

Tectonic inheritance from deformation fabric in the brittle and ductile Southern California crust

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Abstract

Plate motions in Southern California have undergone a transition from compressional and extensional regimes to a dominantly strike-slip regime in the Miocene. Strike-slip motion is most easily accommodated on vertical faults, and major transform fault strands in the region are typically mapped as near-vertical on the surface. However, some previous work suggests these faults have a dipping or listric geometry at depth. We analyze receiver function arrivals that vary harmonically with backazimuth at all available broadband stations in the region. The results show a dominant signal from contrasts in dipping foliation as well as dipping isotropic contrasts from all crustal depths, including from the ductile middle to lower crust. We interpret these receiver function observations as a dipping fault-parallel structural fabric that is pervasive throughout the region. The strike of these structures and fabrics is parallel to that of nearby fault surface traces. We also plot microseismicity on depth profiles perpendicular to major strike-slip faults and find consistently NE-dipping lineations in seismicity shallowing in dip from near vertical (80-85) on the Elsinore Fault near the coastal ranges to 60-65 slightly further inland on the San Jacinto Fault to 50-55 on the San Andreas Fault. Taken together, the dipping features in seismicity and in rock fabric suggest that preexisting fabrics and faults likely act as strain guides in the modern slip regime, with reactivation-like mechanisms operating both above and below the brittle-ductile transition.

Tectonic inheritance from deformation fabric in the brittle and ductile southern California crust

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Key Points:

- Receiver function harmonic conversions show pervasive dipping lithospheric fabric in Southern California
- Seismicity is listric near the brittle-ductile transition on profiles across major transform faults
- Non-optimal dipping fault geometry for strike-slip motion is likely due to tectonic inheritance

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 31 ther inland on the San Jacinto Fault to $50\text{-}55^\circ$ on the San Andreas Fault. Taken together,
 32 the dipping features in seismicity and in rock fabric suggest that preexisting fabrics and
 33 faults likely act as strain guides in the modern slip regime, with reactivation-like mech-
 34 anisms operating both above and below the brittle-ductile transition.

35 **1 Introduction**

36 The San Andreas Fault (SAF) is the main strand of the California plate bound-
 37 ary between the Pacific and North American plates. In central California, strike-slip mo-
 38 tion is focused on the SAF, and observations of deep seismicity and tremor in the lower
 39 crust suggest a vertical fault structure cutting through the entire crust (Becken, Ritter,
 40 Bedrosian, & Weckmann, 2011; Johnson, Shelly, & Bradley, 2013; Nadeau & Dolenc, 2005;
 41 Ryberg, Haberland, Fuis, Ellsworth, & Shelly, 2010; Shelly, 2017; Shelly & Hardebeck,
 42 n.d.). The fault geometry in Southern California is considerably more complex, with a
 43 long restraining bend in the SAF south of which strain is partitioned onto several ma-
 44 jor strike-slip faults along with thrust and normal structures in the region. The main ad-
 45 ditional active strike-slip faults include the San Jacinto Fault (SJF) and Elsinore Fault
 46 (EF) lying between the SAF and the coast to the southwest. The subsurface geometry
 47 of strike-slip faults including the SAF is generally treated as vertical (e.g., Plesch et al.,

2007), but recent work proposed dipping or listric geometry at depth along most of the SAF (Barak, Klemperer, & Lawrence, 2015; Fuis et al., 2017; Fuis, Scheirer, Langenheim, & Kohler, 2012; Qiu, Lin, & Ben-Zion, 2019; Share, Guo, Thurber, Zhang, & Ben-Zion, 2019) and for parts of the SJF (Ross, Hauksson, & Ben-Zion, 2017), with all dips to the northeast except for a SW-dipping segment just south of the central SAF (Fuis et al., 2012).

Nonvertical geometries of strike-slip faults are of interest because ground shaking is increased on the hanging wall compared to a vertical geometry (Fuis et al., 2017; Oglesby, Archuleta, & Nielsen, 2000). Such geometries may contain information on older fault networks that may influence present day deformation behavior. They may also affect the interpretation of geodetic data (e.g., Dair & Cooke, 2009; Fialko, 2006; Lindsey & Fialko, 2013). Development of a nonvertical strike-slip fault has previously been interpreted as recording the reactivation of a prior dipping thrust or normal fault (Avouac et al., 2014; Sato, Kato, Abe, Van Horne, & Takeda, 2015). Dipping geometries on the SJF and EF have also been suggested as a consequence of inheritance from the shallowly dipping Western Salton Detachment fault (Dorsey, Axen, Peryam, & Kairouz, 2012; Mason et al., 2017). Inheritance of preexisting faults from past deformation regimes can influence the evolution of faults that accommodate strain in response to a change in plate motion, here related to subduction of an oceanic spreading ridge (Atwater & Stock, 1998). Inheritance and reactivation can also be related to older rock fabrics or rheological boundaries such as the edges of batholiths. Dip on the SAF may be due to fault localization along the boundary of a previously underthrust mafic Pacific plate under the North American plate in the region of the San Geronio to Cajon passes and southern SAF (Barak et al., 2015; Fuis et al., 2017).

An additional open question is how faults root and are connected between the brittle crust and the ductile lower crust and lithospheric mantle. It is unknown whether major continental transform faults root into localized subvertical shear zones, coalesce into wider regional-scale shear zones, or root into an evenly shared ductile substrate (Bernard & Behr, 2017; Sibson, 1983; Titus, Medaris, Wang, & Tikoff, 2007). As in the brittle case, if earlier strain led to fabric development in the ductile lithosphere, such inherited ductile fabric may lead to mechanical anisotropy and change the deformation response to applied stress. It is unknown how quickly plastic lithospheric fabric may be reset under a new stress field (Vauchez, Tommasi, & Mainprice, 2012). Ductile crustal and lithospheric

81 fabric and shear zone geometry may in turn affect how faults are loaded at brittle depths
 82 (e.g. Bourne, England, & Parsons, 1998; Dolan, Bowman, & Sammis, 2007; England &
 83 Jackson, 1989; Montési, 2004; Platt & Becker, 2010; Roy & Royden, 2000a, 2000b). Fault
 84 geometries near the brittle-ductile transition and deformation fabric at ductile depths
 85 are therefore of particular interest.

86 Our goal in this study is to investigate the geometry of faults and of shear zones
 87 and rock fabric below faults in Southern California using two types of data. The first
 88 data set consists of azimuthally varying arrivals in receiver functions that are generated
 89 from contrasts in dipping shear fabric or dipping isotropic interfaces (section 2). Pre-
 90 vious anisotropic receiver function work focused on the lower crust (Porter, Zandt, & Mc-
 91 Quarrie, 2011) and on integrated shear fabric throughout the crust and uppermost man-
 92 tle (Schulte-Pelkum & Mahan, 2014b) at a selected number of permanent stations. We
 93 present depth-dependent results throughout the lithosphere at all available permanent
 94 and temporary broadband stations. The second data set is the waveform-based, relocated
 95 Southern California catalog of relocations for 1981-2017 by Hauksson, Yang, and Shearer
 96 (2012). We plot profiles showing the fine structure of seismicity across major fault strands
 97 and near the brittle-ductile transition in section 2. We synthesize the findings from seis-
 98 micity and receiver function anisotropy in terms of tectonics, geodynamic memory, and
 99 the influence of geological history on present-day deformation via inheritance of faults
 100 and fabrics in section 4.

101 **2 Receiver function arrivals from anisotropic and dipping contrasts**

102 Previous work using receiver functions in Southern California based on a sparser
 103 subset of stations suggested dipping foliation in the lower crust (Porter et al., 2011) and
 104 in the entire lithosphere (Schulte-Pelkum & Mahan, 2014b). If present day strike-slip
 105 motion controls lithospheric fabric, one might expect vertical faults and vertical shear
 106 planes in this region rather than dipping ones. Crustal layers with vertical foliation or
 107 horizontal lineation generate receiver function arrivals with π -periodic polarity (second
 108 azimuthal harmonic, A_2) changes in backazimuth (Brownlee et al., 2017; Levin & Park,
 109 1998). In surface wave studies, such fabric is referred to as having azimuthal anisotropy
 110 (e.g. Lin, Ritzwoller, Yang, Moschetti, & Fouch, 2011). Figure 1 displays which stations
 111 in the EarthScope Transportable Array show at least one A_2 -periodic arrival with an am-
 112 plitude exceeding 10% horizontal to vertical amplitude ratio within the first 8 seconds

113 (lithospheric depths), requiring a lithospheric layer with strong vertical foliation or hor-
 114 izontal lineation (Schulte-Pelkum & Mahan, 2014b). A strong A_2 signal is plotted as a
 115 station with a cross. The cross bars denote the fast and slow orientation of anisotropy
 116 averaged over depths exceeding 10% signal amplitude, which are degenerate in this anal-
 117 ysis unless additional assumptions are made. The A_2 signal dominates in the northern
 118 Basin and Range and the coastal ranges of Central California, while it is weak (stations
 119 without crosses) in the Snake River Plain, the Colorado Plateau, the southern Basin and
 120 Range, and the Sierra Nevada (Figure 1). Since the transform plate boundary runs through
 121 Southern California into the Central California coastal ranges, it is surprising that South-
 122 ern California shows a weaker azimuthal anisotropy signal than the latter region. Pre-
 123 vious studies (Porter et al., 2011; Schulte-Pelkum & Mahan, 2014b) also suggest dom-
 124 inant dipping rather than vertical foliation in Southern California. In the present study,
 125 we analyze all available permanent and temporary stations from 2004-2017, forming a
 126 denser data set, for signatures of contrasts in dipping foliation and dipping isotropic con-
 127 trasts in receiver functions.

128 **2.1 Receiver function sensitivity to anisotropic and dipping contrasts**

129 Seismic anisotropy is expressed in receiver functions in two fundamentally differ-
 130 ent ways. One is the splitting of an S arrival after conversion from P; the most commonly
 131 used is the P-to-S conversion from the Moho. We do not apply this method here, as it
 132 only provides an integrated measure of anisotropy over the travel path; it is therefore
 133 unlikely to accurately measure anisotropy from shear zones with limited thickness, and
 134 may also not provide an accurate measure of bulk crustal anisotropy because of arrivals
 135 from thin anisotropic layers masquerading as bulk volume splitting (Liu & Park, 2017).

136 A more appropriate method in the presence of intracrustal anisotropic structure
 137 uses the azimuthally varying conversion from a contrast in anisotropy (Figure 2). A con-
 138 trast in anisotropy of about 3% or a change in foliation orientation are sufficient to gen-
 139 erate large-amplitude arrivals comparable to those from an average isotropic Moho con-
 140 trast (Schulte-Pelkum & Mahan, 2014a). The conversions from such a contrast display
 141 characteristic polarity reversals with backazimuth on transverse as well as radial com-
 142 ponents (C. Jones & Phinney, 1998; Park & Levin, 2016; Savage, 1998; Schulte-Pelkum
 143 & Mahan, 2014a). The depth of such a contrast is given by the arrival time of the con-
 144 version in the receiver function (Figure 2), giving this method depth resolution, unlike

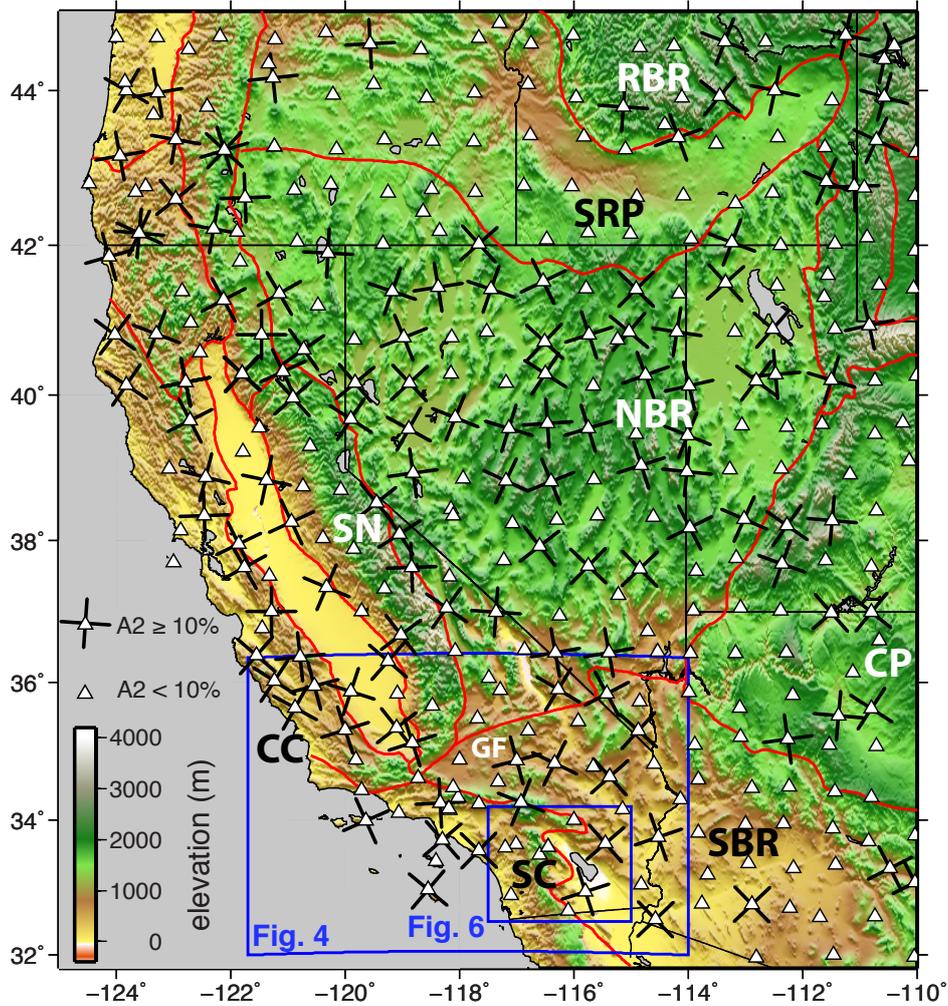


Figure 1. Map of stations in the EarthScope Transportable Array (white triangles) where the A_2 -periodic azimuthal signal from anisotropy exceeds (black bars) or does not exceed (triangle without bars) 10% of the receiver function amplitude. Black bars are fast and slow orientations averaged axially over the parts of the signal exceeding 10%. Red lines are physiographic province boundaries. Blue lines show extents of later figures. RBR-Rocky Mountain Basin and Range; SRP-Snake River Plain; NBR-Northern Basin and Range; SN-Sierra Nevada; CP-Colorado Plateau; CC-Central California; GF-Garlock Fault; SBR-Southern Basin and Range; SC-Southern California. Analysis details in Schulte-Pelkum and Mahan (2014b).

145 splitting methods. Thin shear layers can also be resolved as long as conversions from the
 146 top and bottom of the layer are separated by close to a pulse width of the receiver func-
 147 tion (usually 2-3 km thickness for receiver functions calculated at 1 Hz or higher). In

148 many previous studies, arrivals are modeled directly to arrive at a usually nonunique model
149 of anisotropy (Ozacar & Zandt, 2009; Porter et al., 2011; Vergne, Wittlinger, Farra, &
150 Su, 2003; Zandt et al., 2004). However, the harmonic behavior can be exploited directly
151 (without forward modeling) to systematically separate signal from azimuthal anisotropy
152 (horizontal symmetry axis) in the π -periodic amplitude component (A_2) and dipping fo-
153 liation (plunging symmetry axis) in the 2π -symmetric component (A_1 ; Bianchi, Park,
154 Agostinetti, & Levin, 2010; Bianchi, Piana Agostinetti, De Gori, & Chiarabba, 2008; Brown-
155 lee et al., 2017; Liu & Park, 2017; Long, Ford, Abrahams, & Wirth, 2017; Schulte-Pelkum
156 & Mahan, 2014b). Some studies focus on azimuthal anisotropy using the A_2 component
157 because of an argument that the A_1 component is also sensitive to dipping isotropic con-
158 trasts (Savage, 1998). However, the sensitivity to plunging axis anisotropy is much higher
159 than for horizontal axis anisotropy (Park & Levin, 2016; Schulte-Pelkum & Mahan, 2014b).
160 The method used here solves for strike of a dipping isotropic contrast or strike of the plane
161 perpendicular to the symmetry axis of anisotropy, whether that is a slow or fast axis (Fig-
162 ure 2). This strike and the depth of the contrast in velocity or anisotropy are the most
163 robust observables (Schulte-Pelkum & Mahan, 2014a, 2014b). The sense of dip trades
164 off with whether the anisotropy is best approximated by a fast or slow symmetry axis,
165 and if the stronger anisotropy is above or below the interface (Figure 2). The amplitude
166 of the A_1 arrival depends on the strength of the anisotropic or isotropic contrast as well
167 as on the plunge of the symmetry axis or the isotropic contrast dip, with intermediate
168 foliation dips generating larger conversion amplitudes than subhorizontal or subvertical
169 foliation in the anisotropic case (Brownlee et al., 2017).

170 Figure 2 illustrates how strikes of isotropic dip and of shear fabric foliation result-
171 ing from geological deformation processes are similar in orientation, such that the dis-
172 tinction between isotropic dip and dipping foliation contrast becomes a secondary ques-
173 tion. A distinction between a contrast plunging axis anisotropy at depth versus an isotropic
174 dipping interface can be made by the fact that a polarity-reversed arrival is seen at zero
175 time in the isotropic dip case (Schulte-Pelkum & Mahan, 2014b).

176 **2.2 Data and analysis**

177 We analyze broadband data from stations of the USArray, Southern California Seis-
178 mic Network (SCSN), and Anza networks for the years 2004-2017, as well as from the
179 1997-1998 LARSE II experiment (Murphy et al., 2010; Zhu, 2002) and 2011 Salton Seis-

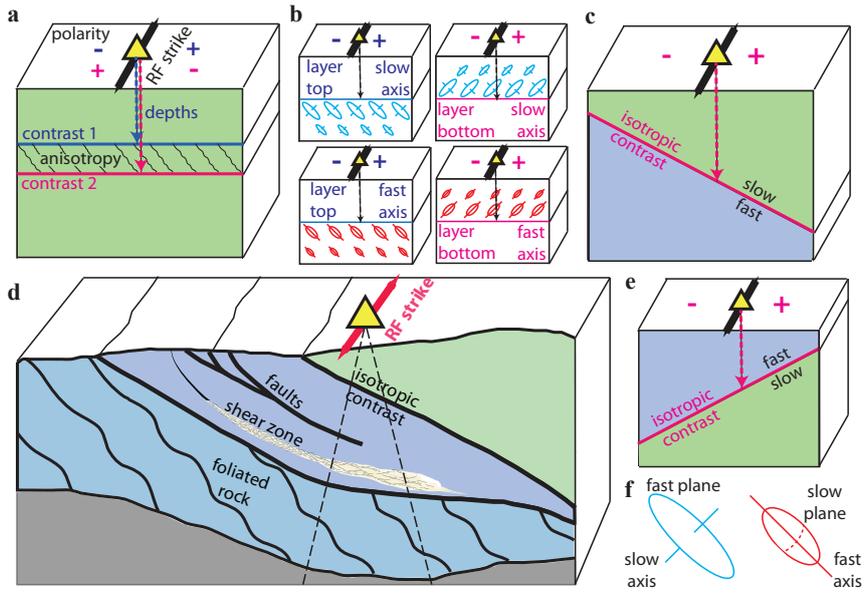


Figure 2. Illustration of sensitivity of receiver function to deformation-related structures in the lithosphere. (a) Horizontal layer with anisotropy (dipping foliation with fast V_p planes shown as wavy lines) embedded in isotropic layer with same average velocity (no isotropic velocity contrast). Receiver functions from different backazimuths at station (yellow triangle) show two arrivals with opposite polarity (+ and - signs) indicating foliation strike shown as black bar, one at a delay time corresponding to the depth of the blue contrast (blue dashed arrow), another from depth of the red contrast (red dashed arrow). (b) Illustration of tradeoffs between sense of foliation dip, anisotropic symmetry (fast- or slow-axis symmetry approximation), and whether an arrival is generated at the top or bottom of a layer with stronger anisotropy, with symbols explained in f. All four cases generate the same polarity reversal pattern. (c) Dipping interface with contrast in isotropic V_s ; black bar is strike registered at station, depth of red interface below station is found via delay time as in the anisotropic case. (d) Conceptual tectonic sketch after Fossen and Cavalcante (2017) and Harms et al. (2004) showing a composite of cases (a) and (b) within a compressional setting: Isotropic contrasts across faults (black lines), localized shear zone (beige), and distributed shear fabric (purple) with strike of these features picked up by receiver functions at station (yellow) from several depths; dashed lines show receiver function sampling cone. (e) Illustration of tradeoff in dip sense in the isotropic contrast case; this case generates the same polarity pattern as case c. (f) Explanation of symbols used in b.

180 mic Imaging Project (Barak et al., 2015). In order to maximize azimuthal coverage, we
 181 calculate receiver functions for P and P_{diff} arrivals from $30\text{-}150^\circ$ epicentral distance for

182 all events with magnitude 5.0 and larger and apply an automated selection and quality
 183 control process (Schulte-Pelkum & Mahan, 2014b). From all events, we retain a mean
 184 of 6% of radial and 3% of the transverse component receiver functions with high signal-
 185 to-noise ratios and stable deconvolutions. We use the time domain algorithm by Ligor-
 186 ria and Ammon (1999) and a Gaussian filter factor of 3. Details of the processing and
 187 quality control methods are as in Schulte-Pelkum and Mahan (2014b).

188 The harmonic behavior is described in Schulte-Pelkum and Mahan (2014a) and the
 189 analysis method in Schulte-Pelkum and Mahan (2014b); we summarize it briefly here us-
 190 ing station LPC near Cajon Pass as an example (Figure 3). After calculating radial and
 191 transverse component receiver functions, we apply slowness corrections so that the ar-
 192 rival times correspond to vertical incidence and amplitudes to a common incidence an-
 193 gles (C. Jones & Phinney, 1998). We then stack radial and transverse components sep-
 194 arately in azimuthal 10° bins (Figure 3a and b). The radial component is a superposi-
 195 tion of the isotropic flat-layered signal (termed R_0) that has no variation with backaz-
 196 imuth plus an azimuthally varying portion. We calculate the average radial receiver func-
 197 tion over all bins as a proxy for R_0 and subtract it from each bin, so that the azimuthally
 198 varying signal remains ($R-R_0$; Figure 3c). Assuming hexagonal symmetry (usually valid
 199 for crustal materials; Brownlee et al., 2017), in the case of plunging axis anisotropy the
 200 $R-R_0$ component will be the same as the transverse component (T) shifted by 90° clock-
 201 wise in backazimuth ($\phi-90^\circ$) as shown in Figure 3d; in the horizontal symmetry axis
 202 case, the azimuthal shift is 45° (Levin & Park, 1998; Schulte-Pelkum & Mahan, 2014b).
 203 To find the plunging axis anisotropy and isotropic dip signal, we stack $R-R_0$ and $T(\phi-$
 204 $90^\circ)$ together in backazimuthal bins (Figure 3e). For all stations with azimuthal gaps
 205 smaller than 90° , we apply a moving window of 0.25 s length and solve for the first az-
 206 imuthal harmonic in amplitude for each time window to obtain an A_1 signal amplitude
 207 and phase (Figure 3e, g). The phase of an amplitude peak is perpendicular to the strike
 208 of the dipping foliation contrast or dipping isotropic interface (Schulte-Pelkum & Ma-
 209 han, 2014a, 2014b). The depth of the contrast is obtained by scaling the arrival time in
 210 the receiver function to depth using a V_p and V_s model between the converter and sta-
 211 tion.

212 In this study, we use the 3-D Southern California Earthquake Center (SCEC) Com-
 213 munity Velocity Model CFM-H (Shaw et al., 2015). The largest amplitude arrival at sta-
 214 tion LPC is at 1.6 s, with an amplitude of 0.2. Amplitudes in time-domain receiver func-

215 tions are absolute ratios of horizontal to vertical component amplitude, and 0.2-0.3 is
 216 the range of a strong Moho arrival amplitude. The strike of this arrival is 114° (green
 217 dots in Figure 3e), parallel to the strike of the San Andreas Fault beneath the station
 218 (Figure 4c). The delay time of 1.6 s results in a depth of the contrast of 12.1 km when
 219 migrated with the 3-D velocity model profile at the station location. Figure 3f shows a
 220 simplified geological cross section (Forand, Evans, Janecke, & Jacobs, 2018) based on sam-
 221 ples from the nearby Cajon Pass borehole (Figure 4c), surface geology, and shallow ac-
 222 tive source reflection results. The profile shows pervasive NE-dipping structures. A con-
 223 trast at 12 km beneath the station may be the downdip extension of a listric geometry
 224 of the San Jacinto Fault Zone or possibly connected to a range-bounding thrust fault fur-
 225 ther southwest (Figure 3f). Other arrivals are less prominent in amplitude (Figure 3g,
 226 delay time and strike marked as black circles in Figure 3e) and correspond to depths of
 227 0.5, 30.5, and 70 km, with similar strikes. We focus on the largest arrival at each sta-
 228 tion as the best determined signal for the remainder of this paper.

229 **2.3 Receiver function results**

230 The amplitude, depth, and strike orientation of inferred foliation or dip from the
 231 largest arrival at each station are shown in Figure 4 on a map of the entire study area
 232 and two magnified subregions, along with azimuthal histograms of the strikes in each map
 233 region. NW strikes dominate the set from the entire area. Given the tradeoffs illustrated
 234 in Figure 2, additional assumptions are needed to infer the dominant dip sense. If we
 235 assume the signal is from either the top of a layer with dipping foliation with slow sym-
 236 metry axis (Figure 2a), for instance a schist or gneiss layer (Brownlee et al., 2017), or
 237 a dipping slow-over-fast isotropic interface (Figure 2c), then a NW strike would imply
 238 a foliation or isotropic contrast dip down to the NE. Larger amplitudes imply a stronger
 239 foliation or isotropic velocity contrast, but can also be due to intermediate foliation dip
 240 angles which generate conversions with larger amplitudes than those from contrast with
 241 subhorizontal or near-vertical foliation (Brownlee et al., 2017). The dominant NW-SE
 242 strike mirrors that of the dominant fault orientation, in particular that of the major trans-
 243 form faults in the region. In the Transverse Ranges (Figure 4, bottom left), fault orien-
 244 tations and receiver functions strikes rotate to E-W. In this region, an assumption of ar-
 245 rivals from the top of a layer with dipping foliation or a slow-over-fast isotropic contrast
 246 implies dips to the north, consistent with the geometry of thrust faulting located in the

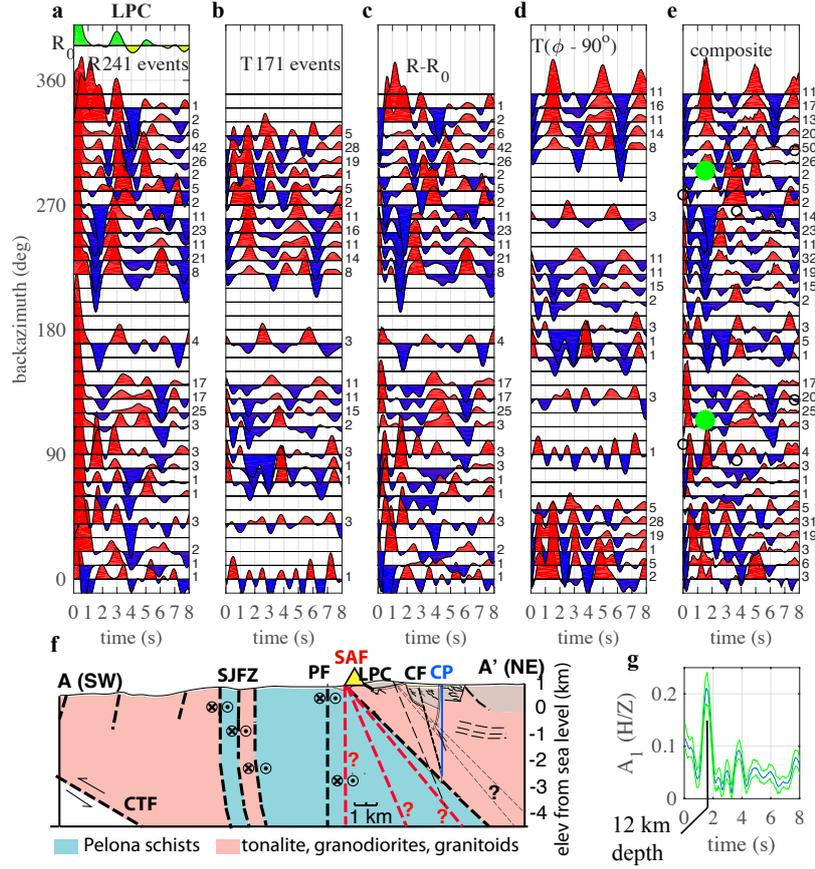


Figure 3. Receiver function analysis for example station LPC near Cajon Pass (Figure 4, bottom right). (a) Radial receiver functions binned by backazimuth after slowness correction; number of events per bin on the right of each trace. Amplitudes are on-normalized, same scale for all subplots. No azimuthal smoothing is applied. Trace on top is average over all azimuthal bins, R_0 . (b) Same as a, but transverse component. (c) Radial component after R_0 was subtracted from each bin trace. (d) Transverse component after shifting traces by 90° in backazimuth. (e) Traces from c and d stacked together. Moving window harmonic analysis to determine amplitude and phase of first azimuthal harmonic A_1 is applied to this set (A_1 amplitude shown below in g). Green dots show strike of foliation inferred from maximum arrival, which parallels the SAF trace near the station (Figure 4). Black circles mark strikes of other maxima. (f) Geological cross section simplified from Forand et al. (2018) (profile location in Figure 4), based on Cajon Pass borehole (CP) samples, surface geology, and shallow reflection work. Blue line is borehole, dashed lines are inferred faults, thin dashed lines metamorphic fabric; three alternative proposed dips were drawn by Forand et al. (2018) for SAF. CTF-Cucamonga Thrust Fault; PF-Punchbowl Fault; CF-Cleghorn Fault. (g) A_1 amplitude and 95% confidence interval on same time scale as e.

247 area. In the entire study region, the depths of the contrasts resulting in the largest A_1
248 arrival at each station are distributed throughout the lithosphere. No systematic changes
249 of strike with depth are apparent.

250 Figure 5 shows the uncertainty in A_1 amplitude and strike based on bootstrap es-
251 timates. Bootstrapping is performed by eliminating azimuthal bins in separate 100 ran-
252 dom instances before fitting the A_1 harmonic in each moving window. Azimuthal bin
253 averages rather than individual receiver functions are eliminated to avoid undue weight-
254 ing of densely sampled backazimuths. The amplitude uncertainty is given as the 95% con-
255 fidence interval, corresponding to two standard deviations if the distribution is normal.
256 The strike uncertainty is given as the standard error of the strike in the bootstrapped
257 distribution using circular statistics.

258 Strike uncertainties are small enough that they plot within the strike arrow for all
259 but a few stations (Figure 5). Amplitude uncertainties vary widely and are higher for
260 temporary stations with poorer azimuthal coverage. However, the similarity of results
261 between neighboring stations with smaller and larger nominal uncertainty, for instance
262 in the temporary more densely sampled linear experiments (e.g. the Salton Sea line in
263 the southern part of the study area), suggests that the uncertainties calculated may be
264 unduly pessimistic. We conclude that the receiver functions suggest dominant dipping
265 fabric that is pervasive geographically and throughout the lithosphere in depth, with fo-
266 liation or dip strikes largely parallel to currently active strike-slip and older inactive thrust
267 or active faults.

268 **3 Seismicity profiles**

269 We compare fault structures as illuminated by microseismicity to the dipping struc-
270 tural fabric imaged at depth using receiver functions by plotting hypocenters on depth
271 profiles perpendicular to major transform faults. We use the entire relocated seismic-
272 ity catalog of Hauksson et al. (2012), extended to the end of 2017. Figure 6 shows the
273 geographic locations of the earthquakes and of seismicity depth profiles. This analysis
274 covers the southern part of the study area that is dominated by strike-slip faulting, ex-
275 tending just into the big bend area to the north that includes thrust components Yang
276 and Hauksson (2013) for the northernmost profile A-A'. Figure 6 displays prominent ge-
277 ographic trends in the depth of seismicity. The maximum earthquake depth gradually

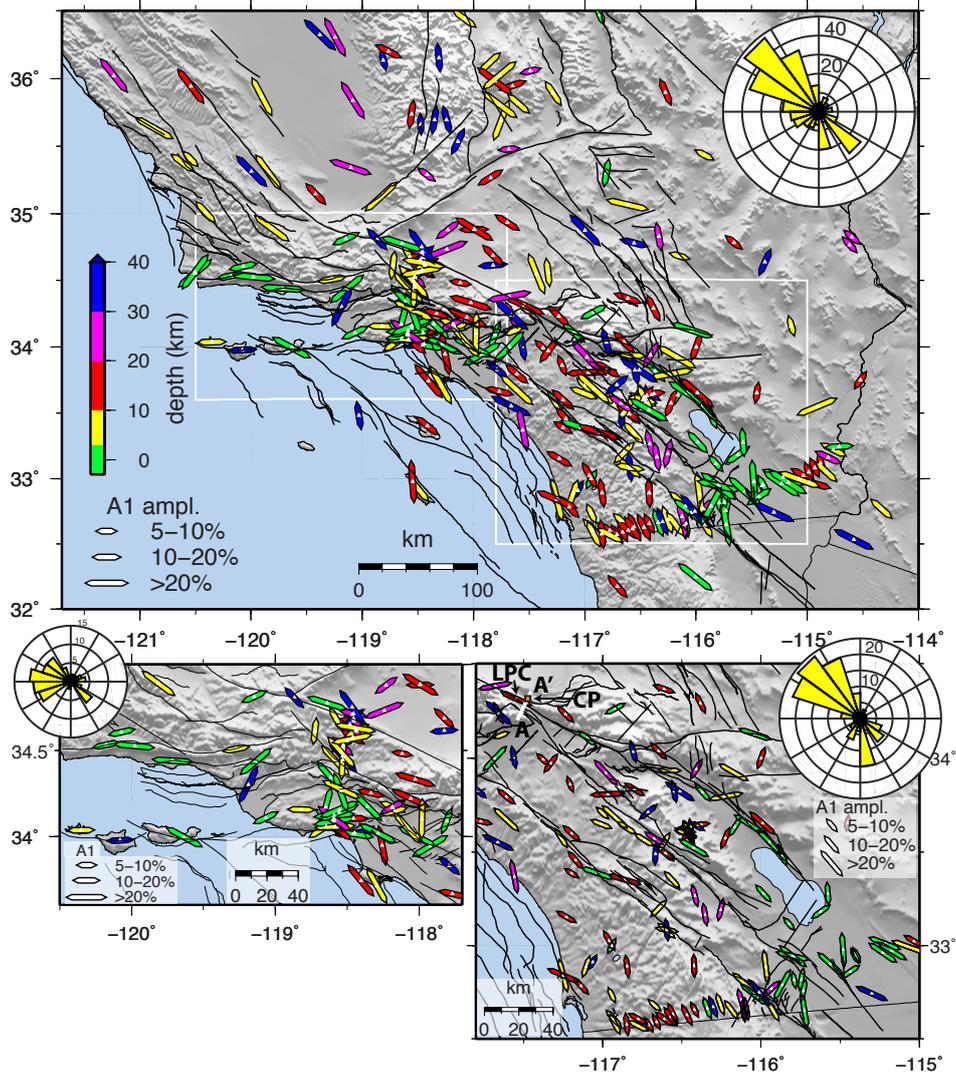


Figure 4. Map showing results for largest A_1 arrival at each station (white dot). See Figure 1 for location. Bar orientation shows the strike of dipping foliation or dipping isotropic interface, length varies with amplitude of the arrival, color indicates depth of the converting contrast. Shading is relief, faults from SCEC CFM-5.2 (Plesch et al., 2007) in black. White outlines are magnified in lower maps. Circular insets are azimuthal histograms of strikes of the bars displayed in each corresponding map (bars on map are shown as axial orientations, while histograms show sense of strike). Numbers in histograms show number of stations in each azimuthal bin. Note the predominance of NW strikes in overall area and in southern SAF - SJFZ - Elsinore Fault area (bottom right) and W strikes in the Transverse Ranges (bottom left). Strikes derived from our analysis are generally parallel those of nearby surface faults (histograms).

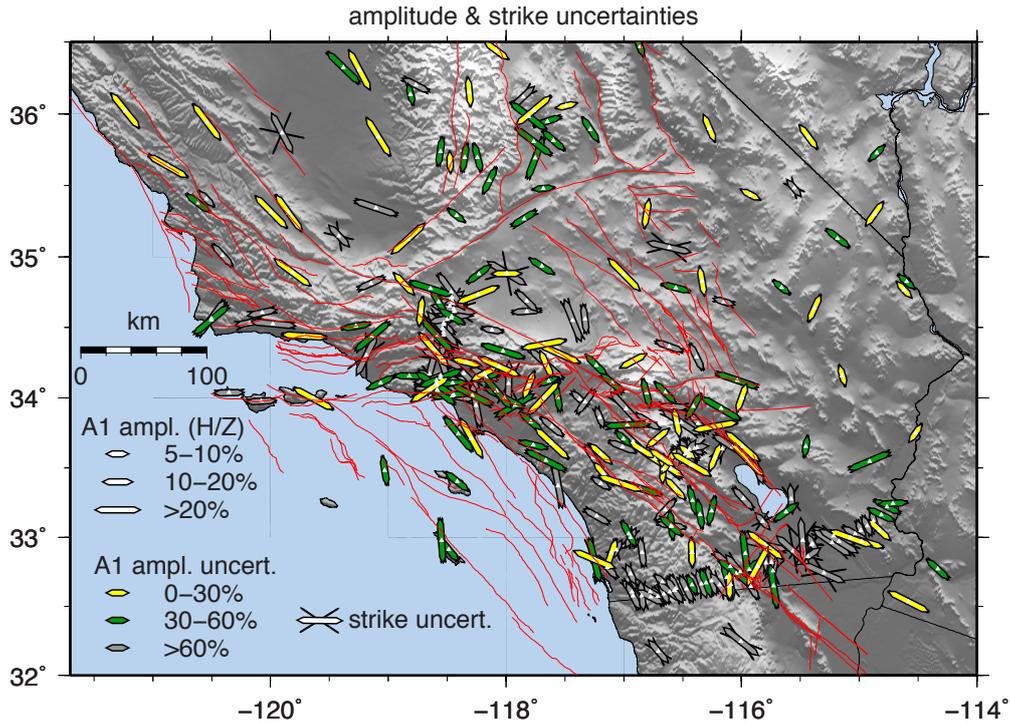


Figure 5. Same as Figure 4, except here color shows amplitude uncertainty based on bootstrap estimation, expressed as percent of the amplitude at each station. Strike uncertainty from bootstrapping is shown as thin grey lines (smaller than width of arrows for most stations).

278 shallows from 20 km at the northwestern end near the San Jacinto mountains, to 10 km
 279 at the southeastern end south of the Salton Sea (supplementary Figure S1). This pat-
 280 tern presumably outlines the brittle-ductile transition, dominantly due to geothermal gra-
 281 dient changes along the profile (Doser & Kanamori, 1986), although other influences on
 282 rheology such as composition and shear weakening are also likely (e.g. Hauksson & Meier,
 283 2019; Shinevar, Behn, Hirth, & Jagoutz, 2018).

284 To better understand the geometry of seismicity across the primary fault zones within
 285 the southern California plate boundary area, we examine a set of seismicity profiles roughly
 286 perpendicular to the San Andreas, San Jacinto, and Elsinore faults. Figure 7a shows a
 287 depth profile of seismicity along the southern SAF (SSAF) using the 1981-2017 SCSN
 288 catalog (top subpanel) and 2008-2017 QTM catalog (bottom subpanel). Relocations for
 289 the two catalogs were done independently using different methods and velocity models.
 290 The SCSN catalog (Hauksson et al., 2012) features a longer time span and captures some

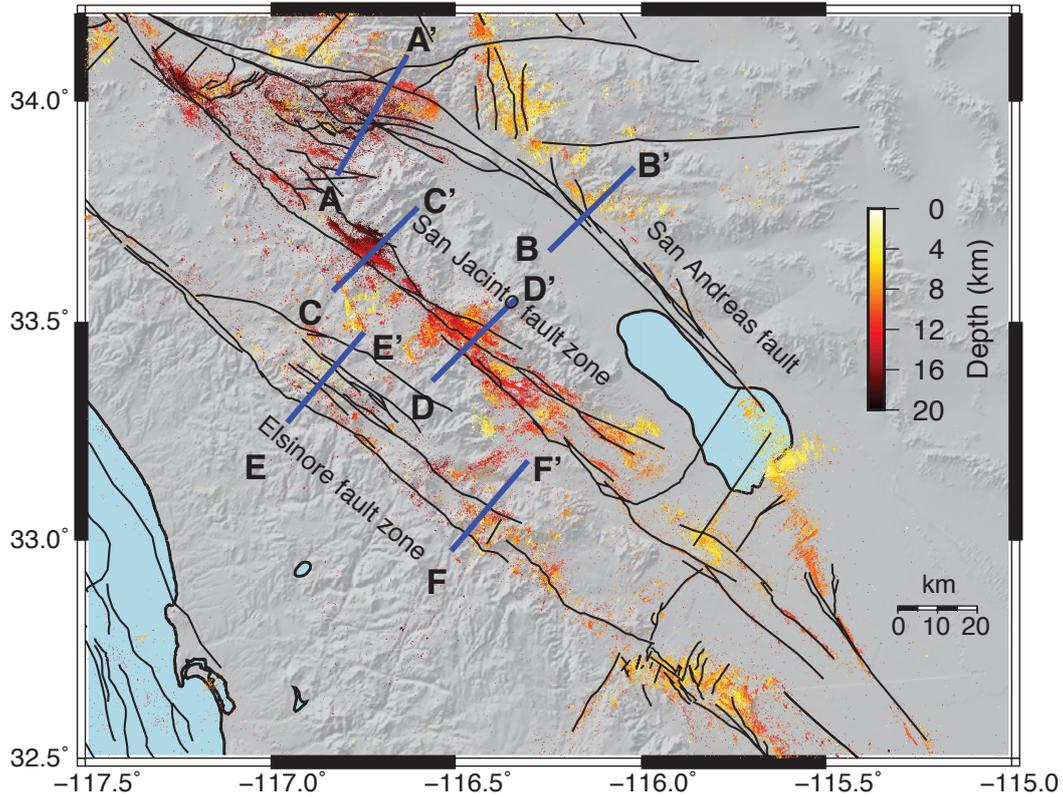


Figure 6. Map of epicenters and seismicity profile locations. Dots are epicenters in SCSN 1981-2017 catalog, color shows hypocenter depth. Relief in greyshade; black lines are fault surface traces from SCEC5 Community Fault Model (CFM, Plesch et al., 2007). Map area is marked in Figure 1 for wider regional context. Blue lines marked A-F are locations of cross-fault depth seismicity profiles in Figure 7.

291 areas that were quiescent in 2008-2017, while the QTM catalog (Ross, Trugman, Hauks-
 292 son, & Shearer, 2019) covers a shorter time span but contains magnitudes down to much
 293 smaller values (nearly complete for events of magnitude above 0.3, compared to above
 294 1.7 for the SCSN catalog). The vertical projections of the surface traces of the Banning
 295 and Mission Creek strands of the SSAF are denoted by the red dashed lines in Figure 7a.
 296 Green dashed lines show the average dip of the fault segments according to the SCEC
 297 CFM5.2 model. There is a clear northeast dipping structure to the seismicity, with dip
 298 values approximately $50\text{-}55^\circ$ over the entire length of the profile. These dip values are
 299 close to that of the 1948 M 6.5 Desert Hot Springs (Richter, Allen, & Nordquist, 1958)
 300 and 1986 M 6.0 North Palm Springs earthquake (L. Jones, Hutton, Given, & Allen, 1986),
 301 which are believed to have occurred on the Mission Creek and Banning Faults, respec-

302 tively. Fuis et al. (2017) argue based on this information and various other data sets and
303 observations that the SSAF in this area as a whole dips 50-60° to the northeast. In ad-
304 dition to these observations, there is a broad area of microseismicity interpreted as the
305 damage zone spanning about 15 km along the profile. A notable feature of this damage
306 zone is that it is asymmetric across the fault. The QTM catalog shows more fine struc-
307 ture as expected from the lower magnitude threshold, while the SCSN catalog show a
308 planar feature predating the QTM catalog in the northeastern portion of the profile.

309 Moving further to the southeast along the SSAF, Figure 7b shows a depth section
310 which also exhibits a northeast dipping structure to the seismicity. This is consistent with
311 observations of head waves propagating along a dipping bimaterial fault in the region
312 (Share & Ben-Zion, 2016). Here, the fault system has a slightly steeper dip than along
313 profile A-A', and while the seismicity is shallower overall, the damage zone here is fully
314 on the northeast side of the SSAF. A NE-dipping SAF was inferred at this location and
315 south along the Salton Sea by Barak et al. (2015) using surface wave tomography.

316 Next, we move to the San Jacinto fault zone (SJFZ), which is located to the south-
317 west of the SSAF. Figure 7c contains a seismicity profile (3) across the Clark strand of
318 the SJFZ through the Hot Springs segment. In this area, the seismicity also defines north-
319 east dipping planes, however they dip about 65-70°, compared with the dips of 50-60°
320 observed for the seismicity on the SSAF. The seismicity has a listric appearance, even
321 more so than in profile Figure 7a.

322 The dipping pattern is also apparent in the complex trifurcation area of the SJFZ,
323 which is located about 40 km to the southeast (Figure 7d). This area was studied in de-
324 tail by Ross et al. (2017), who noted that in addition to the seismicity, the focal mech-
325 anisms at depth showed planes dipping about 70° to the northeast. They further sug-
326 gested that the SJFZ transitions from dipping to near-vertical above around 8 km depth
327 based on the relative positions of the surface traces of the main faults and steeper fo-
328 cal mechanism and seismicity dips in the shallow structure. As with the SSAF, the dam-
329 age zones in these areas of the SJFZ are also predominantly located on the northeast side
330 of the surface trace of the fault. Miocene-age structures that are part of the Colorado
331 River extensional corridor (Shirvell, Stockli, Axen, & Grove, 2009) formed low-angle, east-
332 rooting normal faults. One such fault is the Western Salton Detachment (Mason et al.,
333 2017) that is offset by the SAF, SJFZ, and the Elsinore faults from north to south (Dorsey

334 et al., 2012). Mason et al. (2017) propose a transition from NE-dipping shallow normal
335 faulting to more steeply NE-dipping strike-slip motion in the Pleistocene (8 Ma). Pre-
336 existing extensional structures may therefore act as strain guides during subsequent strike-
337 slip faulting.

338 Finally, we examine two profiles across the Elsinore fault zone (EFZ). While the
339 EFZ has notably less seismicity than the other two faults, the same type of analysis is
340 still possible. Figure 7e and Figure 7f show seismicity profiles across the EFZ which both
341 indicate damage zones containing smaller cracks and faults northeast of the surface fault
342 trace. Figure 7e shows features dipping about $80\text{--}85^\circ$ to the northeast. Figure 7f shows
343 features dipping from vertical to 85° in addition to a subhorizontal to shallowly NE-dipping
344 feature at 3-4 km depth. In both profiles, the damage zone and seismicity are asymmet-
345 ric to the northeast of the surface trace of the fault. The shallowly dipping feature from
346 3-4 km depth in profile F matches the position of the Western Salton Detachment fault
347 shown by Dorsey et al. (2012). As in the case of the SJFZ, Dorsey et al. (2012) show a
348 NE dip on the EFZ consistent with synthetic normal faults in the hanging wall of the
349 top-to-the-east Western Salton Detachment system.

350 The dipping features are unlikely to be artifacts of the relocation procedure because
351 the fabric of the seismicity exhibits these patterns over several different length scales and
352 in both catalogs. These include narrow zones of activity that are relatively isolated, as
353 well as more distributed damage zones that collectively define a persistent dipping pat-
354 tern. While the relocation procedure performs a cluster analysis to identify which events
355 to group together for relocation, there is no reason that the final locations should col-
356 lapse to highly-localized seismicity structures. Neither of the relocation methods impose
357 planar features, which gives confidence to the results. The observations are furthermore
358 generally consistent with regional focal mechanisms.

359 To summarize these results, all three of the major transform fault systems in south-
360 ern California exhibit narrow planar features of seismicity that dip to the northeast. The
361 dip of each system progressively steepens from about 50° in the northeast to about 80°
362 in the southwest. This steepening may also occur toward the southeast, but the evidence
363 for this is weaker from the seismicity. In addition, the damage zones which produced these
364 earthquakes are all strongly asymmetric across each fault system, with most of the earth-
365 quakes occurring on the northeast side. Together, the common patterns for all suggest

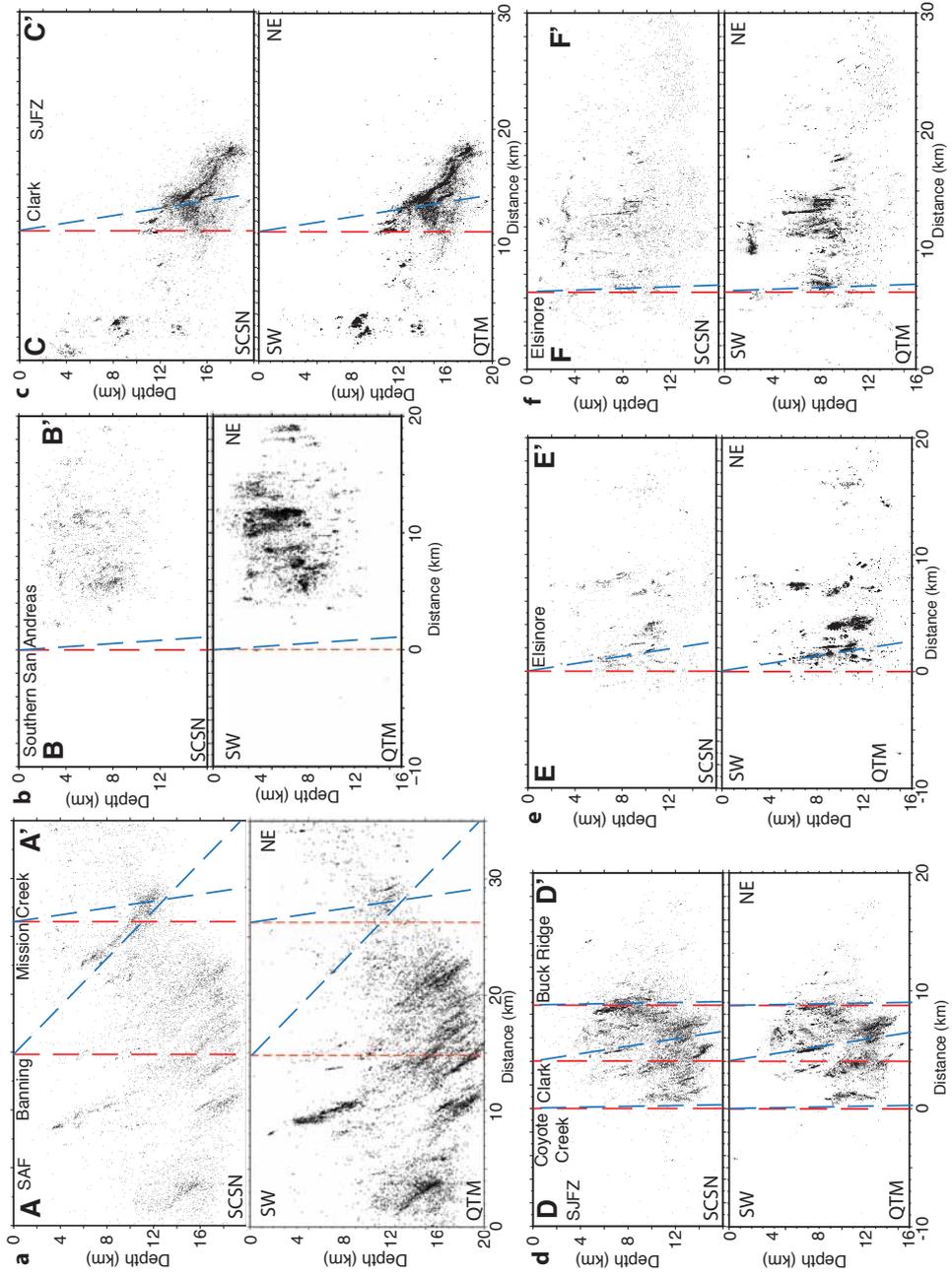


Figure 7. Seismicity profiles, locations marked in Figure 6. Each top subpanel shows locations from the 1981-2017 SCSN catalog, bottom from 2008-2017 QTM catalog. Red dashed lines in profiles show vertical projection of fault surface traces, blue dashed lines show average dips of each fault segment as listed in the SCEC CFM-5 model. There is no vertical exaggeration in the profiles so that any dip angles are true dips. Seismicity shown is from within 5 km of each respective profile.

366 that the observations are related to a regional pattern. Preexisting NE-dipping exten-
367 sional structures may have influenced the development of northeast dip on strike-slip faults
368 (Dorsey et al., 2012; Mason et al., 2017; Shirvell et al., 2009).

369 **4 Discussion: The importance of fault and fabric reactivation and in-** 370 **heritance**

371 The current tectonic regime in Southern California is one of transform motion on
372 the plate boundary and regional N-S compressive stress (Heidbach, Rajabi, Reiter, Ziegler,
373 & WSM Team, 2016; Kreemer, Blewitt, & Klein, 2014; Yang & Hauksson, 2013). There
374 is thrust faulting in the Western Transverse Ranges (Figure 4, bottom left) past the area
375 of profile A to Cajon Pass (Figure 6. In other areas, strike-slip faulting dominates (Yang
376 & Hauksson, 2013). The maximum compressive stress in the crust is generally N-S with
377 some regional variations and rotations with depth (M.-G. P. Abolfathian N. & Ben-Zion,
378 2020; N. Abolfathian, Martinez-Garzon, & Ben-Zion, 2019). This suggests that micro-
379 cracks in the brittle upper crust (above crack closure depths) would be expected to gen-
380 erally align N-S, parallel to the maximum compressive stress. Li and Peng (2017) anal-
381 ysed shear wave splitting from local events within the seismogenic portion in the crust
382 in southern California. They found large deviations between the orientation of maximum
383 compressive stress and the observed fast polarization orientation, with the latter rotat-
384 ing to fault-parallel along large parts of the San Andreas Fault, regions between the SJFZ
385 and Elsinore Faults, the Western Transverse Ranges, and other areas. Li and Peng (2017)
386 concluded that the signal from present-day compression was affected by older structural
387 features generating anisotropy.

388 Our results extend their observations in two aspects. First, the fault-parallel anisotropy
389 is not limited to the seismogenic crust, but extends through all lithospheric depths (Fig-
390 ure 4). Like the shallow results from Li and Peng (2017), the fault-parallel foliation is
391 not limited to the vicinity of major faults but appears pervasive through the entire re-
392 gion. Second, rather than finding vertical foliation or horizontal lineation (i.e., an A_2 sig-
393 nal) and vertical fault traces as one may expect with Andersonian mechanics and a ho-
394 mogeneous crust under dominant transform deformation, our results show dominant dip-
395 ping fabric and faults throughout the region. The strikes of the dominant fabric vary along
396 with the fault strike by geological block. We interpret this as a regional fabric that is

397 not controlled by a present-day region-wide deformation regime, but is dominated by in-
398 herited fabric stemming from the tectonic history of each block and of the entire region.

399 Much like the small-scale examples of inherited fabric and reactivation discussed
400 by Dorsey et al. (2012) and Mason et al. (2017) in the case of the Elsinore and San Jac-
401 into Faults influenced by the Western Salton Detachment Fault, preexisting fabric, fault
402 structures, and shear zones from previous compressional and extensional regimes likely
403 have had a controlling influence on the development and geometry of present-day trans-
404 form faults across the region. Exhumed shear zones in the area show evidence for reac-
405 tivation of Mesozoic thrusts in Cenozoic normal faulting (Goodwin & Wenk, 1995; Todd,
406 Erskine, & Morton, 1988). Langenheim, Jachens, Morton, Kistler, and Matti (2004) pro-
407 pose that the geometry and path of the present-day San Jacinto Fault was dictated by
408 a preexisting physical property contrast, and that earthquakes on the fault continue to
409 nucleate along this discontinuity. Detailed seismic imaging studies show that the main
410 strike-slip faults in Southern California are associated with prominent lithology contrasts
411 (e.g. Fang et al., 2016; Qiu et al., 2019; Share & Ben-Zion, 2016, 2018; Share et al., 2019).
412 Here, we take this concept of inheritance further and propose that the southern Califor-
413 nia region is permeated by a pervasive rock fabric throughout the lithosphere. Present-
414 day faults and deformation align along strikes dictated by this inherited fabric, leading
415 to geometries such as dipping strike-slip faults that would be energetically unfavorable
416 in an isotropic homogeneous medium.

417 Figure 8 illustrates this concept with a loose tie to previous tectonic states and ages.
418 A long history of compression in the Mesozoic and Cenozoic generated NE-dipping thrust
419 fabric and faults (Figure 8, panel 1). The same structures were reactivated in normal
420 faulting during extensional episodes in the Cretaceous and Miocene (Figure 8, panel 2).
421 Miocene initiation of transform motion exploited the same NW-SW striking features and
422 fabric, leading to dipping and listric strike-slip fault structures paralleling the inherited
423 fabric (Figure 8, panel 3). We propose that present-day faulting and deformation are sig-
424 nificantly affected by inherited structure. This observation has broad implications on the
425 modeling and interpretation of present-day continental lithospheric deformation, fault-
426 ing, and fault loading. To treat these processes accurately, it may be necessary to not
427 only consider present-day stress, but also take into account preexisting fabric from past
428 deformation. The pervasive nature of such fabric in southern California implies that rather
429 than considering individual shear zones, it may be sufficient to impose a preferential fab-

430 ric (i.e. anisotropy of viscosity) when modeling deformation processes on a lithospheric
 431 scale. More sophisticated models should also explore the effect of major rheological bound-
 432 aries associated with the Peninsular Ranges batholith and the Orocopia, Pelona, and Rand
 433 schists.

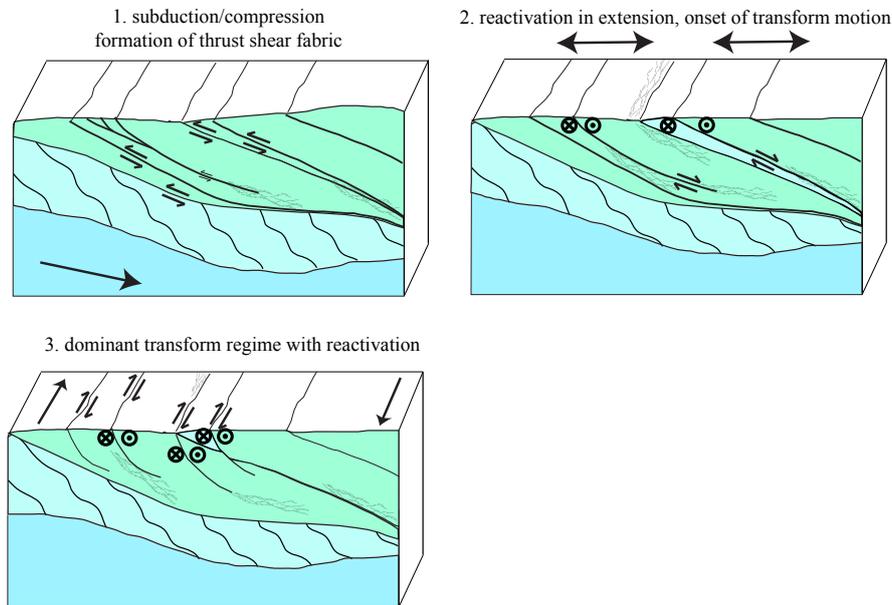


Figure 8. Conceptual illustration of the dominance of inherited structures and fabric on present-day tectonic fabric in Southern California. 1. Compressional regime during Farallon subduction (some reactivation of Mesozoic thrusts in extension). 2. Reactivation of compressional structures during extension; exhumation of some shear zones and deeper units; onset of transform deformation. 3. Dominant transform regime with dipping structures and fabric from the inherited compressional and extensional geometries. Location, strike, and dip of active structures in each regime are influenced by inherited prior structures and fabric.

434 5 Conclusions

435 We analyzed receiver functions for conversions from contrasts in dipping foliation
 436 and from dipping isotropic interfaces and found such contrasts to be pervasive through
 437 lithospheric depths and laterally pervasive in Southern California. The strike of dipping
 438 fabric and contrasts tends to parallel that of surface faults. Seismicity profiles perpen-

439 dicular to major transform faults show dipping or listric features, with dips as well as
440 damage zones preferentially to the northeast. The results are consistent with tomographic
441 and fault zone head wave imaging studies.

442 If deformation in the lithosphere was controlled by present day processes, and if
443 inherited fabric was reset, one may expect vertical strike-slip faults and deformation fab-
444 ric that is strongest in the immediate vicinity of faults. We observe pervasive fabric that
445 is not concentrated around faults and consistently dips to the northeast despite present
446 day dominant strike-slip deformation. Taken together with local geological reconstruc-
447 tions and other observations showing reactivation of inherited deformation features, this
448 leads us to propose that present-day deformation is to a large extent controlled by re-
449 gional lithospheric fabrics inherited from past deformation episodes.

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Supplementary information

This file contains a supplemental figure and its caption. Figure S1 shows a long seismicity profile running SE – NW with a change in the depth of the brittle-ductile transition.

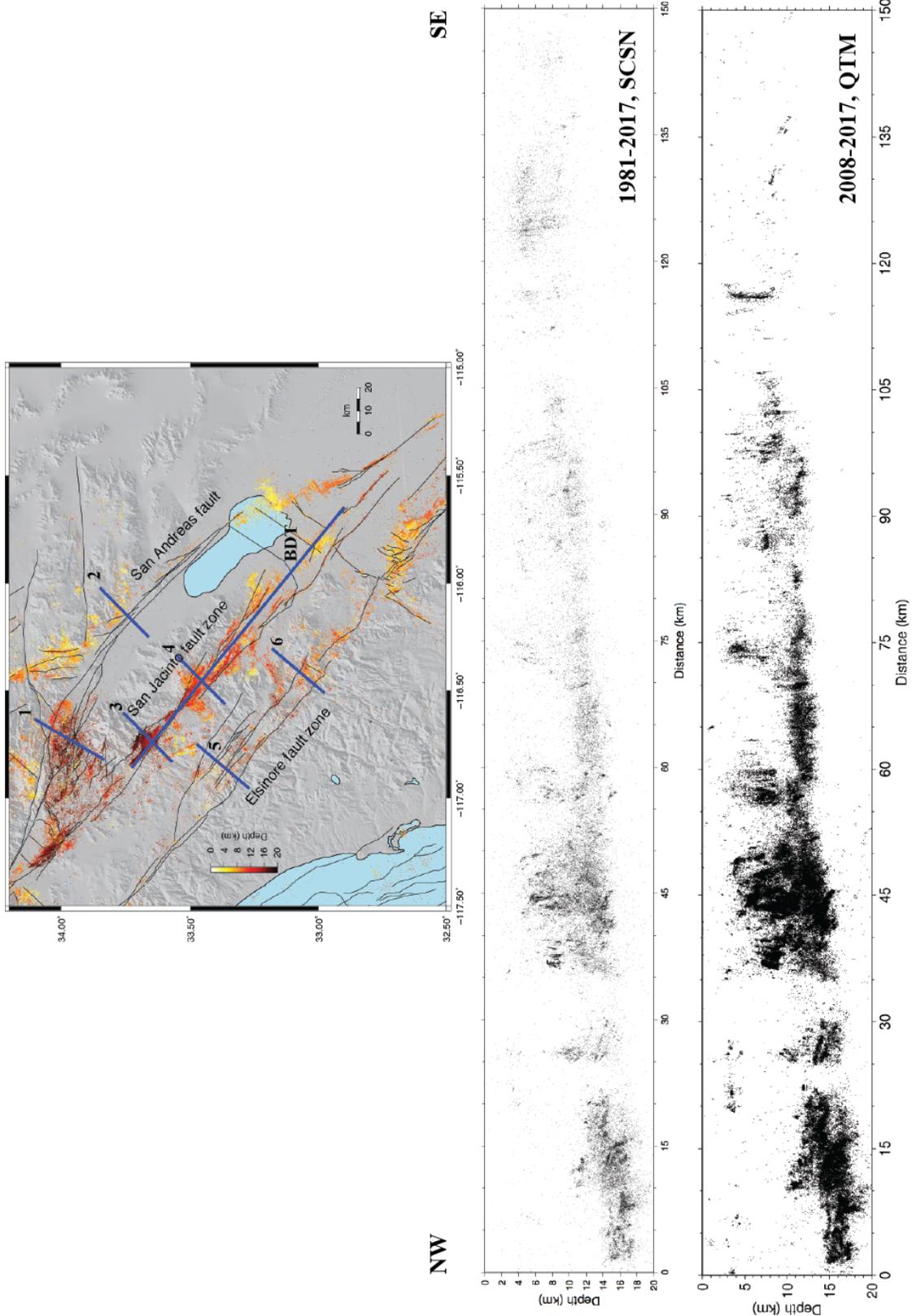


Figure S1: (Top) Map of seismicity as in main text, but with long profile named 'BDT' (brittle-ductile transition) added. (Middle) Seismicity along profile BDT from SCSN catalog (Hauksson et al., 2007 and updates). (Bottom) Seismicity along same profile from QTM catalog (Ross et al., 2019), which has a lower magnitude completeness threshold.