

A Comprehensive Stress Drop Map from Trench to Depth in the Northern Chilean Subduction Zone

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Key Points:

- A comprehensive stress drop distribution with more than 51,000 stress drop estimates for the Northern Chilean Subduction Zone is computed.
- Systematic stress drop variations between upper plate, interface and intermediate depth seismicity are revealed.
- Reliability and comparability of results are increased by using two independent methods, the spectral ratio as well as the spectral decomposition approach.

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Abstract

We compute stress drops for earthquakes in Northern Chile recorded between 2007 to 2021. By applying two different analysis techniques, (1) the spectral ratio (SR) method and (2) the spectral decomposition (SDC) method, a stress drop map for the subduction zone consisting of 51,510 stress drop values is produced. We build an extended set of empirical Green's function (EGF) events for the SR method by systematic template matching. Outputs are used to compare with results from the SDC approach, where we apply cell-wise obtained global EGF's to compensate for the structural heterogeneity of the subduction zone. We find a good consistency of results of the two methods.

The increased spatial coverage and quantity of stress drop estimates from the SDC method facilitate a consistent stress drop mapping of the different seismotectonic domains. Albeit differences of median stress drops are relatively small, strike-perpendicular depth sections clearly reveal systematic variations, with earthquakes within the upper plate, along the interface and at intermediate depth exhibiting distinct values. In particular, interface seismicity is characterized by the lowest observed median value, whereas upper plate earthquakes show noticeably higher stress drop values. Intermediate depth earthquakes show comparatively high average stress drop and a rather strong depth-dependent increase of median stress drop.

Additionally we observe spatio-temporal variability of stress drops related to the occurrence of the two megathrust earthquakes in the study region.

The here presented study is the first coherent large scale 3D stress drop mapping of the Northern Chilean subduction zone. It provides an important component for further detailed analysis of the physics of earthquake ruptures.

Introduction

Stress drop is the seismological parameter that relates seismic moment and average slip to the rupture dimension. It characterizes the earthquake source, and its spatial-temporal distribution and possible correlation with other properties such as earthquake depth, mechanism or stress state may contribute to a better physical understanding of earthquakes.

A number of studies report scale-invariance or self-similarity, i.e a fundamentally constant average stress drop which is then used to obtain rupture size or average slip (e.g., Aki, 1967; Shaw, 2009). From detailed studies of individual events, however, it becomes increasingly clear that stress drops may actually be widely variable, usually ranging between about 0.1 to 100 MPa (e.g., Abercrombie, 1995; Allmann & Shearer, 2009; Bindi et al., 2020). Therefore, a possible scale-invariance over several magnitudes is debated. It is supported by some studies, e.g. Abercrombie (1995); Shearer et al. (2006), and contested by others, e.g. Prieto et al. (2013); Nishitsuji & Mori (2014); Trugman & Shearer (2017) or Bindi et al. (2020), who observe a dependence of stress drop on seismic moment. It has also been pointed out that estimates of the stress drop value strongly depend on the analysis method (Shearer et al., 2019; Neely et al., 2020), the model assumptions and the parameter choices (Kaneko & Shearer, 2014, 2015; Ji et al., 2022), thus complicating a comparison of results from different studies.

Stress drop values of large to megathrust earthquakes have been analyzed in several global studies (e.g., Allmann & Shearer, 2009; Ye et al., 2016), and crustal earthquakes, both of tectonic and induced origin, have been studied intensively in the last years to infer dependencies on depth, mechanism, magnitude and more (e.g., Abercrombie, 1995; Shearer et al., 2006; Hardebeck & Aron, 2009; Abercrombie, 2014; Goertz-Allmann et al., 2011; Trugman & Shearer, 2017; Chen & Abercrombie, 2020; Bindi et al., 2020). Most studies, however, are limited to a highly focused target zone or a relatively small amount of earthquakes usually not exceeding a few tens or hundreds of events, and systematic stress drop studies in particular of entire subduction zones or at least major parts of them are very rare (Allmann & Shearer, 2009; Uchide et al., 2014; Tian et al., 2022). And yet, knowledge of the variability of stress drop across an entire subduction system, which includes conditions for earthquake generation that can vary widely both spatially and temporally within an earthquake cycle, is particularly valuable.

As mentioned above, an important precondition for this type of study is the consistent and careful processing of a large number of events (i.e. many thousands of events), and the Northern Chilean subduction zone with its high seismic activity and dense seismic monitoring provides an ideal setting.

In an earlier study Folesky et al. (2021) implemented and verified a spectral ratio approach (SR) for stress drop estimation for the rupture and aftershock area of the 2014 Mw8.1 Iquique earthquake. Analyzing ~ 600 events, they identified an increase of stress drop with distance from the plate interface, no clear depth dependence and an increase of stress drop with seismic moment. Additionally, they described the spatio-temporal variation of stress drop in association with the Iquique event.

In this work, we expand the analysis to the greater subduction zone in Northern Chile, a region that has been continuously monitored since 2006 by the IPOC network (IPOC, 2006) accompanied by the permanent networks of the CSN (C,C1) and multiple temporary deployments. Based on these data the distributions of several geophysical parameters in this region such as focal mechanisms, stress orientation, fore-, and aftershock distributions, or inter-plate-locking have been studied intensively by various authors (e.g. Schurr et al., 2012; Fuenzalida et al., 2013; Schurr et al., 2014; Ruiz et al., 2014; Li et al., 2015; Cesca et al., 2016; Hoffmann et al., 2018). An earthquake catalog covering the time period from 2007 to 2014 by Sippl et al. (2018) has recently been updated

and extended until 2021 (Sippl et al., 2023). The new version contains over 180,000 events for the time period from 2007 to 2021. Its uniform processing along with the long time period makes it a great basis for a comprehensive, consistently processed stress drop mapping of a subduction zone. It allows for the detailed study of possible regional variations of stress drop as well as time-dependent variations in connection with the occurrence of two megathrust earthquakes, the 2007 $M_W 7.7$ Tocopilla earthquake and the 2014 $M_W 8.1$ Iquique earthquake.

We apply two different methods for calculating the stress drop: 1) the spectral ratio method (SR) and 2) the spectral decomposition method (SDC). Spectral ratio approaches employ empirical Green's functions (EGFs) to eliminate path and site terms in the observed seismogram spectrum and to isolate the event source term (e.g. Hutchings & Viegas, 2012). While this makes them more robust against systematic errors introduced by over- or under- correcting for attenuation structures and radiation pattern, they require the existence of a suitable nearby EGF event which, in practise, is a rather strong limitation. To overcome this problem, we additionally use a second approach which has proven to be better suited for large data sets and which increases significantly the quantity and spatial coverage of stress drop values. The spectral decomposition method (e.g., Prieto et al., 2004; Shearer et al., 2006; Chen & Abercrombie, 2020; Pennington et al., 2021) implicates to first untangle path, site and event terms of multiple earthquakes simultaneously and then produce an EGF-like correction term which can be applied to an event cell rather than just to single events. In this way, the number of stress drop estimates can be multiplied at the cost of a more general correction of the source medium response.

In this study, we use the two methods jointly, which allows to exploit the benefits of both of them, to mutually control results and to produce a spatially inclusive data base of tens of thousands of stress drop values. The resulting distribution covers all the different seismically active segments of the subduction zone from shallow depths close to trench down to depths of about 180 km.

Catalog and Data

We extract the event origin times, P- and S- arrival time picks, event hypocenters and magnitudes from the catalog by Sippl et al. (2023), which is an updated version of Sippl et al. (2018). In the following, we will refer to it as the IPOC catalog. It comprises 182,847 double-difference relocated events which occurred in the time period from 2007 to 2021. For our stress drop study, we use waveforms of in total 23 seismic broadband stations of the Integrated Plate Boundary Observatory Chile (IPOC, 2006). The network extends in N-S direction from 17.6°S to 24.6°S, a trench-parallel length of about 700 km. The corresponding continuous three-component waveform data, sampled at 100 Hz, were downloaded from the EIDA web service of GFZ Potsdam (Bianchi et al., 2015). Based on event location, the authors assign a class to each event of the IPOC catalog, picking from the following options:

- UP** Upper plate seismicity, predominantly crustal events within the South American plate, but also some earthquakes in the uppermost mantle.
- P1** Seismicity at or very close to the plate interface.
- P2** A plate interface- parallel band of seismicity in the oceanic Nazca Plate.
- P3** A second, deeper interface- parallel band of seismicity below P2.
- ID** Intermediate depth seismicity (sometimes called IDE). This class comprises by far the largest amount of events ($\sim 116,000$), and it extends from about 60 km down to 180 km depth.
- MI** Mining events from open pit mines at the surface.

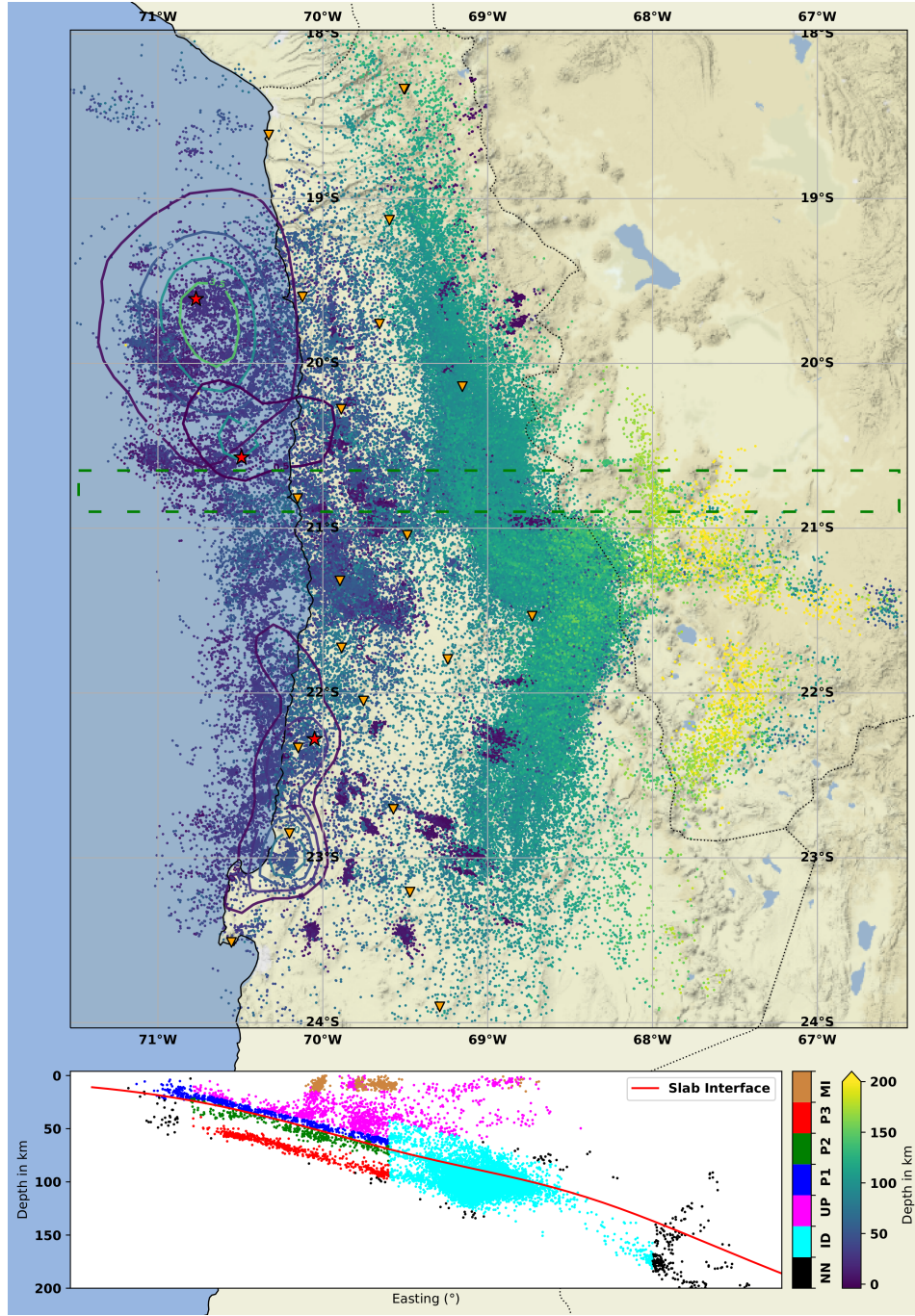


Figure 1. Seismicity from the IPOC catalog by Sippl et al. (2023). Color indicates event depth. Red stars are hypocenters of the $M_W 8.1$ 2014 Iquique event, its largest $M_W 7.6$ aftershock, and the 2007 $M_W 7.7$ Tocopilla event, from North to South, respectively. Their slip contours are taken from Schurr et al. (2014) and Schurr et al. (2012). Orange triangles show the location of the IPOC seismic stations. The bottom panel shows a depth view of selected events from the green dashed box, color-coded by the associated event class as explained in the text. The slab interface is taken from the model of Hayes et al. (2018).

NN Not classified events which are located at the less well constrained edges of the catalog region outside the network. The two biggest subgroups are offshore events and deep event in the east, respectively.

Figure 1 displays the seismicity distribution from the IPOC catalog in Northern Chile, color-coded by depth. It includes a West-East depth view of a catalog slice with coloring according to the event class (green dashed box). For a more detailed description of the event classification, the reader is referred to Sippl et al. (2018).

Stable magnitude estimation is an important prerequisite for consistent stress drop estimates as the moment is an integral parameter in the estimate. Event magnitudes in the IPOC catalog were computed using a technique by Münchmeyer et al. (2020), which applies location-dependent station corrections in order to stabilize the computation against possible reduced or variable station availability.

The spectral ratio technique strongly depends on the availability of suitable event pairs which are used as a target and empirical Green’s function couple. In order to exploit the existing data set at its best, we first perform an exhaustive event search by template matching which complements the IPOC catalog by finding additional, small magnitude EGF candidate events. Template matching uses the cross correlation function to detect waveforms in a continuous dataset that resemble the predefined patterns (i.e. the template waveforms). It is commonly used in seismology to increase the number of mostly weak events missed in the original earthquake catalog. Event seismograms from the catalog serve as templates, and the resulting detections can be assumed to be closely located and to have similar mechanisms as the template events, if cross correlation values are sufficiently high. This method becomes computationally challenging with an increasing volume of continuous waveform data and a high number of template events in the initial catalog. To make the method feasible for our extensive dataset, we applied our own GPU-based template matching code. For each event in the IPOC catalog, template waveforms were extracted for the three closest available stations using the vertical channel only. The minimum length of the template waveform is 15 seconds and increases with hypocentral distance to include both the P and S phases. Additionally, the data was down-sampled to 25 Hz. We define potential pairs of target event and empirical Green’s function (EGF) event, if the normalised cross correlation coefficient is at least 0.70 at minimum two stations, with template waveforms derived from the same master event. In this way, the originally $\sim 180,000$ templates produce 1,836,195 matches, providing a substantial extension of the number of EGFs compared to the initial catalog.

Methods

Source Model and Stress Drop

We compute the stress drop $\Delta\sigma$ assuming the widely used circular source model by Eshelby (1957):

$$\Delta\sigma = \frac{7\pi\mu\bar{D}}{16r} = \frac{7M_0}{16r^3}, \quad (1)$$

where r is the approximate fault radius, \bar{D} is the average slip on the fault, μ is the shear modulus, and M_0 is the seismic moment. In general, slip and fault dimensions are not known, and the stress drop cannot be computed directly (Kanamori & Anderson, 1975). We therefore adopt the approach of Brune (1970), who proposed the following relation between source radius and spherically averaged corner frequency (see also Madariaga, 1976; Kaneko & Shearer, 2014, 2015):

$$f_c = k \frac{\beta}{r}, \quad (2)$$

with the shear wave velocity at the source, β , and a constant k that relates to the spherical average of the corner frequency for a specific theoretical source model. Combination

of equations 1 and 2 leads to the Brune type stress drop (Brune, 1970):

$$\Delta\sigma = \frac{7}{16} \left(\frac{f_c}{k\beta} \right)^3 M_0. \quad (3)$$

The seismic moment M_0 is computed from the refined magnitudes of the IPOC catalog. We use the regional velocity model by Bloch et al. (2014) to compute event location-specific shear wave velocities. We use a k-value of $k_p = 0.32$ for P-wave spectra, a standard value from Madariaga (1976), and we obtain $k_s = 0.265$ by a least square fit of corner frequencies from both P and S phases for identical events, similar to Folesky et al. (2021). In practise, k_s is chosen such that P and S phase-based corner frequencies produce similar stress drop values. According to Kaneko & Shearer (2014) the here obtained k-values correspond to their model of a symmetrical circular rupture with a rupture velocity of $v_r = 0.7\beta$.

The corner frequency f_c is obtained by fitting a spectral model to the earthquake source spectrum. This is not recorded directly, and we only have the observed displacement spectrum $d(f)$, for which we can write

$$d(f) = e(f) * p(f) * s(f), \quad (4)$$

with the earthquake source spectrum, $e(f)$, the path response, $p(f)$ and the site response, $s(f)$. The source spectrum can be expressed as

$$e(f) = \Omega_0 \frac{1}{(1 + (f/f_c)^{\gamma n})^{1/\gamma}}, \quad (5)$$

where Ω_0 is proportional to the seismic moment, n is the high frequency falloff and γ is a constant which is commonly set to $\gamma = 1$ for the Brune (1970) or $\gamma = 2$ for the Boatwright (1980) spectral model, respectively.

Separating the terms in equation 4 in order to apply the spectral model requires careful processing and may be achieved by using one of the following two methods.

Spectral Ratio Approach

For the spectral ratio approach, we use the processing scheme described in detail in Folesky et al. (2021), where a limited region around the 2014 Mw8.1 Iquique earthquake was already investigated.

The general idea of the spectral ratio method (SR) is to use for each target earthquake a smaller event with similar location and focal mechanism as an empirical Green's function (EGF) (Hutchings & Viegas, 2012). By spectral division between target and EGF seismograms the path and site terms which are assumed to be identical are removed (equation 4), leaving essentially the source term of the target event:

$$\frac{d1(f)}{d2(f)} \approx \frac{e1(f)}{e2(f)} = \frac{\Omega_{01}}{\Omega_{02}} \cdot \left[\frac{1 + (f/f_{c2})^{\gamma n}}{1 + (f/f_{c1})^{\gamma n}} \right]^{1/\gamma}. \quad (6)$$

In practice, the similarity of the event locations is ensured by a required minimum cross correlation value between the waveforms of the two events at usually multiple stations. From our template matching results, described in the Catalog and Data section, we select all templates with a catalog magnitude $M \geq 2$ which produced matches having a cross correlation value of $cc \geq 0.70$ at minimum 2 stations ($n=1,836,195$). If at least four picks are available the three-component waveform data are band-pass filtered (0.8-40 Hz) and sliced to maximum 6 s and 10 s phase windows for P and S waves, respectively. A shorter S-P arrival time difference leads to shorter windows, accordingly. We require a $SNR \geq 3$ in the frequency bands 1.5-5, 5-10, 10-15, 15-20, 20-25 Hz, for both template and EGF. Then, the spectral ratio is computed station-wise, and the quotient of the two Boatwright

spectral models, being equation 6 with $n=2$ and $\gamma=2$, is fitted to the data. We require the median of the low frequency plateau to be higher than 5 to ensure sufficient difference of seismic moment between the EGF partners, a necessity for resolving the corner frequency (Abercrombie, 2015). Next, the spectral ratios of all stations are stacked for robustness and resampled to achieve similar weights of low and high frequency content. The fit of the stack to the above Boatwright model quotient yields the corner frequency estimate of the target event. If the value of the corner frequency is within the interval 1 - 30 Hz, the stress drop is computed for this event. We obtain 25,994 P phase estimates and 36,121 S phase estimates which include multiple results for identical target events originating from different EGF events. The procedure, its limitations and error estimates are explained and discussed in more detail in Folesky et al. (2021).

Spectral Decomposition Approach

For the spectral decomposition approach (SDC), we use the decomposition procedure by Chen & Abercrombie (2020) as implemented in Pennington et al. (2021) and called SNSS (Stacking No assumption of Self-Similarity). The approach exploits the redundancy of existing event- station pairs to separate the displacement spectrum into event term, path term and site response (cf. equation 4). A robust iterative stacking procedure as described in Shearer et al. (2006) is used to solve this over-determined problem, which should account for distance- dependent attenuation and site responses. The specific near-source attenuation, however, is assumed to be common for all event source paths, and therefore it needs to be estimated. For this, a so-called global empirical Green's function (gEGF), which is valid for the specific region, is computed. The gEGF can be used later in a similar way to the spectral ratio approach to compute the event- specific source time function properties such as the corner frequency. The construction of the gEGF is done as described in Chen & Abercrombie (2020) and Pennington et al. (2021). It consists of (1) a stacking step, where the phase spectra of all available events from the target region are stacked in 0.2 magnitude units, (2) a misfit computation between the lowest stack of those magnitude bins and different test values of corner frequency put into a spectral Brune model, (3) using the misfit as first gEGF and correction of the stacked spectra of all other magnitude bins, allowing for variable stress drop for each bin, (4) a fit of the now corrected spectra to obtain the overall misfit, (5) defining the final the correction function which produces the lowest misfit with each bin as the gEGF. This global empirical Green's function can then be used to correct all target events in the associated region, after which the individual corner frequency can be computed.

The spectral decomposition method works best for a sufficiently high number of stacked spectra, i.e. many events recorded on many stations in a confined region. The IPOC network consists of 23 stations, but due to the large spatial extent of the network, the weaker events are usually only recorded on a smaller subset of stations. Therefore, a relatively large amount of events is needed for the decomposition and computation of an appropriate gEGF. At the same time, strong variations of ray paths and attenuation are to be expected in a subduction zone. Hence, it makes sense to subdivide the volume into cells with a common gEGF, where the attenuation structure is assumed constant. We define a regular grid over our study region and divide it into cells of $0.5^\circ \times 0.5^\circ \times 20 \text{ km}$. For each cell to be processed, a minimum number of 100 catalog events of $M \geq 1.8$ is required. The waveforms are band- pass filtered (0.8-40 Hz) and sliced into maximum 6 s and 10 s phase windows for P and S waves, respectively, starting 0.1 s before the phase arrival and having shorter windows in case of shorter S-P arrival time differences. We only keep waveforms with a $\text{SNR} \geq 3$ in the frequency bands 1-5, 5-20, and 20-40 Hz. Then, spectra are stacked in bins of 0.2 magnitude units. The bin that is used to compute the initial gEGF ($1.8 \leq M < 2.0$) must contain at least 30 stacked spectra. After the spectral stacking, the result is checked for convergence by examining the misfit grid search results of the different test stress drop values. If a minimum is found within the grid search interval boundaries, the gEGF is used for this region. For all events

with $M \geq 2$ and a minimum of four valid spectra, the individual corner frequencies are computed by correcting the individual event term from the earlier decomposition with the gEGF. The resulting corner frequencies are required to lie between 1 - 30Hz. To apply this procedure to the entire region of cataloged seismicity, the cells are shifted step-wise by 0.25° in horizontal and by 10km in vertical direction. The step size is set to 50% of the cell width to ensure full overlap between cells.

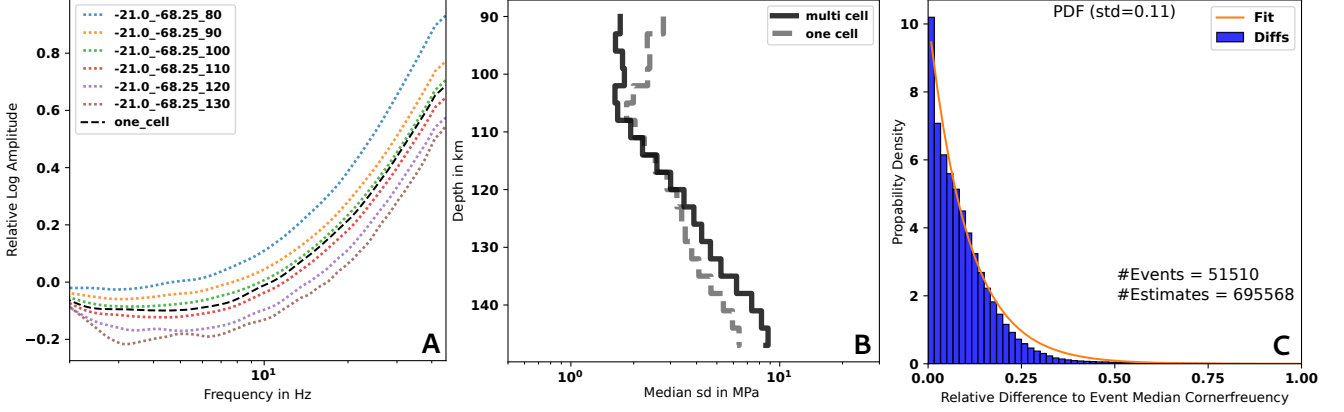


Figure 2. (A) Variation of gEGFs for six grid cells with changing depth. The black dashed line is the gEGF where all events from these cells are decomposed simultaneously in the processing. (B) The corresponding variation of median stress drop with depth for the multi-cell test and the single cell test. Albeit the similar curve shape, note the deviation at shallow depth, where the multi-cell gEGF have higher correction values (blue line in A) and the inverse behavior at greater depths. (C) PDF of the relative differences for all corner frequency estimates for a given event to the event specific median corner frequency, computed for all events of the catalog.

We test the robustness of the multi-cell application of the spectral decomposition method by analyzing the variation of the gEGF with depth for a selected region and comparing it with a decomposition of the events from the same region treated as a single cell. We choose the location 20.5° - $21.0^\circ \times 68.5^\circ$ - 69.0° and process six cells with varying top depths from 80 to 130 km. Figure 2A shows the corresponding gEGF of each cell, together with the single cell gEGF as a dashed line.

A higher value of a gEGF indicates a stronger attenuation correction for this frequency point in the source region. The correction decreases with depth (Figure 2A). Note that the gEGF only represents the spectral correction which is not already captured in the path term (p in equation 4). The single cell gEGF lies in the center of the ensemble of correction functions, representing some sort of average attenuation structure of the region. When used to compute event stress drops and their median, the expected deviations between both scenarios are observed Figure 2B. Compared to the stress drop variation derived from the multi-cell gEGFs (black line), the resulting stress drop variations for the single cell (dashed line) are under-corrected for deeper events and over-corrected for shallower events while the variability of single cell stress drop appears smoother in general. We perform a similar test for a north-south multi-cell crossing the entire catalog in the supplement (Figure S1).

The tests demonstrate that the overall variation of stress drop remains similar, but the smaller scale attenuation heterogeneity is lost by using the single cell version. It also shows that the most populated cells dominate the gEGF. For our target area the great

majority of cells include 100 – 1000 events while several cells have more than 5000 (cf. Figure S2). These exclusively deep cells would dominate the gEGF of possible larger cells, and the characteristics of the shallower events would be lost.

To further test stability, we additionally computed results for a coarser grid with $0.7^\circ \times 0.7^\circ$ cell size and a minimum event number of 400 in each cell. The stress drop distribution is slightly smoother compared to that obtained by using the smaller cells, but very similar in general. To illustrate the similarity we plot the spatial distribution, histogram and depth variability similar to Figures 3&4 in the supplement (Figure S3).

Because of the utilized overlap between grid cells many events have up to 16 independent stress drop estimates (8 for P, 8 for S wave) that are expected to vary slightly due to different gEGFs calculated for each cell. From the total number of estimates (695,568) and the total number of events (51,510) we obtain an average of about 14 stress drop estimates per event. The final stress drop value for each single event is computed from the median of all these estimates. Figure 2C shows that the relative differences between each single event corner frequency estimate to its respective median are small, which demonstrates a very high consistency of corner frequency estimates. This also confirms that the cell-wise gEGFs involve smooth and mostly subtle rather than abrupt changes of corner frequency.

From the performed tests we conclude that the cell-wise application of the decomposition technique works reasonably well in the given setup. In addition, the following section will show the principal similarity between results from our implementations of the SDC approach and the SR method.

Results

Comparison of SR and SDC results

The spectral ratio method produces stress drop estimates for 4,223 events which are a close to complete subset of the 51,510 events with stress drop estimates from the spectral decomposition method. Figure 3 displays map views of stress drop values and their corresponding histograms separately for both methods. Figure 4 displays the overall median stress drop variation against depth for both methods. The evident consistency demonstrates that, qualitatively, the results from both methods are very similar. Also in Figure 4 the similarity is quantified as a cross plot of corner frequencies obtained for common events which have stress drop estimates from both methods. We find that SR corner frequencies are systematically higher than SDC corner frequencies, consistent with the study by Shearer et al. (2019). We obtain a proportionality factor of 0.75 between the two corner frequency families ($fc_{SDC} = 0.75 \times fc_{SR}$). With this, stress drops translate as $\Delta\sigma_{SDC} = 0.42 \times \Delta\sigma_{SR}$. There is no systematic branching or increased deviations at low/high frequency limits observable. Stress drop distributions are, therefore, comparable and differences therein are most likely not caused by the corner frequency estimation method but by the selection of the EGF event(s). Shearer et al. (2019) carefully compared the two approaches and found systematic differences of the calculated stress drops. In order to reconcile the absolute values of both methods, they proposed to fix the corner frequency of the EGF event (fc_2 in eqn 6) in the SR method to an optimized value which is then used for all ratios in the data set. We do not focus on the issue of adjusting the methods as we have demonstrated the principal similarity and especially since almost all SR results have an equivalent from the SDC method.

Summarizing this paragraph, we have shown the high level of similarity of results obtained by both methods. They are qualitatively similar but the number of SDC estimates is significantly higher. The validation of similarity was necessary as we can now focus on the SDC results in greater detail.

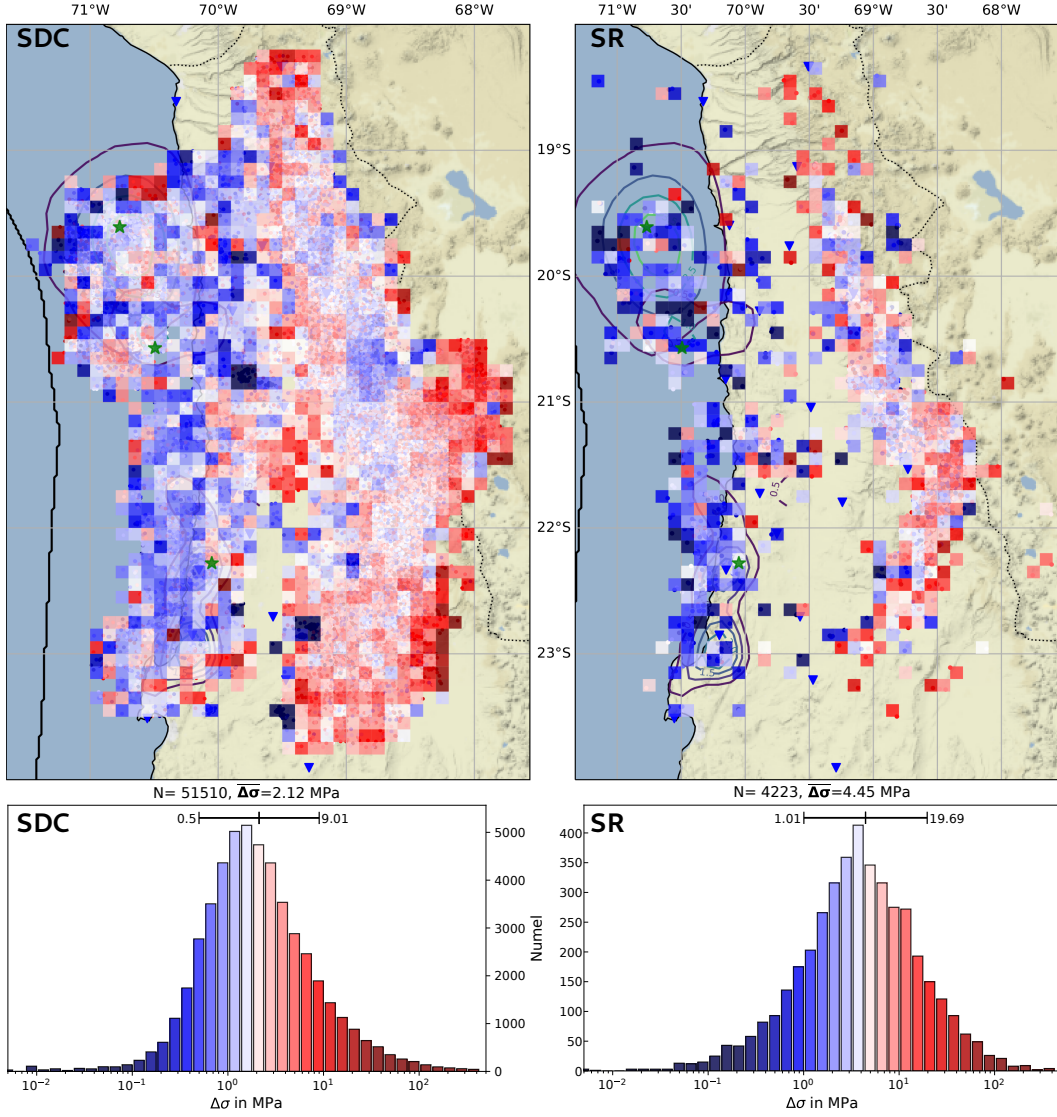


Figure 3. Stress drop distributions in the target region computed by the spectral decomposition approach (SDC) and by the spectral ratio approach (SR). Each $0.1^\circ \times 0.1^\circ$ grid cell represents the median value of all calculated stress drop estimates within the cell. No smoothing is applied. The colors correspond to the histograms on the bottom, which show log-normal distributions in both cases. Stress drop estimates for 51,510 events are obtained with a median of 2.12 MPa for the SDC approach and 4,223 resulting stress drops with a median of 4.45 MPa for the SR method. The standard deviations are indicated above the distributions. Note the good consistency between results of both methods.

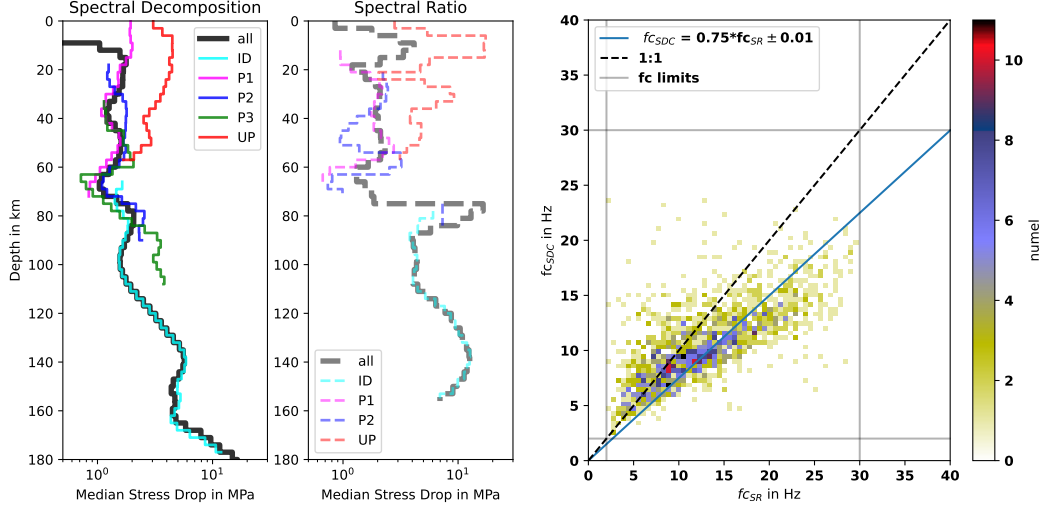


Figure 4. (left) Median stress drop variation with depth for the SDC and SR results, smoothed over three bins (9km), color coded after event classification. SR results scatter significantly more due to the limited event count and sparser coverage. Note the principally good consistency of the curve shapes between results of both methods, for both the overall depth variation and the class wise variations.

(right) Density plot of corner frequency estimates for 1700 events ($M \geq 2.6$), for which stress drop values were found by both methods, the spectral ratio and the spectral decomposition approach. Color indicates event count per cell. The gray lines mark the corner frequency resolution limits. Density is high in the center of the cloud, especially along the regression line, expressing a general agreement between both methods. The obtained relation between results is $f_{CSDC} = 0.75 \times f_{CSR}$ which translates into $\Delta\sigma_{SDC} = 0.42 \times \Delta\sigma_{SR}$.

Stress Drop Estimation Results

The spectral decomposition method yields stress drop values for 51,510 events. For each event, the value is calculated as the median of all available stress drop estimates originating from multiple cells (i.e. different gEGFs) and from both P and S phases. The resulting stress drop distribution is shown in Figure 3, where the event- wise stress drops are averaged on a regular grid using $0.1^\circ \times 0.1^\circ$ cells and color-coded according to the color scheme provided in the corresponding histogram. The stress drop histogram shows a well-defined log-normal distribution with a median value of 2.12 MPa. There is an apparent systematic difference between the stress drops in the western and the eastern part of the study region.

An exemplary west-east side view of the entire depth range from trench to about 180 km depth (Figure 5) better resolves the seismically active domains responsible for the stress drop variation which is evident in the map view. It clearly demonstrates the distinct and systematic differences of stress drop associated with specific regions. The seismicity along the plate interface clearly sticks out with low stress drop values whereas the upper plate crustal events have on average relatively high stress drops. The two interface-parallel bands of seismicity show a mixture of low to medium values, and the intermediate depth seismicity band exhibits a depth- dependent increase of stress drop. Hence, the apparent segmentation of the subduction zone based on deviations in stress drop value corresponds well to the event classification from Sippl et al. (2023). Consequently, we analyze the stress drop variability based on the predefined event classes in the following.

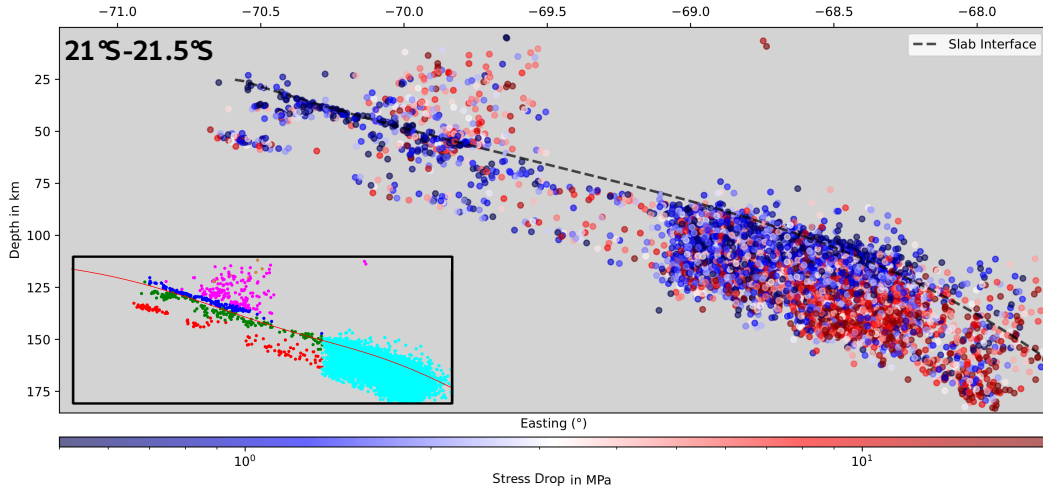


Figure 5. West-east stress drop slice at 21.0°S to 21.5°S . Color indicates single event stress drop value. The inset shows the same section with coloring corresponding to event class similar to Figure 1. Note the clear visual separation of UP and P1 class events. The low stress drop values along the interface are traceable even into the deeper ID cloud. Also, an increasing stress drop with depth in the ID cloud is clearly visible. The slab interface estimate displayed stems from Sippl et al. (2018).

Figure 6 illustrates the class-wise histograms for stress drops, magnitudes, corner frequencies and S- wave velocities. All classes except the mining events (MI) show log-normal distributions of stress drop, ranging generally from 0.1 to 100 MPa, with only few exceptions. Their median values are 1.4 MPa for class P1 events, 1.7 MPa for P2, 2.1 MPa

for P3, 3.3 MPa for UP and 2.3 MPa for class ID events. For each class, the spatial distribution of stress drop in map view is plotted separately in Figure 7.

The spatial availability of stress drop estimates generally follows the variation of seismic activity. In consequence, the results for each event class are limited to cells of sufficient earthquake occurrences. Numbers of upper plate events, for example, are higher in the latitude range of the Iquique event (approx. 20° S), which had a shallow rupture origin and activated a shallower part of the megathrust and many crustal events. In contrast, the smaller Tocopilla event in the south, which had a comparatively deep origin and rupture surface did not have large impact on crustal seismicity. Consequently, fewer UP event stress drops could be computed in the southern region, leading to non-uniform sampled average values of the stress drop distributions. Such sparsity is most apparent in the UP and P3 classes. The interface, in contrast, was activated by both megathrust events (cf. Fig. 7), leading to large amounts of seismicity in the corresponding areas and their surroundings, which enables a good coverage with stress drop estimates in the P1 class. This heterogeneity of existing data points has to be considered when interpreting the overall results.

Next to the SDC results Figure 3 includes the results from the spectral ratio approach. Limited by the number and availability of suitable EGFs, 62,115 stress drop estimates are found for 4,223 events. Again, each final estimate is the median of the stress drop estimates for each target event, which originate from different possible EGFs as well as the use of both P and S phases. The overall spatial distribution of results is similar to the distribution from the spectral decomposition method, and also the stress drop variation appears to be similar, except that a smaller area is covered (fewer estimates) and that the overall median is higher (4.45 MPa). Despite this discrepancy, the median values between estimates of different event classes keep about the same relative differences with P1, P2, P3, UP and ID having average stress drop values of 2.0, 1.9, 1.9, 4.0 and 5.2 MPa, respectively. Note that especially P2 and P3 include results for only few events. An equivalent to Figures 6 & 7 for spectral ratio results can be found in the supplement (Figure S 12&13).

Different profiles (west-east, north-south, depth, distance to interface) for the entire stress drop ensemble along with the median short range stress drop variation is shown in the supplement Figure S 4 as well as for each class separately in Figures S5–11.

Detailed Observations

In this section we describe the main stress drop pattern separately for each event class. The observations mostly relate to the depth- dependent variability (Figure 4), the histograms (Figure 6) or the class- wise maps (Figure 7). More details, such as the variation of the short range median stress drop (for easting, northing, depth and distance to interface) for each separate class is displayed in Figures S5–11.

Spatial Variation

UP Upper plate seismicity shows a comparatively high median stress drop of 3.32 MPa. We observe an increase of stress drop with depth from 2 MPa to about 4.5 MPa in the uppermost 20 km, below which the median stress drop decreases slowly with depth back to about 2 MPa (cf. Figure 4). This clearly distinguishes the upper plate events from all other event classes. Most upper plate stress drop estimates are from the Iquique mainshock area between 19.2 - 20.6°S, and from the seismically highly active region just south of it (cf. Figure 7). The highest median values are found from 20.0-20.2°S (3 MPa) and 21.0-21.6°S (4 MPa). The upper plate events show a steady increase of stress drop with distance to the interface for the first 30 km upward (Figure S5). There are two major continuous patches of increased

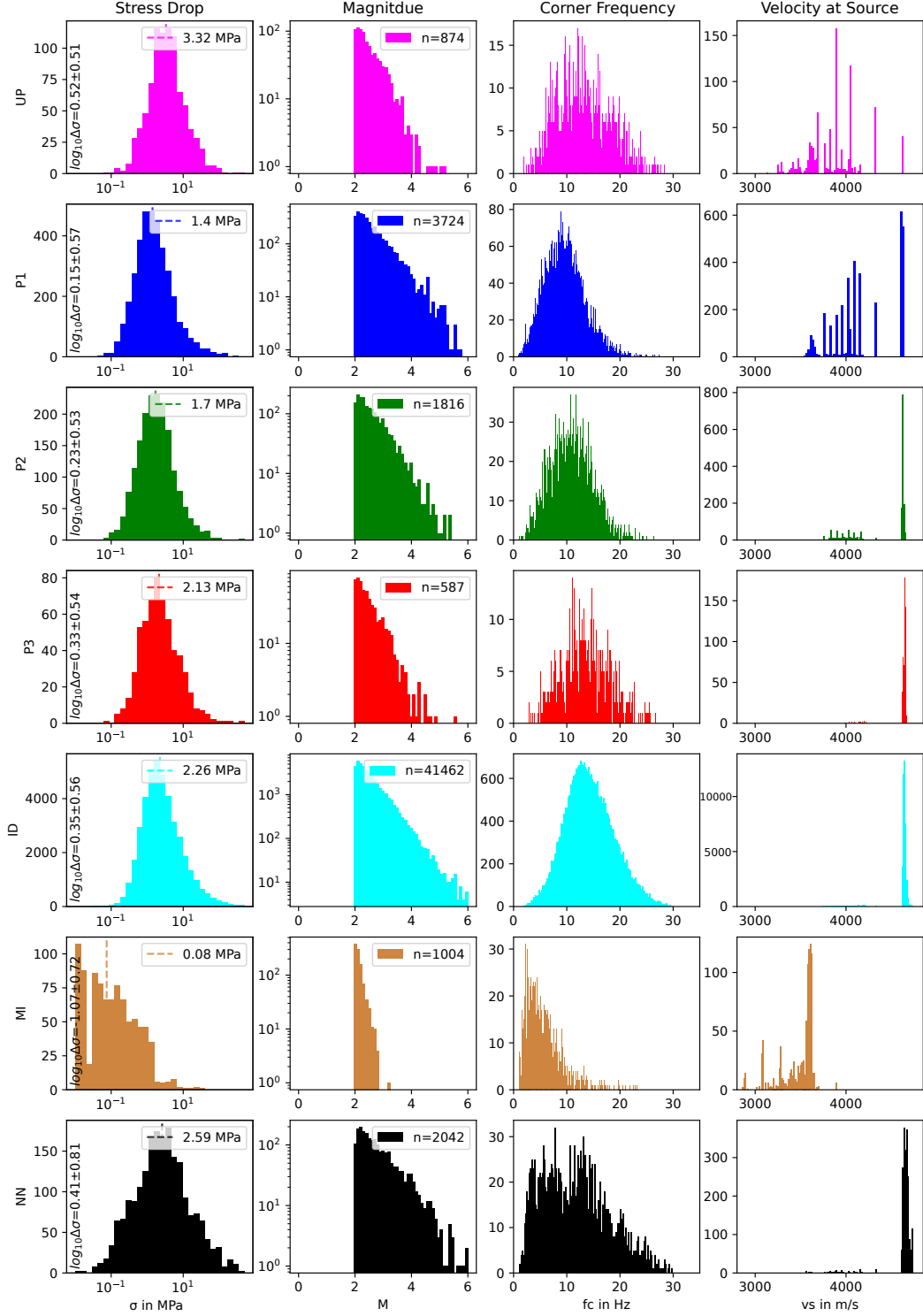


Figure 6. Distributions of stress drop, magnitude, corner frequency and S velocity at the source for UP, P1, P2, P3, ID, MI, and NN classes, from top to bottom. Stress drops were calculated based on the displayed corner frequencies, seismic moments (from the displayed magnitude distribution) and the S wave velocities at the source (eqn 3). Median stress drops are displayed in the stress drops column legends and total event number in the magnitude column legends. Note the difference in median stress drop between classes, e.g. the plate interface (P1) showing the lowest median value and intermediate depth events (ID) or upper plate events (UP) significantly higher median values.

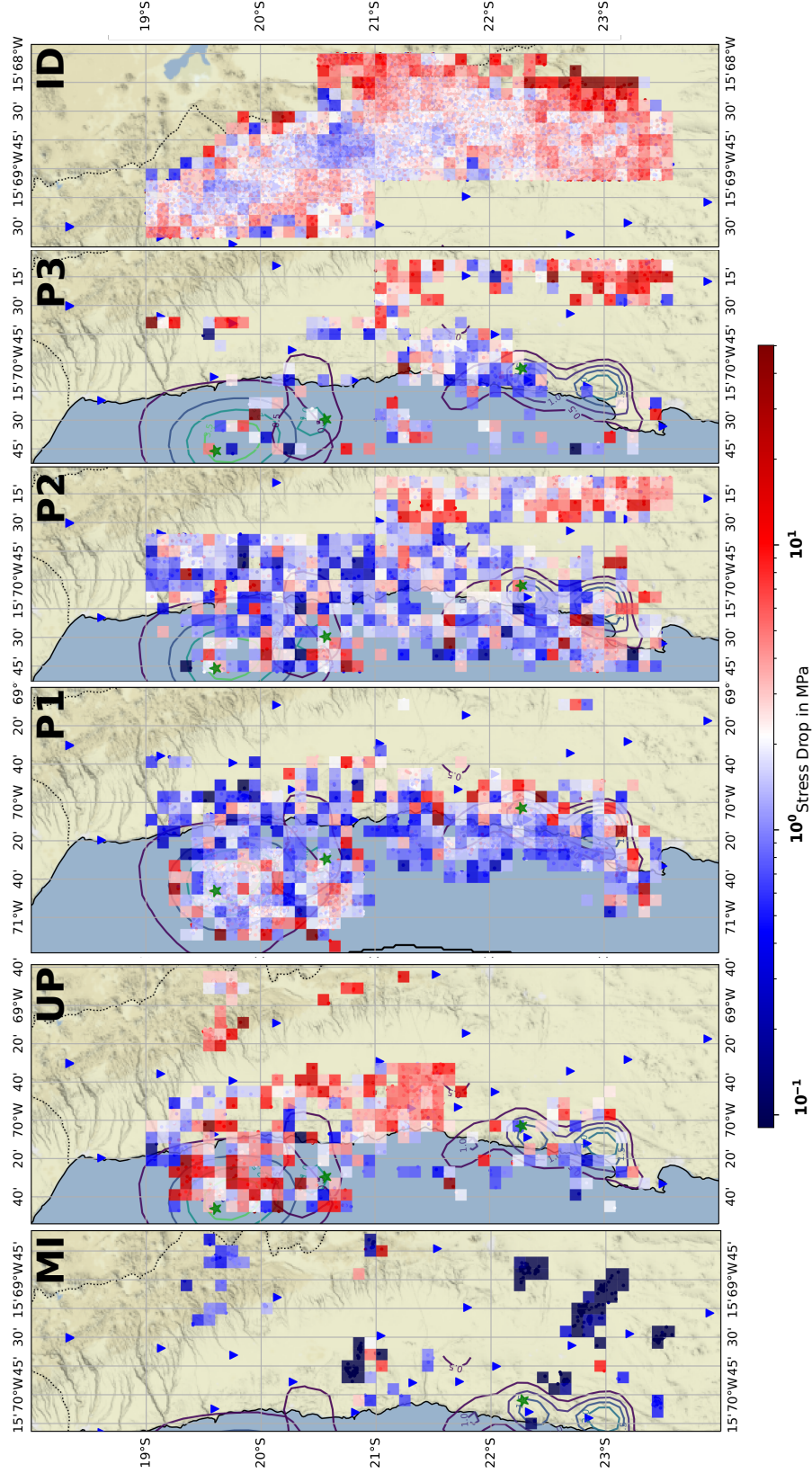


Figure 7. Stress drop distributions in the target region computed from spectral decomposition approach separated into the classes MI, UP, P1, P2, P3, and ID from left to right, respectively. The color scheme is the same as in Figure 3. Histograms and median values for each class of events are displayed in Figure 6. Detailed observations on the spatio-temporal variability of stress drop are found in the text. The map for the NN class is shown in Figure S14.

upper plate stress drops observed (Figure 7). One is located above the main rupture patch of the Iquique event, a region that was struck by the crustal Mw6.7 Iquique foreshock and its aftershocks (Schurr et al., 2014, 2020). The other one is located at 21.0°S-21.7°S, starting from the coast and extending about 70 km towards east. Seismicity rates in this area are reportedly elevated, and Sippl et al. (2018) proposed a correlation to the decreased locking observed by geodetic studies.

- P1** Interface seismicity, labeled as P1, poses the second-largest class of events ($n=3724$). It includes events from near the trench to about 69.6°W inland and almost completely covers the area between 19°S and 23.5°S. It shows the lowest median stress drop of all classes (1.4 MPa). The values found closest to the plate interface are even slightly lower, and increase with distance (cf. Figure S6) both in down and upward direction. The short range median remains stable at 2 MPa down to 15 km depth, and then decreases down to 1 MPa at 40 km. Below at 50 km, a local maximum of 1.8 MPa occurs, followed by values below 1 MPa at around 65 km depth (cf. Figure 4 and Figure S6). The stress drop distribution on the plate interface is dominated by fore- and aftershock sequences of the Iquique earthquake and the Tocopilla earthquake. The stress drop distribution in the Iquique region shows low to average values close to the hypocenter locations of both main shock and aftershock (Figure 7). At the western and northern rim of the main shock, high stress drop values are observed. Just between the main and aftershock slip surfaces, a band of very low stress drop values is located, similar in values to those located at the southern end of the aftershock slip area. The high slip regions of the Tocopilla event are characterized by low to average median stress drop values, while the down dip edge is dominated by high values and the up dip edge is surrounded by low values. The region north of the Tocopilla event rupture area shows a large patch of low stress drop values, covering almost the entire gap northwards to the Iquique event aftershock area. Note that this low stress drop patch directly borders the large high stress drop region to its east observed in the upper plate, as described above. Also note that, although the P1 class is only defined down to a depth of about 65 km, low stress drop values along the interface are observed down to a depth of at least 125 km (see Figure 5).
- P2** The first seismically active band below the interface has a slightly higher median stress drop (1.7 MPa, $n=1,816$) than the interface seismicity. Similar to P1 it covers a large region almost completely. Maybe due to its lower event count the map shows more fluctuations and less well defined low/high valued patches. Still, the stress drops close to the hypocenters of the Iquique event and its main aftershock are recognizable higher than in P1. The low stress drop band between the two mayor events is also indicated, similar to the low stress drops in the seismic gap between the Iquique aftershock and the Tocopilla earthquake, which here extends slightly more to the east. In the central-eastern and southeastern part, some patches of elevated stress drops are observed. The median stress drop depth variation is similar to the P1 curve, having a peak of 2 MPa at about 50 km depth, followed by a decrease. From 65 km on, an increase to about 3 MPa occurs (Figure 4).
- P3** The same variation with depth is seen for the P3 class events, which are part of the second interface parallel band. The limited amount of events ($n=587$) is spatially separated into several areas, like the northern part of the Tocopilla earthquake slip surface, associated with low stress drop values and some deeper and more eastern cells, where median values reach up to 6 MPa. Interestingly, this is a higher value than for the neighboring cells, located in the intermediate depth class (ID). Especially the latter group of comparatively deep events is responsible for the relatively high median value (2.13 MPa) found for this class.
- ID** The ID class (or else called IDEs) is the largest group ($n=41,462$) and shows a higher median of 2.26 MPa compared to the Px groups. It is the only class where the stress drop map is locally fully continuous and many cells exceed 1000 event members. While the ID event stress drop map is dominated by median to high values (Fig-

ure 7) there exist two patches of lower than average stress drop at about 19.5°S and at 20.75°S . Especially the latter region is seismically the most active region in the entire data set (Sippl et al., 2018) and thus, it contributes many low stress drop values to this event class which otherwise would have an even higher median value. In its shallowest part the depth variation is similar to that of the Px group, with a minimum (1.5 MPa) at about 75 km depth, followed by an increase to 2.2 MPa at about 90 km depth and, after some short decrease, a more rapid rise to about 8.5 MPa at a depth of 140 km. Further down the increase of stress drop appears to cease at 175 km below which stress drops increase again up to values exceeding 10 MPa. Note that this class of events shows the highest range of median stress drop variation with depth. Several possible reasons for the increase will be discussed later.

- MI** We obtain stress drop estimates for 1,004 events from the mining class with a very low median stress drop of 0.08 MPa. 876 of them were labeled MI in the IPOC catalog and 128, which initially were classified as UP events, were reassigned to the MI class because of their particularly low stress drops and their clustered and shallow locations. It is interesting to note that only a small fraction of the events exceed stress drops of 1 MPa while many estimates significantly undercut 0.1 MPa. The statistical distribution of the mining-related events stands out against the other classes of tectonic events and deviates strongly from a log-normal shape as seen in Figure 6.
- NN** The locally less well constrained events from the NN class are basically split into two mayor groups. The first group are those events which lie far offshore and have poorly constrained depths. The second group are events located far east of the network mostly at great depth ($> 150\text{ km}$). Here, the location accuracy is also decreased. Hence, the stress drop values suffer from possibly wrongly assigned S wave velocities. For example, for the shallow group, almost exclusively lower plate mantle velocities are used. Should any of them lie at or closely below the interface, their stress drops would be much higher when using lower plate crustal velocity. The stress drop distribution from Figure 6 reflects the separation of stress drops into two groups by a comparatively large standard deviation and a bimodal shape of the corner frequency distribution. Including this class here has two main reasons. One is the additional depth extent (160–180 km) covered almost exclusively by this event group, and the second is the consistency of the processing with the complete IPOC catalog. Similarly as the MI events we do not further discuss their properties or interpret their stress drops. The corresponding stress drop map for this class is shown in the supplement Figure S14.

Temporal Variation

Stress drops do not only vary in space, but also in time (e.g. Allmann & Shearer, 2007). We observe such variation for shallow, predominantly interface and upper plate seismicity in Figure 8. Strikingly, the one hundred event median shows maxima at the occurrence times of the large interface events. We therefore show narrower time frames for spatially limited areas around four major earthquakes: the Mw8.1 Iquique event, its Mw6.7 foreshock, its Mw7.6 aftershock and the Mw7.7 Tocopilla event. All stress drop curves have similar shapes. Initially, at occurrence time, the median stress drop is high, up to multiple times the long term median of the specific region. In the following days, a decline of stress drop values down to values slightly below average is observed, followed by a recovery back to average values. The decay time is longest for the largest of the four and shortest for the smallest main shock.

This behavior has already been described in Folesky et al. (2021) for the Iquique event. There, the variability was attributed to the increased moment rates during the immediate post-seismic interval. We observe the same effect here for three additional large magnitude events. To illustrate the influence of the temporal moment rate variability,

we plot the time- dependent stress drop variation scaled by event moment rate in the supplement (Figure S16).

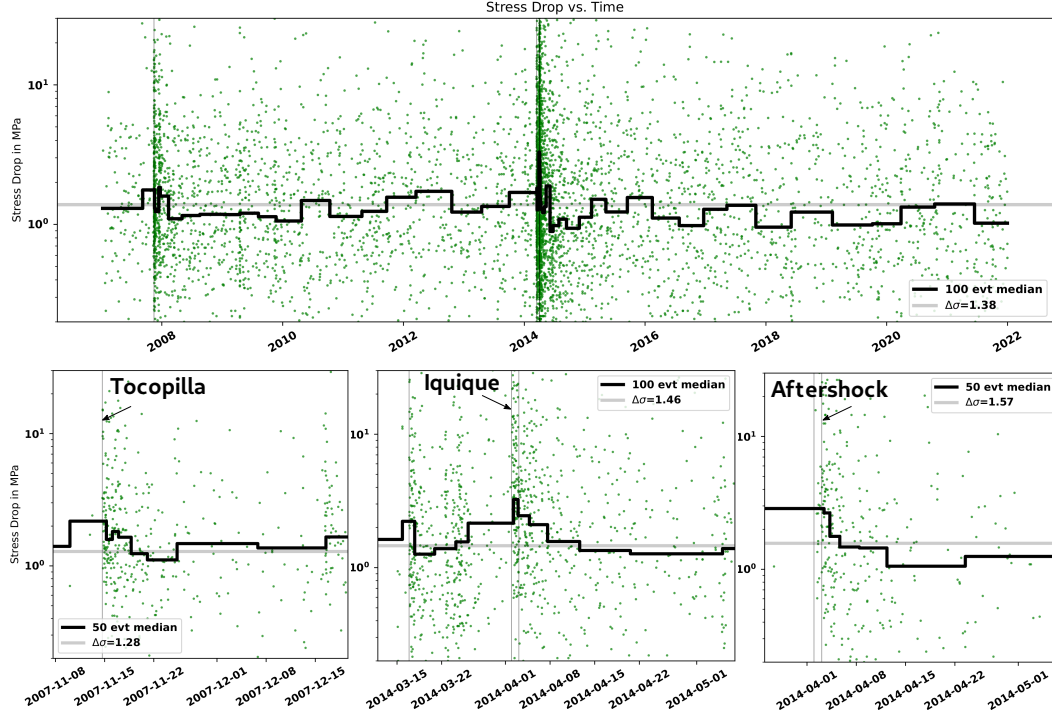


Figure 8. (Top) Stress drop distribution versus time for events limited to latitudes west of 70°W for the entire observation period. The median is computed over 100 events per bin. No smoothing is applied. Grey vertical lines denote the Tocopilla and the Iquique event. (Bottom)(left) Five weeks of stress drops from the Mw7.7 Tocopilla aftershock area ($\text{lat} \leq -70$, $-21.5^{\circ} > \text{lon} > -23.25^{\circ}$). The vertical line is the Tocopilla origin time. (center) Stress drops from the Iquique aftershock area ($\text{lat} \leq -70$, $-19^{\circ} > \text{lon} > -20.25^{\circ}$) 3 weeks before to 5 weeks after the Iquique event. Grey vertical lines denote the Mw6.7 foreshock, the Mw8.1 Iquique event, and the Mw7.6 aftershock, respectively. (right) Stress drops from the Mw7.6 Iquique aftershock area ($\text{lat} \leq -70$, $-20.25^{\circ} > \text{lon} > -20.75^{\circ}$). The second vertical line is its origin time. The grey horizontal line is the long term stress drop median for the corresponding region. For all large events, the stress drop values are elevated close to the origin times of the main shock, followed by a rapid but not instant decline to values below the median, followed by a return to average values within a few days to weeks.

Discussion

In this section, we discuss the results of this study in the context of other stress drop studies. We will highlight the most interesting aspects of stress drop distribution and show possible correlations with other research in the northern Chilean subduction zone. In doing so, we highlight the potential of a comprehensive stress drop map to contribute to interpretations of the tectonic state of, for example, the megathrust, or to questions of source physics and stress distribution in a subduction environment.

The overall range of obtained stress drops in this work is between 0.1–100 MPa, a well accepted corridor for natural seismicity found in multiple studies for many events and a broad magnitude range (e.g. Allmann & Shearer, 2007, 2009; Uchide et al., 2014; Tian et al., 2022). Median values vary significantly less, ranging from 1 MPa for shallow events to 15 MPa for the deepest in this data set, as well as from 1.4 MPa to 3.1 MPa between different event classes. Median values for spectral ratios are on average about twice as high. This shows that although median stress drop is found to differ between different regions, depths, and classes with good consistency, a single event cannot unambiguously be assigned to a class based solely on its stress drop value.

Especially absolute values have to be handled with care. Not only the parameters in the chosen source model but also the methods for determining the corner frequency may introduce additional differences. It is clear that different source model choices, i.e. k -values, also produce large deviations in absolute stress drop (e.g. Kaneko & Shearer, 2015; Ji et al., 2022). Hence, it is beneficial to process many events similarly and interpret results based on relative differences, if possible confirmed by multiple methods (e.g. Pennington et al., 2021).

For example Cabrera et al. (2021) compute stress drops for six large magnitude events located at shallow intermediate depths in northern Chile. They use kinematic source inversion and find absolute stress drops of 7–30 MPa. These values seem relatively low compared to our results of large magnitude intermediate depth seismicity. Five out of six of them are located north of 19°S and therefore not included in the IPOC catalog because the region is no longer covered by IPOC stations. Therefore, a direct comparison is unfortunately not possible. However, since the authors do not provide further details on inversion parameter trade-offs and assumptions of the source model, a direct comparison of absolute values would not have been meaningful anyway.

Derode & Campos (2019) estimated corner frequencies from attenuation-corrected spectra for 96 events in northern Chile. They do not provide values for the stress drop, but note that the corner frequencies of intermediate depth earthquakes are significantly elevated compared to shallow depth earthquakes, which would lead to elevated values for the stress drop, consistent with our results. They emphasize that the stress drop depends strongly on the unknown rupture velocity, which can vary from event to event, which is also true for our study. Folesky et al. (2018) calculated rupture velocities for small to moderate earthquakes in the Iquique region. The resulting average velocity fits well with the assumed value of 0.6β from this study, but the reported scatter of velocity values is quite large, although the study was limited to shallow seismicity. Therefore, we note that variable rupture velocities are an important source of perturbation to consider. For example, the increase in observed stress drop could be explained simply by a change in the assumed average v_r/v_s with depth. We did not observe a change in the f_p/f_s ratios with depth, which would have been a clear indication of a change in rupture velocity (Kaneko & Shearer, 2014, 2015). In this study, similar to Folesky et al. (2021), we observe a clear dependence of stress drop on moment (Figure 9), which contradicts the invariance hypothesis of stress drop scaling for northern Chile. We fit the linear relation ($\log_{10}(\Delta\sigma) = \varepsilon_0 + \varepsilon_1 \log_{10}(M_0)$) to stress drop estimates over separate magnitude bins of 0.5 width (e.g. Trugman & Shearer, 2017) and we obtain $0.46 < \varepsilon_1 < 0.68$. This indicates a relatively strong dependence of stress drop on moment. As reported in Folesky et al. (2021) more restrictive quality criteria for spectra included in the analysis may moderately decrease the ε_1 parameter, but the clear positive dependence remains.

A depth dependence of stress drop was reported in many studies (e.g., Uchide et al., 2014; Boyd et al., 2017; Trugman, 2020), while others find no such evidence (e.g. Shearer et al., 2006). A conclusive explanation of depth dependent stress drop or a convincing correlation to other parameters is still missing. Whether stress drop increases in the crust was recently discussed in detail by Abercrombie et al. (2021), who find that most studies probably overestimate the depth dependency by different ways of under-correcting

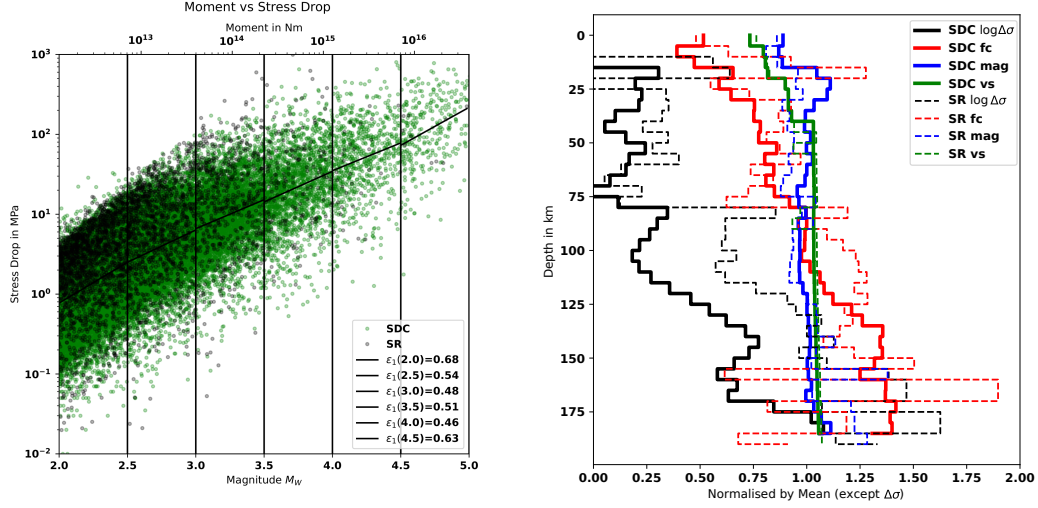


Figure 9. (left) Stress drop against moment with SDC results in green and SR results in black. The black line is the bin wise linear fit of the SDC data. The plots shows a clear increase of stress drop with moment for each single bin. (right) Depth against median log stress drop, normalized corner frequency, normalized magnitude, and normalized S wave velocity, demonstrating that the main driver behind increased stress drop with depth (>80 km) is the increase of corner frequency. Solid lines are SDC medians, dashed lines are SR medians.

the depth dependent attenuation. They state that using a spectral ratio approach with good EGFs should principally be unaffected by such problems. We do not find significant qualitative differences of results of our two approaches concerning this question. And we observe both, a stress drop increase in the crust for $z \leq 20$ km from 2 to 4 MPa and below, a median stress drop that decreases slowly back to about 2 MPa over a range of 40 km (Figure 4 and Figure S6) with minor fluctuation. In northern Chile, continental crustal thickness ranges up to 60 km (Patzwahl et al., 1999) which is also represented in the utilized velocity model (Figure S17). Note that the UP class also contains seismicity from the subduction wedge and that the Moho deepens from west to east with the plate interface deepening. Additionally, stress drop sampling in the upper crust is not uniform but concentrated to several regions. The behavior of the average stress drop median with depth might therefore not be unbiased by the subduction geometry and most likely it is not indicative for intra-plate stress drop variability.

The median value of stress drop combined for all classes shows three relative maxima at 10 km, 55 km and 85 km depth (Figures 4&9). In the corresponding depth range minimum and maximum vary only between 1.25 MPa and 2.5 MPa. At greater depths a consistent increase to values up to 8.5 MPa is observed followed by a plateau another subsequent increase to values above 10 MPa. To investigate the driving parameter for the increase we plot corner frequency, magnitude and S wave velocity against depth (Figure 9). The top 60 km show a complicated behavior of the three curves, but it appears that medium magnitude values remain stable below 40 km and mean velocity is basically constant below 60 km depth. Solely the median corner frequency estimate increases almost steadily with depth causing the rise in stress drop. Therefore, we conclude that the increase of stress drop with depth is mostly driven by an increase of corner frequency. This holds for both the spectral decomposition and the spectral ratio method (with significantly stronger fluctuation), and we consider this result robust.

Interestingly, and different from this observation, the variability of stress drop with time found for the major earthquakes in the catalog region (Figure 8) appears to be driven by increased moment release rates, rather than temporally elevated average corner frequency (cf. Figure S16). Once this effect decays, the overall median stress drop returns to its initial value. Also, on a longer time scale, its alteration is not permanent. For example, the difference of median stress drop for events that occurred earlier than four weeks before the Iquique event and events that occurred later than four weeks after the Iquique event is about 0.03 MPa. For the Tocopilla event, this difference is 0.09 MPa. A similar observation was made in California where the 2004 M6 Parkfield earthquake reportedly did not permanently change the stress drop pattern (Allmann & Shearer, 2007).

To our knowledge, a consistent, continuous and comprehensive analysis of the stress drop pattern over the broad depth range of 0-180 km as performed in the present study has not yet been reported elsewhere. We can therefore only compare with studies that cover subdomains of this range. Allmann & Shearer (2009) studied the distribution of stress drop for global seismicity and found evidence of a downward increase in stress drop from a depth of 30 km. Şen et al. (2015) analyzed the rupture duration of aftershocks from the 2010 Mw8.8 Maule earthquake, Chile, at 35°S and noticed variability in normalized rupture duration with a minimum at 40 km depth. Since the source duration is inverse to the corner frequency, a reduction corresponds to an increase of stress drop. However, in both studies, no results are calculated below 60 km depth. In Japan, Uchide et al. (2014) studied the variation of stress drop in the broader region of the 2011 Tohoku earthquake. Their analysis includes events to a maximum depth of 80 km. They find a sharp increase in stress drop between 30-60 km accompanied by plateaus with constant median stress drop below and above. They have no conclusive interpretation for the increase or the plateaus they observed.

Other studies focus on deeper events, such as Poli & Prieto (2016), who estimate the stress drop for global ID events and deep events based on estimates of the duration of the source time function. They find mean stress drop values of about 10-20 MPa for their shallowest events at about 150 km depth, with no subsequent increase. Given the general variability in stress drop estimates among authors and methods, these absolute values fit well with the estimates that we obtained for the deepest events in Northern Chile, which also indicate a cessation of stress drop increase near the depth limits of our study. Tian et al. (2022) analyze stress drops from intermediate depth to deep events in the Tonga region and find average values around 5.6 MPa at 90 km depth, followed by decreasing median values to a minimum of 3 MPa at 170-250 km and then by an increase back to 6 MPa at greater depth. They report that they can best associate zones of high stress drop with areas of high plate deformation. In northern Chile, the plate undergoes strong geometric changes as it bends at shallow depths (<60 km), stops bending below (60-100 km), followed by bending that ends with a strong downward kink (Sippl et al., 2019, 2023). The strongest geometric deformation is observed between 100-140 km depth (Sippl et al., 2022), where the plate is bent to a significantly steeper subduction angle. As the depth of this bending process varies laterally, an increasing number of high stress drop events is gathered to calculate their median, which could be the cause of the continuous increase in the median stress drop that would otherwise be more abrupt. At about 21°-21.5°S the ID seismicity band is partitioned into a northern and a southern part. At this latitude range the subduction geometry appears significantly perturbed (Sippl et al., 2018) possibly inducing additional zones of increased stress. Interestingly, we observe laterally increased median stress drop in this area (between 21°-22°S, cf. Figure 7). If higher plate deformation is indeed a significant driver behind the stress drop increase one could also suspect that it is responsible for the relative maxima of the median stress drop depth variability at shallower depths.

The depth region of 100–140 km also includes by far the greatest amount of earthquakes in the IPOC catalog, especially between 20°S to 22°S. The initially at shallower

depths observed parallel bands of seismicity dissolve at 80–100 km depth into a 25–30 km single band of very high activity reaching further down. In addition to the above mentioned bending, Sippl et al. (2019) discuss the spatial coincidence of the vanishing velocity contrast between oceanic crust and underlying mantle in receiver function studies and the transition of the cold nose to the hot part of the mantle wedge indicated by seismic attenuation images. To explain the strongly increased earthquake productivity they propose the sudden activation of a kinetically delayed metamorphic reaction with negative volumetric component which further strengthens the local stress level already in place due to the strong slab pull. The increased observed median stress drops in this area might be an expression of the same process which is responsible for the strong increase in seismic activity in this region.

The observed variability of stress drop with depth might also be associated to variable seismic coupling along the interface. The uppermost 70 km are dominated by stress drops from P1 and P2 classes representing interface and upper plane seismicity. At first order the average depth variability curve from these events, shown in Figure 4, seems to correspond to findings of Hainzl et al. (2019) who investigated the relation of aftershock productivity and coupling. The authors compared laterally averaged coupling with depth obtained from three coupling maps (Béjar-Pizarro et al., 2013; Schurr et al., 2014; Li et al., 2015) to aftershock productivity rates, moment release rate and b-values. They report that the shallow region at about 20 km depth shows highest average locking with strong decay down to about 60 km which is reflected in decreasing aftershock productivity rates, moment release rate and b-values, too. They propose a simple interpretation of variable asperity density with depth as a possible source for the variation. The observed scaling of stress drop with moment from our study would support an at least moderate response of the stress drop to the reported variation of moment release with depth, as we observe a moment release maximum at 20–25 km depth with subsequent decrease (Figure 9).

The seismic coupling can also be directly compared to the obtained stress drop distribution along the interface. For example, a very low stress drop region is observed between the rupture patches of the Iquique event and the Tocopilla event (cf. Figure 7) called the *Loa* section of the northern Chilean Margin. One might expect a correlation with low coupling in the same region, but in contrast, several GPS studies find a high coupling (e.g. Schurr et al., 2014; Li et al., 2015; Hoffmann et al., 2018). It is a debated question why neither the Tocopilla earthquake nor the Iquique earthquake have activated this section of the megathrust. One possible explanation might be a creeping behavior, mitigating the potential for large rupture. In California Allmann & Shearer (2007) have reported low stress drop values in the creeping section of the San Andreas fault to the north of the Parkfield event. Their creeping section, however, is separated by a relatively high stress drop barrier from the locked region. Our stress drop map fits into this picture if we consider the few high stress drop values just south of the Iquique large aftershock slip area as indicative for some kind of such barrier. This is supported by observations of repeating earthquakes at that point (Soto et al., 2019; Folesky et al., 2022) which are generally interpreted as indicative for surrounding or adjacent fault creep. Hence, south of the barrier the interface would be considered creeping, also indicated by the relative seismic quiescence, there (Sippl et al., 2023). Further south we observe very low stress drop events which increase in average values until they reach the Tocopilla slip area. From gravity data a locked asperity hosting potential for a large event generally is correlated with decreased values in the residual gravity field (Wells et al., 2003). In contrast to the prior mentioned locking studies Bassett & Watts (2015) report a gravity high for this region which would rather support the interpretation as a creeping segment.

The discussed aspects of the stress drop map obtained here only show possible starting points for linking and deepening the understanding and interpretation of tectonic and geophysical processes in subduction zones, especially in northern Chile. The large ex-

tent and homogeneous processing of our data set lend themselves well to more detailed analysis of lateral or depth profiles, specific regions, or temporal variability. Again, we can only emphasize that stress drop variations should be interpreted as relative variations and absolute values should be treated with caution.

Conclusions

In this study we have computed a comprehensive, consistently processed stress drop catalog for Northern Chile. The data base for this work is the IPOC catalog with over 182,000 events in the time period from 2007 to 2021. We additionally perform template matching using the entire waveform archive to obtain an enlarged set of empirical Green's functions. Stress drops are computed by two different methods. Firstly, a spectral ratio approach (SR), which was recently tested in the Iquique earthquake region, is now applied to the entire northern Chile seismic data providing 4,223 stress drops estimates. While these results show already distinct stress drop patterns for some regions, they are limited by the irregular availability of EGFs. To overcome this restriction and to complement the SR results, we secondly use a modified spectral decomposition approach (SDC). We carefully test and then apply a modification of the standard SDC technique by using multiple well-defined cells to account for the variability of seismic attenuation in the subduction zone. The SDC approach yields 51,510 stress drop estimates improving significantly the continuity and coverage of the stress drop catalog. We find a linear relation of the estimated corner frequencies between both approaches, $fc_{SDC} = 0.75 \times fc_{SR}$, translating into $\Delta\sigma_{SDC} = 0.42 \times \Delta\sigma_{SR}$ for stress drops. Besides the systematic differences in absolute stress drop values, which has also been reported previously in other studies, the two methods produce very similar features.

The SDC- based stress drop distribution represents the first coherent and large-scale stress drop mapping of a subduction zone including several tens of thousands events. We observe small, but systematic differences of median stress drop values between the seismotectonic domains. Interface seismicity is clearly characterized by the lowest median stress drops (1.4 MPa). Also, the two slab-parallel seismicity bands within the subducting Nazca plate exhibit rather low median stress drop values of 1.7 and 2.1 MPa, respectively. Upper plate events, which occur almost exclusively in the continental crust, show higher stress drops with a median of 3.3 MPa for the SDC method and 4.0 MPa when using the SR method. Intermediate depth seismicity is monitored down to a depth of about 180 km. It shows a median stress drop of 2.3 MPa. Two additional classes are treated separately in the analysis. For mining induced events we find particularly small stress drops of about 0.3 MPa. The events from the NN class are poorly constrained; they have the largest location uncertainties and heterogeneous stress drops.

Both interface seismicity and the two seismicity bands in the downgoing slab show only small variations of less than 1 MPa with depth in the interval at 0-80 km depth, i.e., the variability of the median stress drop along the subduction is low. Starting at 80 km depth, and especially also for the intermediate depth earthquakes, we observe a consistent increase of the stress drop from about 2 MPa to 15 MPa.

Data & Resources

Waveform data used in this study was recorded by the seismological CX-net of the Integrated Plate boundary Observatory Chile (IPOC, 2006) using STS-2 broadband seismometers. It was obtained from the EIDA/GEOPHONE web page (eida.gfz-potsdam.de/webdc3/ or geofon.gfz-potsdam.de/waveform/, doi:10.14470/PK615318). Picks, magnitudes and event hypocenter were taken from Sippl et al. (2023). Data processing and figure production were mainly performed using Python3.5.1 (python.org) and packages IPython4.2.0 (Pérez & Granger, 2007), NumPy (Walt et al., 2011), Matplotlib (Hunter, 2007), Ob-

sPy (Beyreuther et al., 2010) and SciPy (Virtanen et al., 2020). Some figures were refined using Inkscape (inkscape.org).

Results from this study are summarized in a table described and made available in the electronic supplement.

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A Comprehensive Stress Drop Map from trench to Depth in the Northern Chilean Subduction Zone

Supplementary Material

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1 This is the supplement file to the manuscript "A Comprehensive Stress Drop
2 Map from trench to Depth in the Northern Chilean Subduction Zone".

3
4 Supplement file 2 (**nc_stress_drop.csv**) comprises the final resulting stress
5 drop table. It consists of the columns [ID, cls, Lon, Lat, Depth, Magnitude,
6 vssource, fc, sd]. The corresponding units are [-,-,°,°,km,-,m/s,Hz,MPa].

7 f_c is the mean f_c from all existing estimates, both P and S phase based and
8 from multiple cells. For this, we have converted all fc estimates to pseudo f_{cp}
9 estimates by applying the relation $f_{cp} = f_{cs} \frac{k_p}{k_s} = f_{cs} \frac{0.32}{0.265} = 1.204 f_{cs}$ from the
10 main manuscript.

11
12

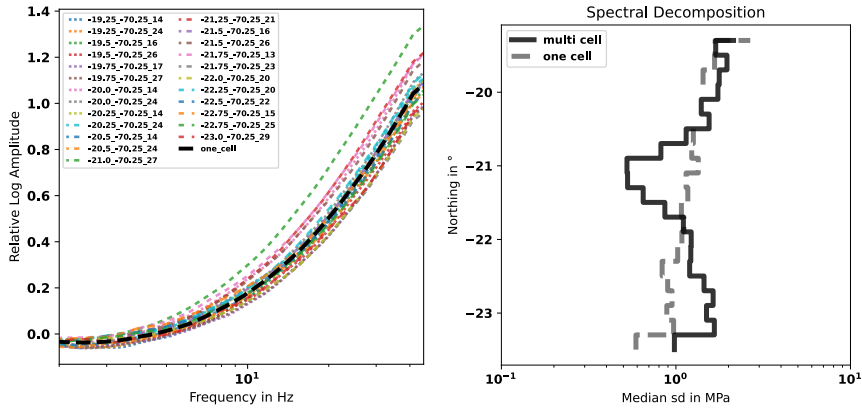


Figure S 1: Similar to Figure 2 main manuscript but for latitudinal variation. (left) Selected multi-cell gEGFs for cells between 70.25°W to 70.75°W and 19.25°S to 23.5°S, and cell top-depth ≤ 25 km. Note, that not all cells are populated due to processing standards. The combined single cell gEGF is plotted as dashed black line. Again, it constitutes a sort of average of all the single gEGFs.

(right) One- and multi-cell median stress drop variation against latitude. Note, that the median stress variation, represented by the -21° gEGFs, is not covered by the single cell gEGF. Here, the global correction appears to be too crude to account for the particular attenuation variation in that region.

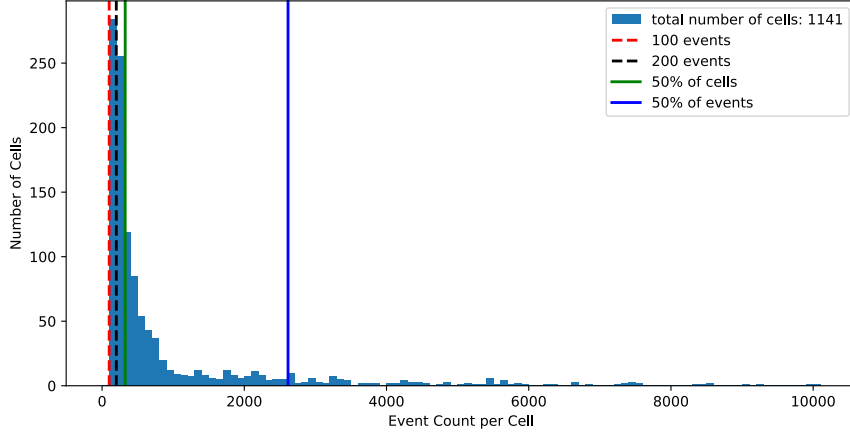


Figure S 2: Events per cell for events and cells that met the processing standards for the spectral decomposition procedure. P and S phase based cells are counted separately. Red line is the minimum number of events in a cell required for processing, $n=100$; black line is for $n=200$. The green line shows the position where the total number of cells is split in two parts. The blue line shows the 50% split for the total event count.

The figure illustrates, that event count varies strongly between cells. On the one hand this shows, that there are many low count cells which might have slightly less stable gEGFs, because sampling is relatively sparse. On the other hand, not using these cells would result in a significant loss of coverage of the stress drop map.

Additionally, the figure shows, that larger cells or even a global gEGF for the entire catalog would be dominated strongly by locations of very high event productivity, that is, in particular the intermediate depth cloud region. Consequently, upper plate events and interface events would undergo a potentially strongly biased correction.

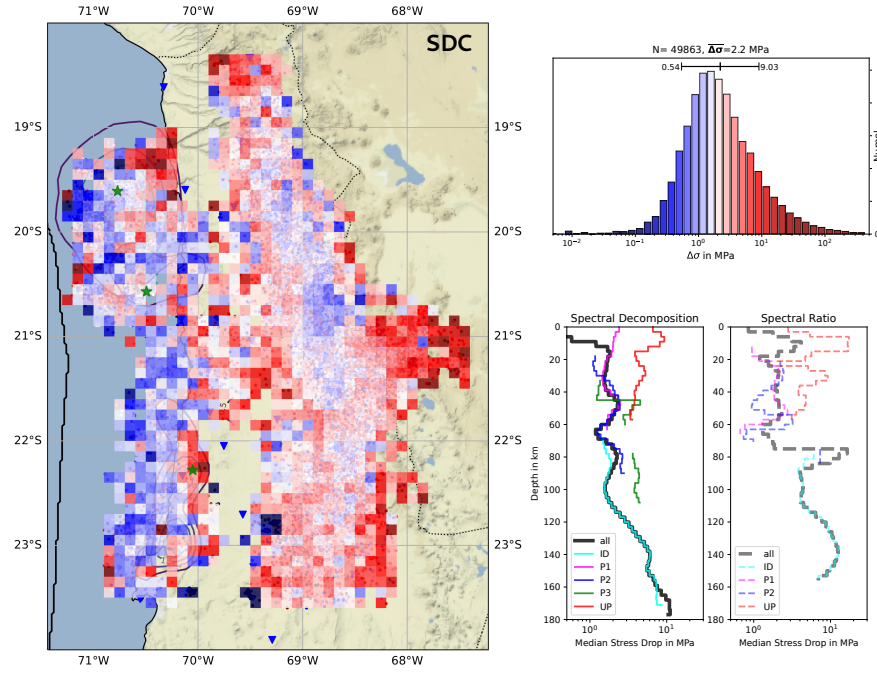


Figure S 3: Stress drop map, histogram and median stress drop variation with depth obtained using the sdc method but an increased cell size of $0.7^\circ \times 0.7^\circ \times 20$ km rather than the $0.5^\circ \times 0.5^\circ \times 20$ km used in the main manuscript. The minimal event count per cell was increased to 400. The spatial distribution reveals only minimal differences. The total number of events is slightly increased to 45825 and the overall median also increased. The most likely reason for this are some additional events of higher stress drop in the rim areas which are processed when using the larger cells but not included in the original, smaller cell processing.

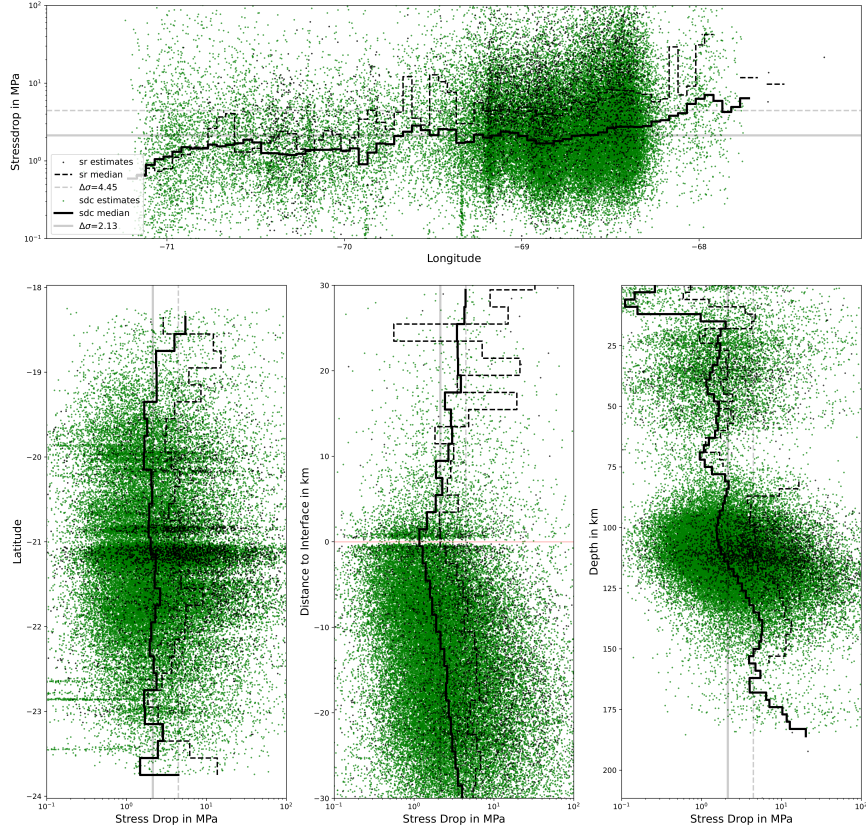


Figure S 4: Stress drop distribution from the spectral decomposition approach. The step plots show the short distance median values with fixed step lengths (no smoothing), the gray lines indicate the overall median. Top: west-east view. Panels left to right: north-south view, distance to interface, and depth view. This figure is mostly dominated by the ID events, which form about 85% of the total result count. The most prominent trends are the increase of stress drop with increase of stress drop with distance to the interface in downward as well as in upward direction, the consistent increase of stress drop with depth starting around 100 km depth, and the increase towards east. A similar figure is provided for each of the event classes including the spectral ratio results in the supplement.

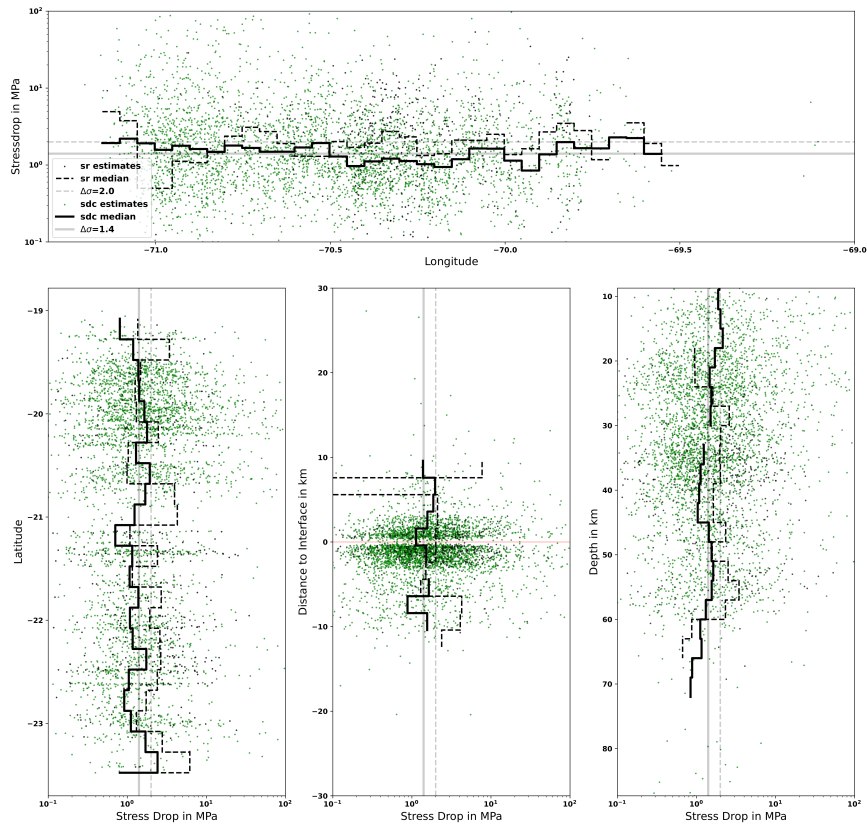


Figure S 6: Stress drops and median values for interface events (P1), only. Similar to Figure S 4.

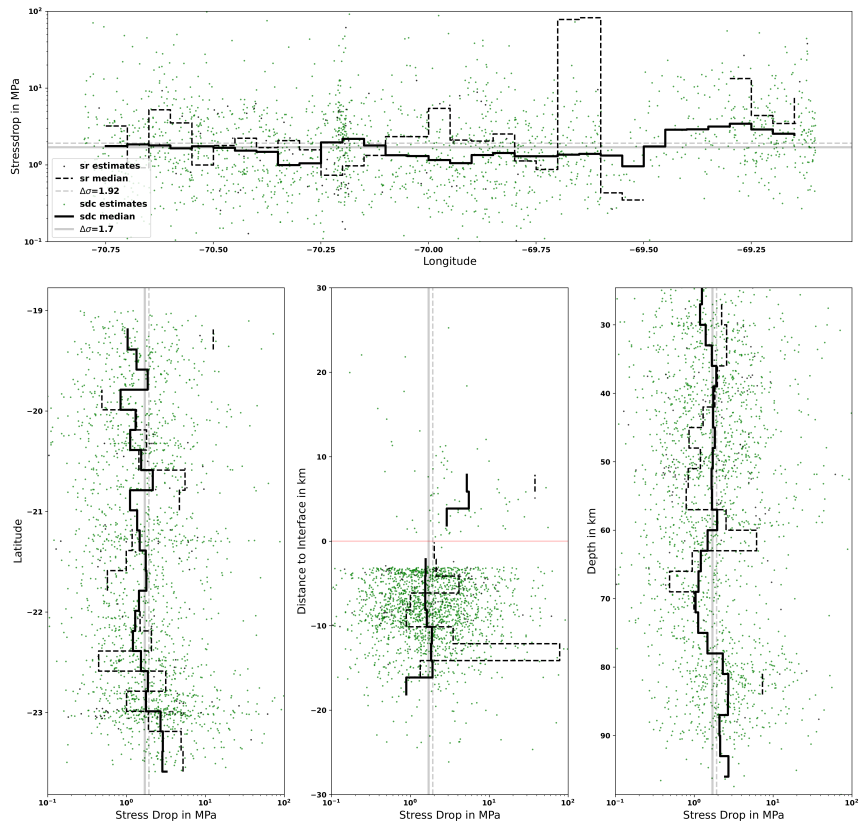


Figure S 7: Stress drops and median values for P2 class event. Similar to Figure S 4.

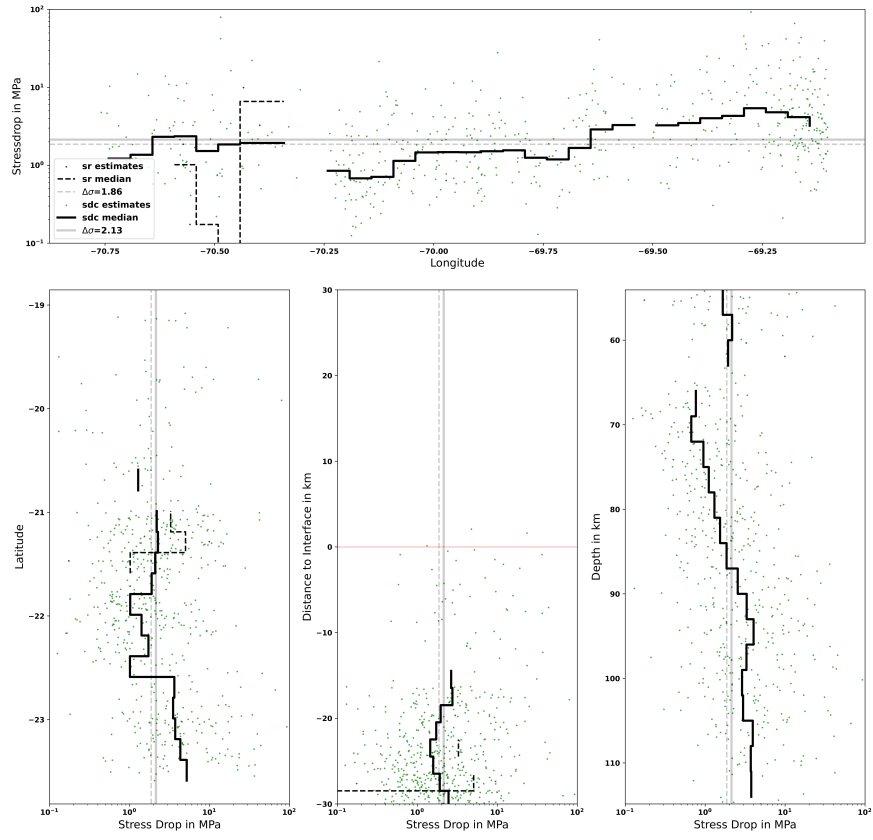


Figure S 8: Stress drops and median values for P3 class event. Similar to Figure S 4.

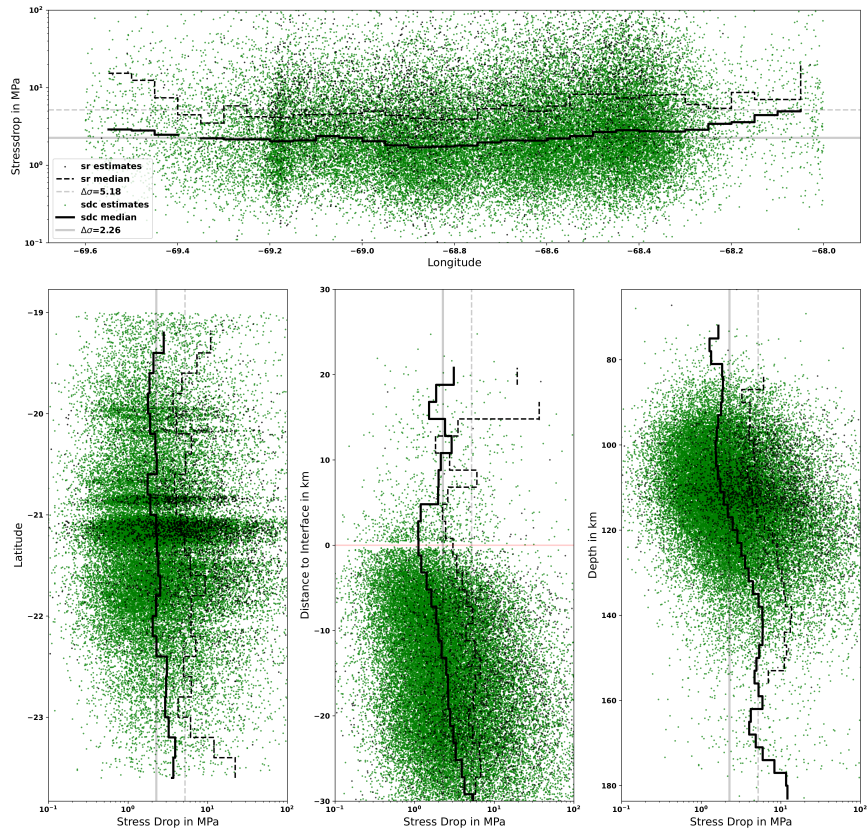


Figure S 9: Stress drops and median values for intermediate depth (ID) class event. Similar to Figure S 4.

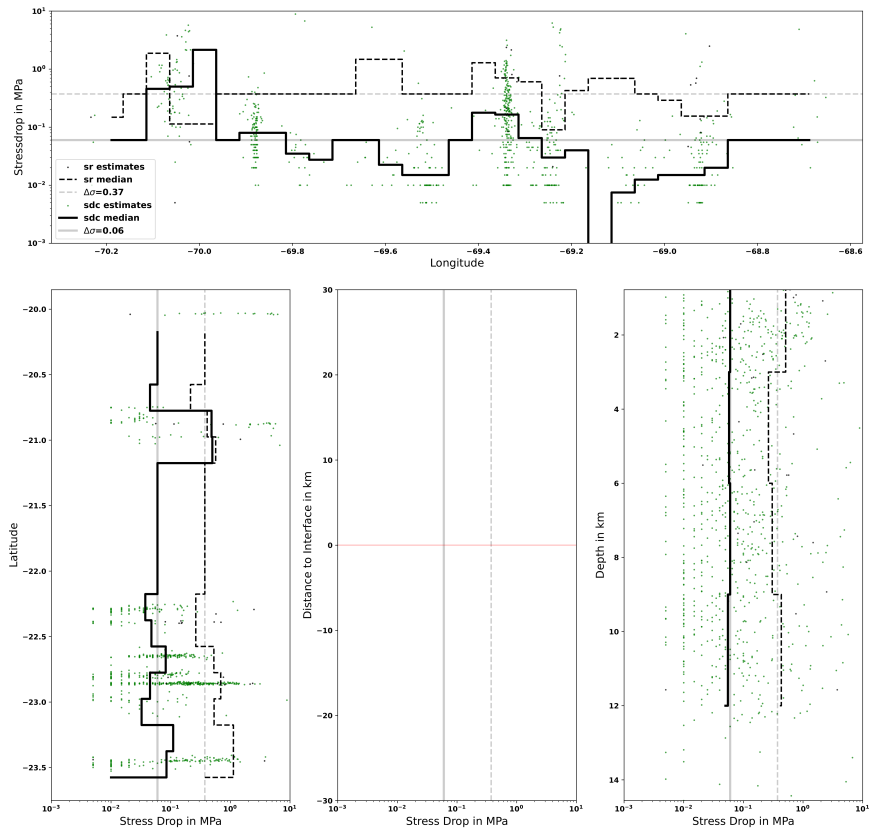


Figure S 10: Stress drops and median values for intermediate depth (MI) class event. Similar to Figure S 4.

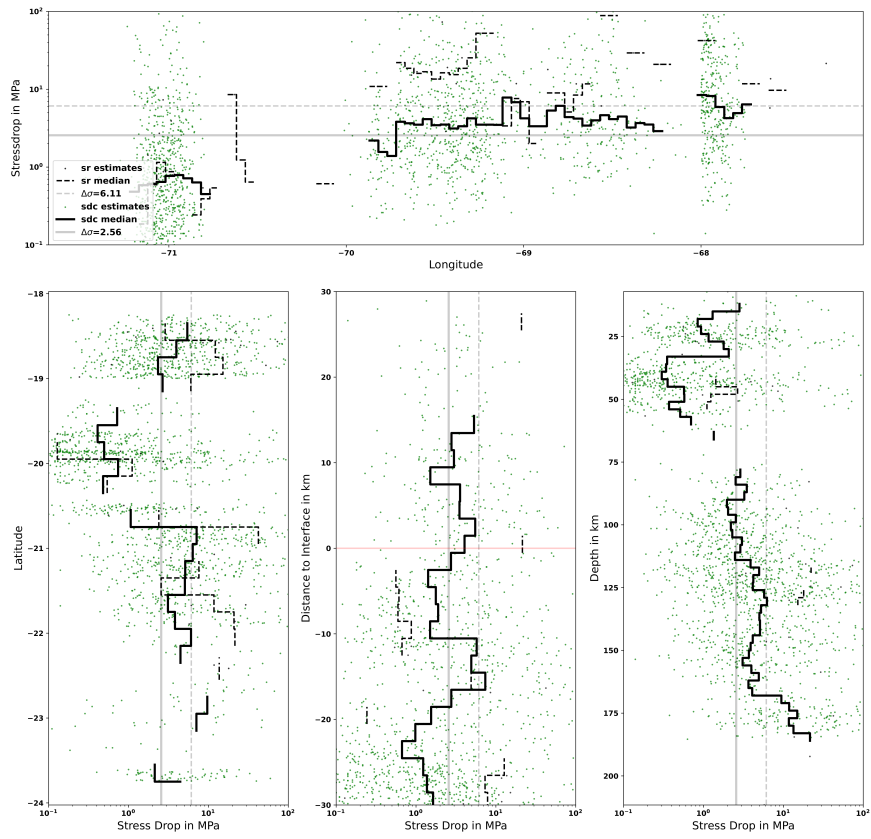


Figure S 11: Stress drops and median values for intermediate depth (NN) class event. Similar to Figure S 4.

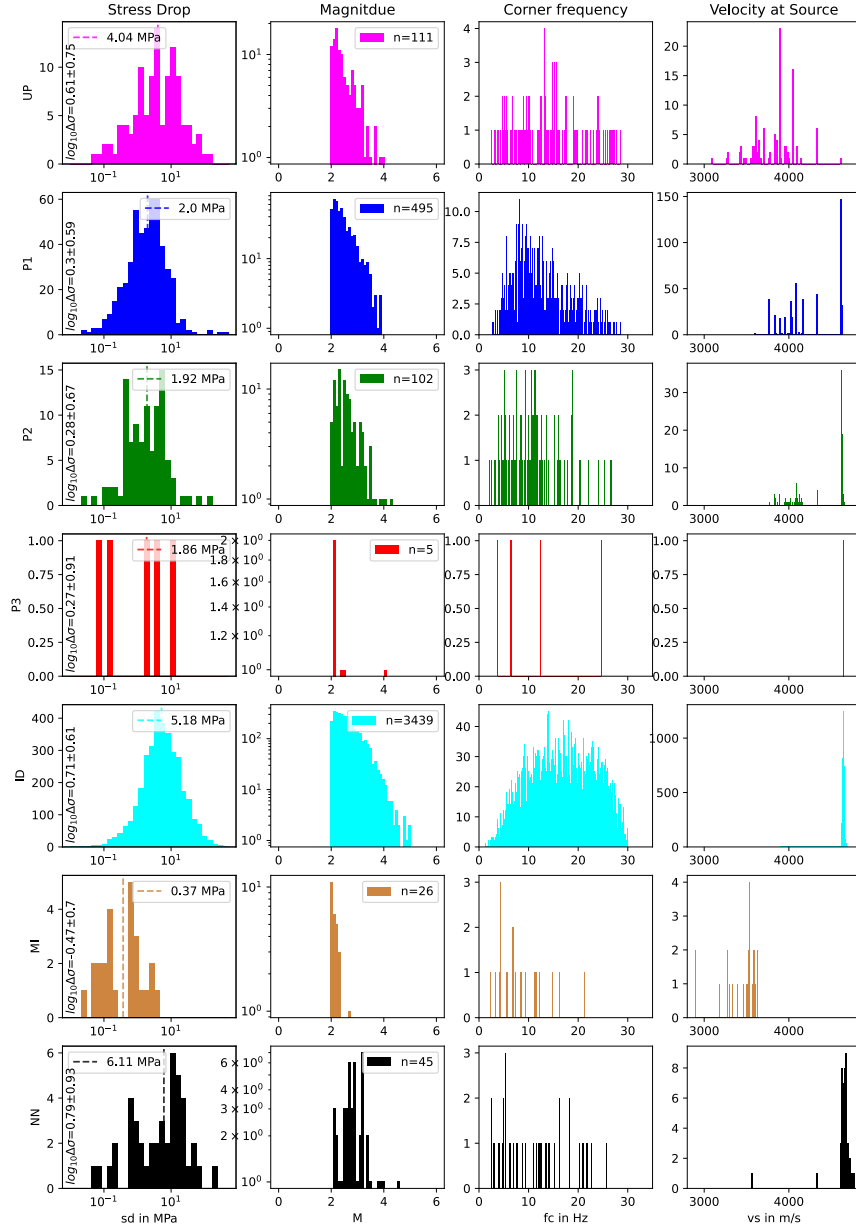


Figure S 12: Stress drop, Magnitudes, corner frequencies and S wave velocities for each of the event classes UP, P1, P2, P3, ID, MI, and NN obtained by application of the spectral ratio method, corresponding to Figure S 13. The distributions appear similar those from Figure 6 in the main manuscript. The relative differences between classes also appear roughly similar to that of the spectral decomposition results, e.g., the P₃ class events have the smallest (except for mining class), UP and ID class events have significantly higher median stress drops.

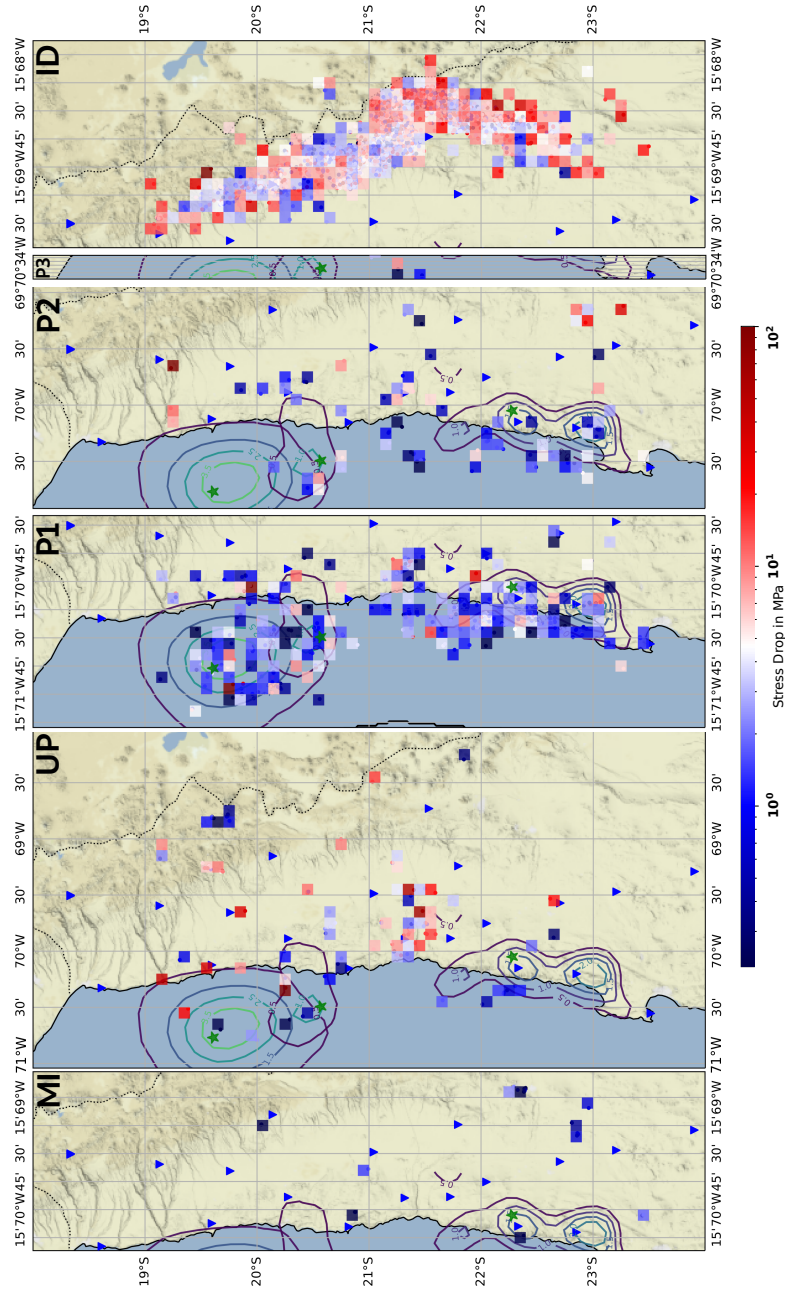


Figure S 13: Stress drop distributions in map view from spectral ratios separated into the event classes MI, UP, P1, P2, P3, ID from left to right, respectively. This figure shows the SR equivalent for Figure 7 from the main manuscript.

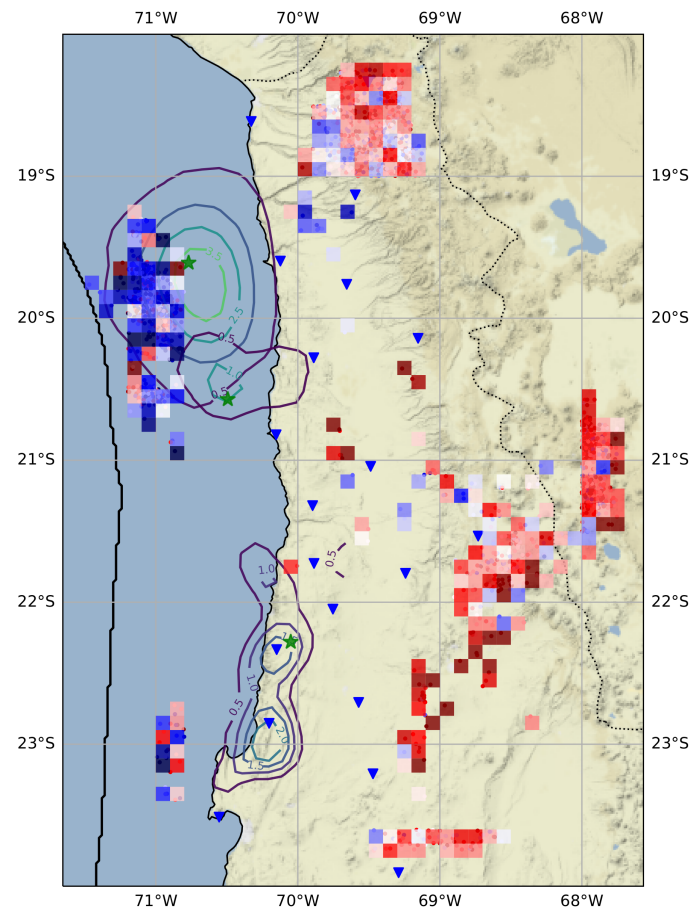


Figure S 14: Stress drop distributions in map view from SDC results for the NN class, which is missing in Figure 7 in the main manuscript.

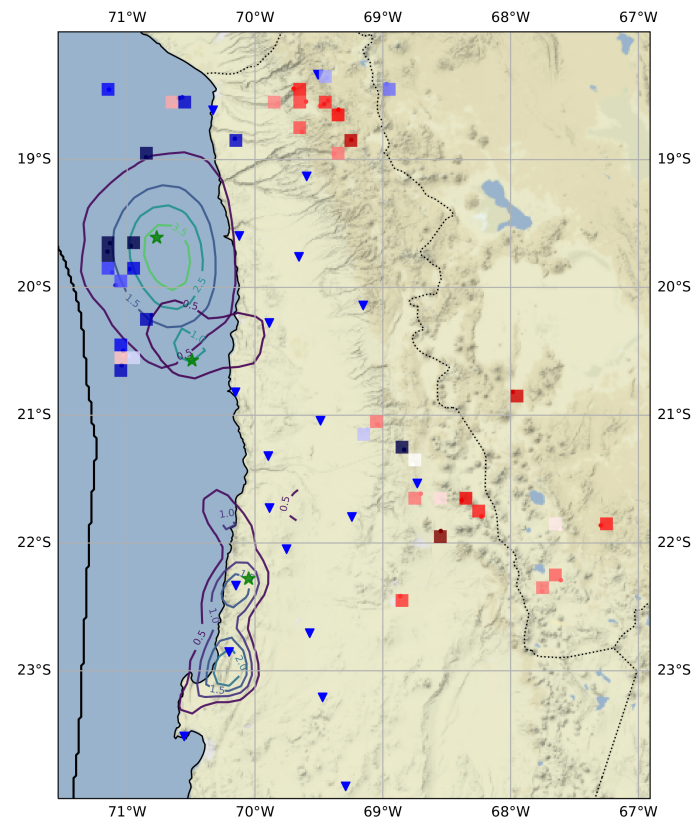


Figure S 15: Stress drop distributions in map view from SR results for the NN class, similar to Figure 14.

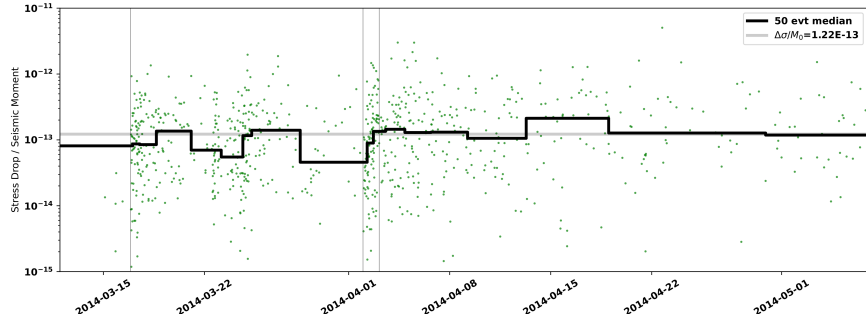


Figure S 16: Five weeks of stress drop estimates scaled by moment in the vicinity of the 2014 MW8.1 Iquique event. Overlain is the 50 event median. The figure indicates that the increase of stress drop close to the origin times of the large events (vertical lines) observed in Figure 8, main manuscript, originates predominately by increased moment rates during those periods.

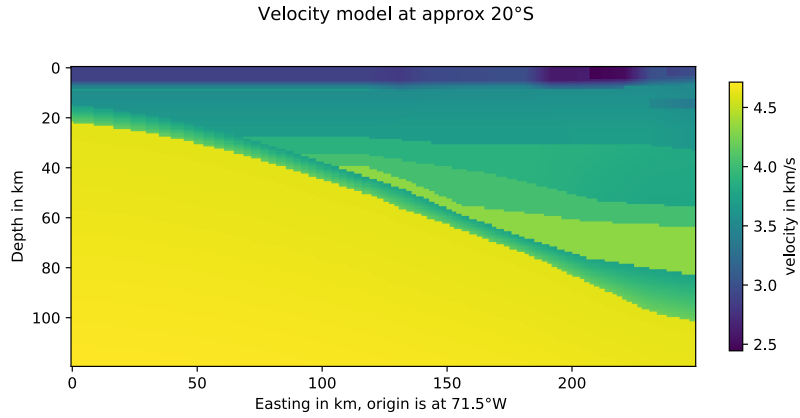


Figure S 17: S phase velocity model slice at 20°S from Bloch et al. 2014.