S5 and statistics of the ratios are shown in Table 2.

Static Stress Change

If aftershocks are primarily triggered by the Coulomb stress change, they should occur within the area with the positive Coulomb stress change. Hence, we compare the positive Coulomb stress change area and the aftershock zone area of each mainshock (Figure 3). We find that the Coulomb stress change area shows a positive correlation with the aftershock zone area (Figure 8), which may support the hypothesis of static stress change being a triggering mechanism for aftershocks. However, this correlation does not necessarily mean causation. Alternatively, the correlation may indicate that both static stress changes and aftershock areas are related to certain mainshock source parameters. The correlation between the Coulomb stress change and the aftershock zone area is not an ideal linear trend either, and the discrepancies may be due to the uncertainty from coseismic slip models and the inclusion of other aftershock triggering mechanisms. In particular, the Loma Prieta aftershock zone area appears to be an outlier, as its aftershock zone area is much larger than that of the other earthquakes given its relatively small Coulomb stress change area. We also show the ratio of Coulomb stress area to aftershock area with magnitude (Figure 9), and the results are inconclusive, with either a slight increase or no change in ratio with magnitude depending on the fitting method used.

Discussion

The value of the aftershock ratio can help reveal important characteristics of the region such as aftershock mechanisms in play, but it cannot be used to precisely predict the mainshock rupture area. One should be careful about drawing conclusions about the rupture area from the aftershock zone area, as (1) there is a broad range of aftershock ratios, (2) even if the ratio is close to 1, the spatial distribution of the aftershocks with respect to the mainshock rupture area may still be different and (3) the aftershock ratios are based on California earthquakes and likely to differ even more for earthquakes outside California. In this section, we discuss the potential causes of large variations in aftershock ratios, including different aftershock mechanisms and uncertainties introduced during our calculations. To further test the method, we also conduct similar analysis on the 2011 Virginia earthquake outside California. Given the large difference in crustal structure between California and the eastern U.S., the Virginia earthquake serves as an extreme contrast to California earthquakes.

Comparison to a non-California earthquake: An example from the Eastern U.S.

We analyze the 23 August 2011 M5.7 Mineral, Virginia earthquake using the slip model from Chapman (2013) and double-difference relocated earthquakes from Wu et al. (2015). We do not use the ANSS catalog that has very few earthquakes over two years (~54). Since the double-difference catalog only has earthquakes starting from 25th August 2011, we cannot calculate aftershock duration, 1-day aftershock zone area or Mc of the background seismicity. Here we assume an aftershock duration of one year and obtain an aftershock ratio of 14.1 for aftershocks within an off-fault distance of 2 km (Figure S4). The ratio of the aftershock zone area to the Coulomb stress change area is 14.3. Since the result is subject to the unknown Mc and uncertainty in aftershock duration, we also investigate their effects by varying Mc in the range of 0 and 1 (the catalog has an Mc of 0) and changing aftershock durations between 5 and 365 days. We find that the aftershock ratio is still much larger than the ratios we obtained for the Californian earthquakes.

The aftershock ratio of 14.1 for the Virginia earthquake is ~3–28 times as large as the ratios we obtained for Californian earthquakes, and the disparity cannot be explained by Coulomb stress change. One possible reason for the large aftershock ratio may be because of the unusually high stress drops of the mainshock (Wu and Chapman, 2017). The high stress drop may have led to a smaller rupture area than California earthquakes with the same magnitude, while the aftershock zone area remains typical for a M5.7 mainshock. This example suggests that aftershock ratios may depend heavily upon tectonic environment and the resulting differences in earthquake characteristics.

Afterslip: An alternative aftershock mechanism

As Coulomb stress change cannot satisfactorily explain the large aftershock zone area of the Loma Prieta mainshock, an alternative mechanism for aftershock generation is afterslip following the mainshock. The variation of geologic conditions in California results in different amounts of afterslip for each earthquake. The central part of the San Andreas Fault exhibits large amounts of aseismic creep (Khoshmanesh and Shirzaei, 2018), whereas the southern portion is locked with significant slip deficit (Fialko, 2006). Though the underlying reason is not well known, some studies suggest that it might be due to the presence of serpentinite at creeping faults in Northern and Central California (Moore and Rymer, 2007). Studies have shown that the Loma Prieta earthquake has afterslip extending around 40-60km towards the southeast along the San Andreas fault (Behr et al., 1990; Pollitz et al., 1998). The shallow afterslip (above 15km depth) was found to have most likely occurred on the Loma Prieta fault (Bürgmann et al., 1997). Although the afterslip was found to be relatively small (less than 1 cm over 4 months), the afterslip area roughly corresponds to the aftershock zone area in our analysis, which extends southwards for 60 km from the mainshock rupture in a shallow region above 15 km depth. Hence, we argue that the afterslip could account for the large aftershock zone area of Loma Prieta.

Afterslip can occur in the surrounding region loaded by mainshock rupture and transfer stresses on faults that promote the generation of aftershocks. It is unfeasible to quantitatively evaluate the contribution by Coulomb stress change and other mechanisms without detailed rupture simulation based on realistic parameters. For mainshocks with observed afterslip, a combination of the static stress change and afterslip instead of the Coulomb stress change alone could contribute to the positive correlation between the Coulomb stress change area and the aftershock zone area. By comparing the ratios from using both entire aftershock duration (Figure 6) and early aftershocks (Figure 7), we observe that the long-term aftershock duration results in relatively larger ratio for the South Napa, Loma Prieta, Brawley, Joshua Tree, Elmore Ranch and Whittier Narrows earthquakes. The larger ratio could be explained by expanding aftershock zones with time caused by postseismic deformation process. In contrast, similar range of aftershock ratios for the other earthquakes support that Coulomb stress change caused by the mainshock rupture plays an important role in aftershock distribution.

We also search for published work on postseismic slip following the studied mainshocks, and seven earthquakes have resolved postseismic slip model (Table S6). For most earthquakes analyzed in the table, the afterslip distribution is similar in extent to our aftershock zone area though their depths may be different, which is consistent with emerging evidences that afterslip could affect the long-term aftershock evolution (Perfettini et al., 2018). To better understand the outlier mainshocks, we could potentially use afterslip models to measure the stress change caused by afterslip, similar to that of Perfettini and Avouac (2004; 2007), to ascertain if it correlates better with their aftershock zone areas. This exceeds the scope of this study and could be a potential work in future with more available afterslip models.

Uncertainty and Limitations

The measurement uncertainties in our calculations include the (1) earthquakes locations, (2) the calculation of Mc, (3) the assumption of threshold β value, (4) the upper limit of the grid and gap size, (5) slip inversion results, and (6) the assumption that the fault plane extends in roughly the same plane outside of the mainshock rupture area.

As shown by the large variations of ratios for different slip models, slip inversion results probably contribute to the largest uncertainty in this study. Thus, we estimate the uncertainty by calculating rupture areas defined by 0.1 and 0.2 of the maximum slip (Table S7), which is shown as error bars in Figures 6 and 7. The results show that the aftershock ratios of 2014 South Napa and 1986 North Palm Springs earthquakes may be biased by slip models that have large uncertainty. Aside from the slip contours, other sources of error from the slip models include the slip inversion parameters used, such as the geometry and orientation of the fault planes, but they are much harder to quantify.

We also examined the uncertainty of earthquake locations for the Parkfield mainshock using the Ji (2004) model. We use location uncertainties of 0.5km, 1km and 2km to randomly vary the locations of all the aftershocks. We generate 10,000 synthetic distributions of aftershocks and find the standard deviations of aftershock zone areas are 0.09, 0.12 and 0.15 respectively, which is about 4.8 to 7.8% of the mean value. As the location uncertainties for most earthquakes are smaller than 2km, we believe that the location uncertainty will not greatly affect the ratios.

We calculate Mc before and after the mainshock and remove earthquakes below Mc. Though this procedure ensures that the seismicity change is not biased by the incomplete catalog, it also removes earthquakes from consideration, which may cause the calculated aftershock zone area to be smaller than the real aftershock zone area. To estimate the impact of removing earthquakes below Mc, we calculate the aftershock zone areas of the Brawley and El Mayor-Cucapah mainshocks using the Quake Template Matching (QTM) Catalog for Southern California (Ross et al., 2019) that has a much lower Mc due to the new detections. The ratios of aftershock zone areas to mainshock rupture areas estimated from this catalog are larger than that calculated from the relocated SCEDC catalog (Table S8, Figure S5). However, they are still within the range of ratios (0.5-5.4) obtained for all the mainshocks.

We limit the calculation of the aftershock zone area by setting a threshold β value of 2. A threshold value of 2 indicates 95% significance of increase in seismicity when we normalize the β value by its standard deviation. The assumption behind the calculation of standard deviation is that each earthquake is an independent event and the probability of an earthquake occurring at any given time is equal. This may not be a valid assumption for earthquakes as the probability of having earthquakes after a mainshock is much higher than before the mainshock, but all metrics for determining aftershock zone area necessarily contain arbitrariness.

We also set an upper limit of the spatial grid and gap to terminate aftershock zone, which is many times of the source dimension but may still violate the observation of the so-called "global aftershock zone" (Parsons and Geist, 2014; Johnson and Bürgmann, 2016). Among our investigated mainshocks, we noticed an increase of microearthquakes within the Geysers geothermal region following the 2014 Napa earthquake (Figure 2), likely triggered by the passing seismic waves (Meng et al., 2014). More recently, Ross et al. (2019) suggested that the 2010 El Mayor-Cucapah earthquake widely triggered events in Southern California. Hence, we are referring to the traditional aftershock zone in this study, where various triggering mechanisms are comparable.

Other Results

Studies have shown that aftershocks tend to be concentrated around the boundary of the mainshock rupture zone, with a deficit in the center regions of higher slip (Carlos Mendoza and Hartzell, 1988; Dietz and Ellsworth, 1990; Wetzler et al., 2018). This is because most of the strain in the regions of higher slip are already released during the mainshock and hence these areas are less able to generate aftershocks. We test this hypothesis using a slightly modified version of the method used in Wetzler et al. (2018), which calculates the distances of aftershocks from the slip contours of several earthquakes and normalizes by the radius of a circle that has an area equal to the area enclosed by the slip contour. As many of our slip contours are elongated, we change the normalization constant to the minor axis of an ellipse fitted to the slip contours (Wijewickrema and Paplínski, 2004), as shown in Figure 10. The distances are then calculated from the closest slip contour (if there are multiple parts) and normalized by the minor axis of the ellipse fitted to that slip contour. Negative distances refer to distances of aftershocks inside the slip contour while positive distances refer to distances of aftershocks outside the slip contour. We use this method to analyze one slip model from each earthquake (list of models in Table S9). We find that most of the aftershocks are located near the slip contours, within a distance of -0.25 to 0.25 the slip contours. Compared to the results obtained by Wetzler et al. (2018), we find more earthquakes located between 0.5 to 1 distance inside the slip contours (Figure 11), probably because we use the minor axis of an ellipse as the normalization constant. But our results still support the notion that there is a deficit of earthquakes in the central regions of the largest slip.

Conclusion

We have developed a consistent method to quantify aftershock zone and mainshock rupture areas, and analyzed 12 mainshocks ($M_w \geq 5.4$) in California. We find that the ratios of aftershock zone areas to mainshock rupture areas lie within a range of 0.5 to 5.4, and can be used as a first order estimate of the mainshock rupture area, especially for early aftershock durations where ratios range from 0.5 to 3.5. Using either the relocated catalog or the ANSS catalog leads to similar patterns of the aftershock zone area. Our results show that Coulomb stress change exhibits a positive correlation with aftershock zone area, suggesting that the mainshock slip contributes to aftershock distribution. Moreover, a combination of different mechanisms should be used to better explain the aftershock zone areas (especially for the entire aftershock duration) for several mainshocks. Further studies should be directed towards understanding how the relationship between the aftershock zone area and the mainshock rupture area varies for earthquakes in different tectonic environments and crustal structures.

Data and Resources

Slip inversion data was downloaded from the SRCMOD website at http://equake-rc.info (last accessed July 2020). Earthquake catalogs were obtained from the NCEDC www.ncedc.org (last accessed October 2019) and SCEDC websites http://scedc.caltech.edu (last accessed October 2019) and the catalog by Dr. Felix Waldhauser, https://www.ldeo.columbia.edu/~felixw/NCAeqDD/ (last accessed March 2019), version NCAeqDD.v201112.1. Earthquakes from the ANSS ComCat Catalog were downloaded from USGS https://earthquake.usgs.gov/earthquakes/search/ (last accessed August 2019) and Coulomb 3 MATLAB codes were downloaded from the USGS website, https://earthquake.usgs.gov/research/software/coulomb/ (last accessed July 2019). We obtained MATLAB codes written by D. Schorlemmer and J. Woessner to calculate Mc from geophysics.eas.gatech.edu/people/bsullivan/tutorial/StatisticalSeismology.htm (last accessed March 2019). The supplementary material contains additional information about individual earthquakes, early aftershock ratios and the results of Coulomb stress change calculations.

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Table 1
Summary of the source properties and ratios of earthquakes.

Earthquake	Date	Location N/W	Magnitude (Mw)	Depth (km)	Slip Inversion References	Data Sources*	Reloc.	ANSS
South Napa (SN)	2014/08/24	38.22/122.31	6.10	11.0	Wei et al (2015)	SGM	1.04	0.92
			6.07	10.0	Gallovič (2016)	SGM	4.16	3.66
Parkfield (Pf)	2004/09/28	35.82/120.37	5.90	8.0	Ji (2004)	SGM, GPS	1.28	2.04
			6.00	8.3	Dreger et al (2005)	SGM, GPS	2.37	2.70
			6.06	8.3	Custodio et al (2005)	SGM	2.36	2.18
Loma Prieta (LP)	1989/10/18	37.04/121.88	6.98	17.6	Zeng and Anderson (2000)	SGM	5.38	4.54
			6.94	17.6	Wald et al (1991)	SGM, TELE	4.19	3.53
			6.96	17.6	Beroza (1991)	SGM	3.82	5.23
			6.91	17.6	Emolo and Zollo (2005)	SGM	3.40	4.69
Brawley Swarm (BS)	2012/08/26	33.02/115.54	5.45	6.4	Wei et al (2013)	SGM, GPS	0.82	2.29
El-Mayor-Cucapah (EMC)	2010/04/04	32.30/115.30	7.35	10.0	Mendoza and Hartzell (2013)	TELE	2.17	2.61
			7.29	5.5	Wei et al (2011)	TELE, SPOT, GPS, INSAR, SAR	1.79	1.74
Hector Mine (HM)	1999/10/16	34.59/116.27	7.24	6.0	Kaverina et al (2002)	SGM, GPS	1.26	1.18
			7.16	15.0	Jonsson et al (2002)	GPS, INSAR	1.49	0.98
Northridge (Nr)	1994/01/17	34.21/118.54	6.71	17.5	Zeng and Anderson (2000)	SGM	2.33	3.01
			6.80	17.5	Wald et al (1996)	SGM, TELE, GPS	1.30	1.88
			6.81	17.5	Hudnut et al (1996)	TRIL, GPS	1.38	1.67
			6.73	17.5	Hartzell et al (1996)	SGM	1.21	1.49
Landers (Ld)	1992/06/28	34.20/116.43	7.20	7.0	Zeng and Anderson (2000)	SGM	1.18	0.88
			7.22	7.0	Hernandez et al (1999)	SGM, GPS	1.80	1.26
			7.29	7.0	Cotton and Campillo (1995)	SGM	1.47	1.15
Joshua Tree (JT)	1992/04/23	34.00/116.32	6.25	12.5	Bennett et al (1995)	TRIL, GPS	2.12	1.96
Elmore Ranch (ER)	1987/11/24	33.08/115.80	6.52	10.0	Larsen et al (1992)	TRIL, GPS	4.30	2.06
Whittier Narrows (WN)	1987/10/01	34.05/118.08	5.89	14.6	Hartzell and Iida (1990)	SGM	1.61	NA
North Palm Springs (NPS)	1986/07/08	34.00/116.57	6.14	11.0	Hartzell (1989)	SGM	0.54	0.79
			6.21	11.0	Mendoza and Hartzell (1988)	TELE	1.87	2.72

*SGM: Strong ground motion, TELE: Teleseismic data, GPS: Global Positioning System, SAR: Synthetic-Aperture Radar, INSAR: Interferometric Synthetic-Aperture Radar, SPOT: Optical imaging from the SPOT-5 satellite.

Table 2
Statistics of the ratios

	Entire Duration		1-Day Aftershocks	
	Reloc. Catalogs	ANSS Catalog	Reloc. Catalogs	ANSS Catalog
Mean	2.18	2.16	1.83	1.70
Variance	1.58	1.90	0.84	1.07
Median	1.80	2.00	1.71	1.75
Mean Absolute Deviation	0.98	1.03	0.77	0.84

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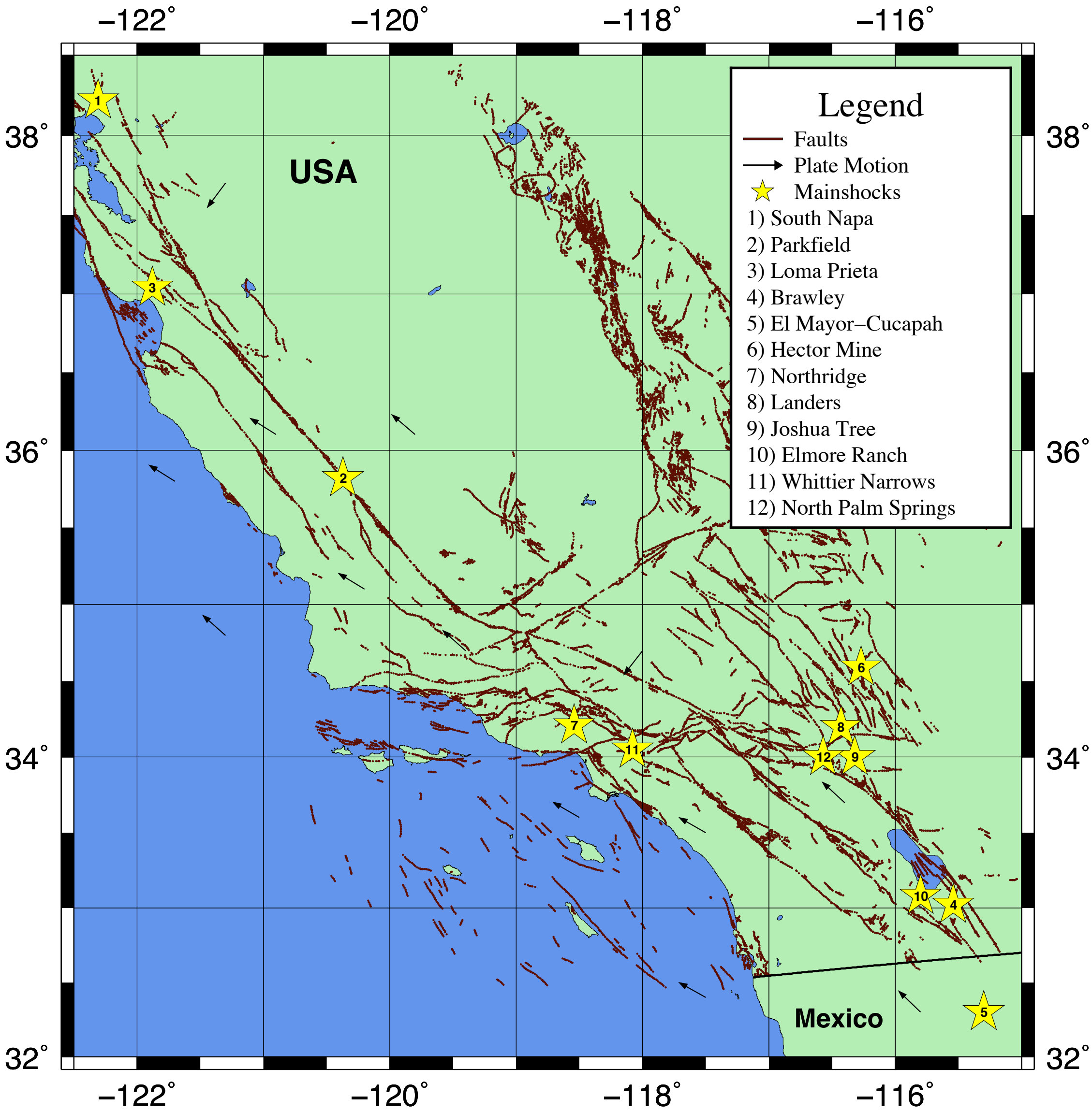

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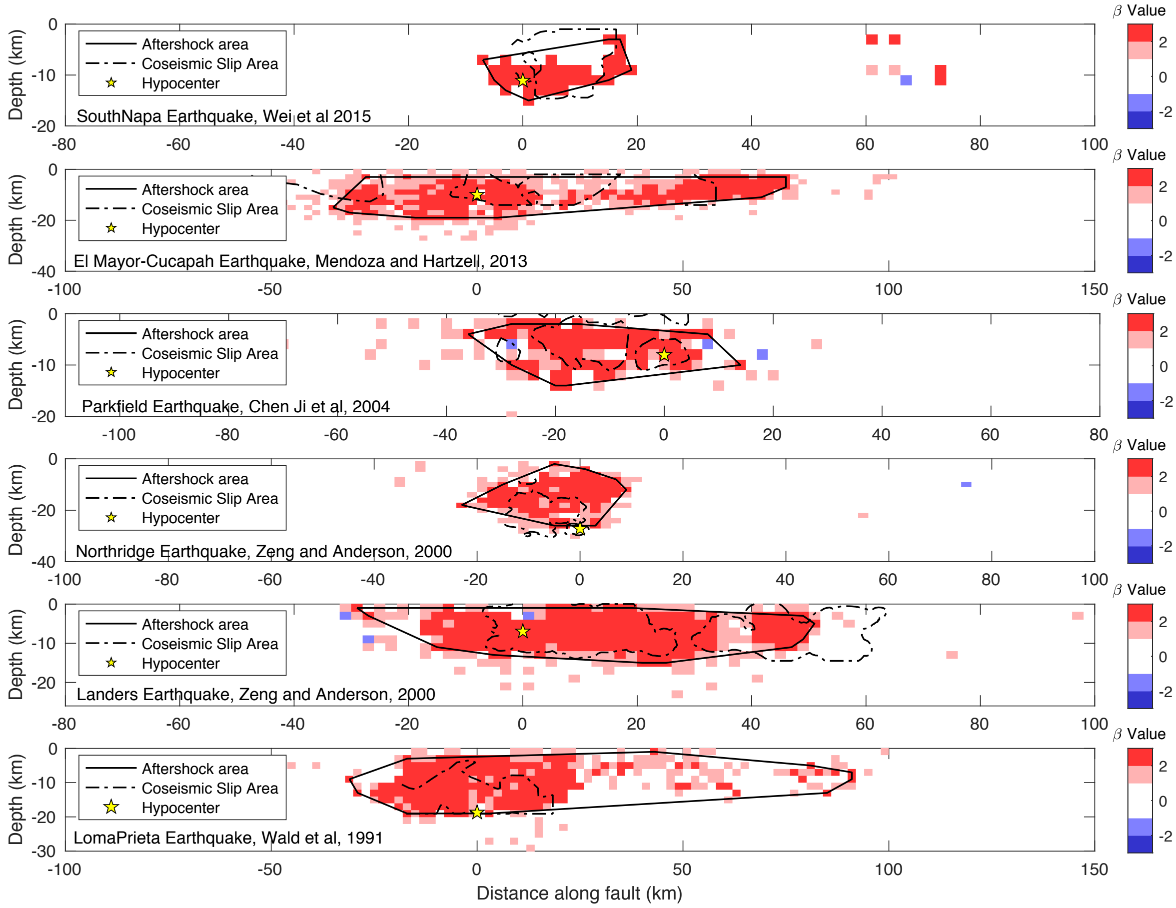


Figure 2. Illustration of how the aftershock zone area of each mainshock is defined using the β values. The diagrams show the fault plane view, with β values of each grid cell calculated from the aftershocks projected on to the fault plane. Aftershocks from the relocated catalogs are used for this figure.

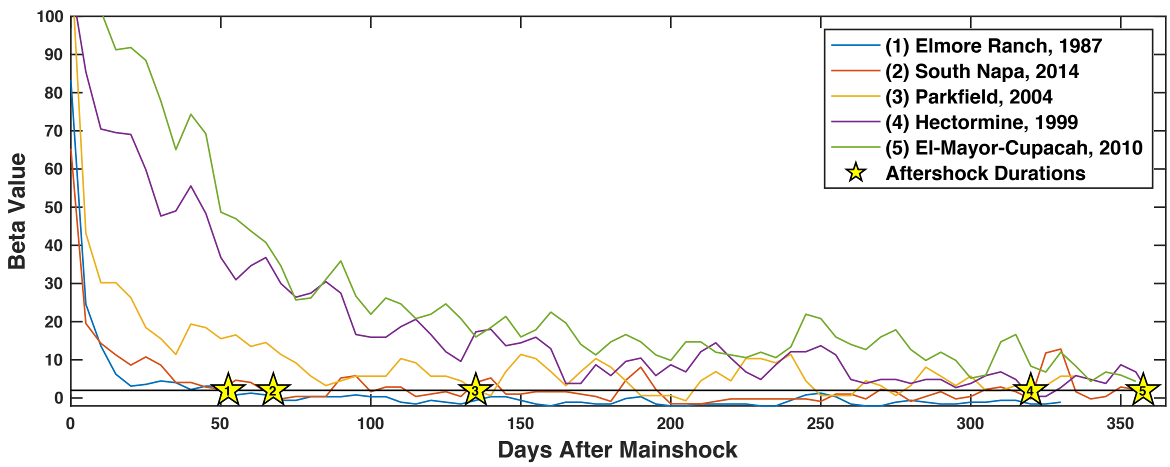


Figure 3. Illustration of how aftershock duration is calculated. The horizontal black line is at a β value of 2, and the aftershock duration is taken to be the end of the time window where the β value first dips below the line (indicated by the numbered stars). If the β value never dips below 2, 1 year is used.

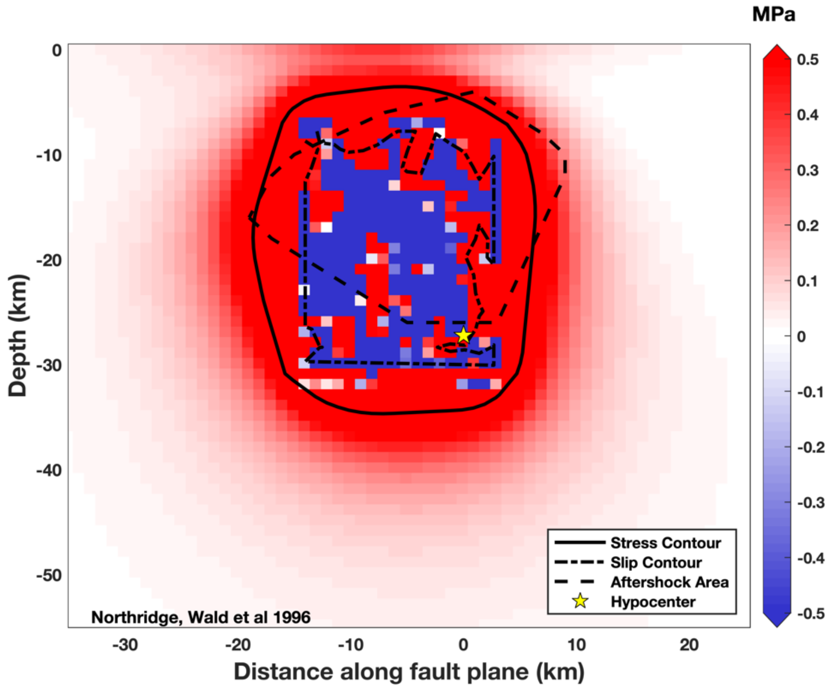


Figure 4. Depiction of how the Coulomb stress change area is calculated. For illustration, a contour is drawn around the boundary of cells with a positive Coulomb stress change of >1 MPa or more. The Coulomb stress change area is given by the sum of the area of these cells.

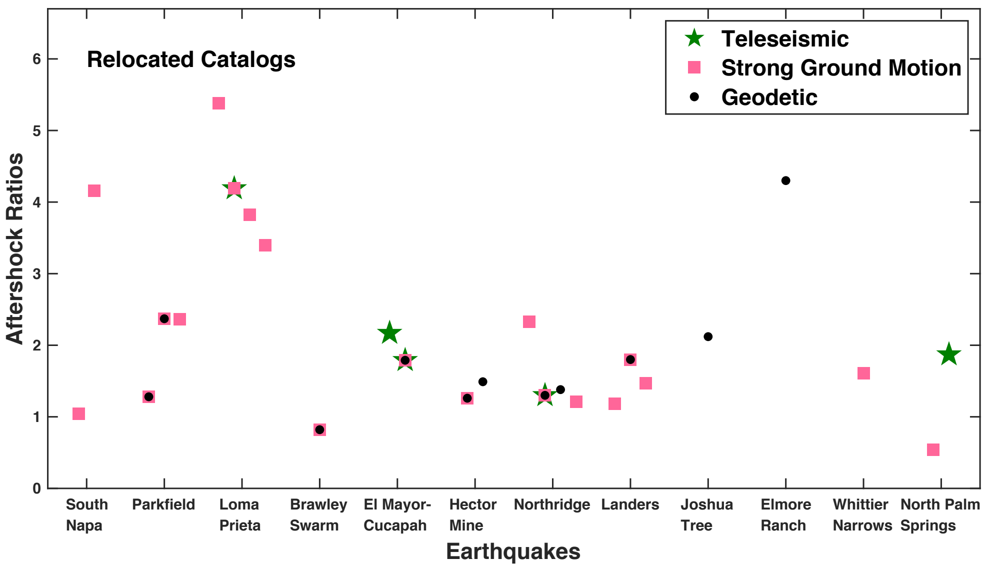


Figure 5. Plot of the data types used for each slip inversion, where the ratios are calculated using the relocated catalogs. SGM: Strong ground motion; Teleseismic: Teleseismic waveform data; Geodetic: GPS, INSAR.

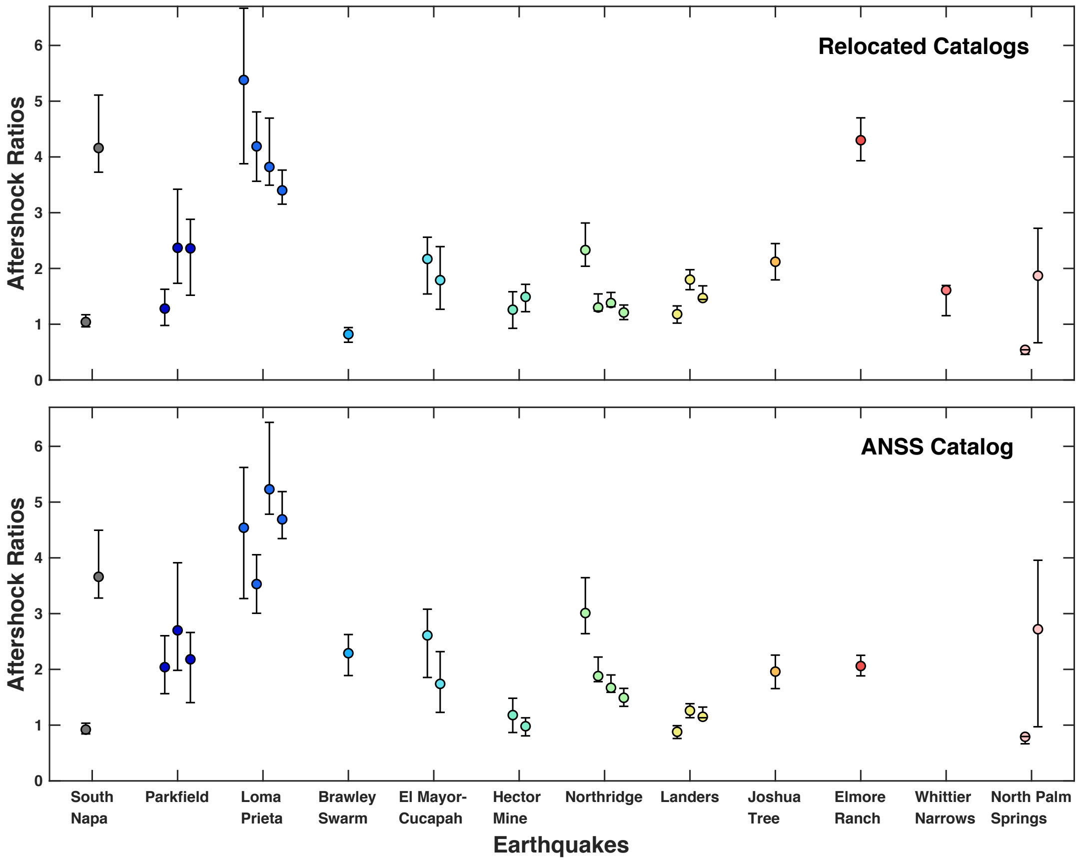
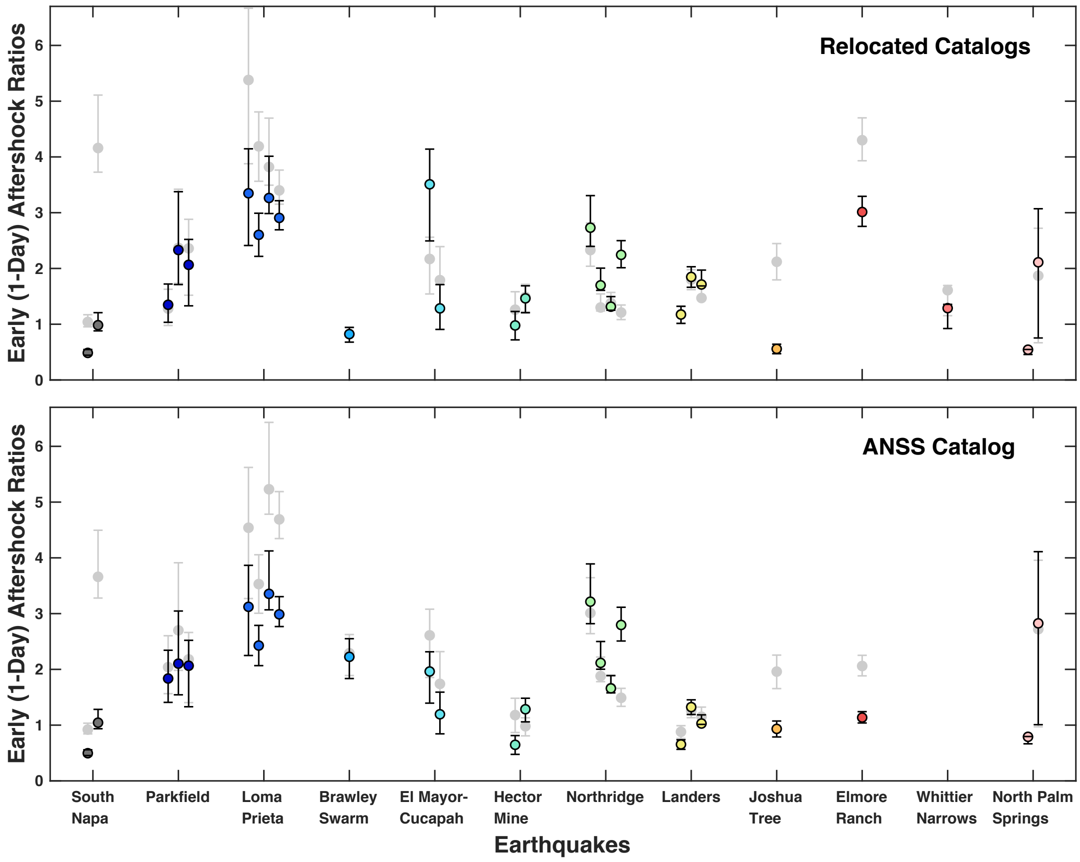
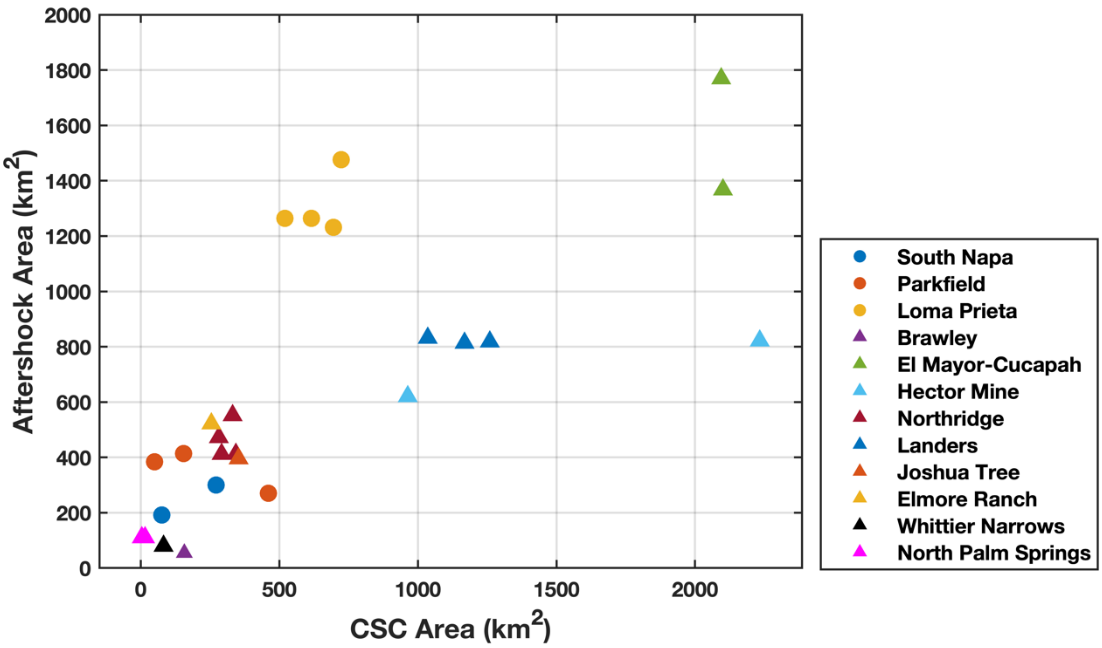


Figure 6. Aftershock zone area ratios for different earthquakes using different earthquake catalogs. If there are multiple slip inversions for the same earthquake, the ratios are slightly offset so that they do not overlap. The error bars are calculated using slip contours of 0.1 and 0.2 of the maximum slip to calculate different rupture areas. (Top) NCEDC data is used to calculate the aftershock zone area for the first 3 earthquakes, while SCEDC data is used for the rest of the earthquakes. (Bottom) ANSS catalog is used. The ratio for the Whittier Narrows (WN) earthquake is not obtained from the ANSS catalog because the data does not yield a robust estimation of the magnitude of completeness.



747
748 **Figure 7**. Aftershock ratios calculated from 1-day aftershock durations for both earthquake catalogs. The
749 aftershock ratios for the whole duration is plotted in the background in light grey for comparison.



750
751 **Figure 8**. Aftershock zone area vs. Coulomb stress change area.

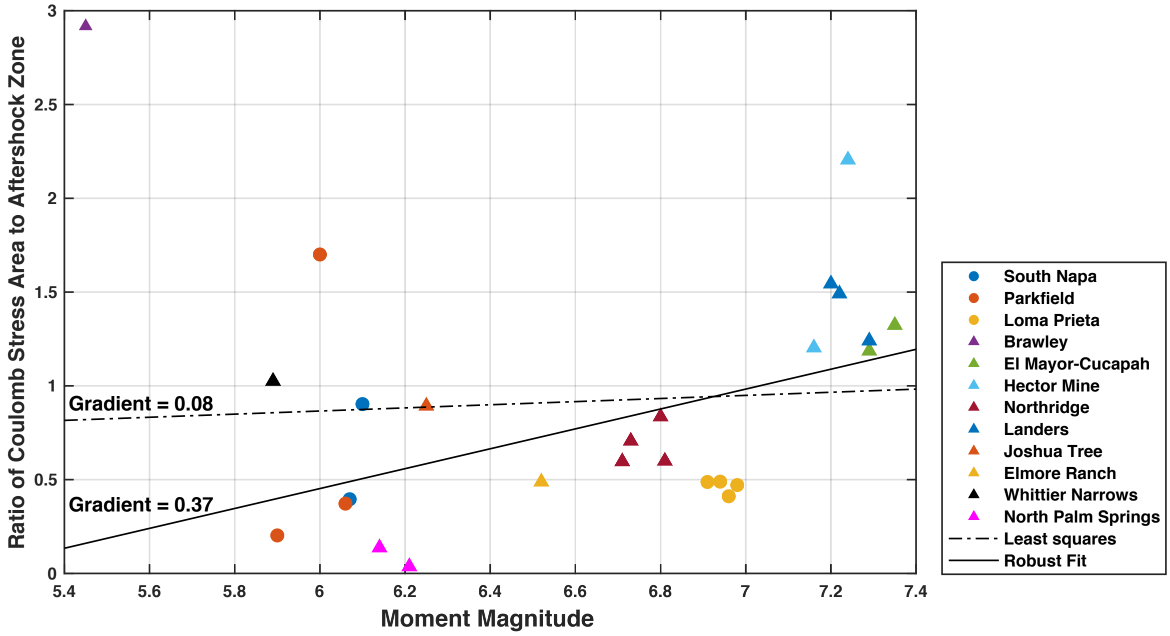


Figure 9. Robust fitting (solid line) and least squares fitting (dashed line) of ratios of Coulomb stress area to aftershock zone area with magnitude

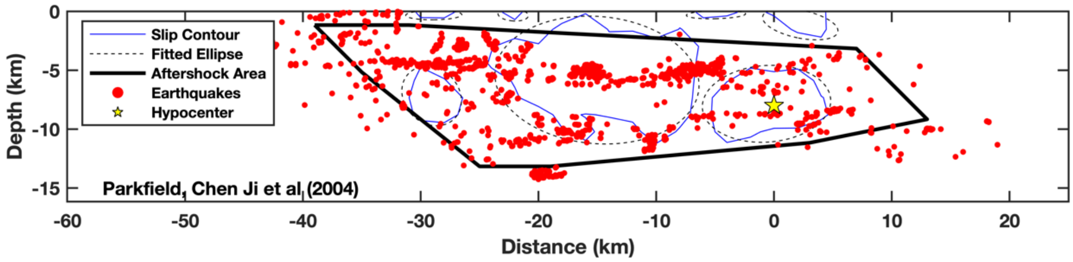


Figure 10. Illustration of how the distances from slip contour are calculated using the Parkfield, Chen Ji et al slip model.

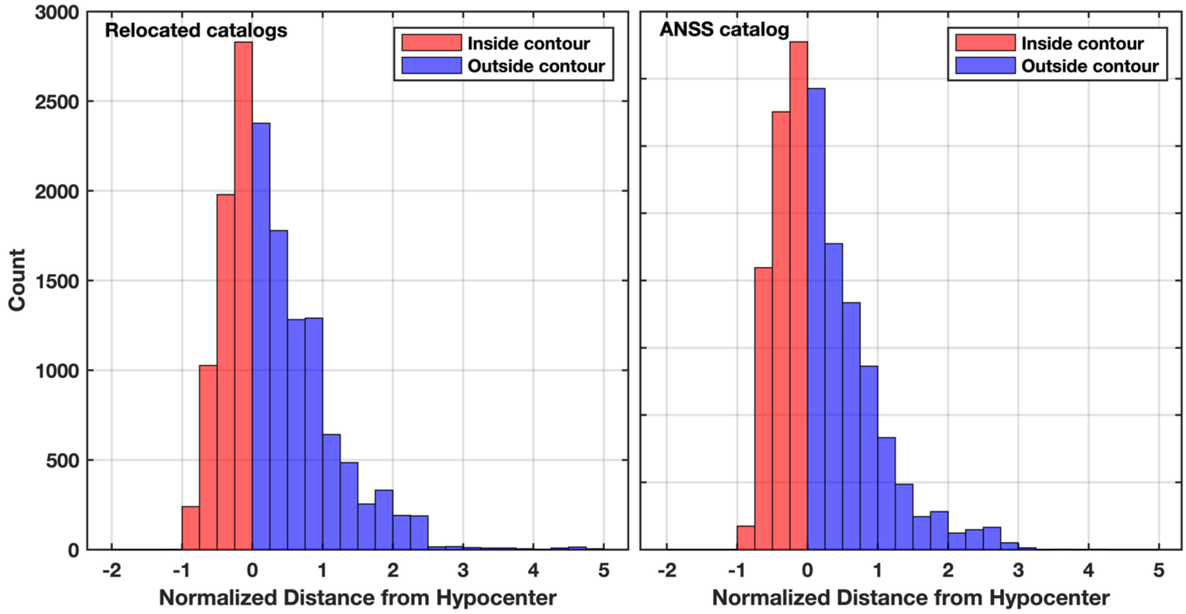


Figure 11. Histogram of aftershock distances from slip contours for all earthquakes using the relocated and ANSS catalogs.