

## **Preliminary Report on the 29 July 2008 Mw5.4 Chino Hills, Eastern Los Angeles Basin, California, Earthquake Sequence**

Egill Hauksson<sup>1</sup>, Karen Felzer<sup>2</sup>, Doug Given<sup>2</sup>, Michal Giveon<sup>1</sup>, Susan Hough<sup>2</sup>, Kate Hutton<sup>1</sup>, Hiroo Kanamori<sup>1</sup>, Volkan Sevilgen<sup>2</sup>, Shengji Wei<sup>1</sup>, and Alan Yong<sup>2</sup>,

<sup>1</sup>Caltech

<sup>2</sup>USGS

Contact: Egill Hauksson: (626)-395-6954 or hauksson@gps.caltech.edu

### **INTRODUCTION**

The 29 July 2008 Mw5.4 Chino Hills earthquake was the largest event to occur within the greater Los Angeles metropolitan region since the Mw6.7 1994 Northridge earthquake. The earthquake was widely felt in a metropolitan region with a population of over 10 million people, and was recorded by hundreds of broadband and strong motion instruments. In this report we present preliminary analysis of the event and discuss its significance within the seismotectonic framework of the northern Los Angeles basin as revealed by previous moderate earthquakes.

The Chino Hills, mainshock-aftershock sequence began at a depth of about 15 km, in the east Los Angeles area at 11:42am (PST) (Figure 1). The epicenter is between two mapped faults: the Whittier fault to the west and the Chino Hills fault to the east. The focal mechanism indicates a mixture of strike-slip and thrust faulting on a west-southwest or a west-northwest striking nodal plane. The mainshock was followed by only two aftershocks with  $M > 3$ ; M3.8 at 11:52 am (PST) and M3.6 at 13:40 pm (PST). In the first two hours, 37 smaller aftershocks were also recorded in the magnitude range of 1.3 to 2.8. By August 14<sup>th</sup>, the Southern California Seismic Network (SCSN), a joint project of Caltech and USGS had recorded  $\sim 150$  aftershocks of  $M \geq 1.0$ . The mainshock was not preceded by foreshock activity.

During the 2008 Chino Hills sequence, the SCSN automatically processed real-time waveform data from 370 stations across southern California. The first location and ML magnitude estimate of 5.6 were released  $\sim 80$  s after the origin time. An updated location and final ML5.8 were released after  $\sim 140$  s. The automatic moment tensor and the Mw estimate of 5.4 were available  $\sim 10$  minutes following the origin time. These SCSN rapid notifications were posted on the Web, and data were made available via <http://www.data.scec.org> and <http://earthquake.usgs.gov>.

The Chino Hills sequence was widely felt across southern California although damage was minimal. Relatively strong shaking was recorded to the north in the Diamond Bar area and to the northwest in the eastern Los Angeles basin, as demonstrated in the ShakeMap (Wald *et al.*, 1999a) and actual strong motion records, which were made available via <http://www.strongmotioncenter.org>. The initial ShakeMap, was available 12 minutes after the origin time; six updates of the ShakeMap followed during the next hour as more data arrived from near real-time stations (Figure 2). The final map included amplitudes from 526 California Integrated Seismic Network (CISN) stations. Over 40,000 people filled out an Internet form, “Did You Feel It” to describe the effects of the earthquake at locations throughout southern California.

Five other moderate-sized mainshock-aftershock sequences have occurred in the general vicinity of the Chino Hills earthquake since 1987. The largest event was the Whittier Narrows earthquake of 1 October 1987, which was located about 30 km west-northwest, and had a magnitude of 5.9. It caused 3 direct fatalities, and over \$358 million in damage. The Whittier Narrows earthquake resulted from thrust faulting on the Puente Hills thrust (Shaw and Shearer, 1999). The other four earthquakes were the M4.6 1989 Montebello, M5.0 1988 Pasadena, M5.2 1990 Upland, and M5.8 Sierra Madre earthquakes. The 1988 Montebello earthquake was caused by thrust faulting, similar to the Whittier Narrows mainshock (Hauksson, 1990). Both the Pasadena and the Upland earthquakes exhibited west-southwest left lateral strike-slip faulting while the Sierra Madre earthquake exhibited thrust faulting (Jones *et al.*, 1990; Hauksson and Jones, 1991; Hauksson, 1994; Shearer, 1997; and Astiz, 2000). Thus the crustal deformation associated with the 2008 Chino Hills earthquake is similar to deformation associated with the previous events.

### **MAINSHOCK MOMENT TENSOR**

A real-time SCSN moment tensor solution (Clinton *et al.*, 2006), which was derived using an automated analysis of waveforms from six stations and the inversion method of Dreger and Helmberger (1993), revealed a strike of  $291^\circ$ , a dip of  $59^\circ$ , and a rake of  $142^\circ$ . The double couple solution was also determined using the “cut-and-paste” technique (Zhao and Helmberger, 1994; Zhu and Helmberger, 1996). In the “cut-and-paste”, we analyzed the broadband waveforms of regional seismograms from 22 stations with epicentral distances between 100 to 200 km and maximum azimuth gap of  $32^\circ$  (Figure 3A). The errors in the strike, dip, and rake of the focal mechanism are about  $\pm 3^\circ$ ,  $\pm 6^\circ$ , and  $\pm 15^\circ$  respectively.

The Green's functions were generated from the FK technique (*Zhu and Rivera, 2002*) and were based on a 1-D southern California velocity model (*Zhao and Helmberger, 1994*). A source depth of 15 km and  $M_w 5.4$  correspond to minimum error between the real and synthetic waveforms (Figure 3B). The high cross-correlation coefficients (Figure 3C) confirm the similarity of the synthetics with the data, both for the Pnl waves and the surface waves. Because the "cut and paste" allows time shifts between portions of seismograms and synthetics, the results of this method are relatively insensitive to minor changes in the velocity model and to possible lateral crustal variations. (*Zhu and Helmberger, 1996*).

The Global CMT solution (<http://www.globalcmt.org/CMTsearch.html>) for this event yields a best double-couple with the nodal planes (strike/dip/rake)  $43^\circ/62^\circ/31^\circ$ ,  $297^\circ/63^\circ/149^\circ$  with  $M_w=5.4$ . Also the W phase inversion (*Kanamori and Rivera, 2008*) using 18 broadband SCSN stations with a pass-band of 50 to 150 s yields a similar mechanism and  $M_w$  ((strike/dip/rake)  $51^\circ/38^\circ/32^\circ$ ,  $294^\circ/71^\circ/123^\circ$ ,  $M_w=5.4$ ). The main difference between the GCMT solution and the WP inversion is in the dip angle of the southeast dipping plane. These solutions represent the long-period characteristics of the source, and the good agreement between the regional, global and long-period solutions suggests that the source was associated with no significant anomalous long-period (up to 100 s) deformation.

## RELOCATIONS AND FOCAL MECHANISMS

We relocated the mainshock and aftershocks, using a 3D velocity model and HypoDD (*Waldhauser and Ellsworth, 2000*). The relocations reveal minimal clustering around either mainshock nodal plane (Figure 4). In addition, the aftershocks occurred in a limited depth range from 13 to 16 km and form a tight sub-horizontal distribution. It is thus not easy to determine from the spatial aftershock patterns which of the two steeply dipping nodal planes ruptured in the mainshock. Further, the mainshock is located at a depth of 15 km, mid-way between the Whittier and Chino faults, and because of its large focal depth, projecting the nodal planes up to dipping surface faults is difficult at best.

In map view, the aftershocks scatter from the Chino fault in the northeast, across the mapped trace of the Whittier fault, and coincide with the Yorba Linda trend to the southwest, as defined by *Hauksson (1990)*. Almost no aftershocks occurred in the immediate vicinity of the mainshock or within a radius of 1.5 km. The seismicity within the Yorba Linda trend extends from a few kilometers depth down to 16 km depth. If the south dipping east-northeast trending nodal plane was the fault plane, it would intersect the surface near the San Jose fault to the north. However, the San Jose fault could not be the causative fault, because it dips steeply to the north, and the Chino Hills sequence is located ~15 km

distance to the south. The 1990 M5.2 Upland earthquake was associated with the San Jose fault but at shallower depth of 5 to 10 km (*Hauksson and Jones, 1991*).

The northeast dipping nodal plane of the mainshock has a very similar strike as the Whittier fault. However, associating the mainshock with the Whittier fault becomes difficult because the fault dip is uncertain, and the aftershock locations do not provide additional constraints. The Whittier fault has a steep dip of  $\sim 80^\circ$  in the near-surface but the dip gradually shallows out to  $\sim 60^\circ$  when it reaches depths of 5 km or greater (*Bjorklund and Burke, 2002*; and *Bjorklund et al., 2002*). Thus some of the aftershocks are clearly located in the footwall block of the Whittier fault, including the largest off-fault aftershock cluster. Both of the two M3.6 and M3.8 aftershocks also formed their own aftershock clusters. We used the method of *Hardebeck and Shearer (2002)* to determine the first motion focal mechanisms for the M3 aftershocks. Their focal mechanisms exhibited strike-slip motion on northwest or northeast striking planes, while the focal mechanisms of the smaller aftershocks exhibit a heterogeneous mixture of strike-slip, and thrust faulting, with some minor normal faulting.

### **MAINSHOCK STRESS DROP**

The 2008 Chino Hills earthquake excited impulsive  $S$  waves that were recorded at many close-in stations. The pulse width,  $\tau$ , is approximately 1 s. Here  $\tau$  is measured from the  $S$  wave pulses observed on the horizontal component records at 9 SCSN non-nodal stations (BRE, PDU, WLT, CHN, RUS, LLS, FUL, DLA, and PSR) and 8 strong-motion records compiled at the Center for Engineering Strong Motion Data. Although the pulse widths of  $P$  waves are equally narrow, we did not measure them because  $P$  waves are relatively small at close-in stations for this event. When plotted on a  $\tau$  vs.  $M_0$  diagram (Figure 5), the data point for this event falls on the trend defined by several Los Angeles basin earthquakes, such as the 1991 Sierra Madre, 1989 Pasadena, and the 1989 Montebello earthquakes which yielded the smallest pulse width for a given magnitude (dashed line in Figure 5). The pulse width for the Chino Hills earthquake is also close to the lower bound of  $\tau$  of the  $\tau$  vs.  $M_0$  data set for eastern North America, western North America, and other continental interiors compiled by *Somerville et al. (1987)*. Since attenuation will broaden the pulse width, the values measured from seismograms should be regarded as upper bound, and the actual pulse width could be even shorter. In general, a small pulse width implies a small source dimension, which in turn implies a large stress drop,  $\Delta\sigma$ . The solid line in Figure 5 is the relationship for a global data set (*Kikuchi, written communication, 2001*; *Kanamori and Brodsky, 2004*). This trend is almost identical to the relationship used by the Harvard and the Global CMT groups, and is given by  $\tau = 2.10 \times 10^{-8} M_0^{1/3}$  ( $\tau$  in s and  $M_0$  in dyne-cm). The

pulse width of the Chino Hills earthquake is about 2.4 times smaller than what the global relationship predicts. Since the commonly used scaling relation gives  $\Delta\sigma \propto \tau^{-3}$ , the data plotted in Figure 5 would suggest that  $\Delta\sigma$  of the Chino Hills earthquake is  $2.4^3$  (~14) times larger than the global average of large earthquakes. The empirical relation given by Cohn et al. (1982),  $\Delta\sigma = M_0\tau^{-3}/10^{15.73}$  ( $\Delta\sigma$  in MP,  $M_0$  in N-m), leads to  $\Delta\sigma \sim 30$  MP for  $\tau = 1$  s. However, since  $\Delta\sigma$  depends on other factors such as the rupture speed and geometry, its exact value is subject to some uncertainty, but the very short pulse width of the Chino Hills earthquake suggests that it is most likely a high-stress drop event.

*Bent and Helmberger* (1989) estimated the stress drop of the 1987 Whittier Narrows earthquake (Mw=5.9) to be about 75 MP. *Dreger and Helmberger* (1991a, b) estimated the stress drops of the 1990 Upland earthquake (Mw=5.5) and the 1991 Sierra Madre earthquake (Mw=5.8) to be 27 and 46 MP, respectively. However, these estimates depend on the assumption on the rupture dimension. Judging from the duration of the source discussed in these papers, these events would probably plot close to the dashed line of the  $\tau$  vs.  $M_0$  diagram (Figure 5). Thus, considering the uncertainties in the stress drop estimates, the 2008 Chino Hills, the 1987 Whittier Narrows, the 1991 Upland, and the 1991 Sierra Madre earthquakes all seem to belong to a group of high stress drop events. Our stress drop results also agree with *Shearer et al.* (2006) who analyzed a large southern California data set and concluded that the range of "stress drop" for southern California is from 0.2 to 20 MP.

### **AFTERSHOCK STATISTICS**

As of August 8, 2008, the aftershock sequence of the Chino Hills earthquake was decaying in time in accordance with the classic modified Omori law ( $R(t) = K(t+c)^{-p}$ ) (*Utsu*, 1961) where  $R(t)$  is earthquake rate,  $t$ , is time since the mainshock, and  $c$  and  $p$  are constants (Figure 6). The value of  $c$  is very small, typically difficult to determine, and unimportant after the immediate post-mainshock period. Hence we focus here on the values of  $K$  and  $p$ . As of August 8, 2008, the best fitting  $p$  value was 1.09, with a 98% confidence range from 0.91 to 1.36, and the best fitting  $K$  value was 3.4, with a 98% confidence range from 2.3 to 5.2, where  $K$  is measured in terms of  $M \geq 2$  aftershocks/day. This  $p$  value is very close to the California state average of 1.08 (*Reasenber and Jones*, 1989) but the  $K$  value is about 20% of the state average for an earthquake of this size (*Felzer et al.* 2003), indicating a low overall rate. As of August 8, only 26  $M \geq 2$  aftershocks in total had been recorded.

Extrapolation of the current decay parameters indicate that if current trends continue 1 to 2  $M \geq 2$  aftershocks may be expected during the month of September 2008 and 2 to 3  $M \geq 2$  aftershocks may be expected for the entire year of 2009. The region of the Chino Hills earthquake had been moderately

active in the past, and so we inspected other local aftershock sequences to see if the low aftershock rates might have been anticipated. Using the region 33.68° to 34.24°N and -117.93° to -117.61°W and the time period January 1, 1984 – June 30, 2008 we identified 213  $M \geq 2$  potential mainshocks that were reasonably isolated (at least 20 days after, or 5 fault lengths away from larger earthquakes). These mainshocks produced  $M \geq 2$  aftershocks (defined as earthquakes occurring within 10 days and 5 fault lengths) at a rate of 84% of the state average, with 98% confidence intervals from 65% to 108%. Thus previous sequences had somewhat low productivity but not nearly as low as that of the Chino Hills sequence. The magnitude-frequency statistics of the Chino Hills sequence were normal, agreeing well with the Gutenberg-Richter magnitude frequency distribution (*Gutenberg and Richter, 1944*) with a  $b$  value of 1.0.

### REMOTELY TRIGGERED EARTHQUAKES

Recent studies (e.g., *Hough, 2005; Felzer and Brodsky, 2006*) have concluded that aftershock/triggered earthquake sequences following small and moderate mainshocks extend to distances of many fault lengths. To explore the spatial distribution of the aftershock sequence of the Chino Hills earthquake, we use the  $\beta$ -statistic to quantify the seismicity rate change at close and regional distances (e.g., *Matthews and Reasenberg, 1988*). The  $\beta$ -statistic, or simply  $\beta$ , is defined as

$$\beta = (N_a - N_e) / \text{sqrt}(u) \quad (1)$$

Where  $N_a$  is the actual number of events in a grid cell,  $N_e$  is the number expected based on the number of earthquakes before the mainshock, and  $u$  is the variance assuming a Poissonian distribution. The  $\beta$ -statistic can be plagued by artifacts; for example, decaying aftershock sequences in a given area will produce a negative  $\beta$ -statistic that is unrelated to a mainshock. It is also not necessarily safe to assume that background seismicity follows a Poissonian distribution. However, the  $\beta$ -statistic provides a simple measure of seismicity rate change relative to a background rate.

To further explore the seismicity rate change away from the immediate epicentral region we average the results azimuthally to obtain  $\beta(r)$  (Figure 7A). We note that, as introduced by *Matthews and Reasenberg (1988)*,  $N_e$  represents a probability density function with equal probability over a +/-0.5 range bracketing whole numbers. Because the expected number of earthquakes cannot be negative, if there are no events in a pre-event window,  $N_e$  is set to 0.25. Thus if there are no events in a grid cell in the post-mainshock window ( $N_a=0$ ),  $\beta$  is not identically zero, but rather slightly negative, in this case approximately -0.7. To calculate  $\beta(r)$  we therefore include only those grid cells for which there is a seismicity change. One cannot ascribe any statistical significance to the very weakly positive  $\beta$  values

between 10 and 50 km (Figure 7B). However, previous results considering  $\beta$ -statistics averaged over large numbers of earthquakes suggest the small increase in seismicity over this distance range might reflect an extended aftershock zone (*Hough, 2005; Felzer and Brodsky, 2006*).

### **STATIC STRESS CHANGE**

We calculate the static Coulomb stress imparted by the Chino Hills earthquake to its aftershocks and to nearby portions of the Whittier and Chino faults, following *Lin and Stein (2004)* and *Toda et al (2005)*. We use both mainshock nodal planes as candidate sources in Figure 8A-B, in each case adopting the same strike, dip, and rake for the source and receiver faults. We note that Nodal plane 1 resembles the strike, rake, and location of the Whittier fault, and Nodal plane 2 would reach the surface near the San Jose fault. There are more aftershocks aligned with nodal plane 1 (Figure 4, section A-B) than with nodal plane 2 (Figure 4, section C-D), and so perhaps the nodal plane 1 is more likely the source. We assume a 300-bar (30 MPa) mainshock shear stress drop, a friction coefficient on receiver faults of 0.4, Poisson's ratio of 0.25, and a  $3.1 \times 10^{11}$  dyne-cm<sup>2</sup> shear modulus. For the map view plots, we calculate the maximum Coulomb stress change over the depth range of the aftershocks, 14-16 km.

Regardless of the source nodal plane, the calculated lobes of increased Coulomb stress appear consistent with the distribution of aftershocks (Figure 8A-B). In cross-section (Figure 8C-D), the majority of aftershocks are seen to lie in regions of calculated Coulomb stress increase, and few aftershocks occur in the lobes of stress decrease. We note that the two aftershocks with focal mechanisms in Figure 4 display vertical strike-slip mechanisms, which differ from the receiver faults used here. The stress increase lobes for such receiver faults are consistent with the aftershock distribution for the nodal plane 2 source.

We resolve the Coulomb stress changes on the projected surfaces of the Whittier and Chino faults in Figure 8E-F, under the assumption that they are right lateral-reverse faults (*Bjorklund and Burke, 2002, Bjorklund et al, 2002; Ludington et al, 2007*). We used a rake of 133°, the same as nodal plane 1. We used a dip of 75°; gentler dips would place the Chino Hills earthquake outside the uplifted wedge between the faults, which we regard as unlikely. For the nodal plane 1 and nodal plane 2 sources (Figure 8E-F), the area of the Whittier fault sustaining a Coulomb stress increase is considerably larger than the area with a stress decrease; the median stress change is +0.15 bars (0.015MPa) over the 15 x 20 km (length x width) fault patch closest to the Chino Hills source. For the Chino fault the changes are more balanced and the median stress change is +0.08 bars (0.008 MPa). Although few of the Chino

Hills aftershocks appear to coincide with these faults, the aftershocks occurring near the Whittier fault lie near the patches with the greatest calculated stress increase.

## **DISTRIBUTION OF INTENSITIES**

The Chino Hills event was widely felt throughout the greater Los Angeles region; as of 12 August 2008, over 40,000 people filled out the Community Internet Intensity Map (CIIM), also known as the “Did You Feel It?” questionnaire (Figure 9). These responses were used to determine Modified Mercalli Intensity (MMI) values for 802 Zip codes (see *Wald et al.*, 1999b; also <http://pasadena.wr.usgs.gov/shake/ca/>.) Because of the dense population of much of the greater Los Angeles region, intensity values averaged within Zip codes provide good spatial correlation to the location of the actual reporting sites (Figure 10A). We use the averages within Zip codes in our subsequent analysis. Our data set includes CIIM results collected as of 12 August 2008, a total of 40,677 individual questionnaires.

The averaged MMI values are first fit by a standard functional form,

$$MMI = A - br - c \log(r) \quad (2)$$

When fitting equation (2) using a least-squares approach, the distribution of average intensities clusters closely around the curve (Figure 10B), corroborating the conclusion that CIIM intensities provide a consistent measure of earthquake ground motions (e.g., *Atkinson and Wald*, 2006).

A further indication of the stability of CIIM intensities is the generally good correspondence between the CIIM distribution and the CISM ShakeMap, which is based on peak amplitudes from instrumental recordings (see <http://earthquake.usgs.gov/shakemap/>; Figure 2). The ShakeMap does, however, reveal somewhat higher intensities to the northwest and southwest of the epicenter than the reported CIIM intensities. That is, at sites where the maps do not agree, observed MMI values were lower than expectations given instrumentally determined peak amplitudes. We suggest this might indicate that the relatively deep event generated less surface wave energy than a shallow event.

Both the ShakeMap (Figure 2) and the CIIM map (Figure 9) that are very well constrained by data show systematic variations that appear to correlate with near-surface geological structure. In particular, the 1- and 3-sec spectral response ShakeMaps (see ShakeMap web site), also capture the amplification at most of the hotspots in the CIIM map. To explore further the distribution of intensities in the CIIM data, we calculate the residuals by subtracting  $MMI_{pred}$  from equation (3) at a given distance from the average  $MMI_{obs}$ , and compare the results to topography. To reduce scatter, we plot residuals at

locations for which four or more reported MMI are available, resulting in a map of residuals with a range of  $\pm 1$  unit (Figure 10A).

Focusing on the greater Los Angeles region, we tentatively identify a number of suggested systematic patterns in the distribution of intensities (Figure 10A). First, intensities are, as expected, generally lower at hard-rock sites, for example on the Palos Verdes Peninsula. Second, intensities are systematically higher in the deepest (central) Los Angeles Basin, consistent with a basin-depth effect (e.g., *Joyner, 2000*). Third, intensities are systematically lower to the west of the Newport-Inglewood fault, a pattern that correlates with subtle topographic relief in proximity to the fault zone and the shallower basin depth west of the fault. Fourth, there is a suggestion that intensities are systematically slightly higher at the western half of the San Fernando Valley, consistent with a shoaling effect at basin edges. Fifth, intensities are systematically higher to the north, northwest, and west, and lower towards the east even at sites in the San Bernardino basin where amplification is expected. These patterns are not easily explained by near-surface geology, and may reflect complex path and source effects.

## **DISCUSSION**

Both north-south thrust and left-lateral faulting on the west-southwest striking faults accommodate ongoing north-south compression of the Transverse Ranges and the compression and lateral block movements within the Los Angeles basin to the south (*Hauksson, 1994*). The moment tensor of the 2008 Chino Hills earthquake, which reveals oblique mixed thrust and left-lateral slip on an east-northeast or alternatively on a north-northwest striking nodal plane, is consistent with previously proposed models of the Los Angeles basin tectonics. Such crustal deformation both accommodates crustal shortening as well as westward translation of crustal blocks within the Los Angeles basin (*Hauksson, 1994; Walls et al., 1998*).

The unusual aspects of the sequence include a large difference between ML and Mw, low aftershock productivity, limited depth range of the sub-horizontal aftershock distribution, and high mainshock stress drop. The mainshock appears to have left behind almost no residual stresses thus causing only a few aftershocks near the mainshock fault plane. Most of the recorded aftershocks, are in three off-fault clusters, suggesting static or dynamic triggering from the mainshock. The high stress drop in the mainshock suggests high stresses at depth, near the brittle ductile transition zone, consistent with previous studies (e.g., *Jones and Helmberger, 1996*). The discrepancy between ML and Mw is further consistent with expectations for a high stress drop event. This and other previous high stress drop

earthquakes that maybe occurring within the northern edge of Peninsular Ranges batholith, suggesting possibly incipient faulting as the San Gabriel range front tectonic deformation migrates to the south.

Using detailed analysis of more than a decade of GPS data, *Argus et al.* (2005) pointed out the discrepancy between the measured geodetic and geological fault slip rates in the northern Los Angeles basin. Because the geologically measured fault slip rates are significantly lower, there may be other crustal deformation process accommodating the tectonic deformation. One possibility is aseismic slip along sub-horizontal decollement surfaces (*Fuis et al.*, 2001). However, there is no clear evidence for aseismic slip, either steady or episodic, along large decollement surfaces. In part, Global Positioning System (GPS) data may not be sampled frequently enough or may not be sensitive enough to detect deep and low amplitude slip events. Geodetic monitoring with higher sampling rates and sensitivity would be required to further explore the possibility that slow slip decollement events are occurring.

## **CONCLUSIONS**

Preliminary analysis suggests that the Mw5.4 Chino Hills earthquake was a high stress drop event, similar to other moderate earthquakes in the region in recent years. The low productivity and nearly complete absence of aftershocks in the vicinity of the mainshock hypocenter further suggests low residual stress near the mainshock rupture plane, which in turn suggests that the mainshock stress drop was nearly complete. However, the sparse aftershock distribution does not illuminate a rupture plane on the Whittier fault or any other mapped late Quaternary fault in the area. The sequence occurred in the limited depth range of 13 to 16 km, and thus provides new information about style of faulting and the state of stress, just above the brittle-ductile transition zone. The observed intensity distribution further implies correlations between intensity/damage and mainshock depth or stress drop. The static Coulomb stress modeling that is based on the moment tensor of the mainshock, suggests that the mainshock caused some loading on portions of the Whittier and Chino faults.

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## Figures

Figure 1. Map of (1981 to 2005) seismicity recorded by the SCSN, and some recent sequences in the Los Angeles Basin, including lower hemisphere focal mechanisms of the moderate-sized mainshocks. The 2008 Chino Hills mainshock is shown as a red star and the aftershocks as red circles. LAX- Los Angeles Airport, MB-Montebello; PA-Pasadena; UP-Upland; WN- Whittier Narrows.

Figure 2. The SCSN/CISN ShakeMap map of Modified Mercalli Intensities for the Mw5.4 Chino Hills mainshock: <http://earthquake.usgs.gov/eqcenter/shakemap/list.php?y=2008&n=sc>, (Wald *et al.*, 1999a). Triangles are seismic stations that reported strong ground motion. Mapped fault traces are from Jennings (1994).

Figure 3. (A) The focal mechanism of the mainshock “Cut-and-Paste” source estimation technique. (B) The misfit variations represented by the fit-error as a function of focal depth. The number above each focal mechanism image is the corresponding Mw. (C) Selected regional broadband displacement seismograms of the Chino Hill event. The synthetics are plotted in red and the data in black. The numbers above the station names are the epicenter distances in km and the numbers below are the azimuths. The first number below each seismogram is the time shift between data and synthetics (positive means synthetic is faster than the data, or synthetic has to shift back to fit the data) and the second number is the cross-correlation coefficient as percentage.

Figure 4. Map and  $\pm 3$  km wide cross-sections of the details of the seismicity. The symbol size is scaled linearly with magnitude. The first motion focal mechanisms for the M3.6 and M3.8 aftershocks are also shown.

Figure 5. Pulse widths of LA basin earthquakes; Adapted from Kanamori *et al.* (1993). The solid sloping line is the relationship for a global data set (Kikuchi, *written communication*, 2001; Kanamori and Brodsky, 2004).

Figure 6. Aftershock rate in terms of  $M \geq 2$  aftershocks per second as a function of time for the first ten days after the Chino Hills mainshock. Aftershock rate is solved from the time separation between successive aftershocks using the nearest neighbor method. The dashed line is the best fit Omori’s law with parameters  $p=1.09$  and  $K = 3.4$  where  $K$  is given in terms of  $M \geq 2$  aftershocks per day.

Figure 7. (A) Map of Beta statistics. (B) Beta versus distance showing only small variations.

Figure 8. (A-B) Maximum Coulomb stress change over the depth range of the aftershocks, using the two nodal plane solutions for both the source and receivers. Green lines indicate surface projection of the source to the ground surface, showing the mainshock nodal planes relative to the Whittier and San Jose faults. Black outlined white circles are aftershocks. (C-D) Static Coulomb stress change cross-sections with the aftershock plots; all aftershocks are projected onto the sections. (E-F) Static Coulomb

stress changes resolved onto the Chino and Whittier fault by using their strike, a 75° dip and 133° rake for the two candidate source faults, for nodal plane 1 and 2 sources. Black lines indicates top of the faults. All calculations using Coulomb 3.2 ([www.coulombstress.org](http://www.coulombstress.org)).

Figure 9. “Did You Feel it”; Community intensity map for the Mw5.4 Chino Hills mainshock: <http://pasadena.wr.usgs.gov/shake/ca/>, (Wald *et al.*, 1999b).

Figure 10. (A) Map consisting of gridded MMI (Modified Mercalli Intensity) residuals, intensity residuals in the greater Los Angeles area on a topographic base for delineating the variations in local terrain. (B) The attenuation curve for modified Mercalli Intensity (MMI) values from the data plotted in the “Did You Feel it Map”.

### 2008 Chino Hills Mw5.4 Earthquake Sequence

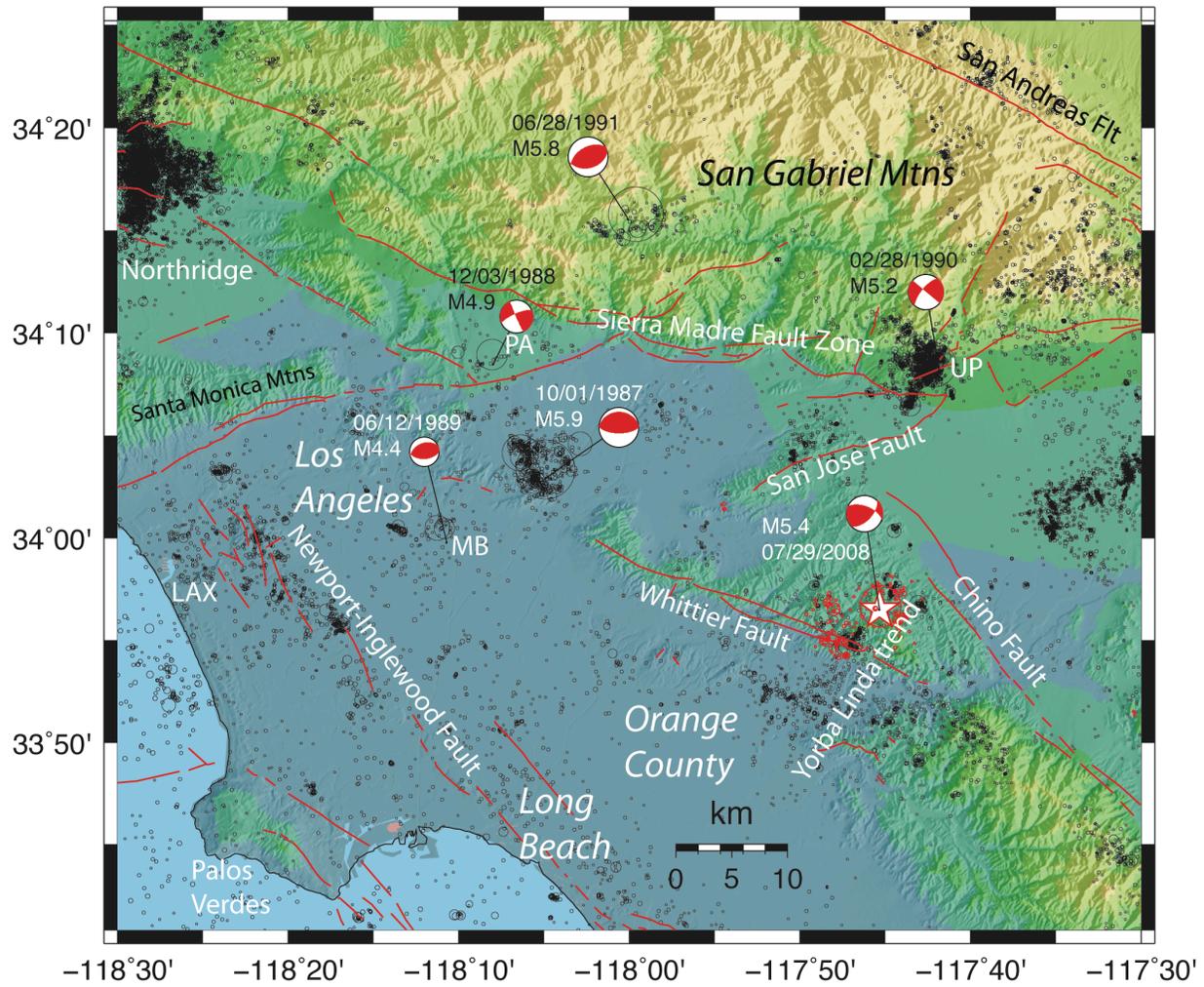
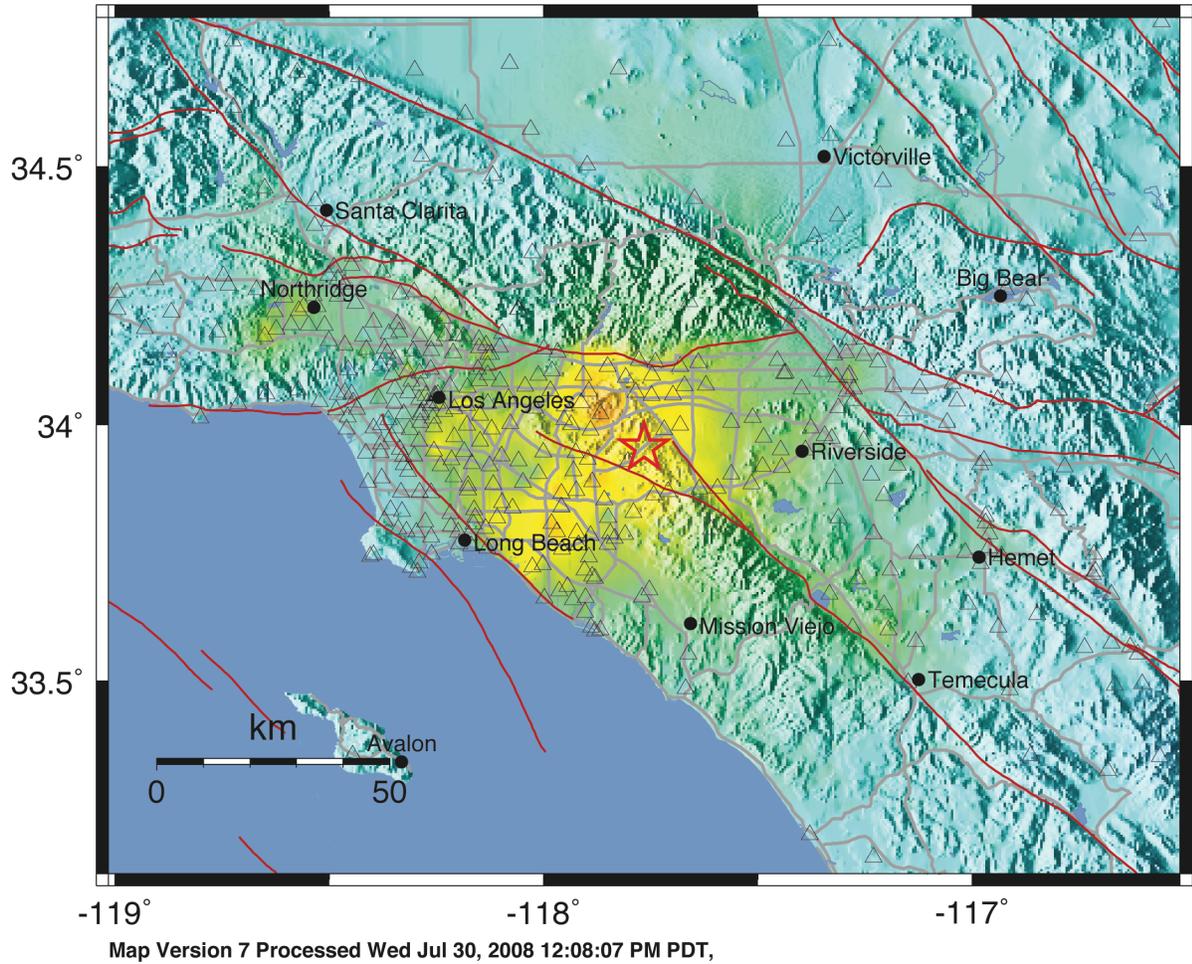


Figure 1

CISN ShakeMap : 4.0 mi SE of Diamond Bar, CA

Tue Jul 29, 2008 11:42:15 AM PDT M 5.4 N33.96 W117.76 Depth: 13.7km ID:14383980

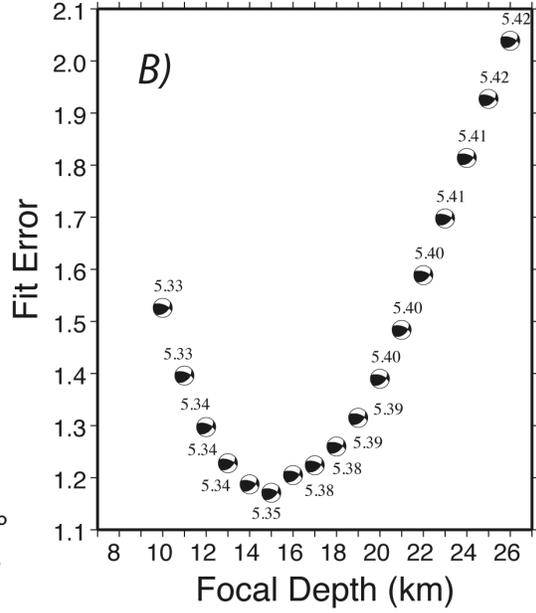
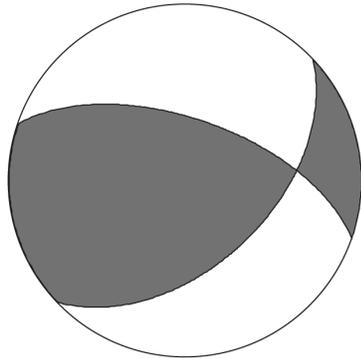


PERCEIVED SHAKING	Not felt	Weak	Light	Moderate	Strong	Very strong	Severe	Violent	Extreme
POTENTIAL DAMAGE	none	none	none	Very light	Light	Moderate	Moderate/Heavy	Heavy	Very Heavy
PEAK ACC.(%g)	<.17	.17-1.4	1.4-3.9	3.9-9.2	9.2-18	18-34	34-65	65-124	>124
PEAK VEL.(cm/s)	<0.1	0.1-1.1	1.1-3.4	3.4-8.1	8.1-16	16-31	31-60	60-116	>116
INSTRUMENTAL INTENSITY	I	II-III	IV	V	VI	VII	VIII	IX	X+

Figure 2

### Mainshock Mw5.4

A)



Strike: 289°; dip: 62°; rake 133°  
Strike: 46°; dip: 50°; rake 38°

C)

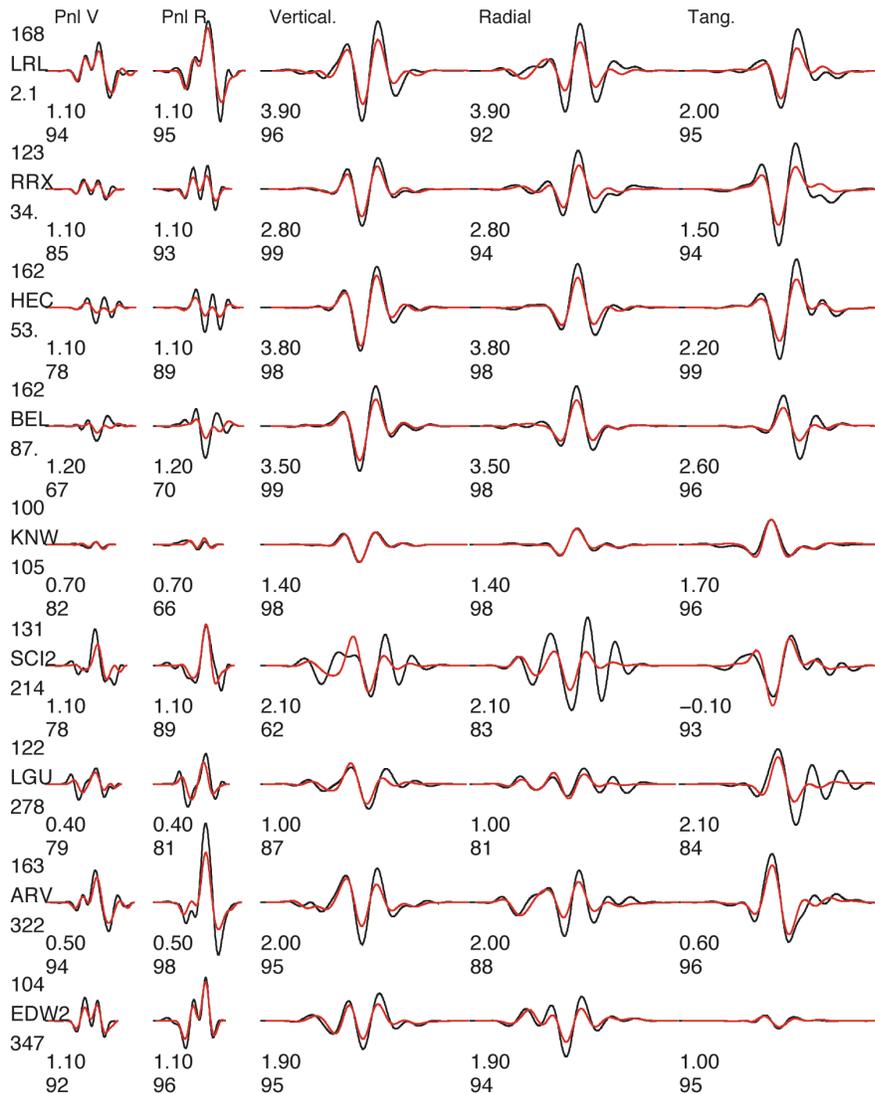


Figure 3

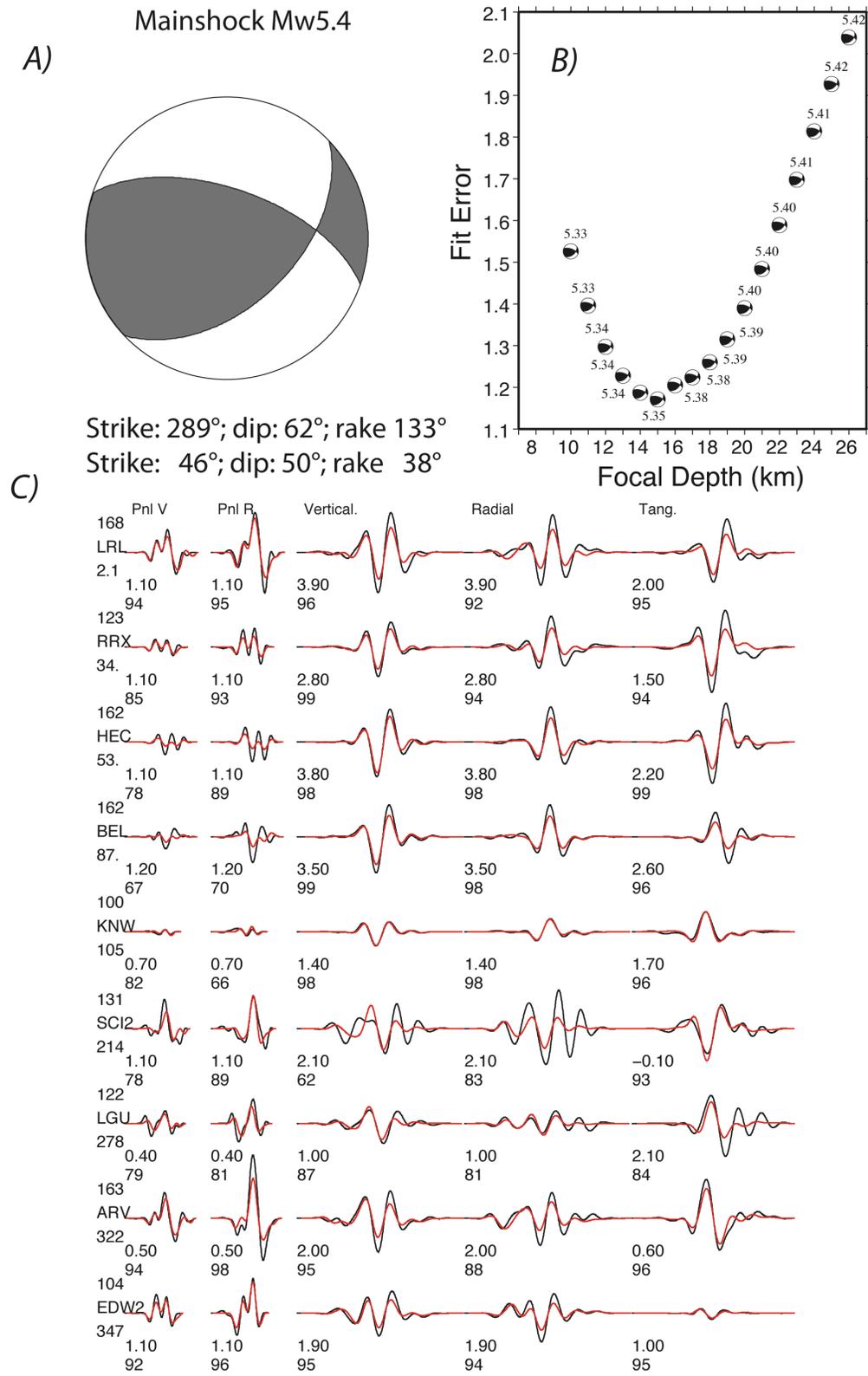


Figure 4

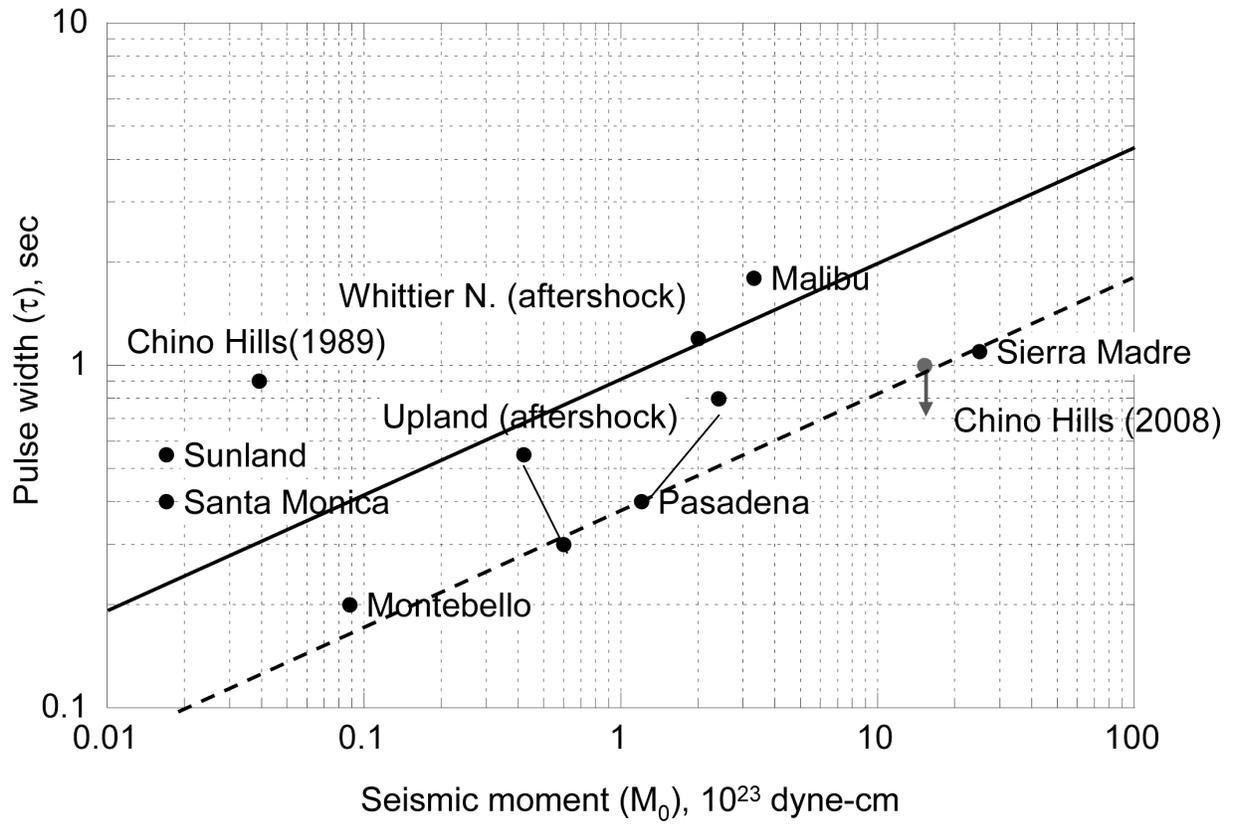


Figure 5

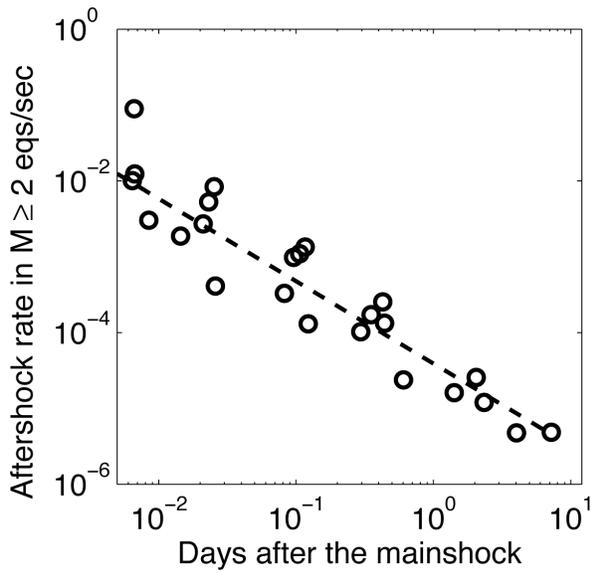


Figure 6

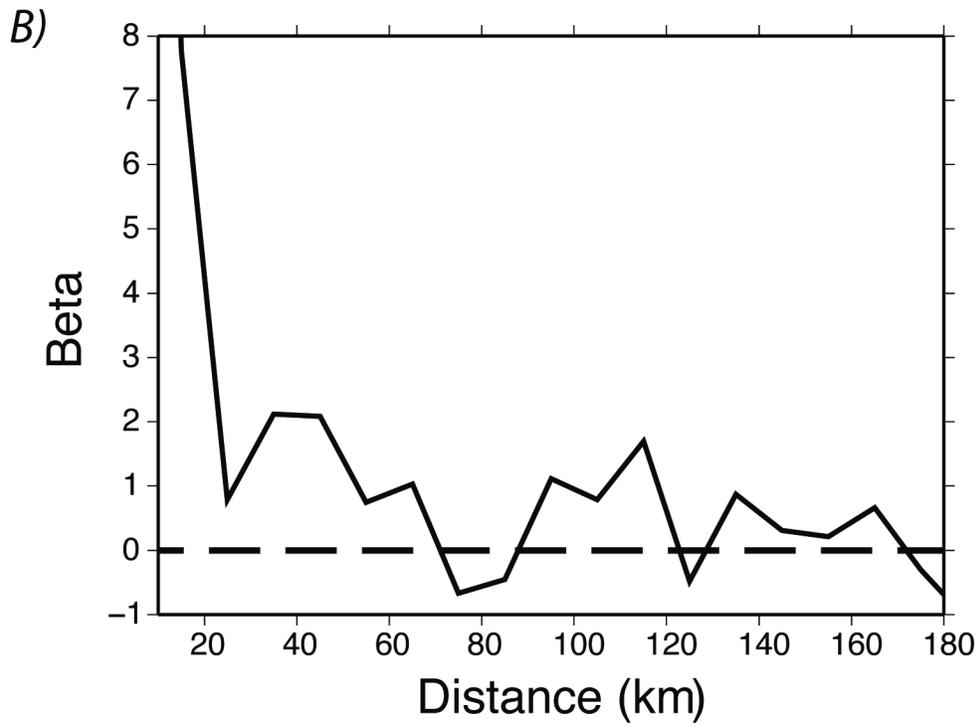
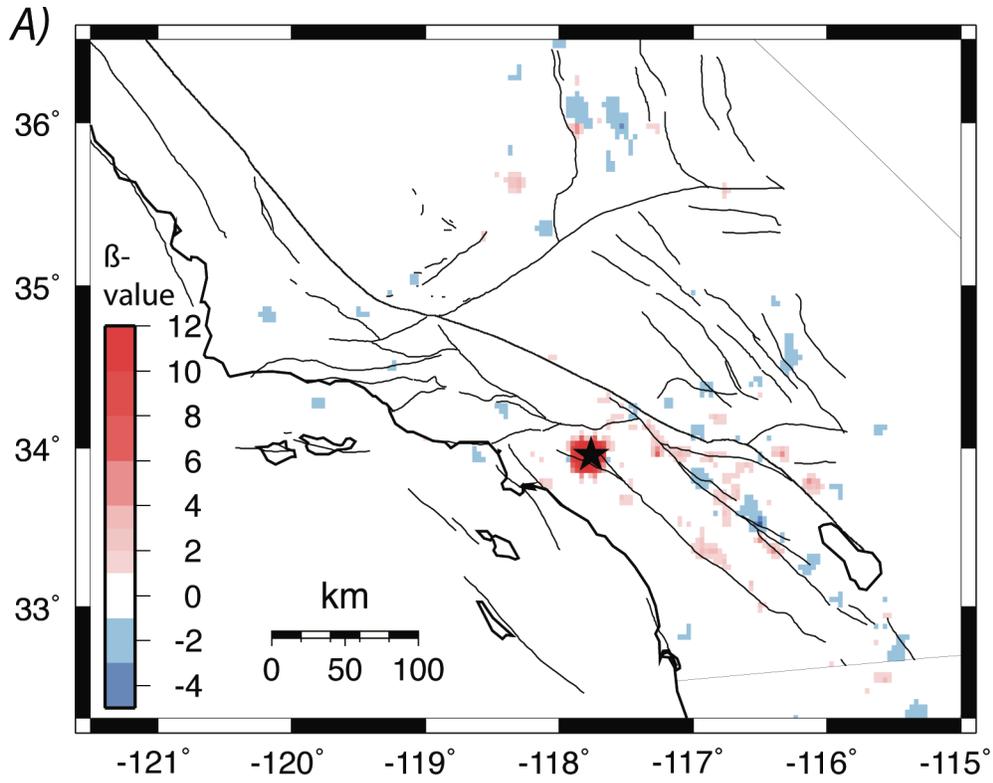


Figure 7

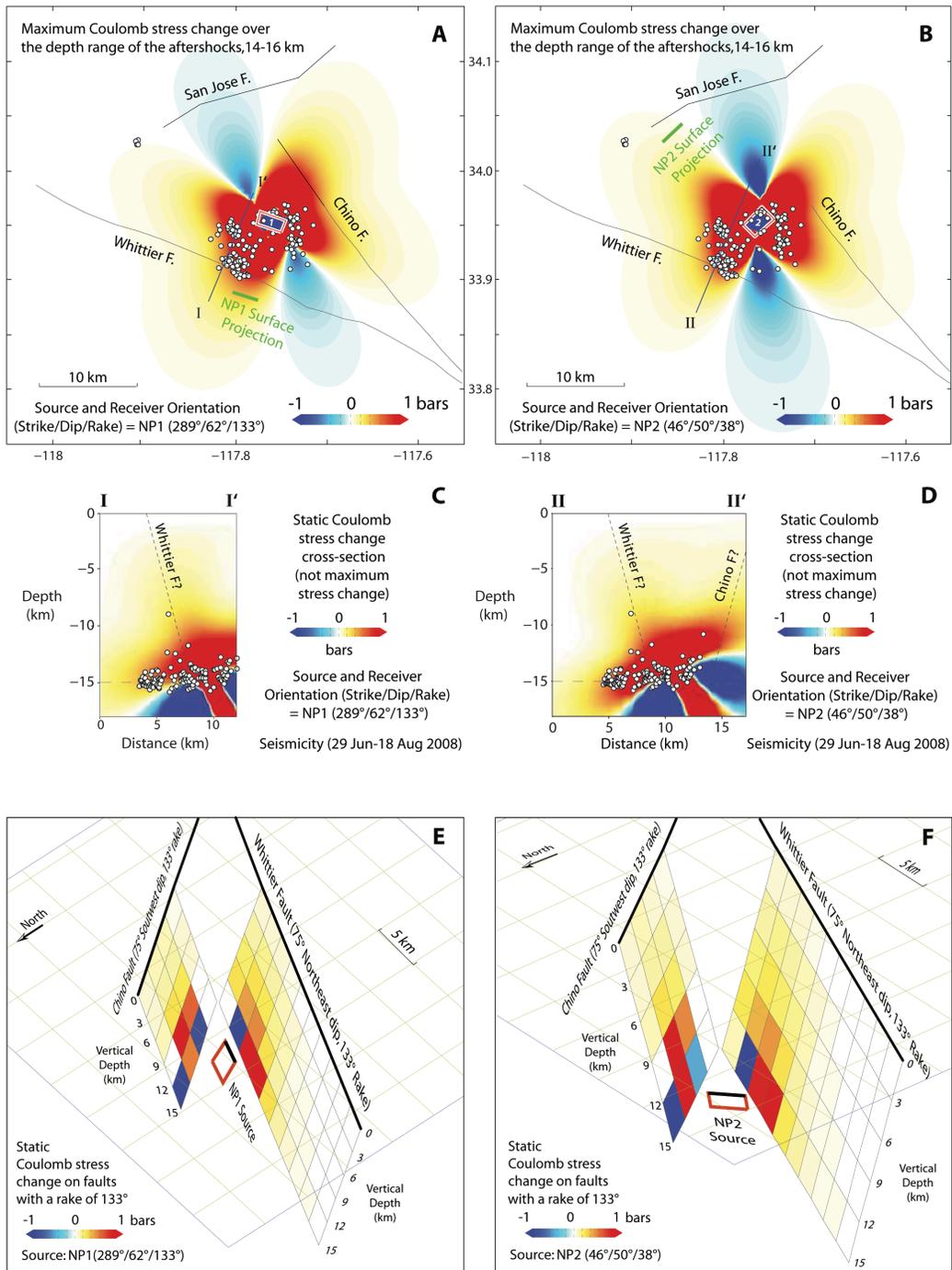
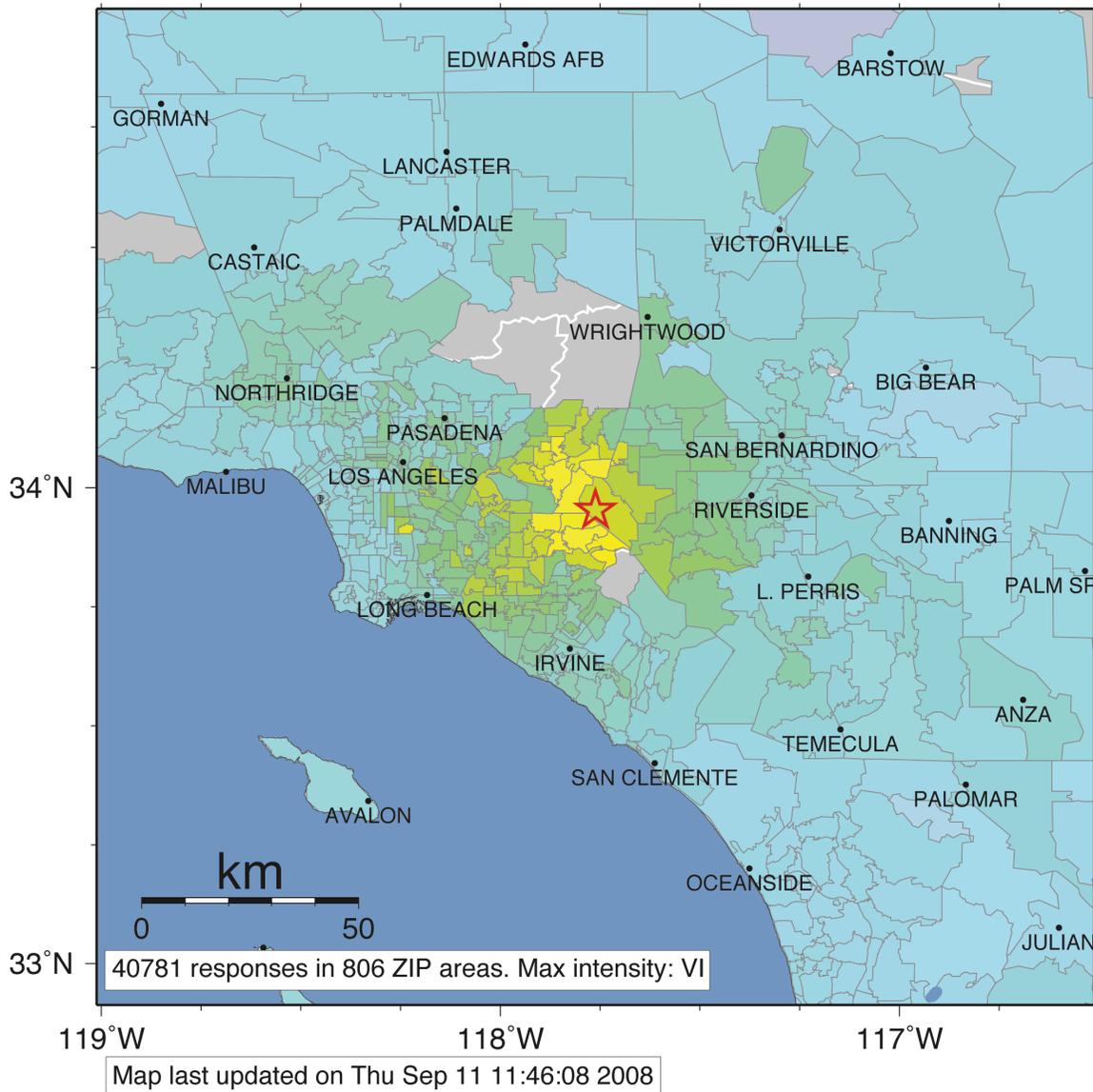


Figure 8

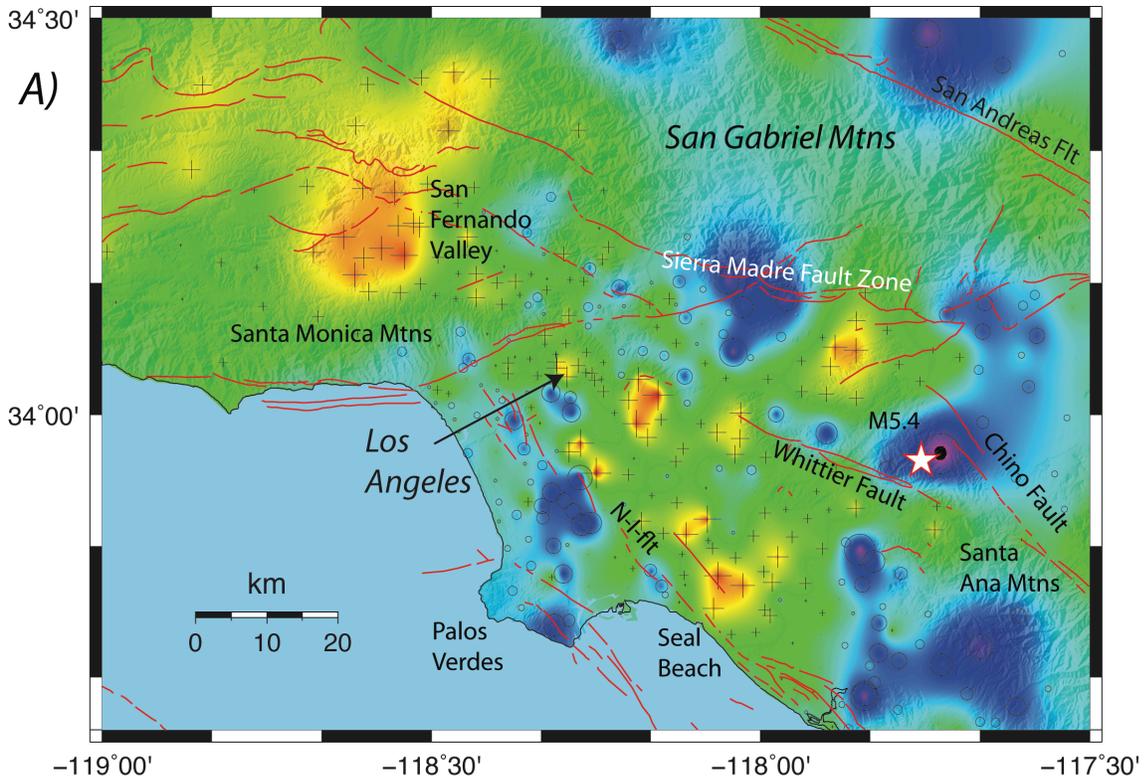
### USGS Community Internet Intensity Map (3 miles SW of Chino Hills, CA)

ID:14383980 11:42:15 PDT JUL 29 2008 Mag=5.4 Latitude=N33.95 Longitude=W117.76



INTENSITY	I	II-III	IV	V	VI	VII	VIII	IX	X+
SHAKING	Not felt	Weak	Light	Moderate	Strong	Very strong	Severe	Violent	Extreme
DAMAGE	none	none	none	Very light	Light	Moderate	Moderate/Heavy	Heavy	Very Heavy

Figure 9



Modified Mercalli Intensity Residuals

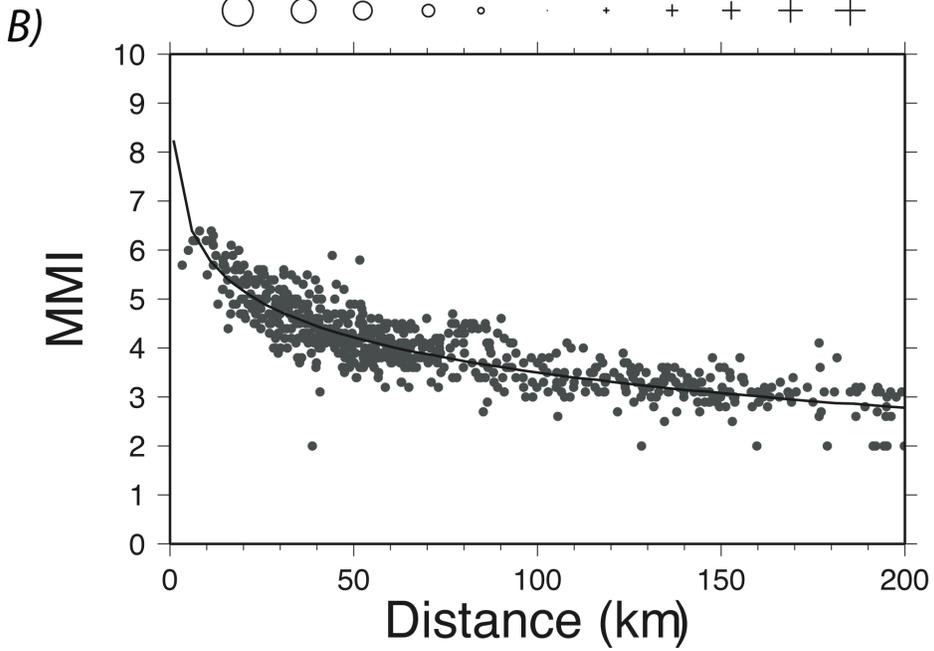
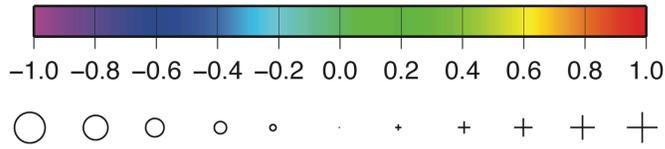


Figure 10