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## 1 Highlights

## <sup>2</sup> Distribution of seismic scatterers in the San Jacinto Fault Zone, southeast of

### <sup>3</sup> Anza, California, based on passive matrix imaging

- <sup>4</sup> Rita Touma, Alexandre Aubry, Yehuda Ben-Zion, Michel Campillo
- A matrix approach allows resolving small-scale heterogeneities such as cracks
- 6 and fractures with an optimal resolution
- Scatterers and reflectors are imaged at the SGB site on the San Jacinto Fault
- 8 Zone, California
- Clear differences are found among different sections of the fault zone in the
   study area

# Distribution of seismic scatterers in the San Jacinto Fault Zone, southeast of Anza, California, based on passive matrix imaging

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#### 15 Abstract

Fault zones are associated with multi-scale heterogeneities of rock properties. Large scale variations may be imaged with conventional seismic reflection methods that detect offsets in geological units, and tomographic techniques that provide average seismic velocities in resolved volumes. However, characterizing elementary localized inhomogeneities of fault zones, such as cracks and fractures, constitutes a challenge for conventional techniques. Resolving these small-scale heterogeneities can provide detailed information for structural and mechanical models of fault zones. Recently, the reflection matrix approach utilizing body wave reflections in ambient noise cross-correlations was extended with the introduction of aberration corrections to handle the actual lateral velocity variations in the fault zone Touma et al. (2021). Here this method is applied further to analyze the distribution of scatterers in the first few kilometers of the crust in the San Jacinto Fault Zone at the Sage Brush Flat (SGB) site, southeast of Anza, California. The matrix approach allows us to image not only specular reflectors but also to resolve the presence, location and intensity of scatterers of seismic waves starting with

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a simple homogeneous background velocity model of the medium. The derived three-dimensional image of the fault zone resolves lateral variations of scattering properties in the region within and around the surface fault traces, as well as differences between the Northwest (NW) and the Southeast (SE) parts of the study area. A localized intense damage zone at depth is observed in the SE section, suggesting that a geometrical complexity of the fault zone at depth induces ongoing generation of rock damage.

<sup>16</sup> *Keywords:* Scattering, Fault zone structure, Passive imaging, Reflection matrix

#### 17 1. Introduction

Earthquakes are among the most destructive natural disasters. Although earth-18 quakes are generally unpredictable, some aspects of their behavior such as the 19 likelihood of being arrested and statistically-preferred propagation direction can 20 be estimated from structural properties of fault zones (see e.g. Chester et al. 21 (1993); Wesnousky (1988); Ben-Zion (2008)). Fault zones that are the structural 22 manifestation of earthquakes evolve during deformation and have generally com-23 plex properties (e.g. Mitchell and Faulkner (2009); Ben-Zion and Sammis (2003)). 24 Characterizing the geometrical and seismic properties of fault zones can provide 25 important information for assessing likely past and future rupture properties. Fault 26 zones have also strong impact on fluid flow in the lithosphere (Knipe et al., 1998). 27 Fault zones are manifested at the surface by several main fault traces that 28 accommodate the bulk of the long term slip. They are characterized by linea-29 ments, topography and various geometrical complexities. The identification of 30 fault traces is done by field observations (major line of fracturing, offset in ge-31 ological units), along with remote sensing techniques that analyze ground de-32

formation after major earthquakes obtained from satellites and aircrafts such as Interferometric Synthetic Aperture Radar (InSAR, Massonnet et al. (1993)) and subpixel correlation of optical images (SPOT, Binet and Bollinger (2005)). Fault zone properties below the surface are obtained by seismic and other geophysical imaging techniques.

Fault zones have hierarchical damage structures that evolve during the fault 38 zone activity and have several general elements (e.g. Chester et al. (1993); Rock-39 well and Ben-Zion (2007)). The principle slip zone is a highly localized thin layer 40 (0.01 - 0.1 m thick) that accommodates most of the fault slip and is characterized 41 by ultra cataclasite rock particles. The principle slip zone is bounded by a core 42 damage zone (inner damage zone) that is typically about 100 m wide and asym-43 metrically located on one side of the slip zone of large faults (Lewis et al., 2005; 44 Dor et al., 2006). The core damage is surrounded by a broader zone of reduced 45 damage intensity (referred to as outer damage zone) that may extend for several 46 km on each side of the fault. Properties of the fault zone damage provide infor-47 mation on statistical tendencies of local earthquake ruptures, operating dynamic 48 stress field, energy dissipation and more (e.g. Manighetti et al. (2005); Mitchell 40 and Faulkner (2009); Xu et al. (2012)). For that reason, a number of seismic and 50 other methods have been developed to provide detailed information on fault zone 51 structures. 52

Among the seismic imaging techniques, reflection seismology is generally pertinent to image planar horizontal layers with a very high resolution, and provide indirect imaging of faults by the offset of sedimentary layers. It relies on the analysis of seismic waves that are sent back towards the surface after being reflected or scattered by subsurface structures with strong impedance contrasts. The

recorded wavefield is composed of reflected waves that result from the interaction 58 of seismic waves with planar reflectors such as layer boundaries, and diffracted 59 waves from small-scale geological objects such as cracks and fractures. To obtain 60 structural information on the subsurface, migration techniques are applied aiming 61 mainly to relocate reflectors and scatterers in depth or in time (see, e.g., review by 62 Etgen et al., 2009). Migration of seismic wavefields requires an accurate velocity 63 model of the Earth. Errors and biases in the velocity model produce artefacts and 64 defocusing due to phase distortions in the migrated images (Moser and Howard, 65 2008). 66

Reflection information is often gathered from seismic surveys where seis-67 mic energy propagating into the medium is generated by man-made sources such 68 as vibrators, explosives, etc. In the last decade or so, passive methods based 69 on the ambient seismic noise have been developed to substitute active imaging 70 techniques (e.g. Campillo and Roux, 2014). Cross-correlation of passive traces 71 recorded at two receivers allows retrieving the Green's function between these 72 two receivers. In other words, the resulting correlation is comparable to the seis-73 mogram that would be obtained at the first receiver if there is a source located at 74 the second receiver's location. This approach is referred to as seismic interferom-75 etry (Wapenaar et al., 2010). Noise cross-correlations are often used to retrieve the 76 surface waves component of the Green's function (Shapiro and Campillo, 2004, 77 and later works). Extracting body waves contributions to the Green's function 78 is more difficult (Poli et al., 2012a). However, it has been shown that ambi-79 ent noise cross-correlations can be used to image deep targets inside the Earth 80 with body wave reflections (Draganov et al., 2007; Poli et al., 2012b). Inspired 81 by works done in ultrasound imaging (Aubry and Derode, 2009) and optical mi-82

croscopy (Badon et al., 2016), a "reflection matrix approach" was introduced to 83 geophysics and used body wave reflection from coda-wave cross-correlations to 84 image the complex medium below Erebus volcano - Antarctica (Blondel et al., 85 2018). The "reflection matrix approach" allows to image both specular reflectors, 86 corresponding to the boundaries between layers of different propagation veloc-87 ities, and non-specular reflections generated by a distribution of localized inho-88 mogeneities. The matrix imaging approach has then been developed to overcome 89 phase distortions for multi-layered media (Lambert et al., 2020a) and strongly 90 heterogeneous media (Badon et al., 2020; Lambert et al., 2021a). Using the same 91 matrix formalism, Touma et al. (2021) analysed ambient noise recorded at a dense 92 array (1108 vertical geophones in 600 m  $\times$  700 m configuration) to image subsur-93 face properties of the San Jacinto Fault zone (SJFZ) southeast of Anza, California. 94 While Blondel et al. (2018) dealt with imaging problems in the multiple scatter-95 ing regime in the case of volcanoes, the main challenge in fault zones imaging is 96 the presence of phase distortions, also referred to as aberrations, due to the strong 97 structural heterogeneities within and around the fault zones. 98

The present paper follows the approach of Touma et al. (2021) to derive more 90 detailed 3D images of the SJFZ at depth and to resolve the shallow structure with 100 unprecedented resolution. Touma et al. (2021) computed noise cross-correlations 101 in the (10-20) Hz frequency range. Whereas the cross-correlations provide a re-102 sponse matrix between sources and receivers located at the surface, the reflec-103 tion matrix contains the response between sources and receivers that are virtually 104 moved inside the medium by performing focusing operations. This process is gen-105 erally known as redatuming (Berkhout and Wapenaar, 1993) and it allows local 106 information on the medium's reflectivity to be retrieved. To project the data in the 107

virtual focused basis, a homogeneous transmission matrix is used with a constant 108 velocity model of 1500 m/s. This velocity was chosen to optimize the focusing 109 as discussed in Touma et al. (2021). The main advantage of this method is that 110 it only requires an approximate estimation of the medium's velocity. Although 111 an incorrect velocity model produces phase distortions (aberrations) in the propa-112 gated data, the reflection matrix approach allows us to account and correct these 113 aberrations through the distortion matrix concept (Badon et al., 2020; Lambert 114 et al., 2020b, 2021b). From that matrix, the distorted component is extracted and 115 is used to virtually focus back waves inside the medium (see Appendix A). As 116 a result of the correction process, the resolution of the final subsurface images is 117 drastically improved and a three-dimensional image of the subsurface reflectivity 118 is revealed. The derived images represent reflectivity maps of the medium beneath 119 the Clark branch of the SJFZ at the Sage Brush Flat site (Ben-Zion et al., 2015; 120 Roux et al., 2016). The site under study is located in the complex trifurcation area 121 southeast of Anza, California (Fig. 1a). The locations of the surface fault traces 122 are derived from recent detailed studies of the surface geological mapping and 123 shallow geophysical imaging (Wade, 2018; Share et al., 2020). The basic goal of 124 this paper is to interpret the obtained scattering images that exhibit features with 125 higher lateral resolution than conventional seismic investigations. The variability 126 and attenuation of scattered intensity within and around the major fault zone are 127 also discussed. 128

#### 129 2. The San Jacinto Fault Zone

The 230 km-long San Jacinto Fault Zone is the most seismically active fault zone in southern California (Hauksson et al., 2012) and is one of several major



Figure 1: (a) Topographic map of Trifurcation area of San Jacinto fault zone. The red square marks the studied area. (b) Map of the geophones of the dense array at SGB site (black dots). The Clark fault traces are represented by the red<sup>7</sup>lines. The blue lines indicate the locations of the cross-sections represented in Fig. 2 and 3.

а

right-lateral strike-slip fault zones over which the North American-Pacific plate 132 boundary is distributed in southern California. The SJFZ branches from the San 133 Andreas fault at Cajon Pass and was formed 1-2 million years ago, presumably 134 in response to the geometrical complexities on the San Andreas Fault in the trans-135 verse ranges (e.g. Matti and Morton (1993)). The SJFZ represents a less mature 136 evolutionary stage in the life of a large continental strike-slip structure than the 137 San Andreas fault. Approximately 24 km of slip has been accommodated by the 138 SJFZ since the latest Pliocene to early Pleistocene (Dorsey and Roering, 2006), 139 with estimated slip rates that vary along strike between 8 - 20 mm/yr (Rockwell 140 et al., 2003; Fialko, 2006). The SJFZ has varying surface complexity and seismic-141 ity along its strike. The Anza section to the northwest of the SGB site consists of 142 a single strand, the Clark fault, with relatively regular geometry and low current 143 background microseismicity (Sanders and Kanamori, 1984). The trifurcation area 144 of the SJFZ where the SGB site is located (Fig. 1) is associated with branching 145 of the Clark fault in the Anza section into three major faults: a continuation of 146 the Clark fault and the Buck Ridge and Coyote Creek faults. The Trifurcation 147 area has a broad zone of high seismicity rates that include five earthquakes with 148 magnitudes around 5 since 2001. The geometrical properties of the seismicity in 149 the trifurcation area are very complex and consist of a diffuse pattern in the top 5 150 km that changes to more localized structures dipping to the NE below 6 km, along 151 with zones of seismicity that are orthogonal to the main strike of the SJFZ (Ross 152 et al., 2017). 153

#### **3. 3D scattering volume**

Migration techniques are known to be powerful tools for imaging strong re-155 flecting boundaries. These boundaries are identified by discontinuities of acoustic 156 impedance in the subsurface and are characterized by specular returns in seis-157 mic records. Less interest has been accorded to the non-specular component that 158 arises from small-scale geological objects (Khaidukov et al., 2004). The energy 159 generated by such small objects is commonly referred to as diffractions. Non-160 specular energy holds valuable information on the local heterogeneities in the 161 medium (Schwarz, 2019, and references therein). While many conventional mi-162 gration techniques, such as Gaussian beam migration and reverse-time migration, 163 honor the diffracted component, such contributions are generally difficult to ana-164 lyze: they are often suppressed due to the processing done in conventional seismic 165 methods or masked by specular reflections whose amplitudes are much larger than 166 the scattered components (Kozlov et al., 2004). They can also be considered as 167 noise in several migration techniques. Keeping the non-specular component in 168 the analysis allows retrieving signatures of localized scatterers such as cracks or 169 inclusions that lack lateral continuity. Imaging such features whose size is of the 170 order and even smaller than the seismic wavelength contributes significantly to 171 seismic interpretation (Schwarz and Krawczyk, 2020). 172

Fault surfaces and zones with increased fractures density are non-specular objects for surface sensors (Kanasewich and Phadke, 1988). Imaging such features is a challenge in most conventional seismic exploration surveys. Several studies have discussed the necessity of distinguishing between diffractions and reflections, and provided techniques to separate them (Khaidukov et al., 2004; Moser and Howard, 2008; Bakhtiari Rad et al., 2018). Diffraction imaging is performed <sup>179</sup> usually by suppressing the specular reflections so that the migrated image contains the diffracting component that have been isolated. The reflection matrix technique allows us to image, without performing any prior filtering, not only specular reflectors but mainly non-specular backscattered energy that directly gives insight into rock properties at the subsurface. The contribution of specular and non-specular features are distinguishable in the derived images.

The constant velocity model of 1500 m/s chosen for the redatuming operation 185 in the initial study of Touma et al. (2021) is highly approximate especially for the 186 deep structure. It was chosen to optimize the focusing at depth with an "apparent" 187 velocity that increases the effective aperture of the geophone array to exploit the 188 actual contribution of multi-scattered paths as discussed in Touma et al. (2021). 189 We recall that a higher velocity will only stretch the detected features vertically. 190 The images are presented as a function of depth for the chosen background veloc-191 ity and two-way travel time. 192

In the subsequent subsections, we provide a detailed description of cross-193 sections taken from the 3D scattering volume. Those images are obtained after 194 correcting the aberrations induced by the mismatch between the velocity model 195 used to perform the focusing operation and the data. Although the images shown 196 in Fig. 2 and Fig. 3 will lead to enlightening interpretations, they should be also 197 taken with caution. First, these images have been corrected from transverse aber-198 rations but axial aberrations can subsist. To cope with this issue, a more accurate 199 2D velocity model can be implemented in the future. Secondly, the aberration 200 correction process applied to these images only identify two isoplanatic patches 201 at each depth. This is probably a perfectible point in such complex heterogeneous 202 media with strong lateral velocity variations. Other less reflective features belong-203

ing to other isoplanatic patches are thus possibly not revealed by our method inFig. 2 and 3.

A compensation method for attenuation, described in Appendix C, is applied to the reflectivity maps. The results are plotted on a Cartesian grid, with the origin located at the center of the array and the x-axis orthogonal to the fault traces. The colors represent the backscattered intensity plotted in dB. The local distribution of heterogeneities and discontinuities of material boundaries are revealed with maximal focused intensity. We first present results associated with shallow materials and then discuss deeper structures.

We compared our results with previous studies investigating the same region. Most of the studies conducted at the SGB site utilize surface waves to resolve shallow features and the velocity distribution below the dense array (Roux et al., 2016; Hillers et al., 2016; Mordret et al., 2019) or fault zone head waves to image lateral variations of lithology and damage across the fault (Ben-Zion et al., 2015; Qin et al., 2018). To our knowledge, the deep scattering structure has not been examined and analyzed yet.

#### 220 3.1. Images of the shallow fault zone

The high frequency cross-correlations of noise recorded by the dense array 221 allow resolving features in the top few hundred meters of the crust. Fig. 2 shows 222 vertical slices of the 3D volume with a close-up view of the first 400 m: two slices 223 perpendicular to the fault (AA', BB') and two slices parallel to the fault traces 224 (CC', DD'). For the sake of simplicity, the cross-sections are labeled North, South, 225 West and East, respectively. The slices are plotted in logarithmic dB scale and 226 reveal the backscattered intensity rising from highly reflective features detected 227 through the aberration correction process. The location of the each cross-section 228



Figure 2: Shallow cross-sections of the 3D scattering volume. Vertical slices oriented perpendicular to the fault traces. North/South denote Northwest and Southeast. The main fault strands are represented by the bold white lines. (a) Profile A-A', (b) Profile B-B'. Vertical slices oriented parallel to the fault traces. West/East denote Southwest and Northeast. (c) Profile C-C', (d) Profile D-D'. The color scale is in dB. The white dashed lines correspond to the spatial extension of the array.

is indicated by blue lines in Fig. 1b. In the first two vertical slices, the white lines
refer to the location of the three main sub-parallel strands represented in red in
Fig. 1b.

Several differences between the results in the slices oriented differently stand 232 up. The first thing to notice are the structural variations across the fault zone 233 in Figs. 2a and b. We observe high intensity of scatterers within the core fault 234 damage zone and reduced scatterers intensity outside. A clear offset of reflective 235 structures is observed around a depth of 150 m in the two perpendicular panels. 236 Figs. 2a and b share the same features although the scattering appears stronger 237 and more extended in the southern profile (B-B'). The intensity below 150 m de-238 creases in the North cross-section while in the South cross-section a high density 239 of scatterers extents to greater depth revealing a localized damage zone around the 240 fault traces (white lines). 241

The offset of the scatterers in Figs. 2a and b is observed mainly below the SW fault trace. Qin et al. (2018) suggested that the SW fault trace is the main seismogenic fault separating two crustal blocks of different seismic properties. Mordret et al. (2019) also showed the presence of a velocity contrast across the SW fault trace. The observed offset in structural properties can be explained by the fact that the SW trace represents the main seismogenic fault.

Figs. 2c and d representing the West and East cross-sections show no clear lateral variations of the subsurface structures. The reflectivity is associated with planar features or layers located on each side of the fault. The high scattering zone to the NE of the surface trace, clearly observed in Fig 2d, is in general agreement with the trapping structure identified by Ben-Zion et al. (2015) and Qin et al. (2018). This zone is characterized by significant low velocities and an intense localized damage producing reflections (Roux et al., 2016; Hillers et al., 2016;
Mordret et al., 2019). The reflective layer SW of the fault observed in Fig 2c coincides with the local sedimentary basin reported by Ben-Zion et al. (2015), Roux
et al. (2016) and Hillers et al. (2016).

Many studies of the San Jacinto Fault Zone observed an asymmetric rock damage across the fault (Lewis et al., 2005; Dor et al., 2006; Qin et al., 2018; Wade, 2018). The damage at the SGB site was shown to be greater on the NE side of the fault. The scattering in Fig. 2a is more pronounced in the NE. Between 150 and 400 m, the scattering zone dips slightly to the NE and is comparable to the shape of the low velocity trough found by Mordret et al. (2019) beneath the fault trace at the same location as profile AA'.

#### 265 3.2. Deeper and larger scale structure

We now investigate deeper sections from the 3D volume. Fig. 3 shows four 266 vertical 4 km deep slices; two slices are oriented perpendicular to the fault, one 267 in the Northwest (EE') of the array and the other in the Southeast (FF'), and two 268 additional slices oriented parallel to the fault, one in the Southwest (GG') of the 269 array and the other in the Northeast (HH'). The locations of the cross-sections 270 are marked by blue lines in Fig. 1b. All vertical cross-sections are presented in 271 the form of animated movies in Animation D.1 (slices perpendicular to the fault 272 traces) and Animation .D.2 (slices parallel to the fault traces). 273

In general, reflectors can only be imaged over a transverse field-of-view limited by the size of the geophones' array. Indeed, in the case of planar interfaces, a part of the reflected wave-field is not captured by the geophone array for large angles of incidence (See supplementary Fig. B.1a). Yet, the 3D scattering images shown in Fig. 3 reveal features that extend well beyond the spatial extension of



Figure 3: Deep cross-sections of the 3D scattering volume. Vertical slices oriented perpendicular to the fault traces. (a) Profile E-E' Northwest (NW) of the array, (b) Profile F-F' Southeast (SE) of the array. Vertical slices oriented parallel to the fault traces. (c) Profile G-G' Southwest (SW), (d) Profile H-H' Northeast (NE). The color scale is in dB. The white dashed lines correspond to the spatial extension of the array.

the array. This observation is made possible due to the intense damage around the fault zone. In such diffusive media, the strong heterogeneities reflect the incident waves in many different directions, and therefore the imaged field-of-view is extended (See supplementary Fig. B.1b). Interfaces between geological layers are also imaged beyond the array extension by means of the waves scattered by the localized heterogeneities that reside at the discontinuity across interfaces.

The comparison between the first two panels (Figs. 3a and b) reveals a clear 285 difference in reflectivity between the Northwest (NW) and Southeast (SE) por-286 tions of the SGB site. Both panels show a broad scattering zone in the shallow 287 crust that has a V shape with about 800 m wide area at z = 500 m (t = 0.7 s) 288 (Figs. 3a and b). The dense distribution of scatterers in the shallow zone results 289 likely from the heavily damaged rocks around the fault traces. In the NW panel, 290 the diffuse damage is less apparent deeper than z = 1000 m (t = 1.3 s). However, 291 the observed back-scattered energy in that section is associated with horizontal 292 reflectors emerging on both sides of the fault. Discontinuous blocks on the right 293 and left side of the fault traces highlight the offset of geological features across the 294 fault. In contrast, the high intensity scattering zone extends deeper than z = 1000295 m (t = 1.3 s) in the SE slice. Around z = 1500 m (t = 2 s), the backscattered 296 intensity reveals a zone that is about 450 m wide. Deeper in the crust, scattering 297 seems to persist in combination with specular reflections arising from discontinu-298 ous deep layers (z = 3300 m, t = 4.4 s). 299

The different scattering zone extensions in Figs. 3a and b are consistent with a change in the nature and structural complexity of the fault zone in the study area. To the NW of the SGB site, the SJFZ occupies a linear valley, whereas, to the SE, it becomes more localized and is associated with a canyon (Sharp, 1967). Recent

geological mapping in the area (Wade 2018) shows multiple fault strands at the 304 SGB site (Fig. 1b). One main fault is mapped at the base of the NW boundary of 305 the SGB basin. To the SE along strike, that fault merges with two other faults and 306 results in a more localized zone that is associated with a higher damage intensity. 307 The reflectivity panels confirm this feature by showing a significant scattering at 308 depth SE of the array related to highly damaged (cracked and crushed) rocks. In 309 the first 2 km (t = 2.7 s), the fault appears to be more localized in the SE generat-310 ing an intense distribution of scatterers. Indeed, the scattered wavefield dominates 311 and the specular component is less apparent in the SE compared with the NW. 312 In the NW, the damage zone is more distributed and less intense. The damage 313 intensity is rapidly decaying with depth, and the spreading of the scattering zone 314 is mainly observed in the first kilometer. At larger depth, specular reflections 315 predominate over the scattered component. 316

The panels oriented parallel to the fault also display a structural difference between the right and left sides of the fault. The scattering appears to be more concentrated in the SW profile (Fig. 3c), whereas the NE profile (Fig. 3d) shows strong continuity of planar boundaries. These observations are also in agreement with (Sharp, 1967, Fig.3) where the SW of the SJFZ at Table Mountain exhibits more complexity than the NE region. The scattering zones in both sections are dipping toward the SE where the fault zone is more localized.

In the following, we will show that the reflection matrix contains much more information than the medium's reflectivity. It also provides a direct insight into the variations of the scattering properties and the attenuation of scattered waves in the fault region.

#### 328 4. Lateral variations of intensity

Seismic waves propagating inside the Earth give direct insight on the nature and properties of rocks. While travelling through complex heterogeneous media, waves suffer from seismic attenuation. In this section we briefly discuss seismic wave attenuation principles and how attenuation is accounted for in our matrix formalism, and therefore in the images obtained.

Seismic attenuation describes the decay of energy experienced by seismic 334 waves while they propagate. Amplitudes are easily altered by several factors such 335 as geometrical spreading, scattering and absorption (intrinsic or anelastic attenua-336 tion) (Shapiro and Kneib, 1993; Knipe et al., 1998). Evaluation of the attenuation 337 due to scattering and intrinsic absorption has been the subject of considerable stud-338 ies in seismology (Aki, 1969; Sato et al., 2012). Estimating attenuation properties 339 provides complementary information to seismic velocity distribution, and can be 340 particularly useful in fault zone studies to obtain a better understanding of rock 341 properties and subsurface structures. Within the framework of the present paper, 342 seismic attenuation is being examined to: (i) compensate the intensity decay in the 343 3D volume (see Appendix C) and (ii) reveal lateral variability of backscattered 344 intensity in the fault zone. 345

The intensity represented in the pixels gives a direct estimate of the scattering properties in the region. To detect possible lateral variability of the energy distribution below the dense array, we divide the area into sub-regions and examine the intensity of temporal decay for shallow and deep parts. In Fig. 4a, the study area is divided into three zones displayed in Fig. 4a1: a zone representing the main fault zone (red shaded area), a region to the SW of the fault zone (blue shaded area) and a region to the NE of the fault zone (green shaded area). For each region, we



Figure 4: Time decay of the backscattered intensity. (a) Lateral variation of the time decay in the shallow crust for the first 400 m. The area is discretized into three zones: red color refer to the fault zone, blue color to the SW and green color to the NE. (b) NW - SE variation of the backscattered intensity time decay in the deep crust up to 4000 m. The area is discretized in two zones: blue color refers to the SE and green color to the NW. (1) Map of the geophones of the dense array at SGB site (black dots). The Clark fault traces are represented by the red lines. The black rectangle corresponds to the area covered by the images. (2) Intensity decay as a function of apparent depth and, equivalently, of two way travel time of the scattering volume.

compute the mean intensity of the pixels located beneath that region. The energy 353 distribution for the first 400 m is presented in Fig. 4a2. The linear regression is 354 also plotted (dashed lines) to visualize the slope of the energy decay. We only 355 show the energy decay corresponding to the images after the correction process. 356 The first noticeable feature is that the fault zone is associated with higher intensity 357 values resulting from non-specular energy transmitted from the localized damage. 358 This is consistent with the overall amplification of seismic waves in low velocity 359 fault zone layers (e.g. Kurzon et al. (2014)). The energy in the fault zone with a 360 steep slope distinguishes itself from the surrounding western and eastern regions 361 where the decay slopes are more gradual. The fault zone is defined by major 362 fracturing and crushed rocks, and consequently it is characterized by a rapid en-363 ergy decay. In the neighboring regions where the damage is more distributed, less 364 attenuation is observed in comparison with the fault zone. These results are con-365 sistent with significantly lower values of attenuation coefficients generally found 366 within fault zones by waveform fitting of trapped waves (e.g. Lewis et al. (2005); 367 Qin et al. (2018)). 368

We also compare the intensity decay between the SE and NW sections 4 km 369 below the array. Fig. 4b2 shows plots of the backscattered intensity averaged 370 across the two regions delineated in Fig. 4b1. We choose to plot the measured 371 intensity values without the linear regression. The discrepancy observed in the 372 backscattered intensity distribution primarily reflects the difference in subsurface 373 structure between the two regions that was highlighted in section 3.2. The fluc-374 tuation of intensity in the NW plot (green dots) is associated with the specular 375 returns at several depths. In other words, high values correspond to the reflective 376 boundaries observed in Fig. 3a, while the blue scatter plot representing the SE of 377

the array decreases smoothly. This is explained by the consistent density of scatterers around the core of the fault zone damage area previously highlighted by the cross section in Fig. 3b.

#### **381 5. Discussion and conclusions**

The presented results provide detailed images of seismic properties in the 3D 382 volume around the San Jacinto fault at the SGB site. We used one month of 383 ambient seismic noise recorded by a dense array deployed at SGB site around the 384 Clark branch of SJFZ. The high frequency seismic data provided by the spatially 385 dense array allows us to resolve features near the surface with high resolution. 386 The reflectivity maps representing slices of the 3D scattering volume are obtained 387 through the reflection matrix procedure developed in Touma et al. (2021). Body 388 wave reflections from ambient noise correlations are used to image the fault zone 389 structure up to 4 km below the surface. These images reveal the backscattered 390 intensity generated by the distribution of heterogeneities in the medium. 391

Fault zones are very complex regions with extensive fracturing and damage 392 that can reach the bottom of the seismogenic zone in some places as seen in to-393 mographic studies around large faults (e.g. Allam et al. (2014)). Tomographic and 394 other imaging studies provide average properties of rock volumes, but do not re-395 solve the presence, location and intensity of scatterers that are imaged with the re-396 flection matrix method. The strong variations of velocities and significant attenua-397 tion in fault zone regions present challenges for conventional imaging techniques. 398 However, our reflection matrix approach allows us to derive the distribution of 399 scatterers inside the medium with an approximate velocity model of the medium. 400 To that aim, a powerful aberration correction process is performed and provides 401

high resolution images of the subsurface. However, associating a reflector with
a specific depth remains dependent on the reference velocity model, so the reflectivity maps are displayed as a function of an effective depth and the observed
two-way travel time.

A significant advantage of the matrix approach is that focusing inside the 406 medium enables the imaging of not only specular reflectors but also of scattering 407 objects such as cracks and fractures. While many methods consider the diffracting 408 and scattering components as noise in the seismic data, and tend to remove these 409 components to image discontinuous layers, the current approach takes advantage 410 of the scattering in the complex fault zone to resolve features of the order of the 411 wavelength. This constitutes one of the main strength of the method. The images 412 in Figs. 2 and 3 show both discontinuities of some layers that are signature of a 413 large fault, along with lateral and axial variations in the backscattered intensity 414 induced by cracks and other small-scale heterogeneities. The axial variations of 415 the backscattered intensity in Fig. 4 also reveal systematic differences in scatter-416 ing properties in the region within and around the surface fault traces relative to 417 the outside volume, as well as differences between the NW and SE portions of the 418 study area. The results are consistent with more localized intense damage zone at 419 depth in the SE section where the SJFZ enters the Horse canyon, and more diffuse 420 rock damage to the NW where the SJFZ is in a linear valley. The higher damage 421 at depth in the SE section also suggests a geometrical complexity at depth lead-422 ing to an ongoing generation of rock damage that is overprinted on older healed 423 damage (Sharp, 1967). 424

Fig. 5 summarized schematically the obtained imaging results for both the inner and outer damage zones in the area. The fault traces in Fig. 1a suggest a

broader and less intense fault zone in the NW than in the SE. Surface observations 427 are consistent with the fact that the main fault and the surrounding core damage 428 are more localized in the SE where the principle slip zone is delineated by in-429 tense damage and fracturing extending down to 3 to 4 kilometers. In the NW, the 430 scattering in the FZ is only observed in the first kilometer indicating a shallow 431 less intense and diffuse damage. The results in Figs. 3c and d show that the dam-432 age distribution is more complex to the SW side of the fault exhibiting a more 433 pronounced outer damage in the SW than in the NE side of the fault. 434

The lateral variations of the fault structure between the NW and SE are consistent with the transition from the Anza section of the SJFZ associated with a single major fault trace, to a complex fault zone in the trifucation area with several traces at the surface. The high scattering zone that extended to depth in the SE part of the SGB site is likely associated with highly damaged fault zone rocks between the sub-parallel strands in the SE.

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![](_page_25_Figure_0.jpeg)

Figure 5: Schematic illustration of the fault zone at SGB site. Black dots refer to the geophones and red lines represent the Clark fault traces.

#### 451 DATA AVAILABILITY

The seismic data used in this study can be obtained from the Data Management Center of the Incorporated Research Institutions for Seismology (IRIS). The facilities of IRISData Services, and specifically the IRIS Data Management Center, were used for access to waveforms, related metadata, and/or derived products used in this study. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience (SAGE) Award of the National Science Foundation under Cooperative Support Agreement EAR-1851048.

#### 459 Appendix A. Imaging and aberration correction

In this section, we review the reflection matrix technique (Touma et al., 2021) applied to obtain the 3D scattering images of the accompanying paper. First we describe the virtual focusing method inside the medium and then we recall the aberration correction process.

Ambient noise cross-correlations between the stations are computed in the 464 10-20 Hz range. These correlations provide the impulse response between vir-465 tual sources (emission) and receivers (reception) at the surface. In other words, 466 the cross-correlations form a response matrix  $\mathbf{K}(t)$  of the underground associated 467 with the geophones' array. This matrix contains considerable of information about 468 the medium, especially the distribution of its local reflectivity. This information 469 is retrieved from the response matrix by performing focusing operations, both in 470 emission and reception. Focusing consists in applying appropriate time delays, at 471 emission and reception, so that the seismic waves resulting from a scattering event 472 at the focal point interfere constructively. The goal of this operation is to virtu-473 ally move the sources and receivers inside the medium onto a virtual plane. It is 474

![](_page_27_Figure_0.jpeg)

Figure A.1: Focused reflection matrix and aberration correction. (a) The response matrix is projected onto a focused basis at depth z both at emission (in) and reception (out). The focusing points synthesise virtual sources and receivers scanning every point in the virtual plane. In the presence of heterogeneities, the waves are distorted and the focusing is not optimal. (b) Confocal imaging principle: The image of the medium is obtained when input and output focusing are performed at the same location in the focused basis. (c) Far-field projection of the focused reflection matrix: For each virtual source, the reflected wave-front is investigated in the far-field. The phase distortions are identical for nearby virtual sources belonging to the same isoplanatic patch. (d) The phase conjugate of the aberration phase law enables a fine compensation of aberrations over the corresponding isoplanatic patch.

commonly referred to as redatuming (Berkhout and Wapenaar, 1993) and can be easily implemented in the frequency domain via simple matrix products. The first step is to apply a temporal Fourier transform to the correlation matrix to obtain  $\mathbf{K}(f)$ .

To perform focusing, we define at each depth a basis of focal points r, which 479 corresponds to the location of virtual geophones inside the medium. A constant 480 velocity model of 1500 m/s is chosen for the transmission matrix  $G_0(f)$  that 481 describes the propagation of waves between the surface and the focused basis. 482 At each frequency, the response matrix is projected into the focused basis both 483 in emission and reception (Fig. A.1a). The broadband focused reflection matrix 484  $\mathbf{R}_{\mathbf{rr}}(z)$  is then computed by calculating the coherent sum of the focused reflection 485 matrices over the considered frequency range. This last operation amounts to a 486 ballistic time gating of singly-scattered waves at time  $t \sim 2z/c_0$ . The coefficients 487  $\mathbf{R}$   $(\mathbf{r}_{out}, \mathbf{r}_{in})$  of this matrix correspond to the responses between a set of virtual 488 sources  $\mathbf{r_{in}} = (x_{\text{in}}, y_{\text{in}}, z)$  and receivers  $\mathbf{r_{out}} = (x_{\text{out}}, y_{\text{out}}, z)$  at each depth z. 489 Among all these cofficients, the diagonal elements ( $\mathbf{r}_{in} = \mathbf{r}_{out}$ ), are of particular 490 interest since they provide a confocal image of the underground (Fig. A.1b). The 491 other coefficients of  $\mathbf{R}_{rr}$  are also useful since the focusing quality can be directly 492 quantified by the spreading of the backscattered energy over its off-diagonal ele-493 ments. In the presence of strong phase distortions (aberrations), resulting from a 494 very complex seismic velocity distribution, each input and output focal spot can 495 actually spread well beyond the diffraction limit, giving rise to a loss of resolution 496 and contrast of the confocal image. Interestingly, the reflection matrix can also 497 be used to retrieve in post-processing the shape of wave-fronts that would allow 498 a perfect compensation for these phase distortions. To do so, one can exploit the 490

spatial correlations that exist between the phase distortions urdergone by the re-500 flected wave-fronts induced by nearby virtual geophones. This coherence area is 501 referred to the isoplanatic patch (Fig. A.1c). The idea is to project the focused 502 reflection matrix at input or output in a basis that maximizes the size of such iso-503 planatic patches in the focal plane. For a multi-layered basis, the most adequate 504 basis is the Fourier basis (k) that amounts to projecting the reflected wave-field 505 in the far-field. The aforementioned spatial correlations are then leveraged by 506 extracting the distorted component of each reflected wave-field. It results in a dis-507 tortion matrix whose singular value decomposition provides a decomposition of 508 the field-of-view into a set of isoplanatic patches with the associated aberration 509 phase transmittances. The phase conjugate of each aberration phase law provides 510 the focusing law that should be used to compensate for aberrations on each iso-511 planatic patch (Fig. A.1d). Such a focusing law is applied here to the reflection 512 matrix from the dual basis to obtain an updated focused reflection matrix. A cor-513 rected confocal image is obtained by considering the diagonal coefficients of this 514 new focused reflection matrix. 515

The whole process can be iterated at input and output in order to refine our estimation of the focusing law over each isoplanatic patch. To that aim, the focused reflection matrix should be projected in the far-field (Fourier basis k) by switching, at each iteration, between input or output.

#### 520 Appendix B. Specular vs. diffuse reflection

The finite size of the geophone array has an impact on the field-of-view that can be imaged by the reflection matrix approach. To quantify this effect, two synthetic tests have been performed: (i) considering planar reflectors at a given depth z = 3600m (specular scattering, see Fig. B.1a1); (ii) considering randomly distributed scatterers at the same depth (diffuse scattering, see Fig. B.1b1). In each case, an homogeneous effective wave velocity  $c_0 = 1500m/s$  is considered. Fig. B.1a2 and B.1b2 show the results of the numerical simulations.

Fig. B.1a2 shows a synthetic confocal image obtained for the specular reflec-528 tor. The field-of-view is limited by the size of the array. The fraction of the inci-529 dent wave-front that spreads beyond the transverse size of the array gives rise to 530 reflected waves that cannot be recorded by the array of geophones (red arrows in 531 Fig. B.1a1). Fig. B.1b2 displays the field-of-view of our matrix imaging method 532 in the diffuse scattering regime. This intensity distribution is obtained by averag-533 ing the synthetic confocal image over a number N = 50 realizations of disorder. 534 Unlike for specular reflectors, incident wave-fronts are reflected in different direc-535 tions (arrows in Fig. B.1b1). Consequently, the field-of-view spreads well beyond 536 the transverse size of the geophone's array. The latter observation explains the 537 wide field-of-view obtained in the 3D-scattering images of the accompanying pa-538 per (Fig. 3). The reflection matrix method mainly images a distribution of small 539 heterogeneities that are located in the damage area at shallow depths (z < 1500m) 540 and that reside between each geological layer at larger depths. 541

#### 542 Appendix C. Time-gain compensation

The energy losses increase with the heterogeneity of the medium. The propagation matrix used to project the raw data to the focused basis accounts for geometrical spreading whereas the effect of absorption and scattering are ignored. In particular, in strong scattering regime, these losses can strongly degrade the contrast of the images at larger depth. To overcome these problems in the imag-

![](_page_31_Figure_0.jpeg)

Figure B.1: Impact of specular and diffuse reflection on the field-of-view of the confocal image. (a) Smooth mirror-like surface. (a.1) Sketch showing the reflected waves traveling with the same angle as the incident waves. Some of the reflected rays are not captured by the array. (a.2) Synthetic confocal image of a specular reflector at z = 3600 m. The white dashed square corresponds to the spatial extension of the array. (b) Random scattering medium. (b.1) Sketch displaying distributed heterogeneities at depth. An incident ray is scattered in all directions enabling a confocal image that extends well beyond the lateral dimensions of the geophone array. (b.2) Mean synthetic confocal image at z = 3600 m estimated from several iterations obtained in the case of a random medium. The white dashed square corresponds to the spatial extension of the array. The imaged field-of view extends well beyond the transverse size of the array.

![](_page_32_Figure_0.jpeg)

Figure C.1: Mean intensity decay as a function of depth with the corresponding linear regression in logarithmic scale and the characteristic length of decay values. (a) Intensity decay of raw images. (b) Intensity decay of images corrected from the aberrating component.

ing process and in order to visualize the entire field-of-view, it is mandatory to compensate for the amplitude drop in the 3D-images (Fig. 2 and 3), especially in the shallow crust. This time gain compensation is done manually by multiplying the intensity profiles by an increasing function with depth, that is in this case the reciprocal of the mean intensity calculated at each depth.

An estimation of the seismic wave attenuation at the SGB site can be directly measured through the backscattered intensity at each depth. The energy is obtained by calculating the mean square of the intensity of the image pixels, i.e. the intensity of the diagonal of the reflection matrix. The amplitude decay of the energy is expected to follow an exponential decaying as a function of depth (Fig. C.1).

<sup>559</sup> Figs. C.1a and b displays the natural logarithm applied to the mean intensity

as a function of depth calculated from the raw reflection matrix and the corrected reflection matrix respectively (aberration correction process disclosed in Touma et al. (2021)). In the logarithmic scale, the mean confocal energy is expressed as a decreasing linear line. The characteristic length of attenuation, that corresponds to the inverse of the slope of the log-energy decay, is also shown.

A change of slope is observed in both plots corresponding to depth of 750 m 565 (t = 1 s). While a pronounced decrease in intensity occurs until 750 m, a more 566 gentle slope is noticed below 750 m. The presence of stratigraphic boundaries, ab-567 sorbing sediments and intense fracturing is a possible cause of the high attenuation 568 observed at shallow layers. Another thing to notice is the increase by around twice 569 the characteristic length values after correction. We recall that the energy decay 570 describes the losses that the seismic waves undergo while propagating inside the 571 medium. In case of strong inhomogeneities and velocity variation, the wavefield 572 is heavily distorted and the focusing inside the medium fails. The increase in the 573 slope provided by the aberration correction process indicates an improvement in 574 the focusing operation and consequently an enhancement of the detection. 575

#### 576 Appendix D. Supplementary Data

577 Supplementary data associated with this article can be found in the online 578 version.

579 Animation D1: Animated vertical slices oriented perpendicular to the fault 580 traces.

581 Animation D2: Animated vertical slices oriented parallel to the fault traces.

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