

1 **1D-velocity structure and seismotectonics of the Ecuadorian margin inferred from the**
2 **2016 Mw7.8 Pedernales aftershock sequence**

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41 ABSTRACT

42 On April 16th 2016 a Mw 7.8 earthquake ruptured the central coastal segment of the
43 Ecuadorian subduction zone. Shortly after the earthquake, the Instituto Geofisico de la
44 Escuela Politecnica Nacional of Ecuador, together with several international institutions
45 deployed a dense, temporary seismic network to accurately categorize the post-seismic
46 aftershock sequence. Instrumentation included short-period and broadband sensors, along
47 with Ocean Bottom Seismometers. This deployment complemented the permanent
48 Ecuadorian seismic network and recorded the developing aftershock sequence for a period
49 of one year following the main-shock. A subset of 345 events with $M_L > 3.5$, were manually
50 picked in the period of May to August 2016, providing highly accurate P- and S-onset times.
51 From this catalogue, a high-quality dataset of 227 events, with an azimuthal gap $< 200^\circ$, are
52 simultaneously inverted for, obtaining the minimum 1D velocity model for the rupture
53 region, along with hypocentral locations and station corrections. We observe an average
54 V_p/V_s of 1.82 throughout the study region, with relatively higher V_p/V_s values of 1.95 and
55 2.18 observed for the shallowest layers down to 7.5 km. The high relative V_p/V_s ratio (1.93)
56 of the deeper section, between 30 km and 40 km, is attributed to dehydration and
57 serpentinization processes. For the relocated seismicity distribution, clusters of events align
58 perpendicular to the trench, and crustal seismicity is also evidenced, along with earthquakes
59 located close to the trench axis. We also compute Regional Moment Tensors to analyze the
60 different sources of seismicity after the mainshock. Aside from thrust events related to the
61 subduction process, normal and strike-slip mechanisms are detected. We suggest that the
62 presence of subducting seamounts coming from the Carnegie Ridge act as erosional agents,
63 helping to create a scenario which promotes locking and allows seismicity to extend up to
64 the trench, along zones of weakness activated after large earthquakes.

65

66 KEYWORDS: 2016 Pedernales earthquake, Ecuador, velocity model, subduction zone,
67 regional moment tensor, aftershock sequence.

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70

71 1. INTRODUCTION

72 The South American subduction zone has a long history of large megathrust earthquakes
73 and in the past decades a variety of different types of seismicity have been identified,
74 including repeating earthquakes, seismic swarms and slow slip events. During the last
75 century, the Ecuadorian-Colombian margin has experienced several major earthquakes
76 (Ramirez, 1968; Kelleher, 1972; Abe, 1979; Herd et al., 1981; Kanamori and McNally, 1982;
77 Mendoza and Dewey, 1984; Beck and Ruff, 1984; Swenson and Beck, 1996). More recently,
78 slow slip events have been observed (Mothes et al., 2013; Vallee et al., 2013; Chlieh et al.,
79 2014; Segovia et al., 2015; Vaca et al., 2018; Collot et al., 2017; Rolandone et al., 2018), in
80 addition to seismic swarms (Segovia, 2001; Segovia, 2009; Vaca et al., 2009) and repeating
81 earthquakes (Rolandone et al., 2018) (Figure 1).

82 In 1906 a megathrust earthquake Ms 8.7 (Kelleher, 1972), re-estimated to Mw 8.8
83 (Kanamori and McNally, 1982) and more recently to Mw 8.4 (Yoshimoto et al., 2017),
84 ruptured the subduction segment along central and northern Ecuador and southern
85 Colombia generating severe destruction which reached up to 100 km inland. The rupture
86 extent of this earthquake is still under discussion. Whilst Kanamori and McNally (1982)
87 suggested a rupture of approximately 500 km length, Yoshimoto et al., (2017) proposed a
88 smaller rupture area mainly located near the trench. The 1906 earthquake also caused
89 permanent coastal uplift and generated a destructive tsunami (Ramirez, 1968; Kelleher,
90 1972; Abe, 1979; Herd et al., 1981; Kanamori and McNally, 1982). Subsequently, smaller
91 earthquakes of diverse magnitudes (between Mw 7.0 to Mw 8.2) ruptured with thrust
92 mechanisms, leading to several authors proposing a segmentation of the subduction zone
93 along the northern Ecuadorian margin (i.e. Marcaillou et al., 2006; Gailler et al., 2007). In
94 1942, a Mw 7.8 earthquake occurred close to the coastal city of Pedernales affecting an area
95 of ~200 km x ~90 km. The epicenter was located at the southern edge of the rupture region,
96 suggesting that the event propagated to the NE (Kelleher, 1972; Swenson and Beck, 1996).
97 In 1958, the Mw 7.7 Colombia-Ecuador earthquake (Kanamori and McNally, 1982) broke a
98 small segment of the 1906 megathrust earthquake in the offshore portion of the Esmeraldas
99 province to the North. Like the 1942 earthquake, the 1958 event showed a rupture
100 propagation in the NE direction (Rothe, 1969; Kelleher, 1972; Mendoza and Dewey, 1984).
101 Finally, in 1979, the largest event of this series of major earthquakes occurred close to the

102 Ecuadorian-Colombian border with magnitude of Mw 8.2 also rupturing in a NE direction
103 along ~230 km of the margin (Herd, 1981; Kanamori and Given, 1981; Beck and Ruff, 1984).

104 Between the equator and south of the Carnegie Ridge (CR), there have been few large
105 subduction earthquakes recorded (Egred, 1968; Dorbath et al., 1990; Bilek et al., 2010) with
106 only two events of Mw>7.0 (in 1956 and 1998) occurring close to Bahia Caraquez (ISC-GEM
107 catalogue, Storchak et al., 2013). However, a high rate of seismicity up to Mw 6.5 has been
108 registered in this region (Vallee et al., 2013) suggesting that the accumulated stresses are
109 released by different mechanisms than in the northern segment (White et al., 2003; Agurto-
110 Detzel et al., 2019). Moreover, the area shows a complex slip behavior, with different types
111 of seismic activity highlighted in Figure 1.

112 The Mw 7.8 Pedernales earthquake occurred in 2016 and led to the largest earthquake-
113 related loss of life (668, Lanning et al., 2016) in Ecuador since the 1987, Mw 7.1 Salado-
114 Reventador event, which occurred in the Andean volcanic arc that killed around 1000
115 people (Bolton et al., 1991; Beauval et al., 2010). The 2016 earthquake also affected the
116 Ecuadorian economy due to widespread destruction of houses, hotels and hospitals
117 (Lanning et al., 2016). One month after the earthquake, a dense temporary network (10-30
118 km station spacing) was deployed in collaboration with the Instituto Geofisico de la Escuela
119 Politecnica Nacional (IGEPN) and several international institutions covering the rupture
120 area. Instrumentation included Ocean Bottom Seismometers (OBS), inland short period and
121 broadband sensors, which complemented the stations of the permanent Ecuadorian seismic
122 network (Alvarado et al., 2018). The combined array recorded the unfolding aftershock
123 sequence up to one year after the mainshock. Although large subduction thrust earthquakes
124 are well documented historically, occurring extensively throughout the entire South
125 American margin, it is only over the last decade that dense monitoring seismic networks
126 have been deployed to record the aftershock sequences of such events. This allows for
127 highly accurate hypocentral relocations enabling detailed investigations into the physical
128 processes occurring within the rupture zone.

129 One of the major limitations in imaging the seismicity distribution along the Ecuadorian
130 margin is the lack of a robust and well constrained velocity model for the area. The
131 unprecedented seismic coverage provided by the international aftershock deployment
132 offers a unique opportunity to better resolve the velocity structure of this region with the

133 Pedernales earthquake sequence. The recorded data will improve our knowledge of the
134 seismotectonic structures, such as the CR and the Atacames seamounts, along with
135 understanding the mechanisms that control subduction processes in the area.

136 Although the Ecuadorian margin has been widely studied, there is still no consensus on a
137 representative one-dimensional (1D) velocity model. An improved 1D regional model
138 together with well located aftershocks recorded by our dense seismic monitoring network
139 will provide a better constrained seismic velocities in the area and, the foundation and
140 starting model for future detailed 3D seismic imaging.

141 Currently, the available velocity models for the Ecuadorian margin are usually focused in
142 small areas or have limited resolution due to the lack of instruments monitoring the
143 offshore and coastal portion of the country; they are also affected by along strike
144 heterogeneities observed in the region. At present, the national network of seismographs of
145 the geophysical institute (RENSIG) from IGEPN uses a five-layer model (called ASW; for
146 detailed values see Font et al. (2013)) and is used by the IGEPN for locating tectonic events
147 in Ecuador. It, therefore, results in large errors in hypocenter location, especially regarding
148 depth of subduction related seismicity (See Supplementary Material 1). Font et al. (2013)
149 built an *a-priori* three-dimensional P-wave velocity model for the Ecuadorian subduction
150 margin (4° N – 6° S, 83° W – 77° W), obtained via a joint compilation of marine seismics,
151 gravimetric data, geological observations and seismicity of the region. The model is an
152 improvement for the forearc and provides better hypocentral solutions for the coastal area
153 but has limited capabilities due to the absence of an S-wave velocity model. Gailler et al.
154 (2007) inverted wide-angle seismic data to describe deep structures present offshore of the
155 Ecuadorian – Colombian border. In the same area, Agudelo et al. (2009) also improved the
156 structural model using a combined data set of multichannel seismic reflection and wide-
157 angle data. Garcia Cano et al. (2014) provide a 3D velocity model encompassing the rupture
158 of the 1958 earthquake. Both models provide insights about offshore velocities. Further to
159 the south of our study area, Calahorrano et al. (2008) developed a velocity model using
160 marine seismic experiments representing the area close to the Gulf of Guayaquil. However,
161 all three models are located outside of the Pedernales rupture area and outside the area
162 covered by our temporary seismic array. Finally, Graindorge et al. (2004) analyzed the
163 velocity structure in the Carnegie Ridge (CR) area using onshore and offshore wide-angle

164 data to calculate a two-dimensional model. Although our array covers a portion of the CR,
165 this model is not appropriate for the whole area covered by the array due to the anomalous
166 structure of this bathymetric feature.

167 Using high quality manually picked arrival times (P and S), we develop a minimum 1D
168 velocity model for P- and S-wave velocities of the area affected by the Pedernales
169 earthquake and post-seismic sequence. Based on a Monte Carlo-like approach in which we
170 use a set of 5000 randomly perturbed velocity models as starting models for our inversion,
171 we obtain a new robust minimum 1D velocity model for our study area. Finally, we analyze
172 in detail the seismicity recorded offshore Bahia Caraquez, close to the trench, by
173 determining Regional Moment Tensors (RMTs) to explore the correlation between the
174 distribution of the events, the seismogenic contact and possible faults that might be
175 re/activated due to the mainshock.

176 2. SEISMO-TECTONIC SETTING OF THE PEDERNALES EARTHQUAKE

177 2.1 Structural setting

178 The Ecuadorian margin is dominated by a convergent regime that involves the subducting
179 oceanic Nazca plate and the overriding continental South American plate. This process
180 occurs with a relative convergence rate of 55 mm/yr (Kendrick et al., 2003) in the N80°E
181 direction. The margin is highly segmented and mainly of erosional type (Collot et al., 2002;
182 Gailler et al., 2007; Marcaillou et al., 2016) creating diverse patterns in seismicity in the
183 subducting and overriding plates as well as along the plate interface. Broadly, the
184 Ecuadorian territory (1°N – 3°S) can be divided into three main domains extending inland
185 from the margin (Jaillard et al., 2000) and our study area comprises mainly along domain (I)
186 and part of domain (II). The profile A, in Figure 1b, shows the marine forearc including a
187 small layer of sediments of 500 m to 1000 m thick. The coastal region is characterized by the
188 presence of bathymetric features such as the aseismic CR. The CR (~280 km wide and 2 km
189 high) is located near the Ecuadorian trench between latitude 0° and 2°S (see Figure 1a). The
190 CR was formed by the Galapagos hot-spot located about 1000 km west of the coastline of
191 Ecuador and is currently subducting in an ENE direction. The marine forearc has a series of
192 seamounts such as the Atacames chain which subduct beneath the South American plate
193 and is inferred to play an important role in the nucleation of large subduction earthquakes
194 (Marcaillou et al., 2016). The coastal region shows accreted oceanic terranes from the late

195 Cretaceous-Paleogene period. The width of this domain (I) varies longitudinally from 50 to
196 180 km, has low relief with less than 300 m altitude, and is filled with sediments from the
197 Andes, creating several alluvial basins. The second domain (II) is the Andean chain that
198 involves the Western cordillera, with deformed rocks from accreted oceanic crust, and the
199 Eastern cordillera (Cordillera Real) which contains metamorphic rocks from the Paleozoic to
200 Mesozoic (Litherland et al., 1994). Both areas are separated with Tertiary to Quaternary
201 volcanic and volcanoclastic rocks that were deposited in the Interandean Valley. At $\sim 3^{\circ}\text{S}$, the
202 Chingual-Cosanga-Pallatanga-Puna (CCPP) fault zone separates the continental plate in two
203 (Alvarado et al., 2016), extending throughout the Puna segment, through the Colombian
204 border and towards Venezuela, creating the North Andean Sliver (NAS) shown in Figure 1. In
205 front of the NAS, the Nazca Plate is less than 26 Ma old (Lonsdale, 2005) and subducts at a
206 relative rate of about 46 mm/yr (Chlieh et al., 2014). More specifically, if we consider the
207 coseismic slip derived by Nocquet et al. (2017) and the aftershock activity analyzed by
208 Agurto-Detzel et al., (2019), the area affected by the Pedernales earthquake extends from
209 Esmeraldas canyon in the north ($\sim 1.5^{\circ}\text{N}$), down to La Plata area in the south ($\sim 1^{\circ}\text{S}$). In this
210 segment, the depth of the trench varies from 3.7 km in the north to 2.8 km at the crest of
211 the CR (Collot et al 2004) in the south. Moreover, the oceanic crust thickness shows an
212 along strike variation from 5 km in the north to 14 km in the south, reaching its maximum of
213 19 km beneath the crest of the CR (Meissner et al., 1977; Calahorrano, 2001; Sallares and
214 Charvis, 2003; Sallares et al., 2005; Graindorge et al., 2004; Garcia Cano et al., 2014).

215

216 2.2 The Pedernales earthquake and its aftershocks

217 On April 16th, 2016, the Mw 7.8 Pedernales earthquake occurred in the central coast of
218 Ecuador, close to the city of Pedernales. This estimated event focal depth is 20 km (Nocquet
219 et al., 2017) and the rupture is characterized as a thrusting mechanism consistent with the
220 megathrust on the plate interface. The Pedernales earthquake ruptured an estimated area
221 of 100 km x 40 km (Nocquet et al., 2017) and propagated along strike from north to south.
222 Although there are different rupture areas proposed for the 1906 earthquake (e.g.
223 Kanamori and McNally, 1982; Yoshimoto et al., 2017), the edges of the 2016 mainshock
224 appear to coincide with the southern boundaries of the 1906 earthquake. This segment of
225 the subduction zone last ruptured in a Mw 7.8 earthquake in 1942. At the southern limit,

226 the rupture coincides with the northern edge of the 1998 Mw 7.1 earthquake in Bahia
227 Caraquez. The Pedernales earthquake affected an area where large megathrust
228 earthquakes, with magnitudes greater than 7, occurred in the past. Its occurrence correlates
229 with a locked area of interseismic coupling (Chlieh et al., 2014; see Figure 1).

230

231 3. NETWORK AND DATA PROCESSING

232 3.1 THE SEISMIC AFTERSHOCK NETWORK

233 Immediately after the 2016 Pedernales mainshock, an international collaboration of several
234 institutions including; IGEPN from Ecuador, the French Centre d'études et d'expertise sur les
235 risques, l'environnement, la mobilité et l'aménagement (CEREMA) and GEOAZUR institutes,
236 Lehigh University and University of Arizona from the United States and the University of
237 Liverpool from the United Kingdom, coordinated a rapid response effort to deploy a
238 temporary seismic array. One month after the Pedernales earthquake, we deployed a dense
239 temporary seismological array of broadband and short period seismic stations equally
240 distributed inland along the area affected by the aftershocks (Meltzer et al, 2019). Ocean
241 Bottom Seismometers (OBS) were installed along the trench to complement the stations
242 deployed on land. Figure 2 shows the spatial distribution of the full array including both
243 temporary and permanent stations consisting of more than 80 stations with a station
244 spacing of approximately 10-30 km. The RENSIG Ecuadorian network is a mix of diverse
245 instrumentation including CMG-3ESP, Trillium compact, Trillium 120p, L4C, TSM-1 Strong
246 Motion instruments. The temporary array consists of: (i) short period Lennartz LE 3D Lite
247 recording at 100 Hz; (ii) intermediate period seismometers CMG-40T recording at 200 Hz;
248 (iii) broadband stations Trillium Compact, Guralp CMG-3T and STS-2 recording at 100 Hz.
249 The temporary array was fully operative from May 2016 to June 2017. OBSs were recording
250 between mid-May to November in 2016, at 100 Hz.

251 3.2 DATA PROCESSING

252 3.2.1 Onset detection

253 Based on the initial catalogue provided by IGEPN ([http://www.igepn.edu.ec/solicitud-de-](http://www.igepn.edu.ec/solicitud-de-datos)
254 [datos](http://www.igepn.edu.ec/solicitud-de-datos)) with 1677 reported seismic events between May 15th and August 26th, 2016, we
255 created a high quality dataset to be used for the 1D inversion process. We first selected

256 aftershocks with local magnitude (M_L) greater than 3.5 located in the vicinity of the
257 temporary array. From this catalogue, P- and S-wave arrival times were manually picked for
258 345 aftershocks using the Seismic Data Explorer (SDX) software package
259 (<http://doree.esc.liv.ac.uk:8080/sdx>). The events were located based on an average 1D
260 velocity model derived from Font et al. (2013); SDX utilizes a modified hypo71 algorithm for
261 hypocenter location (Lee et al., 1972). Following the procedures from Agurto et al. (2012)
262 and Hicks et al. (2014), we assigned pick error categories, referred as weights, from 0 to 4 to
263 describe the quality of the selected arrival times. Each weight corresponds to the following
264 time uncertainties: Weight 0 (< 0.04 s); Weight 1 ($0.04 - 0.1$ s); Weight 2 ($0.1 - 0.2$ s);
265 Weight 3 ($0.2 - 1$ s); Weight 4 (> 1 s).

266 *3.2.2 Minimum 1D model*

267 From the manually picked catalogue, we further filter to ensure only the highest quality
268 events remain for the minimum 1D velocity model inversion. We selected earthquakes with
269 at least 10 P- and 10 S-onset observations and an azimuthal gap $< 200^\circ$. In total we obtained
270 a dataset of 227 events with an average of 21 and 17 P- and S-onset times respectively,
271 which contains 4939 P-phases and 3931 S-phases (See Supplementary Material 2).

272 To minimize the influence of strong topography changes and avoid bias due to the depth of
273 the OBSs, we followed the strategy described by Husen et al. (1999) and Hicks et al. (2014)
274 and set station elevation equal to zero. Station correction terms therefore account for both,
275 the relative elevation and the site-specific velocity differences. Station correction terms
276 were strongly damped (damping = 1000) during the V_p inversion, allowing for greater initial
277 exploration of the V_p parameter space. For the V_s inversion, the damping was decreased
278 (damping = 20), this allows the station-delay terms to absorb errors due to the velocity
279 fluctuations near the stations.

280 We selected five starting models (see Supplementary Material 3) to cover a wide range of
281 plausible cases: (1) the average model derived from Font et al. (2013), selected as it covers
282 our entire study area with a 3D regional model, (2) an average offshore model based on
283 previous studies by Gailler et al., (2007) and Agudelo et al., (2009), and (3-5) modified
284 versions of the ASW model, used by IGEPN (Font et al., 2013). For models 3-5, we vary the
285 Moho depth between the range of 40 to 60 km. To cover the entire range of feasible
286 velocity models, we created 1000 random variations of each reference velocity model,

287 sampling from a uniform distribution with bounds ± 0.5 km/s of the reference model
288 velocity. We, therefore, explore 5000 starting models in total, derived from the five classes
289 of models. From the 5000 starting models, the P-wave velocity model with the lowest
290 overall misfit to the travel time picks (RMS 0.335 s) is assumed as our best Vp
291 representation. The density plots in Figure 3d and Figure 3e demonstrate how the best 200
292 solutions for the inverted Vp and Vs models have a clear convergence towards the best
293 solution between 7.5 km and 40 km depth. The shallowest layers (down to 7.5 km) do also
294 exhibit a convergence towards a minimum velocity solution, however, due to a lack of
295 events occurring in this region, the solution shows a slightly wider range of possible
296 velocities. For the deeper layers (> 40 km), absolute velocities cannot be constrained due to
297 the lack of data and therefore several cluster of solutions can be observed. To detect picking
298 outliers and to estimate an average Vp/Vs ratio for the region, we performed Wadati
299 analysis (Wadati, 1933), displayed in Figure 3a. Onset times show a clear trend for $(t_s - t_p)$
300 as a function of t_p , a linear trend fit provides a Vp/Vs ratio value of 1.82. A reduced Wadati
301 diagram was also used to have a better control on the outliers in our dataset. Onsets greater
302 than 2σ from the Vp/Vs trend were removed (see Figure 3b). The final dataset of arrival
303 times is represented in a histogram in Figure 3c with a mean of $t_s - t_p$ of 0.01 s and a variance
304 of 0.70 s^2 .

305 Finally, using our best Vp model and the Vp/Vs ratio (1.82) obtained via Wadati analysis, we
306 built an initial 1D S-wave velocity model. The inversion for the optimum S-wave velocity
307 model follows the same procedure as for the P-wave velocity model. We perturbed the S-
308 wave velocities of the starting model within a range of ± 0.5 km/s, creating 1000 models
309 where both P- and S-phases were simultaneously inverted for. In this step, we fixed the Vp
310 velocities to avoid changes in the previously obtained Vp model by S-wave observations.
311 The solution with the lowest RMS (0.303 s) was accepted as our final S-wave model along
312 with the final set of station corrections and hypocentral locations. By using a Monte Carlo
313 approach for selecting the starting model, we mitigate the dependency of the final model
314 from starting model as we sample a much larger model space.

315

316 *3.2.3 Regional Moment Tensors*

317 A significant portion of the aftershocks are related to the megathrust faulting displaying
318 reverse solutions consistent with the slip of the plate interface (Agurto-Detzel et al., 2019).
319 We focus our RMT analysis on the seismicity recorded throughout offshore Bahia Caraquez
320 and Cabo Pasado which exhibits various faulting mechanisms. In this area, we observe
321 events distributed close to the trench (5 km – 10 km), however, due to the trade-off
322 between hypocentral depth and origin time, depth location uncertainties in this region are
323 larger than the average. These events could, therefore, be located at the interface, in the
324 upper or in the downgoing plate. To better constrain the focal depth parameter for events
325 in this region, we calculated RMT solutions for 14 events by performing a full waveform
326 inversion using the *isola* package (Sokos et al., 2008; Sokos et al., 2013). We analyzed events
327 reported by IGEPN with local magnitude (M_L) greater than 4.0. The location was fixed to the
328 epicentral locations obtained from our relocation process with VELEST, and the depth is
329 allowed to vary. Green's functions are calculated from a simplified version of our minimum
330 1D model where the shallowest 8 layers are merged into 4 layers (see Supplementary
331 Material 4). Seismic records were bandpass filtered between 0.04 and 0.09 Hz before the
332 inversion, and between 4 to 10 broadband stations were used for the inversion process of
333 each event.

334 4. RESULTS

335 Our final minimum 1D model, represented by 12 layers, with P- and S-wave velocities and
336 the V_p/V_s ratio are given in Table 1 and shown in Figure 4. The model is well constrained
337 between 7.5 km and 40 km, where most of the seismicity is concentrated (see Figure 4).
338 Although, most of the ray paths are concentrated in the shallower layers, we observe some
339 rays down to 40 km depth, which help us to constrain the average of the oceanic and
340 continental Moho depth. At shallower depths, the lack of crustal events (< 7.5 km depth)
341 means we are not able to obtain reliable absolute velocities due to the trade-off with station
342 corrections; at greater depths (30 km – 40 km) the ray path distribution still shows rays
343 passing through, however at depths > 40 km the limited extend does not allow a robust
344 estimation of the average velocities.

345 The obtained model shows a V_p that fluctuates between 4.45 km/s and 5.04 km/s, in the
346 top layers (down to 7.5 km). From 7.5 km depth to 25 km, V_p consistently increases from 6.3
347 km/s to 6.6 km/s. Between 30 km – 40 km the constant increase of V_p (from 7.37 km/s at 30

348 km to 8.04 km/s at 40 km) suggests an area of transition from crust to mantle. At 40 km Vp
 349 reaches 8.04 km/s, which can be related with continental upper mantle velocities. Finally,
 350 the absence of data between 40 km and 60 km depth, does not allow us to resolve the P-
 351 wave velocity, so the resulting values are mainly influenced by our reference model derived
 352 by Font et al., (2013) (see Figure 4).

353 Vp/Vs ratio shows a weighted arithmetic mean of 1.82 for the whole model. In the top
 354 layers (down to 7.5 km depth) we find a high Vp/Vs ratio that varies from 1.97 to 2.18. At 10
 355 km depth, the Vp/Vs ratio decreases to 1.85 down to 30 km depth where it rises to 1.93.
 356 The Vp/Vs ratio is not well resolved at depths greater than 40 km as we do not have enough
 357 data to constrain that area (see Figure 4).

358

Depth (km)	Vp (km/s)	Vs (km/s)	Vp/Vs
-5.00	4.45	2.26	1.97
2.50	4.65	2.32	2.00
5.00	5.04	2.59	1.95
7.50	6.32	2.90	2.18
10.00	6.32	3.42	1.85
15.00	6.64	3.69	1.80
20.00	6.64	3.76	1.77
25.00	6.65	3.80	1.75
30.00	7.37	3.81	1.93
40.00	8.04	4.64	1.73
50.00	8.24	4.64	1.78
60.00	8.45	4.71	1.79

359 Table 1 Minimum 1D model. Absolute velocities and Vp/Vs ratio are listed in terms of depth.

360

361 Statistics for the inversion also show the robustness of our model. Arrival time residuals
 362 were classified according to the weights described in the previous section. Histogram plots
 363 in Figure 5 indicate that most of the P- (99 %) and S- (88 %) phase picks are classified
 364 between weights 0 and 1. For high quality P-wave onsets (weight 0 and 1), the standard
 365 deviations of residuals is 0.24 s and 0.26 s, respectively. For the S phase, the deviation varies
 366 from 0.48 s, for weight 0, to 0.52 s for weight 1. Observations for each station display
 367 residuals with a normal distribution concentrated between -1 to 1 seconds (See
 368 Supplementary Material 5). Overall, the obtained catalogue has a variance of 0.084 s^2 , which
 369 is above the variance based on the accuracy provided by the manual picking of 0.027 s^2
 370 indicating that we are not overfitting our onset time data.

371 Our new minimum 1D-velocity model was then used to relocate the 227 manually-picked
372 events (See Supplementary Material 6 for comparison with the original locations). To test
373 the accuracy of the resulting hypocenters, we performed a stability test by randomly
374 perturbing hypocenters starting locations by between 7.5 km – 12.5 km in latitude,
375 longitude and depth. We then relocated the events using our minimum 1-D velocity model
376 and station corrections. The location of the recovered seismicity shifts only marginally
377 respect to the original positions, leading to an estimation of the relocation uncertainties
378 around 1 km in latitude and longitude, and 2.7 km in depth (See Supplementary Material 7).
379 Overall, the relocated seismicity displays an average error of 1.38 km and 1.57 km for the
380 horizontal and vertical component, respectively. As hypocentral uncertainty is shown to
381 increase with distance offshore, we separately calculate the location errors for the offshore
382 seismicity. For this data subset, we obtain average uncertainties of 1.53 km for the
383 horizontal axis and 2.15 km for the vertical axis.

384 The relocated aftershocks are shown in Figure 6. The seismicity is distributed between 5-10
385 km to 100 km eastward of the trench and between 2 km to 35 km in depth. The most
386 important features are as follows:

387 (1) We observe that our epicentral distribution for the analyzed aftershocks are in
388 agreement with other related studies (e.g. Meltzer et al., (2019), Agurto-Detzel et al.,
389 (2019)). We find that events with local magnitude greater than 3.5 are generally distributed
390 outside the coseismic slip area of the 2016 Pedernales mainshock and skew towards the
391 trench.

392 (2) Most of the seismicity along profiles AA' and BB' is located along the plate interface.
393 Slab1.0 (Hayes et al., 2012) is used as a reference in Figure 6, however a comparison
394 between both Slab1.0 (Hayes et al., 2012) and Slab2.0 (Hayes et al., 2018) is shown in
395 Supplementary Material 8. The northern profile AA' demonstrates the differences between
396 the interface for both models, in the region, Slab2.0 is relatively deeper than the Slab1.0
397 interface; the located aftershocks are distributed between both Slab1.0 and Slab2.0. These
398 relative changes in extensively used slab models further highlight the complexities the
399 Ecuadorian margin along strike.

400 (3) Clustered activity in the overriding plate can be observed in both northern (AA') and
401 southern (BB') profiles.

402 (4) Profiles AA' and BB' (see Figure 6) show that a portion of the aftershocks reach the
403 trench. This has been demonstrated as a robust feature due to specific bias testing of the
404 offshore relocated seismicity.

405 Although, the formal locations errors stated before are less than 3 km in depth, absolute
406 locations errors can be significantly larger. To identify whether the events distributed close
407 to the trench are located in the subducting or the overriding plate, we calculated RMT
408 solutions to better constrain the hypocenter depth. We focus here on the offshore
409 seismicity located close to profile BB' and calculated 14 RMTs. This is an area where Agurto-
410 Detzel et al. (2019) reported a single anomalous strike-slip event at about 10 km depth.
411 Figure 6 shows the solutions superimposed over the relocated aftershock activity. Additional
412 information about the waveform inversion fit and comparison with other MT catalogues are
413 shown in Supplementary Material 9, 10 and 11. We also compare the difference between
414 depths obtained from the velocity model inversion and *isola*. Events with depths between 5
415 km and 15 km, based on arrival times, show a small difference in their depth locations in
416 comparison to the RMT solutions. For the offshore seismicity located with depths > 15 km
417 using travel times, the RMT centroid depths are shallower than those of the hypocenters
418 (see Table 2 and Supplementary Material 12 and 13).

419 The majority of the aftershock RMTs for the Pedernales segment exhibit a thrust mechanism
420 related to the subduction interface (e.g. Agurto-Detzel et al., 2019). However, we also find
421 events that display extensional and strike slip faulting mechanisms, distributed along the
422 marine forearc. Table 2 summarizes our results listing the centroid information, nodal
423 planes, magnitude, variance reduction and double-couple percentage.

#	Lat (°)	Lon (°)	Centroid depth (km)	Arrival times depth (km)	Strike	Dip	Rake	Mag (Mw)	Var (%)	DC (%)	ID	Date
1	0.6330	-80.2077	13.00	8.24	196	72	94	4.4	80	82.3	e0	2016-06-06 16:45:08
2	0.2940	-80.5730	2.03	4.42	342	57	-79	5.1	68	89.5	e3	2016-06-01 10:05:16
3	0.3740	-80.4910	14.03	11.91	174	60	65	4.4	86	61.2	e4	2016-05-31 15:48:11
4	0.1295	-80.1691	25.03	25.53	173	84	81	3.3	28	63.8	e5	2016-05-30 05:48:56
5	-0.2000	-80.5770	5.00	18.37	290	54	-66	3.3	13	71.1	e7	2016-06-21 04:40:24
6	-0.2718	-80.6639	16.03	12.54	49	49	164	3.7	13	31.5	e8	2016-06-02 00:18:59
7	-0.1880	-80.6652	16.03	18.99	175	74	74	3.5	44	62.5	e9	2016-06-01 02:12:22
8	-0.4079	-80.9864	14.02	8.14	247	88	177	3.8	45	76.0	e14	2016-07-07 17:10:13
9	-0.4444	-80.9390	8.03	29.99	143	74	-27	3.9	18	99.1	e16	2016-07-08 04:41:34
10	-0.4268	-80.8981	12.03	7.88	316	89	24	4.6	66	55.6	e17	2016-07-08 07:03:48
11	-0.4047	-80.9177	12.03	22.99	316	86	29	4.2	61	43.3	e18	2016-07-08 07:35:15
12	-0.2629	-80.8675	10.03	22.35	324	50	28	4.0	57	10.4	e19	2016-07-10 06:44:34
13	-0.1967	-80.9427	8.03	-	250	80	35	4.2	23	70.0	e25	2016-05-05 16:06:40
14	0.3242	-80.6818	2.03	4.16	318	61	-110	4.7	57	65.9	e27	2016-06-01 15:00:51

424 Table 2 summary of results for the obtained RMT. Depths calculated based on arrival times inversion were
425 included for comparison.

426

427 5. DISCUSSION

428 5.1 Velocity Model

429 Figure 4 shows the best 1D velocity model. The histograms show depth distribution of the
430 aftershock seismicity and help to demonstrate the resolution of our model at different
431 depths. Although a more in-depth interpretation requires at least a 2D velocity model, our
432 obtained 1D velocity model allows us to observe and discuss depth ranges and to first order
433 the velocity structure with depth. High V_p/V_s ratio obtained in the shallow layers can be
434 related to hydrated sediments, non-consolidated soils and/or fractured oceanic crust
435 (Peacock, 2001; Hacker et al., 2003; Kato et al., 2010; Pasten-Araya et al., 2018). The values
436 in this section (up to 2.2 at 7.5 km depth) may reflect an upward migration of fluids coming
437 from the dehydration of sediments subducting within the Nazca plate. Similar cases have
438 been discussed in other areas such as northern Chile (Husen et al., 2001), New Zealand
439 (Barnes et al., 2009), Sumatra (Collings et al., 2012) and Costa Rica (Bangs et al., 2015). As
440 we are dealing with P- and S- onsets coming from both OBS and inland stations, in the depth
441 range between 10 km and 40 km, the subduction process of the Nazca plate descending
442 beneath the South American plate produces an overlapping of velocities from both plates

443 which do not allow us to address each plate individually in a one-dimensional model.
444 However, there is a constant increase in velocities for both P- and S-wave indicating more
445 consolidated rocks. We also identify another increase in the V_p/V_s ratio ($V_p/V_s = 1.90$)
446 around 30 km depth. A change in V_p/V_s ratio at ~ 30 km depth agrees with the findings of
447 Hacker et al. (2003) and Bloch et al. (2018) who associated these characteristics with the
448 dehydration of the subducting oceanic plate. Between 30 km – 40 km it is not possible to
449 observe a sharp contrast. However, due to the rapid increase of V_p we suggest a transitional
450 area where the oceanic Moho could be located. At depths greater than 40 km, V_p reaches
451 values > 8 km/s that can be associated with velocities of the upper mantle (Gailler et al.,
452 2007; Font et al., 2013; Araujo, 2016).

453 Station correction terms were also calculated in the inversion. Figure 8 shows the delays
454 obtained for both P- and S-wave onsets. EC16 was used as a reference station because of its
455 central location along the array and large number of observations (total= 384, P= 214,
456 S=170). For the P-delay times, the standard deviation for all stations shows that values are
457 concentrated around 0.50 s. Delays for the S phase show a distribution around 0.85 s. In
458 both cases, the obtained values are coherent within the study region following a west to
459 east change of sign that moves from negative to positive. This change is mainly associated
460 with two factors: (1) The eastward dipping of the subducting Nazca plate that produces
461 large residuals in the inversion as VELEST maps the 2D structure into the station correction
462 terms; (2) the variations in topography of the area, from around 3000 meters below sea
463 level at the trench to 700 meters above sea level in the coastal range. If we consider the
464 difference in elevation of the array, a geological interpretation will only be valid for the
465 inland stations as the reference station (EC16) is located in this area, and the delay in the
466 OBS can mainly be explained due to the abrupt changes in topography from sea level to the
467 trench (~ 4000 m depth).

468 Taking this into account, the difference in elevation for the inland seismic stations range
469 from sea level for stations installed close to the coast to 688 m for the stations deployed at
470 the base of the coastal range. Station correction terms in Figure 8, especially for the S-
471 phase, show a NE-SW variation in sign from negative to positive which might be related with
472 changes in the geological conditions of the area. Those changes coincide with the Jama Fault
473 System (JFS) mapped in coastal Ecuador (Reyes and Michaud, 2012) and can be attributed to

474 differences in the properties on both sides of the fault. Along the coast, it is possible to find
475 Cretaceous formations (~89 Ma, Luzieux et al., 2006) with rocks from the oceanic crust that
476 were accreted onto the margin and could explain the negative delay in station terms. To the
477 east side of the JFS, less consolidated formations such as conglomerates, volcanic sediments
478 and alluvial formations from the Miocene, Pliocene and Quaternary, respectively, were
479 mapped (Bristow and Hoffstetter, 1977; Cantalamessa et al., 2005; Cantalamessa et al.,
480 2007) and can be correlated with the positive delay observed in our inversion. This
481 difference in the geological structures is also consistent with the observation that cities far
482 from the epicenter (> 100 km) suffered severe damage due to site effects such as Portoviejo
483 and Chone where the vertical component of PGA reached up to 1.01 m/s² and 1.72 m/s²
484 (IGEPN, 2016) and the delay terms for the S-phase calculated were 2.08 s and 2.57 s,
485 respectively. The role of the fault system in the seismicity of the Central Coastal Ecuadorian
486 margin will be discussed in detail in the next section.

487

488 *5.2 Aftershocks distribution*

489 The majority of our relocated events with local magnitude greater than 3.5 are surrounding
490 the coseismic slip area determined by Nocquet et al (2017). This type of aftershock
491 distribution concentrated on the updip part of the rupture has been previously observed in
492 other megathrust earthquakes along the South American margin such as the Mw 8.8 Maule
493 2010 earthquake (Rietbrock et al., 2012) and the Mw 8.2 Iquique earthquake (Leon-Rios et
494 al., 2016).

495 The distribution of our relocated aftershocks along the margin are consistent with the
496 observations from Meltzer et al. (2019), Agurto-Detzel et al. (2019) that show seismicity
497 streaks aligned perpendicular to the trench (see Supplementary Material 14). The seismicity
498 distribution during the interseismic cycle (Font et al., 2013) is similar to the aftershock
499 sequence of the Pedernales earthquake and suggests that this behavior could be a regular
500 feature of the Ecuadorian subduction zone (see Supplementary Material 14).

501 Cross sections AA' and BB', in Figure 6, show that the depth distribution of the events is
502 consistent with Slab1.0 from Hayes et al. (2012). While the northern section is in a good
503 agreement with the projection of the slab at distances greater than 50 km eastward from

504 the trench, the southern profile shows seismicity that could indicate a shallower plate
505 interface (see profile BB' in Supplementary Material 8). The subduction of the CR helps to
506 explain this distribution due to the addition of a more buoyant oceanic crust which causes
507 the raising of the seismogenic interface zone. This has been previously proposed by Collot et
508 al. (2004) and Gailler et al. (2007) using active seismic methods in the forearc region of the
509 Ecuadorian margin.

510 It is also possible to identify clustered seismicity in both sections, north and south, that are
511 located in the overriding plate. This type of activity might be caused by the activation of
512 crustal faults due to changes in Coulomb stress in the area surrounding the coseismic slip
513 (e.g. Ryder et al., 2012).

514 As previously mentioned, the aftershock seismicity, located in the northern profile AA',
515 occurs very close to the trench, up to ~5 km – 10 km, which we confirmed using the OBS
516 stations offshore. Usually, seismicity along subduction margins does not extend up to areas
517 close to the trench axis. Although this phenomena has been previously observed when an
518 accretionary prism is present in the subduction zone, even in a erosional regime it is
519 uncommon to observe seismicity with such proximity to the trench (i.e. 2010 Maule, Chile
520 (Rietbrock et al., (2012); Lange et al., (2012); 2015 Mw 8.3 Illapel, Chile (Ruiz et al., 2016);
521 2014 Mw 8.2 Iquique, Chile (Leon-Rios et al., 2016); New Zealand (Anderson & Webb, 2010);
522 Japan (Asano et al., 2011)). This activity might be related to (1) intraplate deformation due
523 to the bending of the oceanic crust that can causes seismicity in the subducting slab and/or
524 to (2) the absence of a frontal accretionary prism that allows the locking of the megathrust
525 to extend up to the trench, causing seismicity, even at shallow depths.

526 Finally, the presence of subducting seamounts acting as erosional agents, helps to create a
527 scenario which might promote locking and allow seismicity to extend up to the trench along
528 zones of weakness activated after large earthquakes.

529 *5.3 Deformation in the marine forearc and upper crust*

530 Previous studies by Agurto-Detzel et al., (2019) and Meltzer et al., (2019) have described the
531 overall deformation caused by the Pedernales earthquake. As we now have developed a
532 robust minimum 1D model for the area affected by the 2016 mainshock and also manually
533 determined high precision onset times, we are able to constrain the hypocentral depths in

534 greater detail. We, therefore, will concentrate our discussion on the offshore area close to
535 Cabo Pasado (see Figure 7) and analyze the deformation in the marine forearc and upper
536 crust based on the diverse seismicity found in this area. The distribution and mechanisms of
537 the 14 calculated RMTs in this study are superimposed over our event locations in Figure 6.
538 Thrusting focal mechanisms from the gCMT catalogues (Esktrom et al., 2012) are also shown
539 and highlight the diversity of our obtained solutions.

540 In profile AA' we found two events with extensional components (e3 and e27 in Table 2)
541 occurring in the overriding plate. Our RMT solutions agree with the suggestion by Marcaillou
542 et al. (2016) who proposed that seamounts, part of the Atacames chain, could cause the
543 activation of normal faults in the marine forearc of the overriding plate. Normal faulting in
544 the marine forearc has been previously observed in the South American margin (Ruiz et al.,
545 2014), especially in areas within erosive regime like in central Ecuador, where the overriding
546 plate contains a small accretionary prism and a fractured, eroded and hydrated wedge that
547 is easy to break.

548 Across profile BB', we observed a clear pattern of strike-slip mechanisms distributed along
549 the marine forearc (near the plate interface) and surrounding the scarps in front of Cabo
550 Pasado. This strike-slip sequence lasted for two days between the 7th and 8th July 2016.
551 Figure 7 shows in detail the distribution of those events. This type of seismic activity in the
552 area has been previously observed by Vaca et al., (2017, fig 2.7) close to La Plata island (~
553 1°S).

554 Reactivation of preexisting normal faults in the upper crust after large megathrust
555 earthquakes has been recently observed in many subduction zone settings (i.e. Farias et
556 al.,2011, Ryder et al., 2012, Asano et al., 2011, Kato et al., 2011, Toda et al., 2013). Although
557 normal fault mechanisms are also detected for the Pedernales aftershock sequence we
558 observe a considerable amount of strike-slip seismicity that might be related to the
559 following possible causes:

560 (1) The pattern followed by the strike-slip events is consistent with the projection of the
561 crest of the CR derived by Pilger, 1984 (see Figure 7). Around 80 km to the north of the CR,
562 the forearc was affected by the subduction of the Atacames seamounts (Collot et al., 2005;
563 Marcaillou et al., 2016) causing deformation on the margin and creating two scarps of 25 to
564 40 km wide separated by a shallow promontory (Collot et al., 2009) (see Figure 7). This

565 suggests that subduction of the CR beneath the South American plate might create large
566 scale deformation and reactivate transform faulting in the overriding plate.

567 (2) The location of the offshore Cabo Pasado seismicity is coincident with a rotational block
568 system proposed by Daly (1989) that extends along the whole Pedernales segment. This
569 system was created 45 Ma ago in the Mid-Late Eocene because of the rapid convergence
570 rate between the Nazca and the South American plates. The interaction of the two
571 convergent plates plus the presence of a major shear fault could have caused faulting in the
572 forearc between the trench and the coast, creating several independent blocks in the
573 Pedernales segment. Rapid changes in the relative velocities of these blocks could have
574 caused both faults and blocks to have been rotated within the bounding shear zones,
575 generating zones of strike-slip faulting at the edges of each block. These features remained
576 inactive after the stabilization of convergence from the Late Miocene (~10 Ma) to the
577 present day. Possible reactivation of the strike-slip faults might be related to the occurrence
578 of large megathrust earthquake.

579 (3) Seismicity is associated with the possible extension offshore of the Jama Fault System
580 (JFS) that also coincides with the area where the crest of the CR is located. The JFS could be
581 explained as a large-scale system that was generated to release the stress accumulated by
582 the interaction between the CCPP fault zone and the convergence margin. Collot et al.,
583 (2004) proposed the JFS as an active transcurrent fault that extends offshore which was
584 confirmed by Hernandez et al., (2011) and Michaud et al., (2015). It is possible that the
585 occurrence of a large megathrust earthquake triggered the nucleation of seismicity along
586 this fault.

587 In our opinion, a combination between hypothesis (1) and (3) is most likely to explain the
588 observed activity because large-scale strike-slip system as the JFS normally do not extend in
589 one single line but in several strands, covering an entire region as has been observed inland
590 for the same system. In addition, the interaction between the CR and the upper crust where
591 the JFS extends might contribute to the generation of this type of seismicity. These two
592 factors, in addition to a large megathrust earthquake, such as the 2016 Pedernales event,
593 could have provided the conditions to reactivate a strike-slip fault in the marine forearc.
594 This offshore feature might have an important role as a barrier for large megathrust

595 earthquakes by absorbing the stress released during the reverse coseismic and transform it
596 into strike-slip displacement (e.g. Collot et al., 2004).

597 Whatever the cause is, the occurrence of seismicity associated with the reactivation of a
598 strike-slip fault in the marine forearc is a remarkable observation. Upper crustal activity,
599 related to normal faulting has been often documented: e.g. Pichilemu in central Chile (Farías
600 et al, 2011; Ryder et al., 2012) and also in Japan (Kato et al., 2011). It is therefore interesting
601 to note that strike-slip faults can also be reactivated and should, therefore, be mapped in
602 order to update the hazards map in the region.

603

604 6. CONCLUSIONS

605 We inferred a minimum 1D model for the Pedernales segment based on the aftershock
606 sequence ($M_L > 3.5$) of the Mw 7.8 Pedernales earthquake. The velocity structure for both P-
607 and S-phases was obtained after manually picking 345 events recorded by the permanent
608 Ecuadorian seismic network and a temporary array that includes more than 70 inland and
609 marine stations installed one month after the mainshock. P- and S- arrival times from 227
610 earthquakes with gap $< 200^\circ$ were inverted with VELEST to obtain absolute velocities,
611 hypocentral locations and station correction terms.

612 The obtained minimum 1D velocity model shows a good resolution down to 40 km depth
613 constrained with the ray paths and the earthquake distribution. The area has an average
614 V_p/V_s ratio of 1.82 which varies with depth that may be related to hydration and
615 serpentinization in the downgoing plate.

616 The seismicity is distributed in streaks which align perpendicular to the trench. Cross
617 sections allow us to identify the seismogenic contact between the Nazca and South
618 American plates. Shallow seismicity was observed in both northern and southern segments
619 of the rupture area suggesting the activation of faults in the overriding plate. Some of the
620 observed seismicity reached the trench which suggests the absence of a frontal accretionary
621 prism.

622 Regional moment tensors were calculated to analyze the source mechanism of some events
623 offshore. Although the majority of the events show focal mechanisms consistent with the
624 subduction process, we also observe extensional faulting in the marine forearc that can be

625 associated to the subduction of already-mapped seamounts. We also found strike-slip
626 faulting which might be related to the reactivation of a strike-slip structure after the 2016
627 Mw 7.8 Pedernales earthquake that, in combination with the subduction of the crest of the
628 CR, might cause the nucleation of the observed seismicity along the marine forearc margin.
629 The observation of this type of activity suggests the need to reevaluate the geological
630 structures in the marine forearc and to update the seismic hazard map for this region
631 including the possible scenario of a large strike-slip event.

632

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648

649 References (APA)

650 Abe, K. (1979). Size of great earthquakes of 1837–1974 inferred from tsunami data. *Journal*
651 *of Geophysical Research: Solid Earth*, 84(B4), 1561-1568.

652 Agudelo, W., Ribodetti, A., Collot, J. Y., & Operto, S. (2009). Joint inversion of multichannel
653 seismic reflection and wide-angle seismic data: Improved imaging and refined velocity

654 model of the crustal structure of the North Ecuador–South Colombia convergent margin.
655 *Journal of Geophysical Research: Solid Earth*, 114(B2).

656 Agurto-Detzel, H., Font, Y., Charvis, P., Rietbrock, A., Ambrois, D., Paullato, M., ... Soto-
657 Cordero, L. (2019). Ridge subduction and afterslip control aftershock distribution of the
658 2016 Mw 7.8 Ecuador earthquake. *Earth and Planetary Science Letters*.

659 Agurto, H., Rietbrock, A., Barrientos, S., Bataille, K., & Legrand, D. (2012). Seismo-tectonic
660 structure of the Aysén Region, Southern Chile, inferred from the 2007 M w= 6.2 Aysén
661 earthquake sequence. *Geophysical Journal International*, 190(1), 116-130.

662 Alvarado, A., Audin, L., Nocquet, J. M., Jaillard, E., Mothes, P., Jarrín, P., ... & Cisneros, D.
663 (2016). Partitioning of oblique convergence in the Northern Andes subduction zone:
664 Migration history and the present-day boundary of the North Andean Sliver in Ecuador.
665 *Tectonics*, 35(5), 1048-1065.

666 Alvarado, A., Ruiz, M., Mothes, P., Yepes, H., Segovia, M., Vaca, M., ... & Aguilar, J. (2018).
667 Seismic, volcanic, and geodetic networks in Ecuador: Building capacity for monitoring and
668 research. *Seismological Research Letters*, 89(2A), 432-439.

669 Anderson, H., & Webb, T. (1994). New Zealand seismicity: patterns revealed by the
670 upgraded National Seismograph Network. *New Zealand journal of geology and geophysics*,
671 37(4), 477-493.

672 Araujo, S. (2016). *Travel time tomography of the crust and the mantle beneath Ecuador from*
673 *data of the national seismic network* (Doctoral dissertation, Grenoble Alpes).

674 Asano, Y., Saito, T., Ito, Y., Shiomi, K., Hirose, H., Matsumoto, T., ... & Sekiguchi, S. (2011).
675 Spatial distribution and focal mechanisms of aftershocks of the 2011 off the Pacific coast of
676 Tohoku Earthquake. *Earth, planets and space*, 63(7), 29.

677 Bangs, N. L., McIntosh, K. D., Silver, E. A., Kluesner, J. W., & Ranero, C. R. (2015). Fluid
678 accumulation along the Costa Rica subduction thrust and development of the seismogenic
679 zone. *Journal of Geophysical Research: Solid Earth*, 120(1), 67-86.

680 Barnes, P. M., Lamarche, G., Bialas, J., Henrys, S., Pecher, I., Netzeband, G. L., ... & Crutchley,
681 G. (2010). Tectonic and geological framework for gas hydrates and cold seeps on the
682 Hikurangi subduction margin, New Zealand. *Marine Geology*, 272(1-4), 26-48.

683 Beck, S. L., and L. J. Ruff (1984). The rupture process of the great 1979 Colombia
684 earthquake: Evidence for the asperity model, *J. Geophys. Res.*, 89, 9281-9291.

685 Beauval, C., Yepes, H., Bakun, W. H., Egred, J., Alvarado, A., & Singaicho, J. C. (2010).
686 Locations and magnitudes of historical earthquakes in the Sierra of Ecuador (1587–1996).
687 *Geophysical Journal International*, 181(3), 1613-1633.

688 Bilek, S. L. (2010). Invited review paper: Seismicity along the South American subduction
689 zone: Review of large earthquakes, tsunamis, and subduction zone complexity.
690 *Tectonophysics*, 495(1-2), 2-14.

691 Bloch, W., John, T., Kummerow, J., Salazar, P., Krüger, O. S., & Shapiro, S. A. Watching
692 Dehydration: Seismic Indication for Transient Fluid Pathways in the Oceanic Mantle of the
693 Subducting Nazca Slab. *Geochemistry, Geophysics, Geosystems*.

694 Bolton, P. A. (1991). Local Level Economic and Social Consequences. *The March 5, 1987,*
695 *Ecuador earthquakes: mass wasting and socioeconomic effects*, 100.

696 Bristow, C. R., Hoffstetter, R., Feininger, T., & Hall, M. T. (1977). Léxico estratigráfico del
697 Ecuador. *CENTRE NATIONAL DE LA RECHERCHE SCIENTIFIQUE*.

698 Calahorrano, A. (2001). *Estudio del origen del enjambre sísmico de la zona norte de la ciudad*
699 *de Quito, durante 1998–99. Escuela Politécnica Nacional, Facultad de Geología, Minas y*
700 *Petróleos* (Doctoral dissertation, Tesis de Ingeniera Geóloga, Quito, Ecuador).

701 Calahorrano, A., Sallarès, V., Collot, J. Y., Sage, F., & Ranero, C. R. (2008). Nonlinear
702 variations of the physical properties along the southern Ecuador subduction channel:
703 Results from depth-migrated seismic data. *Earth and Planetary Science Letters*, 267(3-4),
704 453-467.

705 Cano, L. C. G., Galve, A., Charvis, P., & Marcaillou, B. (2014). Three-dimensional velocity
706 structure of the outer fore arc of the Colombia-Ecuador subduction zone and implications
707 for the 1958 megathrust earthquake rupture zone. *Journal of Geophysical Research: Solid*
708 *Earth*, 119(2), 1041-1060.

709 Cantalamessa, G., & Di Celma, C. (2005). Sedimentary features of tsunami backwash
710 deposits in a shallow marine Miocene setting, Mejillones Peninsula, northern Chile.
711 *Sedimentary Geology*, 178(3-4), 259-273.

712 Cantalamessa, G., Di Celma, C., Ragaini, L., Valleri, G., & Landini, W. (2007). Sedimentology
713 and high-resolution sequence stratigraphy of the late middle to late Miocene Angostura
714 Formation (western Borbón Basin, northwestern Ecuador). *Journal of the Geological Society*,
715 *164*(3), 653-665.

716 Chlieh, M., Mothes, P. A., Nocquet, J. M., Jarrin, P., Charvis, P., Cisneros, D., ... & Vallée, M.
717 (2014). Distribution of discrete seismic asperities and aseismic slip along the Ecuadorian
718 megathrust. *Earth and Planetary Science Letters*, *400*, 292-301.

719 Collings, R., Lange, D., Rietbrock, A., Tilmann, F., Natawidjaja, D., Suwargadi, B., ... & Saul, J.
720 (2012). Structure and seismogenic properties of the Mentawai segment of the Sumatra
721 subduction zone revealed by local earthquake traveltimes tomography. *Journal of*
722 *Geophysical Research: Solid Earth*, *117*(B1).

723 Collot, J. Y., Charvis, P., Gutscher, M. A., & Operto, S. (2002). Exploring the
724 Ecuador-Colombia active margin and interplate seismogenic zone. *EOS, Transactions*
725 *American Geophysical Union*, *83*(17), 185-190.

726 Collot, J. Y., Marcaillou, B., Sage, F., Michaud, F., Agudelo, W., Charvis, P., ... & Spence, G.
727 (2004). Are rupture zone limits of great subduction earthquakes controlled by upper plate
728 structures? Evidence from multichannel seismic reflection data acquired across the northern
729 Ecuador–southwest Colombia margin. *Journal of Geophysical Research: Solid Earth*,
730 *109*(B11).

731 Collot, J. Y., Migeon, S., Spence, G., Legonidec, Y., Marcaillou, B., Schneider, J. L., ... &
732 Pazmino, A. (2005). Seafloor margin map helps in understanding subduction earthquakes.
733 *Eos, Transactions American Geophysical Union*, *86*(46), 463-465.

734 Collot, J. Y., Michaud, F., Alvarado, A., Marcaillou, B., Sosson, M., Ratzov, G., ... & Pazmino,
735 A. (2009). Visión general de la morfología submarina del margen convergente de Ecuador-
736 Sur de Colombia: implicaciones sobre la transferencia de masa y la edad de la subducción de
737 la Cordillera de Carnegie. *En Geología y Geofísica Marina y Terrestre del Ecuador*.

738 Collot, J. Y., Sanclemente, E., Nocquet, J. M., Leprêtre, A., Ribodetti, A., Jarrin, P., ... &
739 Charvis, P. (2017). Subducted oceanic relief locks the shallow megathrust in central Ecuador.
740 *Journal of Geophysical Research: Solid Earth*, *122*(5), 3286-3305.

741 Daly, M. C. (1989). Correlations between Nazca/Farallon plate kinematics and forearc basin
742 evolution in Ecuador. *Tectonics*, 8(4), 769-790.

743 Dorbath, L., Cisternas, A., & Dorbath, C. (1990). Assessment of the size of large and great
744 historical earthquakes in Peru. *Bulletin of the Seismological Society of America*, 80(3), 551-
745 576.

746 Dziewonski, A. M., Chou, T. A., & Woodhouse, J. H. (1981). Determination of earthquake
747 source parameters from waveform data for studies of global and regional seismicity. *Journal*
748 *of Geophysical Research: Solid Earth*, 86(B4), 2825-2852.

749 Egred, J. (1968). Breve historia sísmica de la República del Ecuador (1534-1965). *Bol. Bibl.*
750 *Geofís. Oceanogr. Am.*

751 Ekström, G., Nettles, M., & Dziewoński, A. M. (2012). The global CMT project 2004–2010:
752 Centroid-moment tensors for 13,017 earthquakes. *Physics of the Earth and Planetary*
753 *Interiors*, 200, 1-9.

754 Farías, M., Comte, D., Roecker, S., Carrizo, D., & Pardo, M. (2011). Crustal extensional
755 faulting triggered by the 2010 Chilean earthquake: The Pichilemu Seismic Sequence.
756 *Tectonics*, 30(6).

757 Font, Y., Segovia, M., Vaca, S., & Theunissen, T. (2013). Seismicity patterns along the
758 Ecuadorian subduction zone: new constraints from earthquake location in a 3-D a priori
759 velocity model. *Geophysical Journal International*, 193(1), 263-286.

760 Gailler, A., Charvis, P., & Flueh, E. R. (2007). Segmentation of the Nazca and South American
761 plates along the Ecuador subduction zone from wide angle seismic profiles. *Earth and*
762 *Planetary Science Letters*, 260(3-4), 444-464.

763 Graindorge, D., Calahorrano, A., Charvis, P., Collot, J. Y., & Bethoux, N. (2004). Deep
764 structures of the Ecuador convergent margin and the Carnegie Ridge, possible consequence
765 on great earthquakes recurrence interval. *Geophysical Research Letters*, 31(4).

766 Hayes, G. P., Wald, D. J., & Johnson, R. L. (2012). Slab1. 0: A three-dimensional model of
767 global subduction zone geometries. *Journal of Geophysical Research: Solid Earth*, 117(B1).

768 Hacker, B. R., Abers, G. A., & Peacock, S. M. (2003). Subduction factory 1. Theoretical
769 mineralogy, densities, seismic wave speeds, and H₂O contents. *Journal of Geophysical*
770 *Research: Solid Earth*, 108(B1).

771 Hacker, B. R., Peacock, S. M., Abers, G. A., & Holloway, S. D. (2003). Subduction factory 2.
772 Are intermediate-depth earthquakes in subducting slabs linked to metamorphic dehydration
773 reactions?. *Journal of Geophysical Research: Solid Earth*, 108(B1).

774 Herd, D. G., Youd, T. L., Meyer, H., Person, W. J., & Mendoza, C. (1981). The great tumaco,
775 colombia earthquake of 12 december 1979. *Science*, 211(4481), 441-445.

776 Hernández, M. J., Michaud, F., Rivadeneira, M., Barba, D., & Collot, J. Y. (2011). El sistema de
777 fallas de Jama en la plataforma continental (off-shore) del margen ecuatoriano: resultados
778 preliminares a partir de la interpretación de perfiles sísmicos de reflexión.

779 Hicks, S. P., Rietbrock, A., Ryder, I. M., Lee, C. S., & Miller, M. (2014). Anatomy of a
780 megathrust: The 2010 M8. 8 Maule, Chile earthquake rupture zone imaged using seismic
781 tomography. *Earth and Planetary Science Letters*, 405, 142-155.

782 Husen, S., Kissling, E., Flueh, E., & Asch, G. (1999). Accurate hypocentre determination in the
783 seismogenic zone of the subducting Nazca Plate in northern Chile using a combined on-
784 /offshore network. *Geophysical Journal International*, 138(3), 687-701.

785 Husen, S., & Kissling, E. (2001). Postseismic fluid flow after the large subduction earthquake
786 of Antofagasta, Chile. *Geology*, 29(9), 847-850.

787 IGEPN, 2016 informe sismico n18, [https://www.igepn.edu.ec/servicios/noticias/1324-](https://www.igepn.edu.ec/servicios/noticias/1324-informe-sismico-especial-n-18-2016)
788 [informe-sismico-especial-n-18-2016](https://www.igepn.edu.ec/servicios/noticias/1324-informe-sismico-especial-n-18-2016)

789 International Seismological Centre, *On-line Bulletin*, <http://www.isc.ac.uk>, Internatl. Seismol.
790 Cent., Thatcham, United Kingdom, 2015.

791 Jaillard, E., Benitez, S., Berrones, G., Huaman, C., Ordonez, M., & Rivadeneira, M. (1994).
792 Analisis de cuencas sedimentarias e historia precoz de los Andes.

793 Jaillard, E., Hérail, G., Monfret, T., Díaz-Martínez, E., Baby, P., Lavenue, A., & Dumont, J. F.
794 (2000). Tectonic evolution of the Andes of Ecuador, Peru, Bolivia and northernmost Chile.
795 *Tectonic Evolution of South America*, 31, 481-559.

796 Kanamori, H., & Given, J. W. (1981). Use of long-period surface waves for rapid
797 determination of earthquake-source parameters. *Physics of the Earth and Planetary*
798 *interiors*, 27(1), 8-31.

799 Kanamori, H., & McNally, K. C. (1982). Variable rupture mode of the subduction zone along
800 the Ecuador-Colombia coast. *Bulletin of the Seismological Society of America*, 72(4), 1241-
801 1253.

802 Kato, A., Iidaka, T., Ikuta, R., Yoshida, Y., Katsumata, K., Iwasaki, T., ... & Watanabe, T. (2010).
803 Variations of fluid pressure within the subducting oceanic crust and slow earthquakes.
804 *Geophysical research letters*, 37(14).

805 Kato, A., Sakai, S. I., & Obara, K. (2011). A normal-faulting seismic sequence triggered by the
806 2011 off the Pacific coast of Tohoku Earthquake: Wholesale stress regime changes in the
807 upper plate. *Earth, planets and space*, 63(7), 43.

808 Kato, A., & Nakagawa, S. (2014). Multiple slow-slip events during a foreshock sequence of
809 the 2014 Iquique, Chile Mw 8.1 earthquake. *Geophysical Research Letters*, 41(15), 5420-
810 5427.

811 Kelleher, J. A. (1972). Rupture zones of large South American earthquakes and some
812 predictions. *Journal of Geophysical Research*, 77(11), 2087-2103.

813 Kendrick, E., Bevis, M., Smalley Jr, R., Brooks, B., Vargas, R. B., Lauria, E., & Fortes, L. P. S.
814 (2003). The Nazca–South America Euler vector and its rate of change. *Journal of South*
815 *American Earth Sciences*, 16(2), 125-131.

816 Kissling, E., Ellsworth, W. L., Eberhart-Phillips, D., & Kradolfer, U. (1994). Initial reference
817 models in local earthquake tomography. *Journal of Geophysical Research: Solid Earth*,
818 99(B10), 19635-19646.

819 Lange, D., Tilmann, F., Barrientos, S. E., Contreras-Reyes, E., Methe, P., Moreno, M., ... &
820 Beck, S. (2012). Aftershock seismicity of the 27 February 2010 Mw 8.8 Maule earthquake
821 rupture zone. *Earth and Planetary Science Letters*, 317, 413-425.

822 Lanning, F., Haro, A. G., Liu, M. K., Monzón, A., Monzón-Despang, H., Schultz, A., ... & Vera-
823 Grunauer, X. (2016). EERI Earthquake Reconnaissance Team Report: M7. 8 Muisne, Ecuador
824 Earthquake on April 16, 2016.

825 Lee, W. H. K., & Lahr, J. C. (1972). *HYP071: A computer program for determining hypocenter,*
826 *magnitude, and first motion pattern of local earthquakes* (No. 72-224). US Geological
827 Survey,.

828 León-Ríos, S., Ruiz, S., Maksymowicz, A., Leyton, F., Fuenzalida, A., & Madariaga, R. (2016).
829 Diversity of the 2014 Iquique's foreshocks and aftershocks: clues about the complex rupture
830 process of a Mw 8.1 earthquake. *Journal of Seismology*, 20(4), 1059-1073.

831 Litherland, M. (1994). The metamorphic belts of Ecuador. *British Geological Survey, overseas*
832 *Memoir.*, 11.

833 Lonsdale, P. (2005). Creation of the Cocos and Nazca plates by fission of the Farallon plate.
834 *Tectonophysics*, 404(3), 237-264.

835 Luzieux, L. D. A., Heller, F., Spikings, R., Vallejo, C. F., & Winkler, W. (2006). Origin and
836 Cretaceous tectonic history of the coastal Ecuadorian forearc between 1 N and 3 S:
837 Paleomagnetic, radiometric and fossil evidence. *Earth and Planetary Science Letters*, 249(3-
838 4), 400-414.

839 Marcaillou, B., Spence, G., Collot, J. Y., & Wang, K. (2006). Thermal regime from bottom
840 simulating reflectors along the north Ecuador–south Colombia margin: Relation to margin
841 segmentation and great subduction earthquakes. *Journal of Geophysical Research: Solid*
842 *Earth*, 111(B12).

843 Marcaillou, B., Collot, J. Y., Ribodetti, A., d'Acremont, E., Mahamat, A. A., & Alvarado, A.
844 (2016). Seamount subduction at the North-Ecuadorian convergent margin: Effects on
845 structures, inter-seismic coupling and seismogenesis. *Earth and Planetary Science Letters*,
846 433, 146-158.

847 Meissnar, R. O., Flueh, E. R., Stibane, F., & Berg, E. (1976). Dynamics of the active plate
848 boundary in southwest Colombia according to recent geophysical measurements.
849 *Tectonophysics*, 35(1-3), 115-136.

850 Meltzer, A., Beck, S., Ruiz, M., Hoskins, M., Soto-Cordero, L., Stachnik, J. C., ... Mercerat, E. D.
851 (2019). The 2016 Mw 7.8 Pedernales Earthquake, Ecuador: RAPID Response Deployment.
852 *Seismological Research Letters*.

853

854 Mendoza, C., & Dewey, J. W. (1984). Seismicity associated with the great Colombia-Ecuador
855 earthquakes of 1942, 1958, and 1979: Implications for barrier models of earthquake
856 rupture. *Bulletin of the seismological society of America*, 74(2), 577-593.

857 Michaud, F., Proust, J. N., Collot, J. Y., Lebrun, J. F., Witt, C., Ratzov, G., ... & Penafiel, L.
858 (2015). Quaternary sedimentation and active faulting along the Ecuadorian shelf:
859 preliminary results of the ATACAMES Cruise (2012). *Marine Geophysical Research*, 36(1), 81-
860 98.

861 Mothes, P. A., Nocquet, J. M., & Jarrín, P. (2013). Continuous GPS network operating
862 throughout Ecuador. *Eos, Transactions American Geophysical Union*, 94(26), 229-231.

863 Nocquet, J. M., Jarrin, P., Vallée, M., Mothes, P. A., Grandin, R., Rolandone, F., ... & Régnier,
864 M. (2017). Supercycle at the Ecuadorian subduction zone revealed after the 2016
865 Pedernales earthquake. *Nature Geoscience*, 10(2), 145.

866 Pasten-Araya, F., Salazar, P., Ruiz, S., Rivera, E., Potin, B., Maksymowicz, A., et al (2018).
867 Fluids along the plate interface influencing the frictional regime of the Chilean Subduction
868 zone, northern Chile. *Geophysical Research Letters*, 45.
869 <https://doi.org/10.1029/2018GL079283>

870 Peacock, S. M. (2001). Are the lower planes of double seismic zones caused by serpentine
871 dehydration in subducting oceanic mantle?. *Geology*, 29(4), 299-302.

872 Perfettini, H., Avouac, J. P., Tavera, H., Kositsky, A., Nocquet, J. M., Bondoux, F., ... & Soler, P.
873 (2010). Seismic and aseismic slip on the Central Peru megathrust. *Nature*, 465(7294), 78.

874 Pilger, R. H. (1984). Cenozoic plate kinematics, subduction and magmatism: South American
875 Andes. *Journal of the Geological Society*, 141(5), 793-802.

876 Reyes, P., & Michaud, F. (2012). Mapa Geologico de la Margen Costera Ecuatoriana
877 (1500000) EPPetroEcuador-IRD. *Quito Ecuador*.

878 Rietbrock, A., Ryder, I., Hayes, G., Haberland, C., Comte, D., Roecker, S., & Lyon-Caen, H.
879 (2012). Aftershock seismicity of the 2010 Maule Mw= 8.8, Chile, earthquake: Correlation
880 between co-seismic slip models and aftershock distribution?. *Geophysical Research Letters*,
881 39(8).

882 Rolandone, F., Nocquet, J. M., Mothes, P. A., Jarrin, P., Vallée, M., Cubas, N., ... & Font, Y.
883 (2018). Areas prone to slow slip events impede earthquake rupture propagation and
884 promote afterslip. *Science advances*, 4(1), eaao6596.

885 Rothé, J. P. (1969). *The Seismicity of the Earty, 1953-1965: La Séismicité Du Globe 1953,*
886 *1965.* Unesco.

887 Ruiz, S., Klein, E., del Campo, F., Rivera, E., Poli, P., Metois, M., ... & Madariaga, R. (2016).
888 The seismic sequence of the 16 September 2015 M w 8.3 Illapel, Chile, earthquake.
889 *Seismological Research Letters*, 87(4), 789-799.

890 Ruiz, S., Metois, M., Fuenzalida, A., Ruiz, J., Leyton, F., Grandin, R., ... & Campos, J. (2014).
891 Intense foreshocks and a slow slip event preceded the 2014 Iquique Mw 8.1 earthquake.
892 *Science*, 345(6201), 1165-1169.

893 Ryder, I., Rietbrock, A., Kelson, K., Bürgmann, R., Floyd, M., Socquet, A., ... & Carrizo, D.
894 (2012). Large extensional aftershocks in the continental forearc triggered by the 2010 Maule
895 earthquake, Chile. *Geophysical Journal International*, 188(3), 879-890.

896 Sallarès, V., & Charvis, P. (2003). Crustal thickness constraints on the geodynamic evolution
897 of the Galapagos Volcanic Province. *Earth and Planetary Science Letters*, 214(3-4), 545-559.

898 Sallarès, V., Charvis, P., Flueh, E. R., Bialas, J., & SALIERI Scientific Party. (2005). Seismic
899 structure of the Carnegie ridge and the nature of the Galapagos hotspot. *Geophysical*
900 *Journal International*, 161(3), 763-788.

901 Segovia, M. (2001). El sismo de Bahía del 4 de agosto de 1998: Caracterización del
902 mecanismo de ruptura y análisis de la sismicidad en la zonacostera. *Titulo de Ingeniera*
903 *Geologia (tesis): Escuela Politecnica Nacional, Quito, Ecuador.*

904 Segovia, M., Alvarado, A., Collot, J. Y., Sallares, V., & Pazmiño, N. (2009). Breve análisis de la
905 sismicidad y del campo de esfuerzos en el Ecuador. *Geología y Geofísica Marina y Terrestre*
906 *del Ecuador: desde la costa continental hasta las Islas Galápagos*, 131-149.

907 Segovia, M., Font, Y., Regnier, M. M., Charvis, P., Nocquet, J. M., Galve, A., ... & Ruiz, M. C.
908 (2015, December). Intense microseismicity associated with a SSE at La Plata Island in the
909 central subduction zone of Ecuador. In *AGU Fall Meeting Abstracts*.

910 Sennson, J. L., & Beck, S. L. (1996). Historical 1942 Ecuador and 1942 Peru subduction
911 earthquakes and earthquake cycles along Colombia-Ecuador and Peru subduction segments.
912 *Pure and applied geophysics*, 146(1), 67-101.

913 Sokos, E. N., & Zahradnik, J. (2008). ISOLA a Fortran code and a Matlab GUI to perform
914 multiple-point source inversion of seismic data. *Computers & Geosciences*, 34(8), 967-977.

915 Sokos, E., & Zahradník, J. (2013). Evaluating centroid-moment-tensor uncertainty in the new
916 version of ISOLA software. *Seismological Research Letters*, 84(4), 656-665.

917 Storchak, D.A., Di Giacomo, I. Bondár, E. R. Engdahl, J. Harris, W.H.K. Lee, A. Villaseñor
918 and P. Bormann, 2013. Public Release of the ISC-GEM Global Instrumental Earthquake
919 Catalogue (1900-2009). *Seism. Res. Lett.*, 84, 5, 810-815, doi: 10.1785/0220130034.

920 Toda, S., & Tsutsumi, H. (2013). Simultaneous reactivation of two, subparallel, inland normal
921 faults during the M w 6.6 11 April 2011 Iwaki earthquake triggered by the M w 9.0
922 Tohoku-Oki, Japan, earthquake. *Bulletin of the Seismological Society of America*, 103(2B),
923 1584-1602.

924 Vaca, S., Régnier, M., Bethoux, N., Alvarez, V., & Pontoise, B. (2009). Sismicidad de la región
925 de Manta: Enjambre sísmico de Manta-2005. *Geología y Geofísica Marina y Terrestre del*
926 *Ecuador Desde la Costa Continental Hasta las Islas Gapapagos*, 151-166.

927 Vaca, S. (2017). Interaction between tectonics, slow-slip processes and earthquake
928 mechanisms across the North Andean sliver (Ecuador). Thèse de doctorat de Sciences de la
929 Terre et de l'environnement. INSTITUT DE PHYSIQUE DU GLOBE DE PARIS.

930 Vaca, S., Vallée, M., Nocquet, J. M., Battaglia, J., & Régnier, M. (2018). Recurrent slow slip
931 events as a barrier to the northward rupture propagation of the 2016 Pedernales
932 earthquake (Central Ecuador). *Tectonophysics*, 724, 80-92.

933 Vallée, M., Nocquet, J. M., Battaglia, J., Font, Y., Segovia, M., Regnier, M., ... & Yepes, H.
934 (2013). Intense interface seismicity triggered by a shallow slow slip event in the Central
935 Ecuador subduction zone. *Journal of Geophysical Research: Solid Earth*, 118(6), 2965-2981.

936 Wadati, K., & Oki, S. (1933). On the travel time of earthquake waves.(Part II). *Journal of the*
937 *Meteorological Society of Japan. Ser. II*, 11(1), 14-28.

938 White, S. M., Trenkamp, R., & Kellogg, J. N. (2003). Recent crustal deformation and the
939 earthquake cycle along the Ecuador–Colombia subduction zone. *Earth and Planetary Science*
940 *Letters*, 216(3), 231-242.