Stress drop heterogeneity within tectonically complex regions: A case study of the San Gorgonio Pass region, southern California

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Abstract

In general, seismic slip along faults reduces the average shear-stress within earthquake source regions, but individual stress drops during earthquakes are observed to vary widely in size. The details of how crustal and fault properties influence variations in stress drop are poorly understood. To advance our understanding of variations in stress drop, we analyze source parameters of small and intermediate magnitude events within the greater San Gorgonio Pass region, southern California. The tectonics within the region are controlled by a restraining bend of the San Andreas fault system, which results in distributed deformation, and heterogeneous slip along numerous strike-slip and thrust faults.

Stress drops are computed by fitting a Brune-type spectral model to individual event spectra obtained through separating the observed spectra into site, path and source contributions. The latter are obtained by iteratively removing stacked site and path terms and correcting high frequency contributions using a regional empirical Green’s function. The stress drop estimates show strong
regional variations from \( \sim 1 \) to \( \sim 25 \) MPa with a median of 4.8 MPa. We observed anomalously high stress drops (>20 MPa) in a small region between the traces of the San Gorgonio and Mission Creek segments of the San Andreas fault. Detailed analyses of focal mechanisms reveal that stress drops are slightly higher for thrust faulting events (\(~6\) MPa) than for normal events (\(~3.5\) MPa). The estimated stress drops also increase below depths of \(~10\) km and along the San Andreas fault segments, both from north and south, towards San Gorgonio Pass (SGP), showing a negative correlation with geologic slip rates. To test the stability of our results, we conducted a sensitivity analysis of input parameters and event selection criteria, confirming the robustness of the observations. We identified crustal conditions and fault properties that contribute to local variations in stress drop estimates including the style of faulting, changes in average tectonic slip rates, mineralogical composition of the host rocks, as well as the hypocentral depths of seismic events. A detailed spatial mapping of stress drop variations can thus advance the assessment of expected earthquake ground motions.

1 Introduction

The relative motion of tectonic plates generally causes stress to build up along systems of faults. These stresses are released during earthquakes. The spatial variations in absolute stresses during earthquakes can generally not be determined directly, however, the relative decrease in shear-stress can be estimated from the radiated seismic spectrum. Stress drop estimates are based on a deconvolution of the seismic record into source, site and path effects. The seismic moment and corner frequency of the source spectrum can be used to determine rupture dimensions and stress drops if the aspect ratio and propagation speed of the rupture are assumed to be constant (e.g. Eshelby, 1957; Knopoff, 1958; Brune, 1970; Madariaga, 1976; Boatwright et al., 1991).

1.1 Earthquake scaling relations and self-similarity

A detailed description of source parameter variations fundamentally influences our understanding of earthquake physics including expected ground motions (e.g. Hanks and McGuire, 1981) and scaling relations (e.g. Hanks and Thatcher, 1972; Prieto et al., 2004; Walter et al., 2006). Self-similar earthquake scaling requires that stress drops remain constant and fault slip increases as a function of rupture area (e.g. Prieto et al., 2004; Shearer, 2009). As a consequence, physical processes involved
in small and large magnitude earthquakes are inherently similar (e.g. Aki, 1981).

If the corner frequency of the source spectrum increases with magnitude, the energy radiation is partitioned differently over the earthquake frequency spectrum, causing large stress drop events to contain a relatively higher proportion of high-frequency energy. This has large implications for the expected ground motion of a particular size earthquake, i.e., seismic events with relatively high stress drops radiate more high frequency energy than low stress drop events (e.g. Hanks, 1979; Hanks and McGuire, 1981; Heaton et al., 1986).

Some studies of source parameter scaling relations indicate self-similar scaling between corner frequencies and moments for regional data sets and mining induced seismicity (e.g. Abercrombie, 1995; Ide and Beroza, 2001; Prieto et al., 2004; Baltay et al., 2010; Kwiatek et al., 2011) whereas other studies highlight deviation from self-similarity on regional and global scales (e.g. Kanamori et al., 1993; Harrington and Brodsky, 2009; Lin et al., 2012). The assessment of earthquake stress drops over a range of magnitudes is further complicated by near-surface attenuation. Attenuation is especially problematic for small events and high-frequencies, which can cause an artificial break-down of self-similar scaling (Abercrombie, 1995). High frequency attenuation, limited recording bandwidths and low quality records add to the controversy of self-similar source parameter scaling which has not been resolved at present.

1.2 Fault properties, crustal parameters and stress drop variations

In addition to magnitudes, stress drops are influenced by local crustal conditions. For example near Parkfield, seismic off-fault events show largely self-similar scaling whereas some events on the San Andreas fault exhibit the same source pulse width, independent of event magnitudes (Harrington and Brodsky, 2009). The independence of pulse width and earthquake magnitude results in stress drops between 0.18 and 63 MPa. High stress drops for on-fault events were also suggested by Nadeau and Johnson (1998). While the Parkfield studies show locally higher stress drops, a study of earthquakes in southern California found no correlation between stress drop and distance from major faults (Shearer et al., 2006), and a study of global earthquakes with M>5 revealed higher stress drops for intraplate compared to plate boundary events (Allmann and Shearer, 2009). Elevated stress drops for intraplate events may be due to higher crustal strength and stresses far from active faults.

Stress drops may also be sensitive to the type of tectonic regime. For example in southern Cali-
fornia, Shearer et al. (2006) identified higher-than-average stress drops in some regions containing a relatively high fraction of normal-faulting events whereas the mainly reverse-faulting aftershocks of the Northridge earthquake have lower-than-average stress drops. In contrast, the global study of Allmann and Shearer (2009) found higher-than-average stress drops for strike-slip events. Furthermore, stress drops are observed to be lower for regions of relatively high heat flow in Japan (Oth, 2013) and increase with depth, for example, in southern California (Shearer et al., 2006; Yang and Hauksson, 2011) and Japan (Oth, 2013). In addition to fault proximity, tectonic regime, heat flow and depth, stress drops have also been observed to vary as a function of recurrence intervals and loading rates in the laboratory and nature (e.g. Kanamori et al., 1993; He et al., 2003). Slower loading rates and longer healing periods within interseismic periods lead to an increase in asperity strengths and stress drops (Beeler et al., 2001).

In this study, we investigate stress drop variations close to the San Andreas fault in Gorgonio Pass. This region provides an ideal natural laboratory to study stress drop variations because of its high seismic activity, station density and well-studied tectonic setting. We first review the tectonic setting (Section 2), then introduce the method for estimating source spectra and stress drops largely following Shearer et al. (2006) (Section 3). We determine spatial variations in stress drops and assess their reliability (Section 4.1–4.2). We then perform a detailed analysis of crustal parameters that may influence stress drop variations (Section 4.3–4.5). More details about the stress drop computations can be found online in the supplementary material.

2 Seismic data and tectonic setting

2.1 Seismicity catalogs and waveform data

We analyzed seismicity data and seismic waveforms within the greater SGP area (Figure 1). Our analysis was based on three different types of data: (1) A relocated earthquake catalog that improved single event location by using a 3D velocity structure, source-specific station terms and relative travel-time differences from waveform cross-correlations of event clusters (Shearer et al., 2005; Hauksson et al., 2012); (2) focal mechanisms, estimated from first-motion polarities and amplitude ratios of P- and S-waves (Yang et al., 2012); (3) seismic waveforms, obtained from the SCEC data center which we used to determine source spectra and source parameters. We limited our
analysis to events that were recorded at broadband stations. These stations show a largely consistent frequency response from $\sim 0.2$–50 Hz with a sampling frequency of 100 Hz. This wide frequency band is helpful for improving the resolution of corner frequencies and high-frequency fall-offs compared to previous studies. The broadband data are available for a dense array of stations in southern California starting in 2000. We selected a period from 2000 to 2013 because of the availability of relatively homogeneous waveform records, station instrumentation and seismicity catalogs. During this period over $\sim 11,300$ seismic events with magnitudes in the range of $M_L=0$–4.88 occurred within the study region. The largest event occurred near the San Bernardino segment of the San Andreas Fault in June 2005 (see Figure 2).

[Figure 1 about here.]

2.2 Tectonic complexity within the SGP region

The study area is crosscut by several faults that comprise the San Andreas fault system. The San Andreas fault system is characterized by relative structural simplicity within the Coachella segment to the southeast and the Mojave segment to the northwest of SGP (Figure 1). The SGP region itself is marked by complex, distributed crustal deformation. Tectonic slip within this region is accommodated by systems of strike-slip and thrust faults (Allen, 1957). These fault segments include the Garnet Hill and Banning segments to the northwest of the Coachella segment, followed by the San Gorgonio thrust fault (SGF), Wilson Creek and San Bernardino segments and the Mill and Mission Creek segments north of SGP (Figure 2). The Banning segment became seismically less active about 5 Myr ago (e.g. Yule and Sieh, 2003). Consequently, the slip on the San Andreas fault system partially by-passes the SGP region, for example, via the San Jacinto fault to the west (Allen, 1957; Yule and Sieh, 2003; Langenheim et al., 2005; McGill et al., 2013).

The San Andreas fault within the SGP region lacks continuity because the regional deformation is strongly influenced by a restraining step within the Mission Creek section (Figure 2a). As a result, several secondary fault strands exist, which are oriented unfavorably with respect to the tectonic plate motion, leading to large-scale transpressional tectonics (Carena et al., 2004; Langenheim et al., 2005; Cooke and Dair, 2011). This tectonic complexity is also articulated in the distribution of seismic events, which occur preferably off the main fault strands of the San Andreas fault (e.g. Yule and Sieh, 2003). Similarly, the tectonic complexity can be observed in the diversity of focal mechanisms.
which show predominant oblique sinistral slip above 10 km. In contrast, below 10 km depth, oblique strike-slip, normal and thrust faulting accommodate east-west extension and north-south compression (Nicholson et al., 1986). The thrust faulting within the SGP region resulted in a high magnetic anomaly, likely caused by the wedging of Peninsular range rocks underneath Transverse range material and the presence of deep, magnetic rocks of San Bernardino or San Gabriel basement types (Langenheim et al., 2005). The convergence rates within this area are estimated at 1–11 mm/yr (Yule et al., 2001; Langenheim et al., 2005). The long term fault slip rates decrease systematically when approaching the SGP region from the North and South from 24.5±3.5 and 14–17 mm/yr respectively down to 5.7±0.8 mm/yr (Dair and Cooke, 2009; Cooke and Dair, 2011; McGill et al., 2013).

Since the 1940s, three mainshocks close to and above M = 5.0 have been recorded within the study area: 1) the 1986 M = 5.6 North Palm Springs, 2) the 1992 M = 6.4 Big Bear, and 3) the 2005 M = 4.9 Yucaipa earthquake (Figure 2a). In addition, three large earthquakes were recorded nearby, i.e., the 1948 M = 6.0 Desert Hot Springs and 1992 M = 6.1 Joshua Tree earthquakes to the east and the 1992 M = 7.3 Landers earthquake to the northeast.

Seismicity becomes deeper north of the SGF, which dips at about $\sim 55^\circ$ underneath the San Bernardino mountains (Figure 2b). The base of the seismicity beneath the San Jacinto mountains dips gently to the north (Figure 2b). This is followed by an abrupt step in the seismicity base from $\sim 21$ to 13 km below the Mission Creek segment. This step marks the boundary between Peninsular and Transverse range rocks (see also Nicholson et al., 1986; Yule and Sieh, 2003). The depth profile of the relocated seismicity catalog suggests that the seismicity step may be slightly disturbed by the presence of the San Gorgonio thrust, leading to seismically active under-thrusting of Peninsular range rocks beneath the Transverse ranges. Based on mapped surface traces and approximate fault dip angles (Fuis et al., 2012), we connected fault surface expression with deep seismicity clusters at depth (Figure 2). The SGF is approximately co-located with the transition between deep seismicity to the south and shallower seismicity to the north. Faults to the South generally lack seismicity above $\sim 5$ km whereas faults to the North (e.g. Mission and Mills Creek) produce seismic events from shallow depths down to 14–15 km.

[Figure 2 about here.]
3  Method: Source spectra inversions and stress drop estimates

Instead of estimating source parameters from individual spectra, we inverted the entire data set for average event, path and station terms thus diminishing the influence of high-frequency fluctuations, radiation patterns and source directivity (e.g. Andrews, 1986). Amplitude spectra were computed for tapered waveforms within a 1.28 s time window after the P-wave arrival. For the spectral inversions, we required a signal-to-noise ratio (SNR) above 5 within three different frequency bands (5–10, 10–15, 15–20 Hz) as well as at least 5 station picks per event. The observed waveforms are a convolution of source, path and site contributions. The convolution changes to a multiplication in the frequency domain and to a summation in the log-frequency domain:

\[ d_{ij} = e_i + t_{ij} + s_j, \] (1)

where \( d_{ij} \) is the logarithm of the recorded amplitude spectrum, \( e_i \) and \( s_j \) are the event and station terms and \( t_{ij} \) is the travel time term between the \( i^{th} \) event and station \( j^{th} \) (see also Suppl. Fig. S1). All of these terms are frequency-dependent.

The path term was discretized by binning at 1-s intervals according to the corresponding P-wave travel times. This system of equations was then solved iteratively by estimating event, station and path terms as the average of the misfit to the observed spectra minus the other terms (e.g. Andrews, 1986; Warren and Shearer, 2000; Shearer et al., 2006; Yang et al., 2009). For robustness, we suppressed outliers by assigning L1 norm weights to large misfit residuals. The robustness of the spectral inversion method was also verified previously by comparing path terms with expectations from a frequency-independent attenuation model (Shearer et al., 2006) and by analyzing a synthetic data set (Allmann and Shearer, 2007). The spectral-stacking does not take differences in focal mechanisms into account which are a potential source of uncertainty within the source spectra estimates (e.g. Kaneko and Shearer, 2014). However, the differences in corresponding radiation patterns were diminished by stacking spectra from many stations thus averaging over the focal sphere.

We estimated the relative seismic moment, \( \Omega_0 \), for individual source spectra from the corresponding low-frequency contributions by averaging the spectral amplitudes from \( \sim 2–4 \) Hz. This frequency range is above the smallest corner frequencies. We then calibrated the relative moments using the catalog magnitudes, assuming that the low-frequency amplitudes are proportional to
moment, and that the catalog magnitude is equal to the moment magnitude at $M_L = 3$ (see Shearer et al., 2006, for details). The source spectra were then binned according to estimated local magnitude using a 0.2 spacing and corrected using a regional Empirical Green’s Function (EGF) approach. The EGF is estimated by simultaneously fitting a constant stress-drop Brune-type spectral model to the magnitude-binned spectra between 2–20 Hz (Figure S2). The spectral model has the following form (Brune, 1970):

$$u(f) = \frac{\Omega_0}{1 + (f/f_c)^2}$$  \(2\)

where $u(f)$ is the source spectra, $\Omega_0$ is the low frequency spectral amplitude, and $f_c$ is the corner frequency. For a circular, isotropic rupture and constant rupture velocity, the stress drop ($\Delta\sigma$) and corner frequency are related by (Eshelby, 1957; Madariaga, 1976):

$$\Delta\sigma = M_0 \left( \frac{f_c}{0.42\beta} \right)^3$$  \(3\)

where $M_0$ is the seismic moment and $\beta$ is the shear wave velocity. Initially, we assumed a constant reference shear velocity of 3.5 km/s. We then tested the sensitivity of stress drop variations to changes in $\beta$, which is discussed in detail in Section 4.3. Changes in rupture velocities and in the scaling between corner-frequency and rupture extent affect stress-drop estimates strongly due to the cubed dependency on corner-frequencies. For example, Brune (1970) assumed that corner-frequency ($f_c$) and rupture dimension ($r$) are related over $f_c = k\beta/r$, with $k = 0.37$, whereas here we assume a value of $k = 0.32$ for far-field P-wave radiation, based on Madariaga (1976). Nevertheless, these changes in scaling constants alter only the absolute value of stress drops whereas relative changes remain constant so that the in the following described spatial variations in stress drop estimates are not affected.

The source parameters of individual events are determined by fitting a Brune-type spectral model to the source spectra after correcting the high-frequency contributions using the regional EGF. Variations in spectra, corner frequencies and stress drops are shown in Figure S3 for events with similar relative moments. The described method enables us to analyze large seismic data sets in a uniform way to obtain reliable estimates of relative differences in source parameters, e.g. stress drops. The absolute stress drop values may be sensitive to some of the modeling assumptions (e.g. constant rupture velocities in Equation 3, and fixed rupture aspect ratio) which has to be considered when comparing our results to other studies. Furthermore, uncertainties in corner-
frequency estimates, or differences in the assumed relationship between rupture extent and corner-
frequencies have a large influence on stress-drop estimates because of the cubed dependency on $f_c$. 
Nevertheless, overall stress drop variations can be interpreted with greater confidence compared 
to the analysis of individual spectra or event pairs.

4 Results

The average stress drop of the stacked source spectra for the entire region was $\Delta \sigma = 6.1$ MPa (Figure S2b), and the median value of individual events was $\Delta \sigma = 4.8 \pm 0.1$ MPa (Figure S5) assuming log-normal-distributed data. The approximate agreement between these two values is one indication of the robustness of our spectral inversion method. Stress drop estimates based on source spectra of small-magnitude earthquake are generally subject to large scatter. This scatter maybe related to different sources, for example, uncertainties in the spectral inversion, deviations from the applied, simplistic source model, as well as uncertainties in corner-frequency and seismic moments estimates. Nevertheless, part of the variations in stress drops may also have underlying, physical causes which will be investigated in the following. For a more detailed presentation of uncertainties and misfits between observed and modeled spectral shapes see the supporting information in the online version of this article. In the following section, we show spatial variations in stress drop estimates and analyze their robustness.

4.1 Spatial variations in stress-drops

To assess the spatial variations of individual earthquake stress drops, we smooth the results using a spatial median filter for the closest 60 epicenters to a 2-D uniform grid within a maximum area of $r = 5$ km. The maximum kernel width is chosen to avoid associating median stress drops with distant events. The resulting map displays gradual variations in stress drop estimates from values of $\sim 2$ MPa up to $\sim 25$ MPa (Figure 3). The most striking feature in Figure 3 is the region of anomalously high stress drops between the SGF and Mill Creek fault traces. Within this area, stress drops change rapidly (from north to south along longitude = 116.8°W) from $\sim 5$ MPa up to $>20$ MPa and back to $<5$ MPa. In addition, we observe several regions of increased stress drop estimates, for example, located close to the San Jacinto fault [-117.08, 33.9] and south of the San Bernardino segment [-117.05 34.07]. The dark red to orange regions highlight areas with stress
drops between 2 to 8 MPa (see legend in Figure 3).

Before probing different mechanisms that could explain the observed variations in stress drop estimates, we tested the robustness of our results. We started by investigating the difference between the high and low stress drop regions (green and red circle in Figure 3) focusing on the relation between corner frequencies and moment. We created a subset of data containing events within the two regions and performed a separate inversion for source spectra and source parameters. This inversion incorporates the estimation of a local EGF, which accounts for possibly unmodeled lateral variations in attenuation using the regional EGF for the entire study area. In case of systematic differences in source spectra, we expect to observe also systematic differences in corner frequency and stress drops for different magnitude events. Our tests confirmed this expectation so that seismic moment and corner frequency exhibit consistently higher ratios for high compared to low stress drop regions in log-log space (Figure 4). Based on the corresponding stress drop distributions, we compute median values of $\Delta \sigma = 1.4$ MPa and 18.7 MPa assuming log-normal distributed data for low and high stress drop regions. These values are comparable to the values for the same regions in Figure 3.

Following the analysis of corner frequency and moment, we compared the relative frequency content of seismic event waveforms within the low and high stress drop regions. To this aim, we juxtaposed low and high stress drop source spectra after normalizing spectral amplitudes by moment and frequencies by the corner frequency derived from equation 3 based on the regional median stress drop (Figure 5). This re-scaling corrects for differences in moment within the individual regions but also shows the differences in frequency content of individual events, thus providing a qualitative estimate of variations in corner frequency. In the case of constant, estimated stress drops, as observed for the regional source stacks (see Figure S2), the shifted source spectra collapse on the same curve. However, the present data subsets display strong variations within the two different regions: Low stress-drop events have lower corner frequencies and plot further to the left (Figure 5a), whereas high stress drop events exhibit relatively higher corner frequencies and plot further to the right (Figure 5b). Consequently, the relative difference between spectra within the low and high stress drop region further supports the reliability of observed spatial variations in
stress drops. (More details about differences in spectra for events with different stress drops are shown in Figure S3).

[Figure 5 about here.]

4.2 Sensitivity analysis of stress-drop computations

To investigate the dependence of source inversion results on input parameters, we conducted a sensitivity analysis of selection criteria for the input spectra. The details of the sensitivity analysis can be found in the supplementary material. The analysis generally confirmed the relative differences between low and high stress drop regions but also showed that the absolute stress drops may vary as a function of input parameters and connected data selection criteria. Limiting the analysis to records with many station picks had a larger influence on stress drops then choosing only high SNR records. Nevertheless, the sensitivity analysis demonstrated that relative variations in stress drops can be identified reliably if the input parameters are chosen consistently though absolute values may vary.

4.3 Stress drop variations with depth

To test the influence of hypocentral depths and to examine possible lateral variations as a function of depth, we constructed smoothed stress-drop maps for three different depth ranges (Figure 6). Because there are few events above 5 km depth, we chose the first depth layer from 0–10 km, the second from 10–15 km and the third for events from 15–25 km. We observed a systematic difference in stress drops between the depth layers. The shallow events (0–10 km) were dominated by low stress drops, the intermediate depth layer includes some of the high stress drops and the deepest events clearly highlight the area of anomalously high stress drops between the San Gorgonio and Mission Creek fault traces. As expected, the intermediate and the bottom depth layers do not show the low stress drop region towards the north edge of the study region, which was dominated by relatively shallow events (see Figure 2b).

[Figure 6 about here.]

Motivated by the observation of stress drop variations for different depth layers, we probed for a general correlation between focal depths and stress drops. Stress drops for events shallower than
10 km are low, with average values from 2.6–3.0 MPa. At ∼10 km the average stress drops increase abruptly to ∼4.8 MPa. At depths from ∼10–17 km, average stress drops continue to increase gradually up to ∼5.5 MPa before decreasing to 5.3 MPa at 20 km depth. We tested whether these observations could be explained by variations in rupture velocity, assuming that rupture velocity is proportional to S velocity changes with depth. We used a regional velocity model (Langenheim et al., 2005), which has a high velocity anomaly just beneath the SGP region. We corrected our initial stress-drop estimates using two different depth profiles that capture the average seismic velocity changes beneath and outside of the SGP region, including a relatively high velocity zone at about 7–13 km depth (Figure 7b). The results are shown by the round markers in Figure 7a. Including a depth-dependent change in rupture velocity affected the variations in stress drops only marginally. This is expected because most of the variations in seismic velocities are located close to the surface from 0–6 km whereas the largest changes in stress drops are at greater depths. The rupture velocity ($V_r$) would have to change abruptly by a factor of 1.2 near 10 km to compensate the observed increase in stress drop with depth, but the inferred increase in $V_r$ at this depth is only about 3%.

The analysis of stress drop variations with depth revealed large values for relatively deep events (below 10 km). To put this finding into the seismo-tectonic context of the SGP region, we mapped stress drops of individual events along the depth cross-section highlighted in Figure 3. The previous results of lower stress drops above 10 km are supported by the overall stress drop distribution (Figure 8a). However, we also observed a relatively dense cluster of high stress drop events in immediate proximity to the seismicity step extending from the base of the seismicity up to the SGF. This region marks the location of the deepest earthquakes within the study area. The transition to the hanging wall of the SGF is characterized by a noticeable decrease in stress drops. Similarly stress drop decreases to the southwest at greater distances to the seismicity step.

The position of the seismicity step itself is likely connected to relatively strong transpressional tectonics, which can be derived from the motion along the SGF and predominant thrust-type focal mechanisms within the same region (Figure 8b). Although there is an apparent dominance of under-thrusting within this area, we also observed a cluster of normal faulting events (at [38, 16] in Figure 8b) which is in contrast to the overall tectonic regime in this area. Motivated by the observation of both thrust and normal faulting, we searched for a possible correlation between
dominant faulting mechanisms and stress drops in the following section.

[Figure 8 about here.]

4.4 Stress drop variations as function of faulting mechanism

We correlated average faulting mechanisms expressed by their differences in rake angle (Figure 9). These differences can be quantified by normalizing the observed rake angles so that the spectrum of faulting mechanisms can be expressed on a continuous scale from -1 to 1 with normal faulting at -1, strike-slip at 0 and thrust faulting at a value of 1 (Shearer et al., 2006). Stress drops and focal mechanisms show a weak, positive correlation so that normal faulting has relatively lower average stress drops ($\Delta \sigma = 3.5 \pm 0.5 \text{ MPa}$) whereas thrust faulting has higher average stress drops ($\Delta \sigma = 6.0 \pm 0.6 \text{ MPa}$). Strike-slip events represent the predominant type of faulting. Consequently, their median value ($\Delta \sigma = 5.1 \pm 0.6 \text{ MPa}$) is similar to the one observed for the whole region ($\Delta \sigma = 4.8 \pm 0.1 \text{ MPa}$).

[Figure 9 about here.]

4.5 Stress drop variations along the San Andreas fault system

One of the fundamental questions concerning the SGP region is the possibility of large penetrating ruptures that could propagate through the entire region, e.g., from Cajon Pass to the Salton Sea. Using the average fault orientation within the Mojave segment (see Figure 1), we determined variations of stress drop in the proximity of a possible path of such a rupture between the San Bernardino and Garnet hill segment (Figure 10). The stress drops decrease to the southeast of SGP within the area of the Banning and Garnet Hill segments which eventually merge with the Coachella segment of the San Andreas fault. The stress drops also decrease to the northwest of SGP and show consistently lower values outside of the San Gorgonio fault segment.

The stress drop traverse through the SGP passes in immediate proximity to local estimations of geologic slip rates (highlighted by blue squares in Figure 3). Slip rates were previously compiled from many different studies and summarized by Dair and Cooke (2009); Cooke and Dair (2011) as well as by McGill et al. (2013) highlighting a systematic decrease from Cajon Creek (slip rates = $24.5 \pm 3.5 \text{ mm/yr}$) to Cabezon ($5.7 \pm 0.8 \text{ mm/yr}$), which is close to SGP. To the southeast, the slip rates increase again within the Coachella region (14–17 mm/yr) of the San Andreas fault. The average
geologic slip rate on the SGF itself is estimated to be as low as 1.0–1.3 mm/yr (Matti et al., 1992). This shows, that stress drops and slip rates are mostly inversely correlated within the study area.

[Figure 10 about here.]

5 Discussion

5.1 Seismicity and fault orientation

The most prominent feature in the seismicity is a lack of shallow events south of the Mission and Mill Creek segment and a seismicity step close to the down-dip end of the SGF. To the north, we observed more shallow seismicity that extends down to about 14–15 km depth. The latter conforms to the commonly observed depth-extent of the seismogenic zone within southern California. The variations in the maximum depth of seismicity may be related to both topographic and lithologic effects, supported by the sharpness of the transition and the approximate, inverse relationship between surface relief and seismicity base-depth (Magistrale and Sanders, 1996; Yule and Sieh, 2003).

The juxtaposition of different lithologies due to the large displacement along the San Andreas fault system, seems to contribute to the creation of the observed difference in the maximum seismicity depths, moving the brittle-ductile transition to greater depths. The latter may be caused by a difference in plasticity temperature between feldspar-dominated Peninsular range and quartz-dominated Transverse range rocks (e.g. Scholz, 1988; Magistrale and Sanders, 1996). In addition, down-thrusting along the SGF may perturb the geotherm downward which can explain the locally deep earthquakes and base of seismicity. We will explore this question in more detail below within the context of the observed changes in stress drops.

Stress drops within the present study show regional variations between ~1 to ~20 MPa. Similar variations are observed in laboratory earthquake-analog experiments and seismic events at shallow depth in mines. The latter exhibited relatively high displacements and locally-high stress drops of up to 70 MPa (McGarr et al., 1979). Shear stress drops during laboratory stick-slip experiments range from ~1 to more than 160 MPa (e.g. Thompson et al., 2005; Goebel et al., 2012). The laboratory studies also highlight a connection between fault heterogeneity, aftershock duration and stress drop magnitudes so that stress release is higher and aftershock duration shorter for smooth, homogeneous faults in the laboratory (e.g. Goebel et al., 2013b,a).
5.2 Stress drop variations

5.2.1 Focal mechanisms and ambient stress level

Previous investigations of the influence of focal mechanism types on stress drop variations pro-
duced mixed results, supporting higher stress drops for both normal (Shearer et al., 2006), and
strike-slip events (Allmann and Shearer, 2009) or no dependence on focal mechanisms (e.g. Oth,
2013). The southern Californian data set was strongly influenced by the 1994 Northridge sequence
which showed predominant thrust-type events with low stress-drops (Shearer et al., 2006). Our re-
results, on the other hand, revealed higher stress drops for thrust events compared to strike-slip and
normal faulting, which can be understood in the context of large compressive stresses and higher
ambient stress level. A possible reason for the difference between our results and other studies may
be related to the observational scales and the mixture of vastly different tectonic regimes. While our
study concentrated on a small crustal region, others investigated stress drops for all of Southern
California (Shearer et al., 2006), Japan (Oth, 2013) and a global data set Allmann and Shearer (2009),
inevitably mixing seismic events from volcanic activity, off-shore events, induced seismicity, and
other sources. Over these large scales, stress level and faulting mechanics are bound to vary sub-
stantially, which may contribute more extensively to variations in stress drops than the differences
in faulting mechanisms. Furthermore, the rather weak correlation between focal mechanisms and
stress drops within the present study, indicates that the type of faulting is not the only contributing
factor to stress drops variations.

5.2.2 Lithological variations

The large cumulative displacement along the San Andreas fault system results in a juxtaposition of
different lithology in many areas. Within the SGP area, feldspar-dominated Peninsular range rocks
have been moved next to quartz-rich Transverse range rocks (Magistrale and Sanders, 1996) with
very different brittle-ductile transition temperatures (e.g. Scholz, 1988). The difference in lithology
and transition temperatures across the San Andreas fault system (or more precisely across the Mis-
sion Creek segment of the San Andreas fault) not only controls the thickness of the seismogenic
zone but also influences the stress drops within the SGP region. We observed an abrupt variation
in determined stress drops across the Mission Creek segment so that feldspar-dominated rocks to
the south are connected to substantially larger stress drops compared to quartz-rich material to
the north of the Mission Creek segment. Similar observations have been made for mining induced seismicity, for which stress drops are higher in feldspar-dominated diorite dikes compared to the surrounding quartzite host rocks (Kwiatek et al., 2011). Kwiatek et al. observed a maximum difference in stress drop estimates of about one order of magnitude whereas seismic velocities varied by only \( \sim 3\% \). Differences in ‘rock-brittleness’ as a function of temperature also influence frictional properties, specifically, frictional strengths and slip stability (e.g. Tse and Rice, 1986; Blanpied et al., 1995).

Furthermore, the frictional stability, i.e., the degree of velocity strengthening or weakening of material interfaces, is directly connected to stress drop (e.g. Gu and Wong, 1991; He et al., 2003; Rubin and Ampuero, 2005). As a consequence, more ductile material, which favors velocity strengthening behavior, also exhibits relatively lower stress drops compared to more brittle material. This behavior appears to be observable for rocks at varying temperatures (e.g. Blanpied et al., 1995), but also, as in our case, for different rock types (quartz- vs. feldspar dominated) with different brittle/ductile transition temperatures.

5.2.3 Asperity strengths and fault slip rates

The present study revealed a correlation between geologically inferred fault slip rates and stress drops so that the areas of highest stress drops coincide with the lowest slip rates (see Figure 10). Relatively high stress drops are also inferred for large magnitude earthquakes (\( M = 5.5–8.5 \)) for faults with long recurrence intervals and high fault strengths (Kanamori, 1986). Besides studies of large magnitude earthquakes, small-scale laboratory stick-slip experiments highlight a connection between loading rates, recurrence intervals and stress drops. In the laboratory, recurrence intervals of stick-slip events are correlated with fault strengths and stress drops so that longer recurrence intervals due to slower loading rates results in relatively high stress drops (Beeler et al., 2001). Similar results have been obtained for repeating earthquakes which show a higher proportion of high frequency energy radiation if the recurrence intervals between events are long (e.g. Beeler et al., 2001; McLaskey et al., 2012). The connection between earthquake recurrence and stress drops can be explained by increasing strength of load bearing asperities as a function of time. Asperities on a slowly loaded fault undergo relatively longer interseismic healing periods and exhibit higher resistance to shear before failure events occur, releasing a comparably high amount of stored stress. The amount of fault healing is, in addition to loading rates, also sensitive to pressure and temperature conditions at depth, which can significantly influence the distribution of radiated seismic energy.
as a function of frequency (McLaskey et al., 2012). Increased asperity strength due to longer healing periods may also influence the tendency of asperities to fail individually. For instance, ruptures on heterogeneous faults with strong asperities are more likely to be arrested before growing to large sizes (Sammonds and Ohnaka, 1998). The presence of strong asperities and fault heterogeneity may explain the relatively high stress drops of small and intermediate magnitude events that were observed here.

Theoretical considerations of seismic slip on a fault that is governed by rate-and-state friction confirm the dependence of stress-drops on loading rates. In addition, the static stress drop ($\Delta \tau_s$) is sensitive to friction-parameters (e.g. Gu and Wong, 1991; He et al., 2003; Rubin and Ampuero, 2005):

$$\Delta \tau_s = \sigma_n (b - a) \ln \left( \frac{V_{\text{dyn}}}{V_l} \right)$$

where $\sigma_n$ is the normal stress, $b$ and $a$ are material parameters that control the frictional behavior, and $V_l$ and $V_{\text{dyn}}$ are the loading and dynamic slip velocities. The latter occupies values close to $1 \text{ m/s}$. Furthermore, if we assume approximately constant friction and normal stress across the fault, the stress drop changes as a function of loading velocity, $V_l$, so that a decrease in loading rate by a factor of 4–5, as observed in our study, corresponds to an increase in stress drop by factor of $\sim 1.7$. Our results show an increase in stress drop along the San Andreas fault by a factor of 2–3 (see Figure 10), which is slightly higher than predicted from this simple model. This difference can be explained by possible changes in material and frictional properties, which were not considered, but likely also contribute to variations in stress drop. In addition, spatial and temporal heterogeneity in stress-drops may be a result of variations in seismic coupling and transient slip processes before mainshocks, for example, expressed by differences in foreshock and aftershock source spectra in Southern California (Chen and Shearer, 2013).

5.2.4 What is the major controlling parameter of stress drop variations?

We identified four parameters that were connected to variations in stress drops within the SGP region, i.e., the type of faulting, hypocentral depths, geologic slip rates and mineralogical composition of the regional rock types. Our analysis suggests that all four mechanisms contribute to some extent to the creation of the relatively high stress drops between the surface traces of the SGF and the Mission Creek segment. The largest variations in stress drops occurred along fault strike and
in the proximity of the seismicity step at the down-dip end of the SGF. This suggests that average
slip rates and the presence of abrupt lithologic changes exert the strongest control on stress drops.
We hypothesize that relatively slow down-thrusting of feldspar-dominated material in connection
with longer healing periods and increased asperity strengths generally promote high stress drops.

5.3 Implications for seismic hazard and earthquake rupture dynamics

The relatively high stress drops and slow geologic slip rates (e.g. McGill et al., 2013) within the San
Gorgonio pass area suggest locally increased fault strength and long earthquake recurrence inter-
vals. We hypothesize that areas of high stress drop are connected to the failure of individual small
but strong fault patches. Consequently, rupture propagation may be stifled within the SGP area
decreasing the probability of large earthquakes that extend through the SGP. The role of the SGP in
hindering rupture propagation has been recognized previously based on the strongly segmented
fault geometry within the area (Magistrale and Sanders, 1996). The overall deformation along the
San Andreas fault system may increasingly by-pass the SGP region to the north and south-east, for
example, via the San Jacinto fault (McGill et al., 2013).

6 Conclusion

We have analyzed the spatial variation in source parameters of small and intermediate magnitude
earthquakes within the San Gorgonio Pass region. Our analysis revealed a localized region with
relatively high stress drop estimates between the surface traces of the San Gorgonio thrust and
Mission fault. Furthermore, stress drops show a weak correlation with focal mechanism types so
that thrust faults are connected to higher median stress drops than strike-slip and normal faults.
Stress drops increase abruptly below ∼10 km depth and at the interface between Peninsular range
and Transverse range rocks. The latter is likely related to differences in lithology between the
two geological formations, so that feldspar-dominated Peninsular range material favors relatively
larger stress drops whereas quartz-dominated Transverse range rocks exhibit relatively lower stress
drops. Stress drops vary systematically with geologically inferred slip rates along the San Andreas
fault system. Consequently, more rapidly loaded fault segments are connected to lower stress
drops whereas slowly loaded faults create events with higher stress drops. While several factors
may contribute to stress drop variations, our results suggest that within the greater San Gorgonio
area, variations in slip rates and lithology are the predominant mechanisms. Thus, they should also be considered for seismic hazard assessment and ground motion simulations.

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Figure 1: Overview of the major faults and seismicity within the study region. The study region and connected seismicity are highlighted in red. Seismic events within southern California are shown by blue dots. The locations and names of major faults are highlighted by black lines and white font respectively. The inset shows the map location with respect to the Californian state boundaries and the San Andreas fault (SAF).
Figure 2: Seismicity within the SGP region in map view (a) and within a 2 km wide depth cross-section between A and A’ (b). Different fault segments that comprise the San Andreas fault system are labeled in blue. The beach balls in (a) mark the locations and focal mechanisms of the 1992, M6.4 Big Bear, the 1986, M5.6 North Palm Springs and the 2005, M4.9 Yucaipa earthquake. The fault orientations in (b) are constructed using mapped fault traces, approximate dip angles and near-by seismicity clusters. Seismic events are broadly distributed and can only partially be associated with mapped fault traces (e.g. for Banning and Mission Creek fault) highlighting the complexity of the deformation within the area.
Figure 3: Map view of smoothed stress drop variations within the study region. Fault segments of the San Andreas fault system are labeled in blue. The red line from A to A’ marks the location of the depth cross-sections in Figures 2b and 8. The blue squares show the sites of geologic slip rate estimates (see Figure 10 and description for details). Stress drops vary substantially from about 1 MPa to more than 20 MPa (see color-bar).
Figure 4: Corner frequency and seismic moment for events within a high (green circle in Figure 3) and a low stress drop region (red circle in Figure 3). The black, dashed lines highlight constant stress drops from \( \sim 1 \) to \( \sim 20 \) MPa and the green and red lines mark the median stress drops for the two different regions. The two data sets show almost no overlap, which highlights a generic difference between the corresponding stress drops.
Figure 5: Source spectra for events within an area of low (left) and high stress drop corrected for differences in moment by shifting along $t^{-3}$ and colored according to stress drop. The solid, black line highlights a high-frequency fall-off slope of $-2$. High stress drop spectra are generally shifted further to the right due to higher corner frequencies and a smaller proportion of low-frequency contributions compared to the area of low stress drop.
Figure 6: Smoothed spatial variations in stress drop for events within three different depth layers from 0–10, 10–15 and 15–25 km.
Figure 7: Variations in stress drops as function of depth (a). Green dots show individual event stress drops and squares show the binned, median stress drops and bootstrap errors. The latter are shown by horizontal error-bars which are of approximately same extent as the markers. The vertical error-bars highlight the extent of individual depth bins. The circles display stress drops after correcting for a depth dependent rupture velocity using two different 1-D velocity profiles (b) for events beneath (green curve) and outside (red curve) of SGP. The dashed lines in b) show 10th and 90th percentiles.
Figure 8: Same depth cross-section as in Figure 2 bottom, now with events colored and scaled according to stress drop. The background colors depict the spatial distribution of median stress drop, smoothed as in Figure 3. The deep events southwest of the Mission Creek segment are connected to clusters of locally high stress drops whereas events above 10 km seem to be marked by generally shallow stress drops. Focal mechanism solutions for events within the area are shown in the inset. The beach-balls show strike-slip mechanisms in red, thrust in blue and normal faulting in green.
Figure 9: Variations in average stress drops as function of faulting mechanism. The gray dots represent individual event stress drops and the solid line marks the median values for normal (green), strike-slip (red) and thrust (blue) faulting. Average values for these three faulting types are shown at the bottom of the figure. The dashed lines show 10th and 90th percentiles. Normal faulting is generally connected to relatively lower stress drops of $\sim 4$ MPa whereas thrust faulting exhibits higher stress drops of $\sim 7$ MPa.
Figure 10: Changes in stress drop for seismic events along the San Andreas fault segments through the SGP region from the northwest to the southeast within a ~10 km wide zone. The x-axis shows the distance from Cajon pass in kilometers (see Figure 1 for Cajon pass location) for a transect that passes through the sites of geologic slip rate estimates (blue squares in Figure 3). Individual events are marked by gray dots and green line marks the median. The dashed lines show 10\textsuperscript{th} and 90\textsuperscript{th} percentiles. Sites of geologic slip rate estimates: BC: Badger Canon (McGill et al., 2013), Pl: Plunge Creek (McGill et al., 2013), WC: Wilson Creek (Weldon and Sieh, 1985), BF: Burro flats (Orozco and Yule, 2003), Cb: Cabezon (Yule et al., 2001), BP: Biskra Palms (Behr et al., 2010).