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Title
Detailed seismic imaging of the Mw 7.1 Ridgecrest earthquake rupture zone from data recorded by dense linear arrays

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1   **Seismic imaging of the Mw 7.1 Ridgecrest earthquake rupture zone from data**

2   **recorded by dense linear arrays**

3

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21   **Key points**:

22   • Several 1- to 2-km-wide low-velocity zones with more intensely damaged inner
23   cores (0.5-1.5 km wide) are identified beneath each array

24   • An automated detector, based on peak ground velocities and durations of high
25   amplitude *S* waves, identifies fault-zone trapped waves

26   • The *P* wave delay time and *S* wave amplification patterns indicate consistent
27   locations and widths of fault damage zones

28   **Abstract**

29   We analyze seismograms recorded by four arrays (B1-B4) with 100-m station spacing

30   and apertures of 4-8 km that cross the surface rupture of the 2019 Mw 7.1 Ridgecrest

31   earthquake. The arrays extend from B1 in the northwest to B4 in the southeast of the

32   surface rupture. Delay times between *P*-wave arrivals associated with ~1200 local

33   earthquakes and four teleseismic events are used to estimate local velocity variations

34   beneath the arrays. Both teleseismic and local *P* waves travel faster on the northeast than

35   the southwest side of the fault beneath arrays B1 and B4, but the velocity contrast is less

36   reliably resolved at arrays B2 and B3. We identify several 1- to 2-km-wide low-velocity

37   zones with much slower inner cores that amplify *S* waveforms, inferred as damage zones,

38   beneath each array. The damage zones at arrays B2 and B4 also generate fault-zone head

39   and trapped waves. An automated detector, based on peak ground velocities and

40   durations of high-amplitude waves, identifies candidate fault-zone trapped waves

41   (FZTWs) in a localized zone for ~600 earthquakes at array B4. Synthetic waveform

42   modeling of averaged FZTWs, generated by ~30 events with high-quality signals,

43   indicates that the trapping structure at array B4 has a width of ~300 m, depth of 3-5 km,

44   *S*-wave velocity reduction of ~20% with respect to the surrounding rock, *Q*-value of

45   ~30, and *S*-wave velocity contrast of ~4% across the fault (faster on the northeast side).

46   The results show complex fault-zone internal structures (velocity contrasts and low-

47   velocity zones) that vary along fault strike.

48

49   **Plain Language Summary**

50   The 2019 Mw 7.1 Ridgecrest earthquake in the Eastern California Shear Zone generated

51   a vigorous aftershock sequence that provided a wealth of seismic data. We derive

52   subsurface structural properties within and across the Ridgecrest rupture zone from

53   seismic waveforms generated by the earthquake sequence. The data are recorded by four

54   dense nodal arrays that were deployed across the Ridgecrest rupture zone with ~100 m

55   spacing and aperture of a few kilometers. Delay times of *P* wave and amplification of *S*

56   waves are used to infer on several 1-2-km-wide low-velocity zones with more intensely

57   damaged inner cores (0.5-1.5 km wide) beneath each array. Waveform modeling of fault

58 zone trapped waves well-recorded by one array provides geometrical and seismic

59 properties of a coherent waveguide in the damage fault zone structure at that location.

60 The results are complementary to tomographic models that provide a regional context but

61 do not resolve internal structural elements of the Ridgecrest rupture zone.

62

1. Introduction

64   The Mw 7.1 Ridgecrest earthquake of July 5, 2019 and the earlier Mw 6.4 event on

65 July 4 in the southern part of the Walker Lane shear zone (Fig. 1) were felt throughout

66 southern California and produced a vigorous aftershock sequence. These events led to

67 rapid deployments of seismic arrays across and around the Ridgecrest earthquake

68 sequence (Catchings et al., 2020). Kinematic rupture processes of the Mw 6.4 and Mw

69 7.1 events, surface deformation, and properties of the aftershocks show complex patterns,

70 with strong variations both along strike of the rupture zones and in depth (e.g., Chen et

71 al., 2020; Cheng & Ben-Zion, 2020; Jia et al., 2020; Ross et al., 2019; Xu et al., 2020).

72 Data recorded by several dense arrays crossing the rupture zone of the Mw 7.1

73 earthquake can be used to derive high-resolution seismic information on the internal

74 structure of the rupture zone. Detailed imaging of the structure associated with the

75 rupture zone can provide important information on various topics, including initiation and

76 arrest of ruptures (e.g., Aki, 1979; King, 1986), amplification of seismic waves (e.g.,

77 Kurzon et al., 2014; Rovelli et al. 2002; Spudich & Olsen, 2001), interactions of ruptures

78 with fault zone properties (e.g., Ben-Zion & Huang 2002; Brietzke & Ben-Zion 2006;

79 Huang et al., 2014), and properties of earthquake sequences (e.g., Thakur et al., 2020).

80   Several velocity models for the Ridgecrest area provide information for volumes with

81 spatial resolutions ranging from several km (e.g., Lee et al., 2014; Shaw et al., 2015) to

82 about 500 m (White et al. 2020). Structures in the top 1-2 km are poorly resolved in these

83 velocity models due to limitations of the input data. Analyses of seismic data recorded by

84 arrays across faults and rupture zones have proven highly effective in complementing

85 regional velocity models and imaging sharp bimaterial interfaces and damage zones with

86 width of a few tens of meters (e.g., Ben-Zion et al., 2003; Cochran et al., 2009; Li et al.,

87 1994; Peng et al., 2003; Qin et al., 2018; Qiu et al., 2017; Share et al., 2019).

88   In this study, we investigate the seismic and geometrical properties of the damage

89   structure associated with the 2019 Mw 7.1 Ridgecrest earthquake, based on the data

90   obtained from four dense linear seismic arrays (B1-B4; triangles in Figs. 1 and 2) located

91   across segments of the rupture. Analyses of the arrival patterns of *P* waves from both

92   teleseismic and local seismic events across each array helps to detect and constrain

93   properties of velocity contrasts across fault sections and low-velocity zones reflecting at

94   least partially damaged rock. We identified fault-zone trapped waves, i.e., amplified

95   motions of *S* waves associated with core damage zones that are sufficiently coherent to

96   act as a waveguide, at some locations and inverted for average geometrical and seismic

97   properties of the fault-zone waveguide.

98   In the following sections, we describe the deployment and data processing in section

99   2 and present the methodology and results on various aspects of the fault-zone structures

100   from different types of observations in section 3. The imaging results from different

101   phases and analyses are summarized and discussed in section 4. The results show overall

102   complex fault-zone structures that vary along the rupture strike, in general agreement

103   with fault surface traces (Xu et al., 2020), seismic catalog (Ross et al., 2019) and potency

104   of aftershocks (Cheng & Ben-Zion, 2020) in the Ridgecrest area.

105

2. Data & basic processing

106

107   Four linear arrays, with about 100-m station spacing and apertures of 4-8 km, were

108   deployed across the surface rupture of the 2019 Mw 7.1 Ridgecrest earthquake (Fig. 1).

109   The arrays extended from B1 in the northwest to B4 in the southeast of the surface

110   rupture (Fig. 2). In total, the B-arrays consisted of 248 Fairfield and SmartSolo sensors

111   that recorded continuously at 500 Hz for about a one-month period (7/12/2019-8/8/2019).

112   For teleseismic delay time analysis (Section 3.1), we use the Taup toolkit (Crotwell et

113   al., 1999) and velocity model IASP91 (Kennett & Engdahl, 1991) for predictions of *P*-

114   arrival time at each station. The teleseismic earthquakes have epicentral distances

115   between 30-90°, depth > 50 km, and Mw > 6.0. For analysis of local *P* waves (Section

116   3.2), we first extract the seismic waveforms generated by ~1200 local events (red box in

117   Fig. 1) at each station and use the catalog of Hauksson et al. (2012, extended to 2019) for

118   locations. The mean and linear trend are removed from the waveforms, and a bandpass

119   filter between 0.5 Hz and 20 Hz is applied. In the study of fault zone trapped waves

120   (Section 3.3), the north-south and east-west components are rotated to a coordinate

121   system parallel and perpendicular to the fault strike.

122

3. Analysis
123

124   We conduct three types of studies involving different signals and spatial scales to

125   image several components of the fault-zone structure associated with the 2019 Mw 7.1

126   Ridgecrest earthquake beneath the four linear arrays (Fig. 2). We describe the analyses

127   below, starting with large-scale structural features (e.g., overall velocity variations across

128   the fault) and progressing to inner fault-zone components (e.g., geometry and velocity of

129   the damage zone). The results are obtained using teleseismic delay-time analyses (DTA),

130   local P-wave DTA, and analysis associated with FZTWs following the S-wave arrival.

131

132   3.1 Teleseismic delay time analysis

133   During the one-month deployment, teleseismic P waves with sufficient signal to noise

134   ratios (SNR > 5) between 0.5 Hz and 2 Hz were recorded for three events at array B1 and

135   four earthquakes at arrays B2-B4 (Figs. S1- S4). We do not investigate teleseismic S

136   waves since they have SNR < 5.

137

138   *3.1.1 Methodology*

139   As shown in previous studies (e.g., Ozakin et al., 2012; Qiu et al., 2017), there are

140   three contributing factors to travel-time delays observed on a linear array for a

141   teleseismic arrival: the geometry between the incoming plane wave and the array,

142   topography, and the crustal structure beneath the array. To obtain the travel-time delays

143   due to local crustal structures, we first predict the arrival time of the teleseismic P-wave

144   for each station and event pair using the IASP91 model and assume the station is at sea

145   level. Then, teleseismic P waveforms are bandpass filtered between 0.5 Hz and 2 Hz. By

146   aligning the teleseismic P waves with respect to the corresponding predicted arrival time

147   (Fig. 3b) at each station, we remove the delay times associated with the non-vertical

148 incidence angle of incoming waves. This assumes a flat Moho interface beneath the

149 array, which is likely given the small area involved and receiver function results (e.g.,

150 Zhu & Kanamori, 2000).

151   In Qiu et al. (2017), the *P* waveform is first stacked over the entire array for a specific

152 teleseismic event and used as the reference. Then, the arrival-time pattern of *P* waves is

153 extracted from cross-correlations of the *P* waveform recorded at each station and the

154 reference waveform. However, this method is accurate only for arrays with short aperture

155 (e.g., ~500 m in Qiu et al., 2017) when *P*-wave delay times are small (e.g., ~0.01 s). For

156 arrays with long aperture, the *P* waveform recorded at a specific station may be used as

157 the reference. Considering the long aperture of the B-arrays (4-8 km; Fig. 2), we estimate

158 the *P*-wave delay time pattern by cross-correlating waveforms within a narrow *P*-wave

159 window (Fig. 3b) for every pair of stations *i* and *j*. The center of the narrow *P*-wave

160 window is determined based on the array-mean envelope function (Fig. 3b), and the peak

161 frequency of the array-mean *P*-wave amplitude spectrum (Fig. 3c) is used to set the

162 window width to be twice the dominant period. To further enhance the *P*-wave signals,

163 we apply another filter with a narrower frequency band (black dashed lines in Fig. 3c) to

164 the teleseismic data prior to the cross-correlation.

165   Let $\tilde{t}_{ij}$ be the time delay corresponding to where the cross-correlation function

166 between *P* waveforms recorded at the *i*- and *j*-th stations reaches the maximum,

$$\tilde{t}_{ij} = \tilde{T}_i - \tilde{T}_j, \tag{1a}$$

167 where $\tilde{T}_i$ and $\tilde{T}_j$ are the teleseismic *P*-wave travel times at the *i*- and *j*-th stations,

168 respectively. Since the mean of the arrival time pattern $\tilde{T}_i$ has no significance for our

169 imaging, we can ignore the constant $\tilde{T}_j$ in equation (1a) and the delay time pattern is

170 given by $\tilde{T}_{i\_j} = \tilde{t}_{ij}$ when the *j*-th station is fixed and set to be the reference. We note that

171 the delay time patterns obtained by using different reference stations should be

172 consistent, i.e., $\tilde{T}_{i\_j_1} - \tilde{T}_{i\_j_2}$ is a constant. Thus, we can minimize the measurement error

173 by averaging equation (1a) over *j*,

$$\tilde{T}_i = \sum_{j=1}^{N} \tilde{t}_{ij} \Big/ N + \sum_{j=1}^{N} \tilde{T}_j \Big/ N = \sum_{j=1}^{N} \tilde{t}_{ij} \Big/ N + C, \tag{1b}$$

174   where N is the number of stations and C is a constant. We again ignore the constant C

175   and remove the effect of un-modeled topography from the teleseismic P-wave delay time

$$T_i = \sum_{j=1}^{N} \tilde{t}_{ij} \Big/ N - \Delta h_i / v_{\text{corr}}, \qquad (2)$$

176   where v_{corr} is the P-wave velocity (Vp) and $\Delta h_i = h_i - \sum_{j=1}^{N} h_j / N$ represents the

177   relative topography, with h_i indicating the elevation at the i-th station (Figs. 2b-e). Here,

178   for the topographic correction, we assume a vertical-incidence angle for the incoming P

179   wave and a constant Vp that is likely representative of velocity structures averaged over

180   the top 1-2 km (e.g., Park et al., 2019). Since the velocity structure at the shallow section

181   is poorly constrained by existing velocity models, we use two constant Vp values, 2 km/s

182   and 4 km/s, to estimate the lower and upper limits of v_{corr}.

183

184   *3.1.2 Results*

185   Coherent P arrivals, with different peak frequencies, are observed crossing the array

186   for the four events in Figure S1. Although the frequency content of the P waveforms is

187   different between events, the obtained arrival patterns prior to topographic correction are

188   in general consistent (e.g., higher velocity in the northeast beneath array B4 in Fig. 4d).

189   The black curves in Figure 4 depict the teleseismic P-arrival patterns for each array

190   averaged over all events with the standard deviation of the mean giving the uncertainty,

191   and the delay times after correcting the array topography are illustrated as colored dashed

192   curves. Features of delay-time patterns associated with a velocity contrast across the fault

193   (i.e., a smoothed step function as in Fig. 6d of Qiu et al., 2017) and a low-velocity zone

194   (i.e., a mountain-shaped function as in Fig. 6b of Qiu et al., 2017) are both observed in

195   the results after the topographic correction.

196   Delay-time patterns resolved at arrays B1 and B4 yield clear velocity contrasts across

197   the fault, with the southwest block having later arrivals indicating slower velocity (~0.15

198   s and ~0.25 s in P-wave arrival time; Figs. 4a and 4d). Topographic corrections have

199   minor effects on the resolved arrival-time patterns at both arrays, as differences between

200   the two dashed curves are negligible in Figures 4a and 4d. Compared to the pattern

201   dominated by the velocity contrast, the time delay associated with low-velocity zones is

202   less obvious and comparable to the level of uncertainties (Figs. 4a and 4d). In contrast,

203   the dominant feature in arrival patterns resolved at arrays B2 and B3 yields several ~1-

204   km-wide low-velocity zones that generate a maximum time delay of ~0.04 s (Figs. 4b-c):

205   two centered at about 2.5 km southwest and 0.5 km northeast of the center of array B2;

206   one centered at ~0.5 km southwest of the B3 array center.

207   Although the topographic correction has negligible effects on the low-velocity zones

208   resolved at arrays B2 and B3, the time delays associated with velocity contrasts across

209   the fault beneath the two arrays are much weaker and vary significantly with the Vp used

210   in the correction (Figs. 4b-c). Different from the other three arrays, the polarity of the

211   velocity contrast at array B3 is flipped for arrival patterns resolved using Vp of 2 km/s

212   and 4 km/s. This is likely due to the combination of the larger difference in topography

213   and smaller velocity contrast at arrays B2 and B3. Therefore, we do not discuss the

214   velocity contrasts resolved from teleseismic *P* waves recorded at arrays B2 and B3 in

215   later sections.

216

217   3.2 Local *P*-wave delay time analysis

218   *P* waves from local earthquakes recorded by the B-arrays are observed at higher

219   frequencies (peaks at ~8 Hz; e.g., Fig. 5a) compared to those of teleseismic events

220   (between 0.5-2 Hz; e.g., Fig. 3). Thus, higher resolution images of local fault zone

221   structures can be achieved by analyzing arrival times of direct *P* waves from local

222   earthquakes across each array.

223

224   *3.2.1 Methodology*

225   Compared with teleseismic arrivals, the effect of source-receiver geometry on *P*

226   waves for local earthquakes recorded by an array requires additional processing than the

227   plane wave correction used in section 3.1.1. To extract the variations in *P*-wave arrival

228   times associated with local fault-zone structures, we first suppress the contributions from

229   source-receiver geometry and topographic variations by dividing the time axis of the *P*

230   waveform recorded at the *i*-th station for event *j* with its corresponding propagation

231   distance H_{ij} (e.g., from Fig. 5a to 5b). Here, we use the hypocentral distance to

232   approximate H_{ij}. *P*-wave picks, $s_{ij} = t_{ij}/H_{ij}$ in units of slowness (e.g., Figs. 5b and S5b-

233   S7b), are then picked via the short-term-average/long-term-average (STA/LTA)

234   algorithm (Allen, 1978) using waveforms within the slowness range of 0.15-0.25 s/km (to

235   exclude the effect of *S* waves). The *P*-wave SNR is calculated as the ratio between the

236   maximum and root mean square amplitudes of waveforms in slowness windows from s_{ij}

237   to 0.25 s/km and 0.15 s/km to s_{ij}, respectively. *P*-wave picks with SNR less than 10 are

238   not used, and events are excluded if less than 80% of the array shows good quality *P*-

239   wave picks (SNR > 10).

240   The array-mean slowness $\bar{s}_j = \sum_{i=1}^{M} s_{ij}/M$, with *M* being the number of stations, can

241   vary significantly with focal depth and epicenter location (due to 3-D velocity structures).

242   Therefore, we use relative slowness, $\hat{s}_{ij} = s_{ij}/\bar{s}_j$ (e.g., Qiu et al., 2017; Share et al.,

243   2017) to characterize statistical features of the local structure-related *P*-wave arrival

244   pattern using all available events. We can also estimate the local structure-related *P*-wave

245   arrival pattern in delay time, δt_{ij} for station *i* and event *j*, as $\delta t_{ij} = \left(s_{ij} - \bar{s}_j\right) \cdot H_{ij}$. Since

246   the existing velocity models indicate that structures in the Ridgecrest area are highly

247   heterogeneous, we only analyze *P* waves from events with depth > 5 km and close to

248   each array (red box in Figs. 5c and S5c-S7c) to ensure that the hypocentral distance is a

249   good approximation of the propagation distance H_{ij} and the resolved delay time pattern is

250   representative of local structures beneath the array.

251   Based on previous fault zone studies (e.g., Qiu et al., 2017; Share et al., 2017), the

252   observed *P*-wave travel time at station *i* for a near-fault event *j* mainly consists of two

253   components:

$$t_{ij} = s_{ij} \cdot H_{ij} \approx \bar{s}_j \cdot H_{ij} \cdot (1 + \eta_i) + \Delta t_{ij}. \tag{3}$$

254   The first term of equation (3) indicates the time delay associated with the cross-fault

255   velocity contrast beneath the array, and, ideally, η_i is a step function, i.e., equals $-\eta/2$

256   and $\eta/2$ ($\eta < 1$) for stations on the faster and slower crustal blocks, respectively. The

257   second term Δt_{ij} represents the contribution of local structures beneath the array at

258   shallow depth (e.g., fault damage zone, sedimentary basin). Let θ_j be the average incident

259   angle of *P* waves from event *j* to the array, Δt_{ij} is inversely proportional to $\cos \theta_j$, i.e.,

260     $\Delta t_{ij} \approx \tau_i/\cos \theta_j$, where τ_i is the delay time of a vertically incident *P* wave associated

261   with shallow structures beneath the array at station *i*.

262   Following the derivation in Text S1, the arrival patterns averaged over all near-fault

263   and close-to-array events are approximately given by:

$$\hat{S}_i = \sum_{j=1}^{N} \hat{s}_{ij} \Big/ N \approx 1 + \eta_i + \tau_i / (\tilde{t}_i \cos \tilde{\theta}) \approx 1 + \eta_i + \tau_i / (\tilde{t} \cos \tilde{\theta}), \tag{4a}$$

264   in relative slowness, and

$$\delta T_i = \sum_{j=1}^{N} \delta t_{ij} \Big/ N \approx \eta_i \cdot \bar{t}_i + \tau_i / \cos \tilde{\theta} \approx \eta_i \cdot \bar{t} + \tau_i / \cos \tilde{\theta}, \tag{4b}$$

265   in delay time, after dropping the higher order terms. $\tilde{t}_i$ and $\bar{t}_i$ are the harmonic and

266   arithmetic means of *P*-wave travel time t_{ij}, respectively, over all events. N is the number

267   of events and $\tilde{\theta}$ is the mean incidence angle averaged over all events. $\tilde{t}$ and $\bar{t}$ denote the

268   array-mean travel times of $\tilde{t}_i$ and $\bar{t}_i$, respectively. It is interesting to note that the shape of

269   arrival patterns $\hat{S}_i$ and δT_i is the same, i.e., $\delta T_i / (\hat{S}_i - 1)$ is a constant, when $\tilde{t} \approx \bar{t}$.

270

271   *3.2.2 Results*

272   Figure 6a shows the results of statistical analysis on the local structure-related *P*-wave

273   arrival pattern in relative slowness ($\hat{S}_i$ in eq. 4a) and delay time (δT_i in eq. 4b) for array

274   B1 using 189 near-fault events with depth > 5 km outlined by the red box in Figure 5c.

275   The relative slowness patterns are averaged over all analyzed events and the standard

276   deviation of the mean is used to estimate the uncertainty. The small error bars and

277   confined width of histograms suggest that the mean relative slowness curve is

278   representative of the patterns observed from all analyzed events. The mean pattern in

279   delay time (δT_i) with small uncertainties (Fig. 6a) is observed to have the same shape as

280   that of the relative slowness, i.e., $\delta T_i / (\hat{S}_i - 1)$ is approximately a constant. This is

281   consistent with equation (4) when $\tilde{t} \approx \bar{t}$.

282   Good agreement between the mean patterns of relative slowness and delay time with

283   small uncertainties is also observed at arrays B2-B4 in Figures 6c, 6e, and 6g. Consistent

284   with the teleseismic *P*-wave arrival time pattern shown in Figure 4, we observe the

285   features of delays in local *P*-wave arrival time associated with a step-function-like and

286   several mountain-shaped components at each array (left panels of Fig. 6) that likely

287   correspond to a velocity contrast across the fault and low-velocity zones, respectively.

288   Similar mean patterns of relative slowness are also observed for subsets of events

289   grouped according to different narrow ranges of depth (Figs. S8a-S11a). The polarity of

290   the velocity contrast is the same for arrays B1 and B4 (Figs. 6a and 6g), with the

291   southwest block being slower, consistent with results of the teleseismic delay time

292   analysis (Figs. 4a and 4d). However, delay time patterns at arrays B2 and B3 show the

293   opposite polarity (Figs. 6c and 6e), i.e., the northeast block having slower velocity. This

294   is consistent with results of the teleseismic delay time analysis before the topographic

295   correction (Figs. 4b-c).

296   We note that delay times associated with shallow materials (τ_i in eq. 4a) could also

297   yield a step-function-like pattern (e.g., variations in sediment thickness across fault).

298   Thus, the amplitude of the step-function-like component, $\tilde{\eta}$, is an upper limit of the

299   magnitude of the cross-fault velocity contrast. We measure the amplitude of $\tilde{\eta}$ as the

300   difference between relative slowness values averaged over stations on the southwest and

301   northeast edges of the array (left panels of Fig. 6), and the values are 5.3%, 0.8%, 3.9%,

302   and 8.8% at the sites of arrays B1-B4, respectively. Following equation (4a), if the

303   observed step-function-like pattern is due to variations in shallow materials, i.e.,

304    $\tau_i/(\tilde{t}\cos\tilde{\theta})$, the amplitude of $\tilde{\eta}$ should decrease with depth as the mean incidence angle

305    $\tilde{\theta}$ and travel time $\tilde{t}$ increase with propagation distance. This contradicts the observations

306   that $\tilde{\eta}$ increases with depth at array B2 (Fig. S9a) and only slightly varies with depth at

307   arrays B1 (Fig. S8a), B2 (Fig. S10a), and B4 (Fig. S11a), suggesting the observed step-

308   function-like pattern is likely a good approximation of the cross-fault velocity contrast, if

309   the contribution due to elevation variation is negligible.

310   To further analyze the effect of array topography in the step-function-like component

311    $\tilde{\eta}$, we first model the delay times of cross-fault velocity contrast (i.e., $\eta_i \cdot \bar{t}$ in eq. 4b) with

312   a smoothed step function (Text S2). The modeled delay times of cross-fault velocity

313   contrast consist of three components: a linear trend within a transition zone between two

314   groups of stations at the southwest and northeast edges with constant values (eq. S5 in

315   Text S2; Fig. 6). We then illustrate the location of the resolved transition zone beneath

316   each array in Figure 2. At array B2, the span of the transition zone covers the area with a

317   steep slope in the array topography (Fig. 2c). Considering the large variation in elevation

318   (> 100 m) and the small amplitude of $\tilde{\eta}$ (0.8%), we conclude that the modeled cross-fault

319   velocity contrast may be dominated by the residual in topographic correction rather than

320   structure-related delay times. In contrast, for the other three arrays, the topography

321   variation is less significant and the amplitude of $\tilde{\eta}$ is much larger (> 4%), indicating the

322   array topography likely has negligible contribution to the modeled cross-fault velocity

323   contrast.

324   The width of these transition zones and their locations relative to fault surface traces

325   provide additional information on the internal structures of the fault zone. At array B1,

326   the southwest edge of the transition zone correlates well with the surface trace of the Mw

327   7.1 earthquake (Fig. 2b), which is indicative of asymmetric rock damage offset to the

328   northeast (faster side). At array B3, the northeast edge of the transition zone agrees well

329   with the surface trace of the Mw 6.4 earthquake. The data recorded by array B3, crossing

330   the surface ruptures of both the Mw 6.4 and Mw 7.1 events, likely detect the velocity

331   contrast across the fault that hosted the Mw 6.4 earthquake, with the northwest side being

332   higher in velocity. An asymmetric rock damage offset to the faster side (northwest) is

333   also observed beneath array B3 (Fig. 2d). At array B4, the transition zone is much wider

334   and almost covers the entire array. This is consistent with the fact that the array is at the

335   southeast end of the Mw 7.1 earthquake rupture, where the rupture zone has several

336   surface traces and is less localized (Fig. 2e).

337   Low-velocity zones that further delay the *P*-wave arrivals are also observed at each

338   array, in addition to the pattern associated with the cross-fault velocity contrast (left

339   panels of Fig. 6). To highlight contributions from these low-velocity zones (i.e.,

340    $(\tau_i/(\tilde{t}\cos\tilde{\theta})$ in eq. 4a), the right panels of Figure 6 show the mean delay time pattern by

341   subtracting the modeled $\eta_i \cdot \tilde{t}$ from δT_i (eq. 4b). A heat map of delay times δT_i after

342   subtracting the modeled $\eta_i \cdot \tilde{t}$ is also shown. Similar delay patterns of τ_i reflecting

343   shallow materials beneath the array are seen using events within different narrow ranges

344   of depth (Figs. S8b-S11b). The range of delay patterns related to the major low-velocity

345   zones underneath each array are outlined with green dashed lines in Figure 6. The slower

346   inner cores (with peak delay time > 0.03 s) of these low-velocity zones are marked by the

347   red dashed lines.

348   Consistent with the teleseismic *P*-wave arrival patterns obtained at arrays B2 and B3

349   (Figs. 4b-c), we observe low-velocity zones with comparable widths centered at the same

350   locations. In addition, we retrieve higher resolution images of the narrower and slower

351   inner cores that correlate well with locations of the local peaks identified in the

352   teleseismic *P*-wave delay time patterns (Figs. 4b-c). This is likely due to the shorter

353   wavelength of local seismic *P* waves that can provide high-resolution images of internal

354   fault-zone structures (e.g., Dahlen et al., 2000). Delay patterns related to low-velocity

355   zones are also observed in the results for arrays B1 and B4 (Figs. 6b and 6h), which are

356   hard to identify in the teleseismic delay time analyses (Figs. 4a and 4d) due to large

357   uncertainties (i.e., insufficient number of analyzed events).

358   We note that consistent patterns of the cross-fault velocity contrast and low-velocity

359   zones are observed (with different amplitudes) in results obtained using events with large

360   along-fault distances (Figs. S8c-d, S9c-d, S10c-d, and S11c-d). To better visualize the

361   locations of these major low-velocity zones with respect to the array configurations and

362   fault surface traces, we depict the core and entire range of these zones in Figure 2. In

363   general, locations of the identified low-velocity zones correlate well with surface traces

364   of the 2019 Ridgecrest earthquake sequence (Fig. 2) inferred from Xu et al. (2020). This

365   suggests that these low-velocity zones are likely indicative of fault damage zones or

366   sedimentary basins around fault segments.

367

368   3.3 Fault-zone trapped waves

369   A low-velocity fault-damage zone that is sufficiently uniform over a given distance

370   can act as a waveguide and generate, in addition to delay times and motion amplification,

371   trapped waves resulting from constructive interference of critically reflected phases

372   within the waveguide (e.g., Ben-Zion & Aki 1990; Igel et al., 1997; Jahnke et al., 2002).

373   Such waves have been observed at many locations, including the San Jacinto fault zone

374   (e.g., Lewis et al., 2005; Qin et al., 2018; Qiu et al., 2017; Share et al., 2019; Wang et al.,

375   2019), the Parkfield section of the San Andreas fault (e.g., Li et al., 1990; Lewis and Ben-

376   Zion, 2010; Ellsworth & Malin, 2011), and various other faults in California, Japan, Italy,

377   Turkey, and other places. Catchings et al. (2016) used peak ground velocities of *P* and *S*

378   waveforms recorded by cross-fault linear arrays to infer the location and width of the

379   West Napa-Franklin fault zone. Similarly, we find fault-damage-zone-related

380   amplification in data recorded by the B-arrays (e.g., Figs. 7 and S12) and use such

381 amplification to detect FZTW candidates. In this section, we first infer the location and

382 width of fault damage zones that produce FZTWs using waveforms for the fault-parallel

383 component, and then use waveforms of these candidates recorded by array B4 to invert

384 for properties (e.g., width, velocity, and attenuation) of the local fault-zone waveguide.

385

386 *3.3.1 Methodology*

387   Figure 7a shows *S* waveforms recorded on the fault-parallel component of array B4

388 for an example event (Fig. 1). Following Ben-Zion et al. (2003) and Qiu et al. (2017, Fig.

389 S6), several preprocessing steps are applied to the data prior to FZTW analyses. These

390 include instrument response removal, integration to displacement seismogram, tapering

391 between 1 s before and 2 s after the *S* pick, and convolution with $1/t^{1/2}$ to convert a point-

392 source response to that of an equivalent *SH* line dislocation source (e.g., Igel et al., 2002;

393 Vidale et al., 1985). Compared to the raw recordings (Fig. S12), clearer Love-type

394 waveform sections with large amplitudes and lower frequency are found at a group of

395 stations (stations B416-423; Fig. 7a) in the southwest part of the array.

396   Figure 7b displays distributions of peak ground velocities (PGV) and root mean

397 square amplitudes (RMS) of the fault-parallel-component *S* waveforms, normalized by

398 the maximum value of the entire array. Large values of PGV and RMS are seen at

399 stations with FZTW, with considerably higher amplitudes than at the rest of the array. We

400 define the likelihood of a FZTW recorded by a station as the multiplication of PGV and

401 RMS (Fig. 7b). The likelihood curve measured for each event is normalized by the

402 maximum value of the entire array. After averaging the likelihood curves over all

403 analyzed events, we highlight the array sections that have high likelihood of recording

404 FZTWs (Fig. 8); these sections can be used to infer the location and width of fault-zone

405 waveguides.

406   Since FZTWs are observed in *S* waveforms recorded at stations B416-B423 of array

407 B4 for the example event in Figure 7a, we can also identify candidate events with similar

408 good-quality FZTWs through template matching, i.e., cross-correlating the fault-parallel

409 component *S* waveforms recorded by stations B416-B423 for each event with those of the

410 example event. The trapped waves of candidate events (Fig. 9a) that yield cross-

411 correlation coefficients higher than 0.85 (e.g., Fig. 9b) are averaged (Fig. 9c) and inverted

412   for properties (e.g., width, shear velocities, and attenuation) of the average fault-zone

413   waveguide using a genetic inversion algorithm (e.g., Ben-Zion et al., 2003; Lewis et al.,

414   2005; Qiu et al., 2017).

415   We test a total of 10,000 models (50 generations and 200 models per generation) to

416   obtain a good estimate of the fault-zone parameters in the inversion. Parameters of the

417   best-fitting model and the 2,000 models (investigated in the last 10 generations) are

418   extracted from the inversion. Because there are strong trade-offs between model

419   parameters governing FZTWs (e.g., Ben-Zion, 1998; Jahnke et al., 2002; Peng et al.

420   2003), a successful inversion not only yields good waveform fits but also shows

421   consistency between parameters of the best-fitting model and peaks of the probability

422   density distributions of parameters developed in the last 10 generations. Additional

423   details on the method can be found in section 3.4 of Qiu et al. (2017) and Ben-Zion et al.

424   (2003).

425

426   *3.3.2 Results*

427   Figure 8 shows the distributions of FZTW likelihood values as a function of station

428   location, estimated at arrays B1-B4 for all events within the red box shown in Figure 1.

429   The likelihood values averaged over all analyzed events are representative of the in-situ

430   amplification in *S*-wave. A good correlation is found in locations between local maxima

431   of the likelihood curve and slower inner cores of the low-velocity zones identified from

432   the local *P*-wave delay time analysis (Section 3.2; Fig. 8). This suggests that the observed

433   inner cores of these low-velocity zones not only significantly delay the *P* arrivals but also

434   amplify the incoming *S* waves.

435   However, not all stations within the inner cores of low-velocity zones identified in

436   Figure 6 yield the same peak value of FZTW likelihood (Fig. 8). At array B4, the location

437   of the candidate fault-zone waveguide (B417-B422; Figs. 7a and 8d) is in good

438   agreement with the inner core of low-velocity zones that yields the highest likelihood

439   value. We do not detect candidate FZTW at stations within the other low-velocity zones

440   with much lower peak likelihood values. This is consistent with detailed fault-zone

441   studies at Parkfield (Lewis & Ben-Zion, 2010), the rupture zone of the 1992 Landers

442   earthquake (Peng et al., 2003), and fault zones in Japan (Mamada et al. 2004; Mizuno et

443   al. 2008). These studies showed that various sections of fault zones produce delay times

444   and other signals of damaged rocks but are either too heterogeneous or have significant

445   segmentation between sources and receivers to generate trapped waves (e.g., Igel et al.

446   1997, 2002; Jahnke et al. 2002).

447   Not all analyzed events show *S*-wave amplification patterns that are consistent with

448   the averaged curve. We first detect candidate events that show high FZTW likelihood

449   values at stations B417-B422 (Fig. 8d) by cross-correlating the likelihood pattern

450   measured for each event (e.g., black curve in Fig. 7b) with that of the mean (e.g., red

451   curve in Fig. 8d). More than 600 events with cross-correlation coefficients higher than

452   0.95 are identified as FZTW candidates. We further select 33 events (Fig. 9a) that

453   produce high-quality FZTWs (e.g., Fig. 9b) from the candidate events, through template

454   matching of fault-parallel component *S* waveforms at stations B416-B423 (using FZTWs

455   shown in Fig. 7a as the template; Section 3.3.1). These high-quality FZTW candidates

456   show a consistent source-receiver path (top inset of Fig. 9a), indicating the depth of the

457   fault-zone waveguide is likely shallower than 5 km, and there is an optimal range of

458   incidence angle for injecting seismic energy into the fault damage zone beneath array B4

459   (e.g., Fohrmann et al. 2004).

460   Compared to FZTW observed from each candidate event (e.g., Fig. 9b), the stacked

461   recordings (Fig. 9c) yield much higher SNRs and can thus provide more reliable and

462   robust estimations of the average fault-zone waveguide properties. Figure 10 presents the

463   inversion results from modeling the stacked waveforms shown in Figure 9c (in red). The

464   best-fitting model yields good waveform fits (Fig. 10a) and suggests a simplified (i.e.,

465   vertical rectangular-shaped) fault-zone waveguide with a width of ~280 m, Q value of

466   ~30, and *S*-wave velocity ~80% of the surrounding host rocks (Fig. 10b). The estimated

467   propagation distance inside the waveguide is ~5.4 km (Fig. 10b). Because this includes a

468   propagation component along-strike (i.e., non-vertical incidence angle), it suggests a

469   waveguide depth of ~4 km (= $5.4/\sqrt{2}$). The estimated average *S*-wave velocity in the

470   host rock is ~4.1 km/s, with the northeast block being ~4.2% faster, consistent with

471   results from the *P*-wave delay time analysis at array B4 (~3% velocity contrast across

472   LVZ#5 with higher velocity in the northeast; Fig. 6g). The parameters of fault-zone

473   models from the last 10 generations (2,000 models) are shown in Figure 10b, along with

474   the corresponding probability density functions computed as the frequency of each

475   parameter value weighted by the fitness values (Ben-Zion et al., 2003). The good

476   waveform fits, combined with the consistency between best fitting parameters and peaks

477   of the probability density functions (Fig. 10b), suggest that the best-fitting model

478   parameters provide robust estimates of the average properties of the fault-zone

479   waveguide.

480   In addition to FZTWs, we detect in data of stations B420 and B422 (Fig. S13) clear

481   fault-zone head waves (FZHWs) that refract along a bimaterial fault interface (e.g., Ben-

482   Zion, 1990) and arrive ~0.1 s earlier than the direct *P*-wave. The early-arriving FZHWs

483   are inferred from horizontal particle motion analysis modified from the method of Bulut

484   et al. (2012). It is important to note that in the polarization analysis we only focus on the

485   rotation of the horizontal particle motion between FZHW and direct *P*-wave. This is

486   because the polarization direction of horizontal particle motion for the direct *P* wave at

487   stations inside a damage zone (Fig. S13) can deviate significantly from the source-

488   receiver back azimuth. Since the differential time between the FZHW and direct *P*-wave

489   decreases significantly from the northeast (B422) to the southwest (B420) over a short

490   distance (~0.1 km), the observed FZHW is likely traveling along a local interface that is

491   associated with the edge of the damage zone (e.g., Qiu et al., 2017) on the northeast side

492   between stations B422 and B423. Similar FZTW and FZHW signals are also clearly

493   observed in the data of array B2 (e.g., Fig. S14) but not for arrays B1 and B3.

494

495   **4. Discussion**

496   We analyze systematically time delays of *P* arrivals from teleseismic and local

497   earthquakes, *S*-wave amplification, and fault zone trapped waves in data recorded by four

498   long-aperture (4-8 km) arrays at different locations along the rupture zone of the 2019

499   Mw 7.1 Ridgecrest earthquake (Fig. 2). The analyses allow us to derive information on

500   velocity contrast interfaces and low velocity damage zones at scales of ~100 m associated

501   with the Ridgecrest rupture zone. The results are complementary to tomographic velocity

502   models that provide a regional context but do not resolve the internal structural elements

503   of the Ridgecrest rupture zone imaged in this work.

504   The delay times of *P*-waves from both teleseismic and local earthquakes, after proper

505   corrections for propagation and topography effects, show clear and consistent velocity

506   contrasts across the rupture zone, with a higher velocity in the northeast at arrays B1

507   (Figs. 4a and 6a) and B4 (Figs. 4d and 6g). Combining the delay times inferred from

508   teleseismic *P* waves in Figure 4 and the amplitudes of cross-fault velocity contrast in the

509   left panels of Figure 6, we can estimate the depth extent of the contrast interface beneath

510   each array. For a constant velocity contrast across the fault with an amplitude of η

511   extending to a depth of H_c, the time delay between teleseismic *P* arrivals, assuming a

512   vertical angle of incidence, at stations on different sides of the fault is given by

$$\Delta t \approx \eta \cdot H_c / \overline{V_p},$$ (5)

513   where $\overline{V_p}$ is the average *P*-wave velocity of the upper crust. We set $\overline{V_p} = 5.6$ km/s as

514   inferred from the mean of the average slowness $\bar{s}_j$ (eq. 3) over all event-array pairs. The

515   H_c beneath arrays B1 ($\Delta t = 0.15$ s from Fig. 4a and $\eta = 5.3\%$ from Fig. 6a) and B4

516   ($\Delta t = 0.25$ s from Fig. 4d and $\eta = 8.8\%$ from Fig. 6g), computed using equation (5), are

517   both ~16 km, suggesting a deep fault interface that likely extends to the seismogenic

518   depth (10-12 km) after taking measurement uncertainties (Figs. 4a and 4d) and non-

519   vertical angle of incidence into account.

520   We do not estimate the depth H_c for arrays B2 and B3 as the teleseismic *P*-wave

521   delay times Δt are highly dependent on the topographic correction (Section 3.1.2; Figs.

522   4b-c). The cross-fault velocity contrast observed in local *P*-wave delay time patterns

523   resolved at arrays B2 and B3 (Figs. 6c and 6e) suggests a higher velocity in the

524   southwest, the opposite of the polarity obtained at arrays B1 and B4 (Figs. 6a and 6g).

525   Further analysis of the transition zones (Figs. 2b-e) resolved from the modeled cross-fault

526   velocity contrasts (left panels of Fig. 6) indicates that: the observed velocity contrast

527   beneath array B2 is likely an artifact due to insufficient topographic correction; array B3

528   may detect the velocity contrast across the rupture zone of the Mw 6.4 earthquake; and

529   the broader transition zone beneath array B4 suggests the fault zone at the southeast edge

530   of the Mw 7.1 earthquake rupture is likely not as well localized (e.g., several rupture

531   surface traces; Fig. 2) as those beneath the other three arrays.

532   Symmetry properties of fault damage zones with respect to the main slip surface can

533   provide information on the direction of earthquake ruptures (e.g., Ben-Zion & Shi, 2005;

534   Dor et al., 2006; Mitchell et al., 2011; Xu et al., 2012). Statistically preferred rupture

535   direction is expected for prominent bimaterial faults (e.g., Ampuero & Ben-Zion, 2008;

536   Andrews and Ben-Zion, 1997; Shlomai & Fineberg, 2016; Weertman, 1980). However, a

537   prominent large-scale velocity contrast is not expected across the Ridgecrest rupture zone

538   based on the regional velocity models (Lee et al., 2014; Shaw et al., 2015; White et al.,

539   2020) and total offset across the structure. The transition zones of the cross-fault velocity

540   contrast resolved at arrays B1 (Fig. 2b) and B3 (Fig. 2d) are asymmetrically distributed to

541   the faster side of the fault surface trace, which is likely indicative of asymmetric rock

542   damage offset to the faster side of the rupture zone for the Mw 7.1 and Mw 6.4

543   earthquakes, respectively. Since the transition zones resolved at arrays B2 and B4 are not

544   well resolved, we do not discuss results obtained at these two locations (Figs. 2c and 2e).

545   Good agreement between locations of slower inner cores of the low-velocity zones

546   identified from delay time analyses and the group of stations with amplified *S* waveforms

547   (Fig. 8) is found beneath all four arrays. Combined with the narrow width (300-400 m;

548   right panels of Fig. 6) of the inner cores, these zones that both significantly delay the *P*

549   arrivals and amplify the incoming *S* waveforms are likely representative of localized

550   damage zones, rather than variations in sedimentary thickness that usually yield much

551   wider patterns of *S*-wave amplification. Waveform modeling of FZTWs detected at array

552   B4 yields good waveform fits and an average waveguide with fault-zone parameters

553   comparable to those inferred from previous studies in the SJFZ (Qin et al., 2018; Qiu et

554   al., 2017; Share et al., 2017, 2019): width of ~300 m, Q of ~30, *S*-wave velocity

555   reduction of ~20% inside the damage zone, and depth extent of 3-5 km (Fig. 10).

556   To further investigate the structures that produce the observed low-velocity zones, we

557   map the inner core and entire span of these zones together with locations of fault surface

558   traces (or their extrapolations) and the surface geology compiled by the USGS in Figure

559   2. The low-velocity zone (LVZ#1 in Fig. 6b) identified at array B1, in contrast to the

560   resolved transition zone (Fig. 2b), is on the southwest (slower) side of the surface trace of

561   the main rupture zone. There are two well-separated inner cores (Fig. 2b) that yield both

562   large delay times (Fig. 6b) and *S*-wave amplitudes (Fig. 8a). The inner core in the

563 southwest (Fig. 2b) is likely associated with the damage zone of past earthquake ruptures,

564 since it is centered on a surface displacement mapped before the 2019 Ridgecrest

565 earthquake. The northeast inner core (Fig. 2b), however, is likely representative of the

566 damage zone of the Mw 7.1 Ridgecrest earthquake ruptures, as it is centered on the main

567 rupture zone.

568   At array B2, two low-velocity zones (LVZ #2 and #3) are identified from Figures 6d.

569 Different from the other three arrays, the surface geology beneath array B2 consists of

570 two different types of rocks (alternating playa and igneous rocks; Fig. 2c), which likely

571 explains the weak cross-fault velocity contrast resolved at this location (Figs. 4b and 6c).

572 Both low-velocity zones are centered on the surface traces (or their extrapolations) of the

573 main rupture zone, with the slower inner cores close to the fault traces. Good agreement

574 between locations of low-velocity zones and surface traces of the 2019 Ridgecrest

575 earthquake ruptures is also observed beneath arrays B3 (Fig. 2d) and B4 (Fig. 2e). Since

576 the locations of these low-velocity zones correlate well with the fault surface traces rather

577 than the surface geology, we conclude that: these low-velocity zones with slower inner

578 cores that considerably amplify *S* waveforms depict the fault damage zones of the Mw

579 7.1 earthquake rupture at arrays B2 and B4, and ruptures of both the Mw 6.4 and Mw 7.1

580 earthquakes at array B3.

581   It is interesting to note that we find a clear and coherent secondary phase between *P*

582 and *S* arrivals in waveforms recorded by part of array B4 for more than 10 events located

583 beneath the array. Figure S15b shows such signals for an example event (circle in Fig.

584 S15a). The strong phases are visible at stations B423-B457 (blue curve in Fig. S15b) and

585 correlate well with the shape of the direct *P*-wave but with the opposite first-motion

586 polarity (e.g., Fig. S15d) suggesting the signals are likely reflected or converted waves.

587 Since the array-mean travel time of the direct P-wave is between 1-1.5 s, based on the

588 ~0.5 s differential time between the direct P and the strong secondary waves (e.g., black

589 and blue dashed lines in Fig. S15d), the reflection or conversion interface should be a few

590 kilometers away from the array and is thus likely associated with the Garlock fault in the

591 south rather than the rupture zone of the Mw 7.1 Ridgecrest earthquake (Fig. S15a).

592 Detailed analysis of these candidate reflected/converted phases may require more

593 sophisticated techniques, such as seismic migration of the arrival times and waveform

594   simulation with accurate source information (e.g., focal mechanism) for the
595   understanding of the wave amplitudes. As these secondary signals detected in Figure S15
596   are outside the scope of this paper, additional analysis of these phases will be the subject
597   of a follow-up study.

598

599   **5. Conclusions**

600   The rupture zone of the Mw 7.1 Ridgecrest earthquake is shown to have seismic
601   velocity contrast interfaces and damage structures with significant along-strike variations
602   of local damage zones. The results are in overall agreement with and complementary to
603   the mapped fault surface traces in the region (e.g., Xu et al., 2020) and previous analyses
604   of aftershocks in the Ridgecrest sequence (e.g., Ross et al., 2019; Cheng & Ben-Zion,
605   2020).

606   Delay times of *P* waves from teleseismic and local earthquakes recorded by arrays B1
607   and B4 resolve velocity contrasts ranging from 5%-9% in Vp across the Ridgecrest
608   rupture zone, which extend to a depth of ~16 km, with the northeast being locally faster.
609   Data recorded by array B3, crossing the surface ruptures of both the Mw 6.4 and Mw 7.1
610   events, likely detects a ~4% velocity contrast in Vp across the fault that hosted the Mw
611   6.4 earthquake, with the northwest side having higher velocity. Asymmetric rock damage
612   concentrated on the stiffer side of the fault is seen beneath arrays B1 and B3.

613   Low-velocity zones that further delay the *P* waves of local seismic events are
614   centered on mapped surface traces of faults associated with both past and the 2019
615   Ridgecrest earthquake ruptures. Considerable amplification is seen consistently in *S*
616   waveforms recorded at stations within the slower inner core of the identified LVZ. Clear
617   FZTWs are identified at arrays B2 and B4, and inversion of high-quality FZTWs at array
618   B4 indicates a waveguide with average properties comparable to those found in previous
619   studies at sites of the San Jacinto fault zone. Phases identified as FZHWs, associated with
620   the northeast boundary of the fault zone waveguide, are observed at arrays B2 and B4.
621   Coherent candidate reflected or converted waves are observed in data recorded at array
622   B4. Additional analysis of these phases can provide additional constraints on prominent
623   impedance interfaces associated with the Garlock fault and/or the Ridgecrest rupture
624   zone.

625

Data Availability

627   The digital data are available in mseed day volume format, with each component in a

628   separate volume. The data samples are 4 byte floats and consistently sampled at 500

629   samples/second. Data described in this report are available from the IRIS Data

630   Management Center (https://ds.iris.edu/ds/nodes/dmc/data/). An accompanying report for

631   data acquisition is available from Catchings et al. (2020). Surface traces of the 2019

632   Ridgecrest earthquake sequence shown in this study were determined by Xu et al. (2020).

633   The geologic data depicted as the background of Figure 2 were obtained from the USGS

634   website (https://mrdata.usgs.gov/geology/state/state.php?state=CA).

635

655 Earthquake Center (based on NSF Cooperative Agreement EAR-1600087 and USGS

656 Cooperative Agreement G17AC00047).

657

658 **References**

659 Aki, K. (1979). Characterization of barriers on an earthquake fault. *Journal of*
660   *Geophysical Research*. https://doi.org/10.1029/JB084iB11p06140

661 Allen, R. (1978). Automatic earthquake recognition and timing from signal traces.
662   *Bulletin of the Seismological Society of America*. 68 (5):1521-1532.

663 Ampuero, J. P., & Ben-Zion, Y. (2008). Cracks, pulses and macroscopic asymmetry of
664   dynamic rupture on a bimaterial interface with velocity-weakening friction.
665   *Geophysical Journal International*, *173*(2), 674–692. https://doi.org/10.1111/j.1365-
666   246X.2008.03736.x

667 Andrews, D. J., & Ben-Zion, Y. (1997). Wrinkle-like slip pulse on a fault between
668   different materials. *Journal of Geophysical Research B: Solid Earth*.
669   https://doi.org/10.1029/96jb02856

670 Ben-Zion, Y. (1990). The response of two half spaces to point dislocations at the material
671   interface, *Geophys. J. Int., 101*, 507-528.

672 Ben-Zion, Y. (1998). Properties of seismic fault zone waves and their utility for imaging
673   low-velocity structures. *Journal of Geophysical Research: Solid Earth*.
674   https://doi.org/10.1029/98jb00768

675 Ben-Zion, Y., & Aki, K. (1990). Seismic radiation from an SH line source in a laterally
676   heterogeneous planar fault zone. *Bulletin - Seismological Society of America*.

677 Ben-Zion, Y., & Huang, Y. (2002). Dynamic rupture on an interface between a compliant
678   fault zone layer and a stiffer surrounding solid. *Journal of Geophysical Research*.
679   https://doi.org/10.1029/2001jb000254

680 Ben-Zion, Y., & Shi, Z. (2005). Dynamic rupture on a material interface with
681   spontaneous generation of plastic strain in the bulk. *Earth and Planetary Science*
682   *Letters*. https://doi.org/10.1016/j.epsl.2005.03.025

683 Ben-Zion, Y., Peng, Z., Okaya, D., Seeber, L., Armbruster, J. G., Ozer, N., et al. (2003).
684   A shallow fault-zone structure illuminated by trapped waves in the Karadere-Duzce
685   branch of the North Anatolian Fault, western Turkey. *Geophysical Journal*
686   *International*. https://doi.org/10.1046/j.1365-246X.2003.01870.x

687 Brietzke, G. B., & Ben-Zion, Y. (2006). Examining tendencies of in-plane rupture to
688   migrate to material interfaces. *Geophysical Journal International*.
689   https://doi.org/10.1111/j.1365-246X.2006.03137.x

690 Bulut, F., Ben-Zion, Y., & Bohnhoff, M. (2012). Evidence for a bimaterial interface
691   along the Mudurnu segment of the North Anatolian Fault Zone from polarization
692   analysis of P waves. *Earth and Planetary Science Letters*.
693   https://doi.org/10.1016/j.epsl.2012.02.001

694 Catchings, R. D., Goldman, M. R., Li, Y. G., & Chan, J. H. (2016). Continuity of the
695   west napa-franklin fault zone inferred from guided waves generated by earthquakes
696   following the 24 august 2014 Mw 6.0 south napa earthquake. *Bulletin of the*
697   *Seismological Society of America*. https://doi.org/10.1785/0120160154

698 Catchings, R. D., M.R. Goldman, J.H. Steidl, J.H. Chan, A.A. Allam, C.J. Criley, Z. Ma,

D.S. Langermann, G.J. Huddleston, A.T. McEvilly, D.D. Mongovin, E.M. Berg, and Y. Ben-Zion, 2020. Nodal Seismograph Recordings of the 2019 Ridgecrest Earthquake Sequence, *Seism. Res. Lett.*, doi: 10.1785/0220200203.

Chen, K., Avouac, J. P., Aati, S., Milliner, C., Zheng, F., & Shi, C. (2020). Cascading and pulse-like ruptures during the 2019 Ridgecrest earthquakes in the Eastern California Shear Zone. *Nature Communications*, *11*(1), 3–10. https://doi.org/10.1038/s41467-019-13750-w

Cheng, Y. and Y. Ben-Zion, 2020. Variations of earthquake properties before, during and after the 2019 M7.1 Ridgecrest, CA, earthquake, *Geophys.*, *Res.*, *Lett.,* 47, e2020GL089650, doi: 10.1029/2020GL089650.

Cochran, E., Li, Y. G., Shearer, P. M., Barbot, S., Fialko, Y., & Vidale, J. E. (2009). Seismic and geodetic evidence for extensive, long-lived fault damage zones. *Geology*. https://doi.org/10.1130/G25306A.1

Crotwell, H. P., Owens, T. J., & Ritsema, J. (1999). The TauP Toolkit: Flexible Seismic Travel-time and Ray-path Utilities. *Seismological Research Letters*. https://doi.org/10.1785/gssrl.70.2.154

Dahlen, F. A., Hung, S.-H., & Nolet, G. (2000). Frechet kernels for finite-frequency traveltimes - I. Theory. Geophysical Journal International, (141), 157–174. https://doi.org/10.1007/978-3-319-06540-3

Dor, O., Rockwell, T. K., & Ben-Zion, Y. (2006). Geological observations of damage asymmetry in the structure of the San Jacinto, San Andreas and Punchbowl faults in Southern California: A possible indicator for preferred rupture propagation direction. *Pure and Applied Geophysics*. https://doi.org/10.1007/s00024-005-0023-9

Ellsworth, W. L., & Malin, P. E. (2011). Deep rock damage in the san andreas fault revealed by P- and S-type fault-zone-guided waves. *Geological Society Special Publication*. https://doi.org/10.1144/SP359.3

Fohrmann, M., Igel, H., Jahnke, G., & Ben-Zion, Y. (2004). Guided waves from sources outside faults: An indication for shallow fault zone structure? *Pure and Applied Geophysics*. https://doi.org/10.1007/s00024-004-2553-y

Hauksson, E., Yang, W., & Shearer, P. M. (2012). Waveform relocated earthquake catalog for Southern California (1981 to June 2011). *Bulletin of the Seismological Society of America*. https://doi.org/10.1785/0120120010

Huang, Y., Ampuero, J. P., & Helmberger, D. V. (2014). Earthquake ruptures modulated by waves in damaged fault zones. *Journal of Geophysical Research: Solid Earth*. https://doi.org/10.1002/2013JB010724

Igel, H., Ben-Zion, Y., & Leary, P. C. (1997). Simulation of SH- and P-SV-wave propagation in fault zones. *Geophysical Journal International*. https://doi.org/10.1111/j.1365-246X.1997.tb05316.x

Igel, H., Jahnke, G., & Ben-Zion, Y. (2002). Numerical simulation of fault zone guided waves: Accuracy and 3-D effects. *Pure and Applied Geophysics*. https://doi.org/10.1007/s00024-002-8724-9

Jahnke, G., Igel, H., & Ben-Zion, Y. (2002). Three-dimensional calculations of fault-zone-guided waves in various irregular structures. *Geophysical Journal International*. https://doi.org/10.1046/j.1365-246X.2002.01784.x

Jia, Z., Wang, X., & Zhan, Z. (2020). Multifault models of the 2019 Ridgecrest sequence highlight complementary slip and fault junction instability. *Geophysical Research*

745   *Letters*. https://doi.org/10.1029/2020GL089802

746 Kennett, B. L. N., & Engdahl, E. R. (1991). Traveltimes for global earthquake location
747   and phase identification. *Geophysical Journal International*.
748   https://doi.org/10.1111/j.1365-246X.1991.tb06724.x

749 King, G. C. P. (1986). Speculations on the geometry of the initiation and termination
750   processes of earthquake rupture and its relation to morphology and geological
751   structure. *Pure and Applied Geophysics*. https://doi.org/10.1007/BF00877216

752 Kurzon, I., Vernon, F. L., Ben-Zion, Y., & Atkinson, G. (2014). Ground Motion
753   Prediction Equations in the San Jacinto Fault Zone: Significant Effects of Rupture
754   Directivity and Fault Zone Amplification. *Pure and Applied Geophysics*.
755   https://doi.org/10.1007/s00024-014-0855-2

756 Lee, E.-J., Chen, P., Jordan, T. H., Maechling, P. B., Denolle, M. A., & Beroza, G. C.
757   (2014). Full-3-D tomography for crustal structure in Southern California based on
758   the scattering-integral and the adjoint-wavefield methods. Journal of Geophysical
759   Research: Solid Earth. https://doi.org/10.1002/2014JB011346

760 Lewis, M. A., & Ben-Zion, Y. (2010). Diversity of fault zone damage and trapping
761   structures in the Parkfield section of the San Andreas Fault from comprehensive
762   analysis of near fault seismograms. *Geophysical Journal International*.
763   https://doi.org/10.1111/j.1365-246X.2010.04816.x

764 Lewis, M. A., Peng, Z., Ben-Zion, Y., & Vernon, F. L. (2005). Shallow seismic trapping
765   structure in the San Jacinto fault zone near Anza, California. *Geophysical Journal
766   International*. https://doi.org/10.1111/j.1365-246X.2005.02684.x

767 Li, Y. G., Leary, P., Aki, K., & Malin, P. (1990). Seismic trapped modes in the Oroville
768   and San Andreas fault zones. *Science*. https://doi.org/10.1126/science.249.4970.763

769 Li, Y. G., Aki, K., Adams, D., Hasemi, A., & Lee, W. H. K. (1994). Seismic guided
770   waves trapped in the fault zone of the Landers, California, earthquake of 1992.
771   *Journal of Geophysical Research*. https://doi.org/10.1029/94jb00464

772 Mamada, Y., Kuwahara, Y., Ito, H., & Takenaka, H. (2004). Discontinuity of the
773   Mozumi-Sukenobu fault low-velocity zone, central Japan, inferred from 3-D finite-
774   difference simulation of fault zone waves excited by explosive sources.
775   *Tectonophysics*. https://doi.org/10.1016/j.tecto.2003.09.008

776 Mitchell, T. M., Ben-Zion, Y., & Shimamoto, T. (2011). Pulverized fault rocks and
777   damage asymmetry along the Arima-Takatsuki Tectonic Line, Japan. *Earth and
778   Planetary Science Letters*. https://doi.org/10.1016/j.epsl.2011.04.023

779 Mizuno, T., Kuwahara, Y., Ito, H., & Nishigami, K. (2008). Spatial variations in fault-
780   zone structure along the Nojima fault, central Japan, as inferred from borehole
781   observations of fault-zone trapped waves. *Bulletin of the Seismological Society of
782   America*. https://doi.org/10.1785/0120060247

783 Ozakin, Y., Ben-Zion, Y., Aktar, M., Karabulut, H., & Peng, Z. (2012). Velocity contrast
784   across the 1944 rupture zone of the North Anatolian fault east of Ismetpasa from
785   analysis of teleseismic arrivals. *Geophysical Research Letters*.
786   https://doi.org/10.1029/2012GL051426

787 Park, S., Tsai, V. C., & Ishii, M. (2019). Frequency-Dependent P Wave Polarization and
788   Its Subwavelength Near-Surface Depth Sensitivity. Geophysical Research Letters.
789   https://doi.org/10.1029/2019GL084892

790 Peng, Z., Ben-Zion, Y., Michael, A. J., & Zhu, L. (2003). Quantitative analysis of seismic

791   fault zone waves in the rupture zone of the 1992 Landers, California, earthquake:
792   Evidence for a shallow trapping structure. *Geophysical Journal International*.
793   https://doi.org/10.1111/j.1365-246X.2003.02109.x

794 Qin, L., Ben-Zion, Y., Qiu, H., Share, P. E., Ross, Z. E., & Vernon, F. L. (2018). Internal
795   structure of the san jacinto fault zone in the trifurcation area southeast of anza,
796   california, from data of dense seismic arrays. *Geophysical Journal International*,
797   *213*(1), 98–114. https://doi.org/10.1093/gji/ggx540

798 Qiu, H., Ben-Zion, Y., Ross, Z. E., Share, P. E., & Vernon, F. L. (2017). Internal
799   structure of the San Jacinto fault zone at Jackass Flat from data recorded by a dense
800   linear array. *Geophysical Journal International*, *209*(3), 1369–1388.
801   https://doi.org/10.1093/gji/ggx096

802 Ross, Z. E., Idini, B., Jia, Z., Stephenson, O. L., Zhong, M., Wang, X., et al. (2019).
803   Hierarchical interlocked orthogonal faulting in the 2019 Ridgecrest earthquake
804   sequence. Science. https://doi.org/10.1126/science.aaz0109

805 Rovelli, A., Caserta, A., Marra, F., & Ruggiero, V. (2002). Can seismic waves be trapped
806   inside an inactive fault zone? The case study of Nocera Umbra, Central Italy.
807   *Bulletin of the Seismological Society of America*.
808   https://doi.org/10.1785/0120010288

809 Share, P. E., Ben-Zion, Y., Ross, Z. E., Qiu, H., & Vernon, F. L. (2017). Internal
810   structure of the San Jacinto fault zone at Blackburn Saddle from seismic data of a
811   linear array. *Geophysical Journal International*, *210*(2), 819–832.
812   https://doi.org/10.1093/gji/ggx191

813 Share, P. E., Allam, A. A., Ben-Zion, Y., Lin, F.-C., & Vernon, F. L. (2019). Structural
814   Properties of the San Jacinto Fault Zone at Blackburn Saddle from Seismic Data of a
815   Dense Linear Array. *Pure and Applied Geophysics*, *176*(3), 1169–1191.
816   https://doi.org/10.1007/s00024-018-1988-5

817 Shaw, J. H., Plesch, A., Tape, C., Suess, M. P., Jordan, T. H., Ely, G., et al. (2015).
818   Unified Structural Representation of the southern California crust and upper mantle.
819   Earth and Planetary Science Letters. https://doi.org/10.1016/j.epsl.2015.01.016

820 Shlomai, H., & Fineberg, J. (2016). The structure of slip-pulses and supershear ruptures
821   driving slip in bimaterial friction. *Nature Communications*.
822   https://doi.org/10.1038/ncomms11787

823 Spudich, P., & Olsen, K. B. (2001). Fault zone amplified waves as a possible seismic
824   hazard along the Calaveras fault in central California. *Geophysical Research Letters*.
825   https://doi.org/10.1029/2000GL011902

826 Thakur, P., Huang, Y., & Kaneko, Y. (2020). Effects of Low-Velocity Fault Damage
827   Zones on Long-Term Earthquake Behaviors on Mature Strike-Slip Faults. *Journal of*
828   *Geophysical Research: Solid Earth*. https://doi.org/10.1029/2020JB019587

829 Vidale, J., Helmberger, D. V., & Clayton, R. W. (1985). Finite-difference seismograms
830   for SH waves. *Bulletin of the Seismological Society of America*.

831 Wang, Y., Allam, A. A., & Lin, F.-C. (2019). Imaging the Fault Damage Zone of the San
832   Jacinto Fault Near Anza With Ambient Noise Tomography Using a Dense Nodal
833   Array. *Geophysical Research Letters*. https://doi.org/10.1029/2019GL084835

834 Weertman, J. (1980). Unstable slippage across a fault that separates elastic media of
835   different elastic constants. *Journal of Geophysical Research*, *85*(B3), 1455–1461.
836   https://doi.org/10.1029/JB085iB03p01455

837 White, M. C., Fang, H., Catchings, R. D., Goldman, M. R., Steidl, J. H., & Ben-Zion, Y.
838   (2020, 07). Detailed traveltime tomography and seismicity around the 2019 M7.1
839   Ridgecrest, CA, earthquake using dense rapid-response seismic data. Poster
840   Presentation at 2020 SCEC Annual Meeting.
841 Xu, S., Ben-Zion, Y., & Ampuero, J. P. (2012). Properties of inelastic yielding zones
842   generated by in-plane dynamic ruptures-II. Detailed parameter-space study.
843   *Geophysical   Journal   International*.   https://doi.org/10.1111/j.1365-
844   246X.2012.05685.x
845 Xu, X., Sandwell, D. T., & Smith-Konter, B. (2020). Coseismic displacements and
846   surface fractures from sentinel-1 InSAR: 2019 Ridgecrest earthquakes.
847   *Seismological   Research   Letters*,   *91*(4),   1979–1985.
848   https://doi.org/10.1785/0220190275
849 Zhu, L., & Kanamori, H. (2000). Moho depth variation in southern California from
850   teleseismic receiver functions. *Journal of Geophysical Research: Solid Earth*,
851   *105*(B2), 2969–2980. https://doi.org/10.1029/1999JB900322

852   Figure 1. Location map for the 2019 Ridgecrest earthquake sequence (colored

853   circles, square, and stars) and four linear arrays (B1, B2, B3, and B4 as red, green,

854   blue, and purple triangles, respectively) analyzed in this study. The catalog of

855   Hauksson et al. (2012, extended to 2019) is used for earthquake locations, with color

856   representing the focal depth (colorbar). Epicenters of the 2019 Mw 6.4 and Mw 7.1

857   Ridgecrest earthquakes are marked as stars. Bandpass filtered waveforms, fault zone

858   head waves, and fault zone trapped waves of an example event (orange square)

859   recorded at array B4 are shown in Figs. S12, S13a, and 7a. Fault surface traces are

860   depicted as black lines with ruptures of the 2019 Ridgecrest earthquake sequence

861   being thicker. Near-fault seismic events outlined by the red box are analyzed in

862   sections 3.2 and 3.3. The background gray colors indicate the local topography.

863   WLSZ – Walker Lane Shear Zone; ECSZ – Eastern California Shear Zone; EF –

864   Elsinore Fault; GF – Garlock Fault; SAF – San Andreas Fault; SJF – San Jacinto

865   Fault.

866   Figure 2. Zoomed-in maps of all the Ridgecrest linear arrays (triangles) in (a) and

867   B1-B4 in (b)-(e), respectively. Color of the map background depicts the surface

868   geology (i.e., rock types; legend in (a)) extracted from the USGS website (see Data

869   Availability). The station elevation is illustrated by the color of the triangles in (b)-

870   (e). The green circle and black lines denote the center of the array and surface traces

871   of faults associated with the 2019 Mw 7.1 and Mw 6.4 Ridgecrest earthquakes (red

872   and blue stars), respectively. The green bars outline the span of the low-velocity

873   zones identified in the local *P*-wave delay time analysis (Section 3.2), whereas the red

874   bars illustrate the region of the slower inner cores (Fig. 6). The blue arrows depict the

875   extent of the transition zone pointing from the slower to faster crustal blocks (black

876   solid curves in Fig. 6).

877   Figure 3. (a) Locations of array B4 (triangle) and four teleseismic events analyzed

878   in this study (stars), with the red star indicating the event analyzed in (b)-(c). (b) *P*

879   waves recorded on vertical-component sensors of array B4 for the target teleseismic

880   event are illustrated in colors with red and blue indicating positive and negative

881   values. The *P*-wave arrival time predicted from the model IASP91 is used to align the

882   *P* waveforms and is set to be zero in the time axis. The *P* waveforms are bandpass

883   filtered twice. After applying a bandpass filter between 0.5 Hz and 2 Hz, the array-

884   mean envelope function and an array-mean *P*-wave pick are computed and depicted

885   as the curve and the vertical solid line in black, respectively. (c) Amplitude spectrum

886   averaged over the entire array with the red star and horizontal dashed lines indicating

887   the peak frequency and median of the amplitude spectrum between 0.5 Hz and 2 Hz,

888   respectively. Then, a second bandpass filter between the frequency range outlined by

889   the vertical dashed lines in (c) is applied. The red dashed curve in (b) depicts the

890   teleseismic *P*-wave delay times, measured using the *P* waveforms tapered between

891   the vertical dashed lines (± one dominant period relative to the array-mean *P*-wave

892   pick).

893   Figure 4. Teleseismic *P*-wave delay times for arrays (a) B1, (b) B2, (c) B3, and

894   (d) B4. The colored stars indicate *P*-wave delay times measured from different

895   teleseismic events and are labeled in the legend by the corresponding peak frequency

896   of the array-mean *P*-wave amplitude spectrum. The black dots depict the delay-time

897   pattern averaged over all teleseismic events, with error bars representing the standard

898   deviation of the mean. The blue and red dashed curves illustrate the delay times after

899   a topographic correction, assuming constant *P*-wave velocities of 2 km/s and 4 km/s.

900   The green bars outline the span of the low-velocity zones identified in local *P*-wave

901   delay time analysis (Section 3.2), while the red bars illustrate the region of the slower

902   inner cores (Fig. 6).

903   Figure 5. (a) *P* waveforms of an example local seismic event, shown as the blue

904   star in (c), recorded on vertical-component sensors of array B1. Waveform at each

905   station is normalized by its corresponding maximum amplitude and bandpass filtered

906   between 0.5 Hz and 20 Hz. Red stars denote the automatic *P* picks. (b) Same as (a)

907   but shown in slowness domain, i.e., the time axis of each station is divided by the

908   corresponding hypocentral distance. (c) Distribution of seismic events that pass the

909   selection criteria (Section 3.2.1) for array B1 (red triangles) are shown as circles with

910   color representing the focal depth. Events used in the delay time analysis (Section

911   3.2) for array B1 are outlined by the red box (Area1) and the results are shown in

912   Figs. 6a-b. Data from events outside the red box (Area2 and Area3) are also analyzed

913   (Figs. S8c-d). The black lines represent fault surface traces. Earthquakes that are

914   excluded from the delay-time analysis at array B1 are shown as grey dots.

915   Figure 6. Statistical analysis of local *P*-wave arrival patterns for all four arrays.

916   (a) Red dots illustrate the *P*-wave relative slowness variation within array B1,

917   averaged over 189 local seismic events with depth > 5 km and along-fault distance <

918   5 km relative to the array. The error bars represent a range of three standard

919   deviations of the mean about each respective mean value. The solid black lines depict

920   the contribution associated with a *P*-wave velocity contrast (~5.3% faster in the NE)

921   across the fault beneath the array. The corresponding mean *P*-wave delay time

922   variation is illustrated as the blue curve with dashed blue curves indicating the

923   uncertainties. (b) The *P*-wave delay time variation after removing the modeled cross-

924   fault velocity contrast (the solid black curve in (a)) is shown in blue for array B1. The

925   green dashed vertical lines outline the identified low velocity zone (LVZ), where *P*

926   waves are significantly delayed with respect to the background level (black dashed

927   line). The red dashed vertical lines denote the inner core of the LVZ, i.e., the

928   amplitude of the blue curve is large whereas its gradient is relatively small. The gray-

929   scale heat map of values for relative slowness and delay times after the cross-fault-

930   velocity-contrast correction obtained at each station for all analyzed events is

931   illustrated in (a) and (b), respectively. PDF – Probability Density Function. Results

932   for arrays B2, B3, and B4 are shown in (c)-(d), (e)-(f), and (g)-(h), respectively.

933   Figure 7. (a) Fault zone trapped waves (FZTWs) following the *S*-wave arrivals for

934   an example event (square in Fig. 1) observed at the fault-parallel component of array

935   B4. The waveforms are preprocessed following the steps of Figure S6 of Qiu et al.

936   (2017). The blue bar outlines the stations with FZTWs. (b) Red dots and blue stars

937   denote the distributions of normalized peak ground velocities (PGV) and root mean

938   square amplitudes (RMS) of the *S* waveforms shown in (a). The black curve

939   represents the likelihood of FZTW, i.e., the normalized multiplication of PGV and

940   RMS values, and is used to identify FZTW candidates.

941   Figure 8. *S*-wave amplification analysis for arrays B1-B4 in (a)-(d). Gray-scale
942   heat maps of FZTW-likelihood values (e.g., black curve in Fig. 7b) computed at each
943   station in the array for all analyzed events are shown. Red dots depict the likelihood
944   curve averaged over all analyzed events. Error bars represent a range of two standard
945   deviations of the mean about each respective mean value. The bottom green bar
946   marks the low velocity zone (LVZ) identified from the local *P*-wave delay time
947   analysis (green dashed lines in Fig. 6), whereas the red bar illustrates the slower inner
948   core of the LVZ (red dashed lines in Fig. 6). The blue bar in (d) outlines the stations
949   with FZTWs at array B4 (blue bar in Fig. 7a).

950   Figure 9. (a) Locations of earthquakes (gray dots) analyzed in section 3.3. FZTW
951   candidates identified through template matching, with correlation coefficient greater
952   than 0.85 marked as stars and color representing the focal depth. Red triangles denote
953   the location of array B4. The along-fault cross section of seismicity (dots and stars)
954   and array B4 (triangle) are shown in the top inset. (b) FZTWs recorded at stations
955   B416-B423 for nine high-quality candidate events (red) with the highest correlation
956   coefficients. The template waveforms (Fig. 7a) are shown in black. The array-mean *S*
957   pick and correlation coefficient of each candidate event are labeled in the top left. (c)
958   Comparison between FZTWs of the reference event (in black) and those averaged
959   over all the high-quality candidate events (in red) observed between stations B416-
960   B423.

961   Figure 10. Inversion results for FZTWs observed between stations B416-423,
962   averaged over candidates shown in Figure 9a. (a) Comparison between synthetic
963   waveforms (red) computed using the best-fitting model parameters (black dots in (b))
964   and the observed FZTWs (in black). (b) Fitness values of fault-zone model
965   parameters from the last 10 generations of the inversion (green dots). The best-fitting
966   parameters (black circles) are displayed in each panel and used to generate the
967   synthetic waveforms shown in (a). Black curve indicates the probability density
968   function of model parameters shown as green dots.
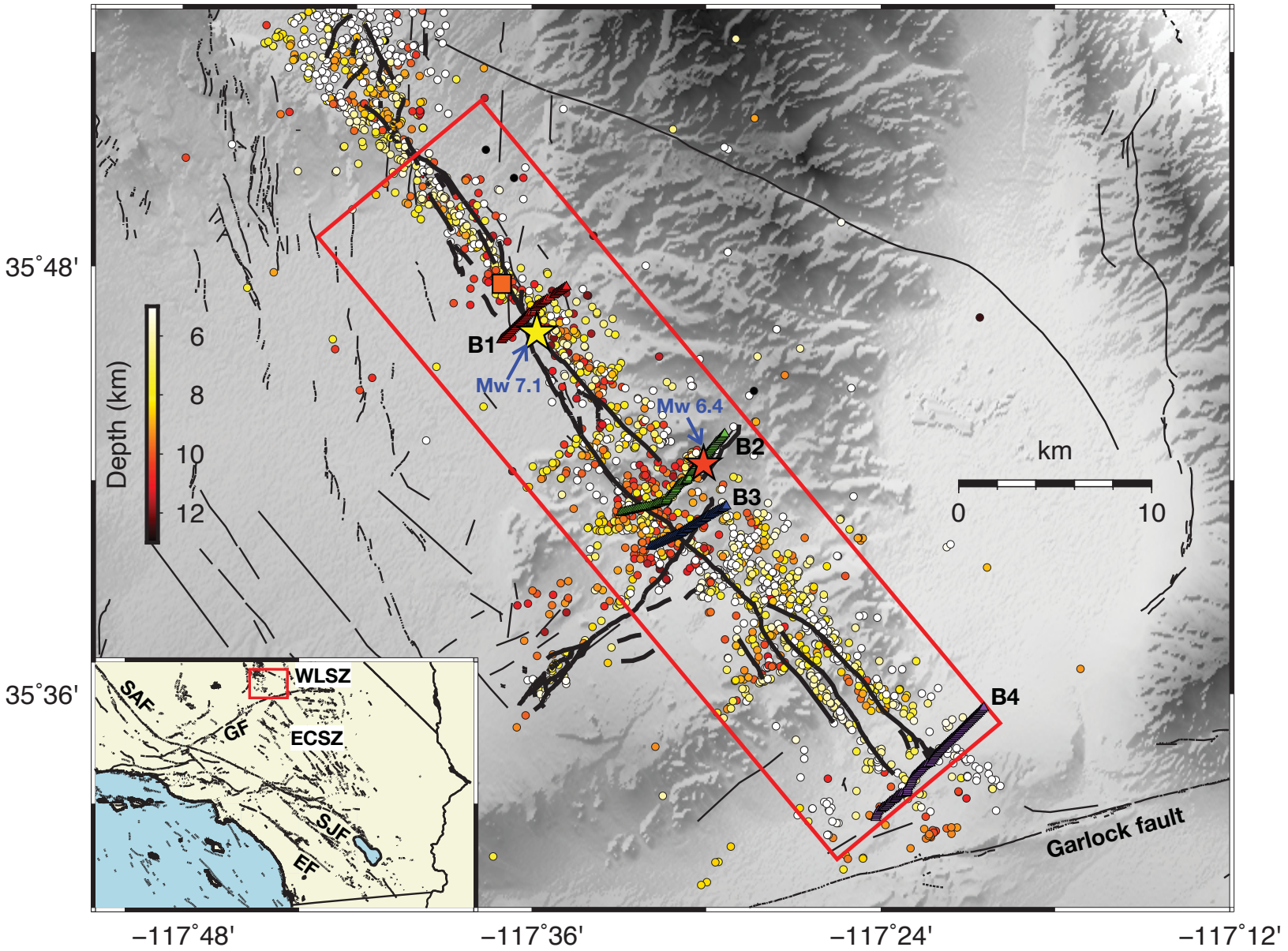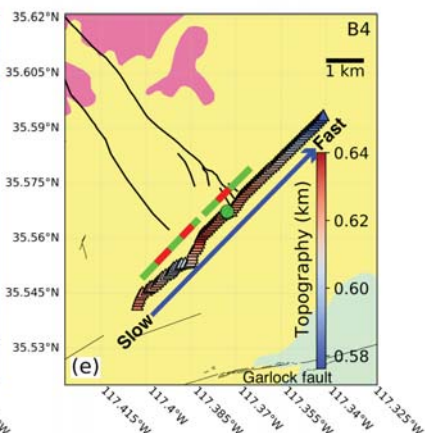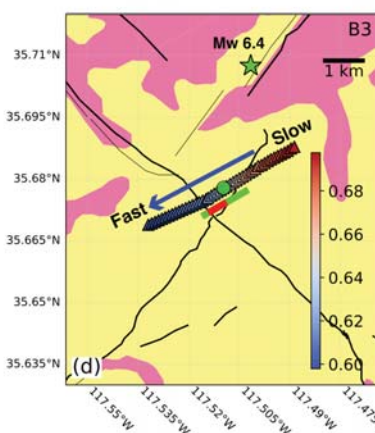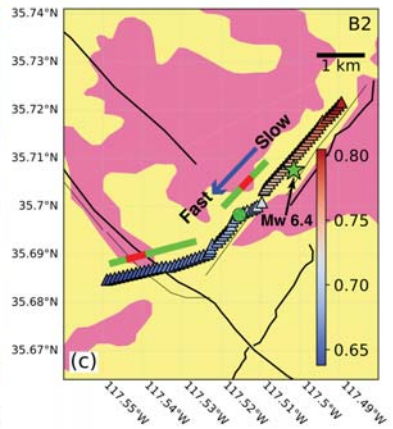
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Figure 1.

Depth (km)
6
8
10
12

35°48'
35°36'

WLSZ
SAF
GF
ECSZ
SJF
EF

B1
Mw 7.1
Mw 6.4
B2
B3
B4

Garlock fault

km
0   10

−117°48'   −117°36'   −117°24'   −117°12'

Figure 2.

Predominant rock type
- Granodiorite
- Schist
- Water
- Sandstone
- Conglomerate
- Alluvium

LVZ
Inner core

Topography (km)

Figure 3.

(a)

(b)

(c)

Figure 4.

Figure 5.

(a)

(b)

(c)

NE

SW

Distance to array center (km)

Time to origin (s)

Slowness (s/km)

Area1

Area2

Area3

B1

Example event

Depth (km)

Figure 6.

Figure 7.

Fault parallel component

FZTW detector

NE

SW

(a)

(b)

Distance to array center (km)

Time (s)

Normalized amplitude

- PGV
- RMS
- PGV x RMS

Figure 8.

Figure 9.

Figure 10.

(a) Waveform comparison showing observed (Obs, black) and synthetic (Syn, red) traces for stations B416–B423, plotted against Time (s).

(b) Fitness and probability density distributions for inverted parameters: FZ shear velocity (%), QS_SW shear velocity (km/s), FZ width (km), FZ Q, Propagation distance in FZ layer (km), and QS_NE shear velocity (%).