

The structure of Mediterranean arcs: new insights from the Calabrian Arc subduction system

Prada, M.¹, Ranero, C.R.², Sallares, V.¹, Grevemeyer, I.³, Roberto de Franco⁴, Gervasi.,
A.^{5,6}, Zitellini, N.⁷

(1) Barcelona-CSI, Inst. de Ciències del Mar-CSIC, Barcelona, Spain

(2) Barcelona-CSI, Inst. de Ciències del Mar, ICREA at CSIC, Barcelona, Spain

(3) GEOMAR Helmholtz Centre for Ocean Research Kiel, Germany

(4) Istituto di Geologia Ambientale e Geoingegneria IGAG-CNR, Milano, Italy

(5) Istituto Nazionale di Geofisica e Vulcanologia, Osservatorio Nazionale Terremoti,
sede di Rende, Italy

(6) Dipartimento di Biologia, Ecologia e Scienze della Terra, Università della Calabria,
Rende, Italy

(7) Istituto Scienze Marine ISMAR-CNR, Bologna, Italy

Abstract

The formation of Cenozoic mountain belts in the Mediterranean realm was preceded by tens of millions of years of subduction, forming volcanic arcs, and frontal contractional systems. In addition, subduction usually involves slab rollback and formation of oceanic backarcs. Although such structure must have influenced the orogeny of Mediterranean mountain belts, no active analog has been mapped with modern crustal-scale seismic methods. Here, we study the entire Calabrian subduction system to map the structure resulting from Tethys lithosphere subduction and slab roll back, in a process that must be akin to that operating during a phase of the formation of the Mediterranean orogenic belts. We present a crustal-scale cross section of the entire Calabrian subduction system

24 obtained from on- and off-shore wide-angle seismic data. The 2D P-wave velocity section
25 shows spatially abrupt (< 5 km of profile distance) structural and petrological transitions
26 from the Ionian sedimentary wedge and Calabrian arc, to the rifted NW Calabrian margin,
27 where the Quaternary Aeolian arc is emplaced. The margin, then, transitions northwards
28 into the Marsili backarc region, where exhumed mantle and localized volcanism occurred
29 during its formation. This complex structure implies rapid temporal and spatial changes
30 between magmatic and amagmatic processes, and between compressional and extensional
31 regimes during the evolution of this subduction system. We find that some terranes
32 involved in the Alpine orogeny share petrological and tectonic similarities with some
33 domains of the Calabrian subduction system. Based on the results of this study we propose
34 the Calabrian Arc system as an analog for the subduction structuration that preceded the
35 formation of Alpine orogenic systems.

36 Keywords: Calabrian Arc, subduction; Mantle exhumation; Travel-time tomography;
37 Wide-angle seismic data; Alpine orogeny; Mediterranean active arcs

38

39 1 Introduction

40 The Mediterranean realm includes a system of Cenozoic arcuate orogenic belts formed
41 during the Alpine Orogeny (Dewey et al., 1973) (Fig. 1a). Mantle tomography images
42 and plate reconstructions support that the collision events that formed these orogens were
43 preceded by subduction of Tethys lithosphere (e.g. Dewey et al., 1973; Schettino and
44 Turco, 2011). The evolution of the subduction systems must have often included slab
45 retreat (e.g. Handy et al., 2010), causing the migration of volcanic arcs, extension of the
46 overriding plate and formation of backarc basins possibly similar to the Tyrrhenian (Prada
47 et al., 2015, Loreto et al., 2020) and the Alboran-South Balearic basins (Gomez de la Peña
48 et al., 2018), and forearc contractional systems (e.g. Stampfli et al., 1998; Polonia et al.,
49 2011; Marroni et al., 2017), determining thereby, the pre-collisional structuration of the
50 lithosphere. The Gibraltar and Calabrian backarcs are no longer opening, and their
51 domains are currently being inverted, which may indicate a current initial phase of a
52 collision (Giaconia et al., 2015; Zitellini et al., 2020). Yet, the structure of Alpine-type
53 belts has been associated to the imbrication of the domains formed during continental
54 rifting (e.g. Reston and Manatchal 2011; Mohn et al., 2014), and the major lithospheric-
55 scale structuration created during subduction has been largely ignored, possibly because
56 of the fragmentary geological record in mountain belt that makes reconstructions
57 disputable, and the lack of present-day potential analogs.

58 Reconstructing the structure of past subduction systems from present-day orogens is a
59 challenging task because of the tectonic and metamorphic overprinting that terranes suffer
60 during orogeny, and the incomplete outcrop information. An alternative way to infer such
61 structure is by exploring the present-day subduction of the Tethys lithosphere in the
62 Mediterranean, as occurs under the Calabrian, Gibraltar and Hellenic arcs (Fig. 1a)

63 (Spakman and Wortel, 2004; Müller et al., 2008). Possibly similar to Alpine subductions
64 (e.g. Maffione and van Hinsbergen, 2018), the upper plate of these present-day
65 subduction systems has been shaped by slab retreat, backarc opening, and accretionary
66 forearc systems. Therefore, understanding of the forearc-to-backarc structure of present-
67 day subduction systems in the Mediterranean may provide insights into the processes that
68 shaped Alpine systems before collision.

69 In this work, we focus on the Calabrian subduction system, which includes the Tyrrhenian
70 backarc basin, the Aeolian volcanic arc, the NW Calabrian margin, the Calabrian arc and
71 the contractional Ionian wedge (Fig. 1a and 1b). Vintage (i.e. 1970) refraction studies
72 provided 1D crustal velocity information across the subduction system, bringing first-
73 order approximations on the crustal structure of the Calabrian arc (e.g. Cassinis et al.,
74 2003). Regional earthquake tomography of the Calabrian region has provided insights on
75 the structure of the slab and the petrological nature of the deep lithosphere and
76 asthenosphere (e.g. Caló et al., 2009). However, these studies do not provide enough
77 resolution to assess in detail the structure and petrological nature of shallow lithospheric
78 domains (i.e. crust and uppermost mantle) across the subduction system.

79 To assess the forearc-to-backarc structure of this system, we present a detailed P-wave
80 velocity (V_p) cross section of the entire subduction system using travel-time tomography
81 from wide-angle seismic data (WAS) (Fig. 1) acquired in 2015 during the CHIANTI
82 (Calabrian arc Hazards in IoniAN and TyrrhenIan seas) experiment (Fig. 1b). The 425
83 km-long model provides unprecedented information of the structure and petrology of
84 shallow lithospheric domains across the Calabrian subduction system from the forearc to
85 the backarc (Fig 1b).

86 2 The Calabrian Arc subduction system

87 Calabria has earthquakes activity up to ~500 km depth under the northern margin of
88 Calabria, delineating a ~70° NW dipping slab (Spakman and Wortel, 2004; Chiarabba et
89 al., 2005). Slab rollback has been driving backarc basin formation by upper plate
90 extension, similar to other regions of the Western and Central Mediterranean since the
91 Oligocene (~30Ma; Malinverno and Ryan, 1986; Schettino and Turco, 2011). After the
92 opening of the Liguro-Provençal basins, the east-southeast rollback of the Apennines-
93 Calabrian subduction system initiated the Tyrrhenian backarc basin in the early Tortonian
94 (~11 Ma) (Kastens and Mascle, 1990). During the Messinian (5.3-6 Ma) the potential
95 under-thrusting of the continental lithosphere of Adria stopped slab rollback and backarc
96 extension in the northern Tyrrhenian (Faccenna et al., 2001).

97 In contrast, the central and southern subduction front migrated towards the southeast
98 (Spakman and Wortel, 2004), opening the central and southern Tyrrhenian Basin, and
99 forming the Magnaghi and Vavilov basins during the Pliocene, and the Marsili basin
100 during the Quaternary (Beccaluva et al., 1990; Kastens and Mascle, 1990) (Fig. 1b).
101 Ocean Drilling Program (ODP) Leg 107 and Deep Sea Drilling Program (DSDP) Leg 42
102 recovered mid-ocean ridge basalts in the Magnaghi and Vavilov basins (Dietrich et al.,
103 1977; Beccaluva et al., 1990). Further basalts were also sampled at the top of the basement
104 in the Marsili basin (Kastens and Mascle, 1990) (Fig. 1b). Consequently, these basins
105 have commonly been interpreted as regions where break up led to oceanic spreading.
106 However, recent V_p models across the Vavilov and Magnaghi basins support that their
107 basement is mainly composed of exhumed mantle rocks, and that basaltic ridges in the
108 area are localized magmatic events (Prada et al., 2015, 2016). The occurrence of exhumed
109 mantle in these two basins is further supported by the drilling of serpentinized peridotites

110 at ODP site 651(Bonatti et al., 1990) (Fig. 1b).

111 The southeastward migration of the subduction system resulted in rifting of the NW
112 Calabrian margin and emplacement of the Quaternary Aeolian volcanic arc at < 1.2 Ma
113 (Argnani and Savelli, 1999), the tectonic Calabrian Peninsula, formed by Paleozoic to
114 Eocene metamorphic rocks covered by Cenozoic sedimentary units (Vitale et al., 2019),
115 and the Ionian contractional wedge (Polonia et al., 2011) (Fig. 1a).

116 3 Acquisition, processing and picking of wide-angle seismic data

117 The CHIANTI seismic experiment was conducted onboard the Spanish R/V Sarmiento
118 de Gamboa. The offshore data used in this study were acquired along two offshore
119 profiles, WAS1 in the Ionian Sea, and WAS2 in the Tyrrhenian backarc region (Fig.1b).

120 The offshore data were recorded by 14 and 17 LC2000 Ocean Bottom Seismometers
121 (OBS) of the Spanish National Research Council (CSIC) along profiles WAS1 and
122 WAS2, respectively. OBS spacing along both lines was ~ 10 km. The seismic source was
123 generated by two arrays of 9 airguns each, which released a total volume of 4760 cu. in.
124 (78 l), and worked at 15 m of depth to provide a range of frequencies between 7-45 Hz,
125 optimum for crustal-scale WAS experiments. Shot interval along both lines was 90s
126 (~230 m at 5 knots). Land stations in Calabria were installed by the University of
127 Calabria. In this study, we use 31 OBS records with useful data and 3 out of the 5 land
128 stations installed in Calabria, which recorded the shots along both WAS1 and WAS2 (Fig.
129 1).

130 We relocated all OBSs using an in-house Metropolis-Hastings algorithm that minimizes
131 the misfit function between the observed and synthetic first arrival travel-times of the
132 water wave to find the most likely location of the OBS. We applied a bandpass filter (1-
133 5-18-25 Hz) to all seismic records to enhance the seismic signal and manually picked

134 travel-times of first arrival and critical reflections. In the contractional wedge, near offset
135 (< 15 km) first arrival times from OBS record sections of profile WAS1 define a first
136 refracted seismic phase with an apparent velocity of 3-4 km/s that corresponds to refracted
137 P-waves through the shallow sedimentary sequence of the contractional wedge or P_s (Fig.
138 2a-b). Between 15-20 km offset a second refracted seismic phases with an apparent
139 velocity of 5-6 km/s can be identified in those OBS deployed near the shore (Fig. 2b). We
140 interpreted this latter phase as refracted P-waves through the uppermost basement of the
141 contractional wedge or P_b . Some sections show a prominent critical reflection at 15-20
142 km of offset, defining the transition between refracted phases P_s and P_b (Fig. 2a-b). We
143 interpreted this phase as P-waves reflected at the top of the acoustic basement or P_sP (Fig.
144 2a-b). The land record sections of line WAS1 only allowed the identification of seismic
145 phases P_s and P_sP , up to ~40 km offset (Fig. 2h).

146 From SE to NW in the NW Calabrian margin, OBS records along line WAS2 display a
147 first refracted seismic phase with apparent velocity of 6 km/s. This seismic phase is
148 interpreted as a P-wave refracted within the crust (P_g). At further offset (> 40 km offset),
149 we identified a second refracted phase, in this case with a faster apparent velocity of ~8
150 km/s. We interpret this phase as a P-wave refracted within the uppermost mantle (P_n).
151 The transition from P_g to P_n is marked by the arrival of critical reflections at the crust-
152 mantle boundary (P_mP). An additional reflected phase can be identified in some receivers
153 located in the NW Calabrian margin. This seismic phase arrives at shorter offsets than
154 P_mP 's, at 30-20 km on average (P_gP in Fig. 2c), indicating the presence of a major intra-
155 crustal interface below the margin. Towards the NW, in the Marsili basin, OBS receivers
156 do not display clear P_g and P_mP phases, just a single prominent refracted phase with
157 apparent velocity of ~8 km/s. These seismic records resemble to ocean bottom
158 hydrophone data recorded in the exhumed mantle region of the Magnaghi and Vavilov

159 basins during the MEDOC WAS experiment (Prada et al., 2016). For layer-stripping
160 modelling purposes, we also assigned this phase the P_n label. Land station records of line
161 WAS2 show clear P_g and P_mP phases (Fig. 2g).

162 In total, we picked 2837 P_s , 857 P_sP and 170 P_b from WAS1 record sections, and 4378
163 P_g , 675 P_gP , 1090 P_mP , and 3081 P_n from WAS2 records. Picking uncertainty was set
164 between 30 ms and 100 ms based on the amplitude ratio of a 250 ms-long window before
165 and after the picked travel-time, following the approach of Zelt and Forsyth (1994).

166 4 Methods: travel-time tomography

167 We invert for V_p and the geometry of the different seismic interfaces (i.e. Moho) using
168 the joint refraction and reflection travel-time tomography code TOMO2D (Korenaga et
169 al., 2000). The starting velocity model is parameterized as a regular grid hanging from
170 the seafloor, with 250 m node spacing both vertically and horizontally. Regularization
171 parameters are defined by horizontal and vertical correlation lengths (CL) that increase
172 from top to bottom in the velocity grid. In this study, we set 0.2 km and 1 km for the
173 vertical and horizontal CL at the top of the model, respectively, and 10 km and 20 km at
174 the bottom of the model, located at 50 km depth.

175 We have followed a layer-stripping strategy as in Prada et al. (2015) to resolve the sharp
176 velocity contrast attributed to the top of the acoustic basement in the contractional wedge,
177 and the crust-mantle boundary of the NW Calabrian margin. This way, along WAS1 we
178 built the model following a two-step inversion, each step consisting of 10 iterations. In
179 the first step, we inverted P_s and P_sP travel-times. The corresponding solution is included
180 and damped in the starting model for the second step, in which we inverted P_b travel-
181 times together with P_s , and P_sP travel-times. We followed a similar approach to build the

182 model along profile WAS2. First, P_g and P_mP travel-times were inverted to solve the V_p
183 structure of the crust beneath the NW Calabrian margin, as well as the Moho geometry.
184 In a second step, we included and damped the solution of the first inversion within the
185 starting model for the second inversion step, in which we added P_n travel-times to the
186 previous dataset. Once both models (i.e. WAS 1 and 2) were obtained, travel-times of
187 refracted phases of land stations in both profiles were jointly inverted to better constrain
188 the velocity structure of the Calabrian arc using crossing rays. P_gP travel-times along
189 WAS2 were inverted using the final velocity model as a reference to retrieve the geometry
190 and location of the intra-crustal reflector beneath the margin (Fig. 3). Acceptable ray
191 coverage beneath profile WAS1 in the Ionian is restricted to the first 10 km of depth of
192 the model, while tomographic constraints along WAS2 can reach down to 15-20 km of
193 depth (Fig.3b). The final root mean square (RMS) of refracted and reflected travel-times
194 for the final velocity model (Fig. 3) is ~60 ms and ~80 ms, respectively.

195 4.1 Uncertainty and resolution of the tomographic model

196 The uncertainty of the model parameters was assessed by means of a Monte Carlo
197 analysis (e.g. Korenaga et al., 2000). We followed the same layer-stripping strategy used
198 to obtain the preferred model to estimate the parameters uncertainty of each layer of the
199 model. We tested 100 realizations, each of them using a 1D V_p -depth profile and a flat
200 reflector as initial model, and a travel-time dataset perturbed with random Gaussian noise
201 (up to ± 100 ms). To generate the starting model, we have applied a random perturbation
202 of $\pm 10\%$ and $\pm 6\%$ for crustal and mantle velocities, respectively, of the reference velocity
203 model for each layer (Supplementary Table 1). A perturbation of ± 4 km respect the
204 reference depth for each reflector was applied to generate each starting interface
205 (Supplementary Table 1). The set of correlation lengths were also randomly perturbed in

206 each realization. This way, we have used for each CL: 3 ± 2 for the top horizontal CL,
207 15 ± 5 for the bottom horizontal CL, 0.3 ± 0.2 for the top vertical CL, and 4 ± 1 for the bottom
208 vertical CL.

209 All realizations converge after 20 iterations with an overall RMS of travel-times of 80 ms
210 (Fig. 4a). We have computed the standard deviation of V_p and depth of each reflector to
211 quantify the range of variation (uncertainty) attributed to all random errors. These values
212 show that the model presents standard deviations $< \pm 0.1$ km/s in most of the model,
213 indicating a satisfactory control of seismic velocities. As expected, higher values (± 0.2 -
214 ± 0.3 km/s) are observed in regions with a sharp vertical velocity contrast or limited ray
215 coverage, like at the lower part of the wedge (Fig. 4). The uncertainty of each reflector
216 ranges between ± 0.15 km in areas well covered by reflected rays, and ± 4 km towards the
217 edges of each reflector, where ray coverage is limited (Fig. 3b).

218 We have also evaluated the resolution of the tomographic model by performing a
219 checkerboard test, following the approach in Zelt (1998). This way, a total of 72
220 checkerboard-type models with nine different squared cell sizes (4, 6, 8, 10, 12, 14, 16,
221 18, and 20 km), and with a maximum velocity perturbation of 5% have been tested. For
222 each cell size, eight different patterns were created with positive and negative polarity,
223 with and without 45° rotation, and with and without a diagonal shift of half of the cell
224 size. Each checkerboard pattern was added to the preferred tomographic model to create
225 each target model and compute the synthetic travel-times. Before inverting each synthetic
226 dataset, we added a random Gaussian noise, with a standard deviation corresponding to
227 the picking uncertainty, to each travel-time, as in the Monte Carlo analysis.

228 After inverting each dataset using the preferred model as starting model, we compute the
229 semblance between the retrieved pattern and the corresponding target pattern. As in Zelt

230 (1998), we consider the pattern to be well retrieved for a semblance > 0.7 (red contour
231 Fig. 5a and 5b). We then compute the average semblance for each cell size and combine
232 them all in a resolution map that depicts the maximum resolution depth of each cell size
233 (Fig. 5c). The map shows that the minimum cell size that can be retrieved is 4 km along
234 profile WAS2, while it is 6 km along WAS1. The resolution depth of the largest cell size
235 (20 km) ranges between 15-20 km of depth beneath the arc/backarc, and 10 km of depth
236 beneath the contractional wedge. The differences in resolution between the arc/backarc
237 and the wedge is caused by the lower amount of travel-times (rays) observed in the
238 seismic records of the Ionian region (Fig. 3b).

239 5 Results

240 From SE to NW, the final tomographic model shows the V_p structure of the main crustal
241 domains of the Calabrian subduction system, namely the Ionian contractional wedge, the
242 Calabrian Arc, the NW Calabrian rifted margin, the volcanic arc (at ~ 290 km in Fig. 3),
243 and the Marsili backarc basin in the Tyrrhenian (Fig. 3).

244 In the Ionian, the V_p model delineates the Pre-Messinian wedge towards Calabria, across
245 the Inner Plateau and the Squillace basin (Fig. 1b and 3). The shallow velocity structure
246 across the Pre-Messinian and Inner Plateau sectors, which have distinct seafloor
247 morphology, indicates a continuous crustal domain. The V_p of the shallowest ~ 3 km
248 displays a gradual increase towards the coast (km 0-175 in figure 3a). V_p of $< \sim 3$ km/s
249 correspond to low consolidated sediment, particularly thick under OBSs 7-6, 3, and 1
250 (Fig. 3). The shallow V_p structure in the wedge has good ray coverage (Fig 3b) and
251 uncertainties < 0.1 km/s (Fig. 4), with well-resolved lateral changes of ~ 10 km in width
252 (Fig. 5). Underneath, V_p steadily increases from 4.0 km/s to 5.0-6.0 km/s with depth (Fig
253 3a). $V_p > \sim 4$ km/s imply a well consolidated material, and V_p 5.0-6.5 km/s at 10-15 km

254 depth indicate crystalline rocks with little porosity (Fig. 3), although with depth ray
255 coverage and resolution decrease and uncertainty increases (fig. 4). Near the coast, the
256 ~4-5 km deep Squillace basin (beneath OBS 17-16 in Fig. 3) V_p increases from 1.70 to
257 4.0-4.5 km/s at 6-7 km depth, overlaid by $V_p > 5$ km/s that supports shallow continental
258 basement rocks. A relatively continuous intra-wedge regional wide-angle reflector
259 roughly follows V_p 3-3.5 km/s, shallowing from 10 km depth beneath OBS 17 to ~5 km
260 depth beneath OBS 7-4. The V_p does not abruptly change across the reflector, which
261 may indicate a rapid decrease in porosity and/or fracturing with depth rather than an
262 abrupt boundary.

263 The structure beneath mainland Calabria is partially resolved with rays from both Ionian
264 and Tyrrhenian marine shots. V_p increases from ~4.0-4.5 km/s at ~1 km depth to 6.5-7
265 km/s at 20-22 km depth, above the Moho mapped by P_mP reflections in a sector of NW
266 Calabria (at 220 km in Fig. 3). These values are in agreement with Moho depth from the
267 1979 Calabria transect (Cassinis et al., 2003), and receiver functions (Piana-Agostinetti
268 et al 2009). The Calabria crust V_p range and low vertical velocity gradient ($\sim 0.1 \text{ s}^{-1}$) match
269 typical continental crystalline crust V_p distribution (Christensen and Mooney, 1995) as
270 shown in Fig. 6.

271 The NW Calabria rifted margin continental crust thins from ~20 km under the coastline
272 to 10 km at the NW edge (km ~325 in Fig. 3). Along this margin, V_p increases from 1.8
273 km/s to 3.5 km/s at ~1.5 km below seafloor and then abruptly to 5.0 km/s at 2 km below
274 seafloor roughly marking the basement top (Fig. 3). The basement V_p is laterally
275 homogeneous and with a gentle vertical gradient from ~ 5 to ~7.0 km/s near the Moho.
276 An intra-basement reflector delineates a slight decrease in vertical gradient for $V_p > 5.7$ -
277 6.0. The basement velocity under Lametini volcano of the Aeolian arc (Fig. 1b) is well-

278 constrained and comparatively higher, of 6 km/s at ~1 km beneath top basement to ~7.0-
279 7.3 km/s in the lowermost crust (km ~290 in Fig. 3). The margin uppermost mantle V_p
280 increases from 7.7-7.8 km/s to 8.2 km/s in 5 km under the Moho (Fig. 3).

281 The backarc region displays an abrupt lateral change in V_p structure at km ~330 (Fig 3a).
282 The Marsili basin has no Moho boundary (no P_mP reflections, Fig. 2e-f) and the transition
283 to mantle V_p occurs with comparatively steep vertical velocity gradients of $\sim 1 \text{ s}^{-1}$ for V_p
284 of $\sim 4.5\text{-}7.0$ km/s reaching $V_p \sim 7.5\text{-}8.0$ km/s at ~ 5 km below seafloor. The general Marsili
285 basin V_p structure is locally interrupted by a gentler vertical velocity gradients of the two-
286 layer V_p structure under the 3-km-high Marsili volcano. The upper layer increases rapidly
287 from 3 km/s near the surface to 6 km/s at 6-7 km of depth, with a 0.6 s^{-1} gradient (Fig. 3).
288 The second layer V_p increases from 6 km/s to ~ 8 km/s with a 0.2 s^{-1} gradient (Fig. 3). The
289 V_p model beneath small seafloor ridges under OBS 28 and 29 contains local sub-vertical
290 anomalies with relatively lower V_p (Fig 1 and 3).

291 6 Discussion

292 6.1 Geological interpretation

293 6.1.1 Ionian wedge

294 The upper 4-5 km of the internal wedge is characterized by low velocity regions (i.e. <
295 2.5 km/s). These low velocity anomalies may indicate comparatively higher porosity in a
296 region of fluid overpressure. Some of these low velocity anomalies (i.e. OBS 6-7 in Fig.
297 1 and 3) are spatially coincident with mud volcanoes mapped with multibeam bathymetry
298 (Gutscher et al. 2017). that are associated with fluid and mud expulsion (Loher et al.,
299 2018). Previous tectonic interpretation based on MCS data shows that low velocity
300 anomalies beneath OBS 6 to 1 are spatially coincident with compressional structures

301 (Polonia et al., 2011) (Fig. 1b). Thus, we interpret that these low velocities may indicate
302 regions of the wedge where active shortening is causing fluid outflow through thrust
303 faulting.

304 6.1.2 The NW Calabrian margin

305 The basement of the NW Calabrian margin and that beneath Calabrian exhibits a similar
306 vertical V_p profile than the global continental crust compilation of Christensen and
307 Mooney (1995) (Fig. 3 and 6). The Moho beneath the NW Calabrian margin shallows
308 from 20-22 km of depth beneath Mainland to ~10 km of depth, reflecting the result of the
309 Late Miocene backarc extension of the margin (~11-7 Ma ago; Kastens and Mascle,
310 1990).

311 The intra-crustal reflector modeled at 7-8 km depth roughly delineates the V_p ~5.8-6.0
312 km/s (Fig. 3). The boundary does not mark an abrupt change either in V_p or in gradient
313 with depth, and we speculate that it may represent a transition with depth in rock
314 fracturing and associated fluid percolation, possibly related to a change from shallower
315 fault-controlled brittle deformation to deeper plastic lower crust deformation. A similar
316 result is found in the Galicia Interior basin (Perez-Gussinyé et al., 2003), where the
317 resulting reflector delineate the 6.2-6.3 km/s V_p at 9-10 km depth in a ~15-km-thick
318 basement. There, coincident seismic images and thermal modelling support that the
319 reflector marks the boundary between a faulted upper crust and a ductile lower crust at
320 the time of rifting.

321 6.1.3 The volcanic arc

322 The NW Calabrian margin is characterized by $V_p > 7$ km/s beneath the volcanic arc (Fig
323 .3). These velocities are anomalously high for extended continental crust settings, which
324 typically range between 6.69-6.93 km/s (Christensen and Mooney, 1995). To assess the

325 petrological nature of these rocks, we compare the vertical velocity structure of the
326 Aeolian arc with (1) the vertical V_p structure of the NW Calabrian margin at 275 km of
327 profile distance, (2) the empirical V_p -depth distribution of different rock-types of the Izu-
328 Bonin-Mariana arc in the Pacific (Kitamura et al., 2003), and (3) the V_p -depth reference
329 function for continental crust (Christensen and Mooney, 1995) (Fig. 3a). The V_p -depth
330 distribution of the arc reveals a compositional differentiation from felsic/intermediate (i.e.
331 6.0 to 6.8 km/s; Christensen and Mooney, 1995) in the upper crust to mafic material in
332 the lower crust (Fig. 3a), where the V_p -depth distribution coincides with the V_p -depth
333 distribution for gabbroic rocks of Izu-Bonin-Mariana arc (Fig. 6a; Kitamura et al., 2003).

334 6.1.4 Marsili backarc basin

335 In the backarc region we compared average vertical velocity distribution at either side of
336 the Marsili volcano, with reference V_p -depth functions from modern tomographic studies
337 for oceanic crust (Grevemeyer et al., 2018a), exhumed mantle regions from the central
338 Tyrrhenian and Gulf of Cadiz (Prada et al., 2015), as well as exhumed mantle regions
339 from ultra-slow spreading centers (Grevemeyer et al., 2018b). Oceanic crust and exhumed
340 mantle references have different V_p -depth trends reflecting their petrology (Fig. 6b). The
341 oceanic crust reference has two-layer structure associated to the petrological
342 differentiation between the widespread basaltic layer 2 and the gabbroic layer 3 (e.g.
343 Grevemeyer et al., 2018a), while the exhumed mantle field appears as a continuous,
344 comparatively steeper velocity gradient of $0.6-0.7 \text{ s}^{-1}$ of the upper $\sim 5-6$ km of basement,
345 reflecting the decreasing degree of serpentinization with depth (e.g. Prada et al., 2016).
346 The comparison with our observations supports that the basement of the Marsili backarc
347 basin is made of exhumed mantle similar to the Magnaghi and Vavilov basins in the
348 Tyrrhenian.

350 Previous regional studies interpreted that the Marsili basin is floored by oceanic crust
351 produced by ultrafast oceanic spreading (e.g. Nicolosi et al., 2006; Manu-Marfo et al.,
352 2019). However, our results show that the structure that can be related to oceanic
353 spreading, that is the Marsili volcano and neighboring ridges, are isolated velocity
354 anomalies in the tomographic model, and that the bulk of the basin has a velocity gradient
355 matching the exhumed mantle V_p -depth reference. This result, together with similar
356 observations from the Magnaghi and Vavilov basins (Prada et al., 2016), indicate that
357 mantle exhumation and emplacement of localized oceanic ridges and large-volcanic
358 edifices (Fig.1b) have been recurrent events during the opening of the Tyrrhenian
359 backarc.

360 The occurrence of exhumed mantle in the Tyrrhenian basins results paradoxical.
361 Stratigraphic record analysis of the Marsili Basin allowed to infer that the basin opened
362 ~2 Ma ago (Kastens and Mascle et al., 1990; Argnani and Savelli, 1999), so that mantle
363 exhumation should have occurred at a rate of 3.5-4.5 cm/yr (Kastens and Mascle et al.,
364 1990), although magnetic anomaly analysis suggests a much faster rate (~19 cm/yr;
365 Nicolosi et al., 2006). This rate is exceptionally fast considering models of partial
366 decompression melting, which suggest that mantle unroofing should not occur at rates
367 $> \sim 2$ cm/yr (e.g. Grevemeyer et al., 2018b). Geochemical analyses of serpentinized
368 peridotites from site 651 show that these rocks are strongly depleted in lithophile elements
369 during recurrent partial melting events (Bonatti et al., 1990). This suggests the presence
370 of a depleted mantle source, which could restrict the production of partial decompression
371 melting during lithospheric extension (Perez-Gussinyé et al., 2006).

372 6.2 The forearc-to-backarc structure of the Calabrian arc: resemblance with
373 Alpine subduction systems

374 The 450 km long transect displays the contrasting V_p structures of the Ionian contractional
375 wedge, the 20-22 km thick continental crust of the Calabrian Arc, the NW Calabrian rifted
376 margin, locally modified by ~20 km-wide volcanic arc activity, and the exhumed mantle
377 domain of the Marsili backarc basin, locally intruded by volcanic ridges (Fig. 7). These
378 domains represent terranes made of distinct petrological suites and formed by contrasting
379 deformation regimes, bounded by <4-5 km wide transitions, near the resolution limit of
380 our data (Fig. 5). This complex terrane structuration involves a tectonic and magmatic
381 evolution not fully understood. The NW Calabrian margin integrates the evolution of
382 extended continental crust possibly by a Miocene to younger rift (Kastens and Mascle et
383 al., 1990), modified by the emplacement of the volcanic arc, and the development of the
384 Plio-Quaternary Paola basin, currently in a forearc position, with up to 5 km infill
385 (Zitellini et al., 2020) that extends towards the NW of our transect (Fig. 1 and 3). The
386 backarc Marsili basin is possibly Quaternary (Kastens and Mascle et al., 1990), and
387 formed by amagmatic extension leading to mantle exhumation and the subsequent
388 intrusion of a ~20 km-wide volcano and localized ridges. This basin is surrounded to the
389 north, east and south by continental crust and volcanic complexes related to the
390 subduction (Fig 1b), while towards the northwest, the basin leads to the Vavilov Basin,
391 where exhumed mantle and localized oceanic ridges were emplaced during its formation
392 (Prada et al., 2016).

393 This complex structuration resulted from subduction and rollback of the Jurassic Tethys
394 lithosphere, which is the same geodynamic process that controlled the subduction phase
395 of possibly many Alpine systems (e.g. Schmid et al. 1996; Wortel and Spakman, 2000;
396 Maffione and van Hinsbergen, 2018). The complex structure of the Calabrian subduction

397 system is possibly not unique or anomalous, and similarly complex structures might have
398 occurred at overriding plates of ancient Alpine subduction systems.

399 General proposals indicate that the structural inheritance of rifted margin architecture is
400 key to explain orogenic configuration (e.g. Reston and Manatchal 2011). More
401 specifically Alpine rock units originally from the Jurassic Tethys (Froitzheim and Eberli
402 1990) are interpreted to signify a structural inheritance that significantly influenced the
403 orogenic development of Alpine belts (e.g. Mohn et al., 2014). Based on those concepts,
404 it is proposed that closure of suspected narrow oceans lacks significant subduction-related
405 magmatism, leaving little trace in mountain belts like the Alps and Pyrenees. It is argued
406 that those orogens develop fundamentally by processes in which the original rifted
407 margins structure controls the final orogen architecture (e.g. Chenin et al., 2017).
408 Although the Pyrenees are overwhelmingly interpreted as the closure of a narrow rifted
409 basin, the Alps are debated and the subduction phase may have lasted 25 Ma, from Early
410 Paleocene to at least Late Eocene, with >500 km of subduction (Schmid et al., 1996),
411 similar to the present-day subduction systems in the Central Mediterranean (e.g. Faccenna
412 et al., 2001). Therefore, such subduction system could have created a structuration similar
413 to the Calabrian Arc system. Furthermore, at a larger scale, opening during Tethys-Ionian
414 slab subduction and rollback since ~25 Ma triggered the formation of the passive margin
415 of the Gulf of Lions, the Ligurian basin and the Corsica-Sardinia continental block
416 (Schettino and Turco, 2011), which adds further complexity to a pre-collision scenario.

417 The geological record of several Alpine-type mountain belts retain unequivocal evidence
418 of terranes formed during subduction , sharing similarities with domains of the Calabrian
419 arc. Conspicuous units found in Alpine belts such as the Apennines, Alps, Hellenides,
420 Dinarides, and Taurides are the remnants of forearc accretionary systems (e.g. Stampfli

421 et al., 1998; Okay et al., 2006; Marroni et al., 2017; Maffione and van Hinsbergen, 2018),
422 and volcanic arcs (Sharkov et al., 2014). But the most distinctive units implying
423 subduction are serpentinized mantle rocks accompanied by basaltic rocks found in
424 Alpine-type ophiolites (e.g. Moghadam and Stern, 2015). Although some mantle rocks
425 from the Alps are related to mantle exhumation in Tethys rift context (e.g. Froitzheim and
426 Eberli 1990), the majority of these ophiolites have geochemical signatures that support a
427 supra-subduction origin and are spatially associated to magmatic rocks with Island arc
428 affinity (e.g. Dilek 2003; Moghadam and Stern, 2015; Maffione and van Hinsbergen,
429 2018), similar to Tyrrhenian samples (Beccaluva et al., 1990). In particular, assemblages
430 similar to Tyrrhenian serpentinized mantle rocks are described in the east and west Vardar
431 Ophiolites in Albania (Maffione and van Hinsbergen, 2018), the Tsiknias ophiolites in
432 Greece (Lamont et al., 2020), and Iran (Moghadam and Stern, 2015). This latter suit of
433 serpentinized mantle rocks found in Mesozoic ophiolites, formed prior to the construction
434 of the Zagros belt, share a peridotite spinel composition with Tyrrhenian peridotites
435 drilled at ODP leg 107 (Fig. 1). Both rocks present high values of Cr# (> 40) and low
436 values of Mg# (< 60), indicating a similar degree of depletion during melt extraction at
437 convergent margins (see fig. 7 in Bonatti et al., 1990; fig. 13 in Moghadam and Stern,
438 2015).

439 Based on these observations, it is likely that the pre-collisional structure of many Alpine
440 orogens from South Europe to the Middle East that experienced extended subduction
441 periods, possibly accompanied by slab retreat, was formed by neighboring terranes with
442 distinct petrological and tectonic characteristics separated by spatially abrupt boundaries,
443 resembling to the structuration mapped here across the of the Calabrian Arc (Fig. 7).

444 7 Conclusions

445 The first modern WAS transect across the Calabrian Arc system provides a 2D V_p
446 tomographic image of the entire system along 450 km from forearc to the backarc region
447 from joint inversion of refracted and reflected P-wave travel-times.

448 The Calabrian system structure is characterized by spatially abrupt changes (<4-5 km
449 width) between domains formed by different processes and differing petrological and
450 tectonic structures.

451 The frontal region where contractional tectonics dominate includes the offshore internal
452 wedge formed mainly by consolidated rocks covered by 1-3 km of sediment with $V_p <$
453 3.5 km/s. The V_p model of the wedge displays gentle undulations defining local low
454 velocity anomalies (< 2.5 km/s) suggesting the presence of high porosity regions of the
455 wedge. These anomalies are spatially coincident with thrusting and mud volcanism on
456 surface, suggesting that thrusting leads to fluid outflow to the seafloor. The transition
457 between the inner wedge and Calabria occurs under the Squillace basin, with an abrupt
458 shoaling of the consolidated basement rocks.

459 The Calabrian Peninsula has a 20-22 km thick crust with a typical continental crust
460 velocity structure. The continental Calabrian crust was extended and thinned in the
461 Miocene to 16-9 km across ~100 km of the NW Calabrian margin. The rifted margin
462 structure was subsequently modified during Plio-Quaternary time by magmatic arc
463 activity across a ~20 km wide area and the development of the 3-5 km thick Paola forearc
464 basin. The continental crust abruptly thins over ~5 km width to the Marsili basin.

465 The 200 km wide Marsili basin has a vertical velocity gradient typical of exhumed mantle
466 domains, similar to Vavilov and Magnani basins in the central Tyrrhenian. Local velocity

467 anomalies under Marsili volcano and low ridges indicate that basaltic features are isolated
468 intrusions in the basement of the basin, which is inferred to be composed by serpentized
469 peridotite.

470 Although it has been long speculated that orogen formation is strongly influenced by the
471 inherited structure of rifted continental margins, most conceptual models and
472 reconstructions have not accounted for the subduction phase that possibly preceded
473 collision in most of the Alpine-type orogens from South Europe to the Middle East. We
474 show the first detailed structure of an analog of a system created by Tethys slab
475 subduction and rollback, which has created an unanticipated complex pattern of large-
476 scale lithospheric domains, some of which are similar to terranes found in Alpine-type
477 mountain belts.

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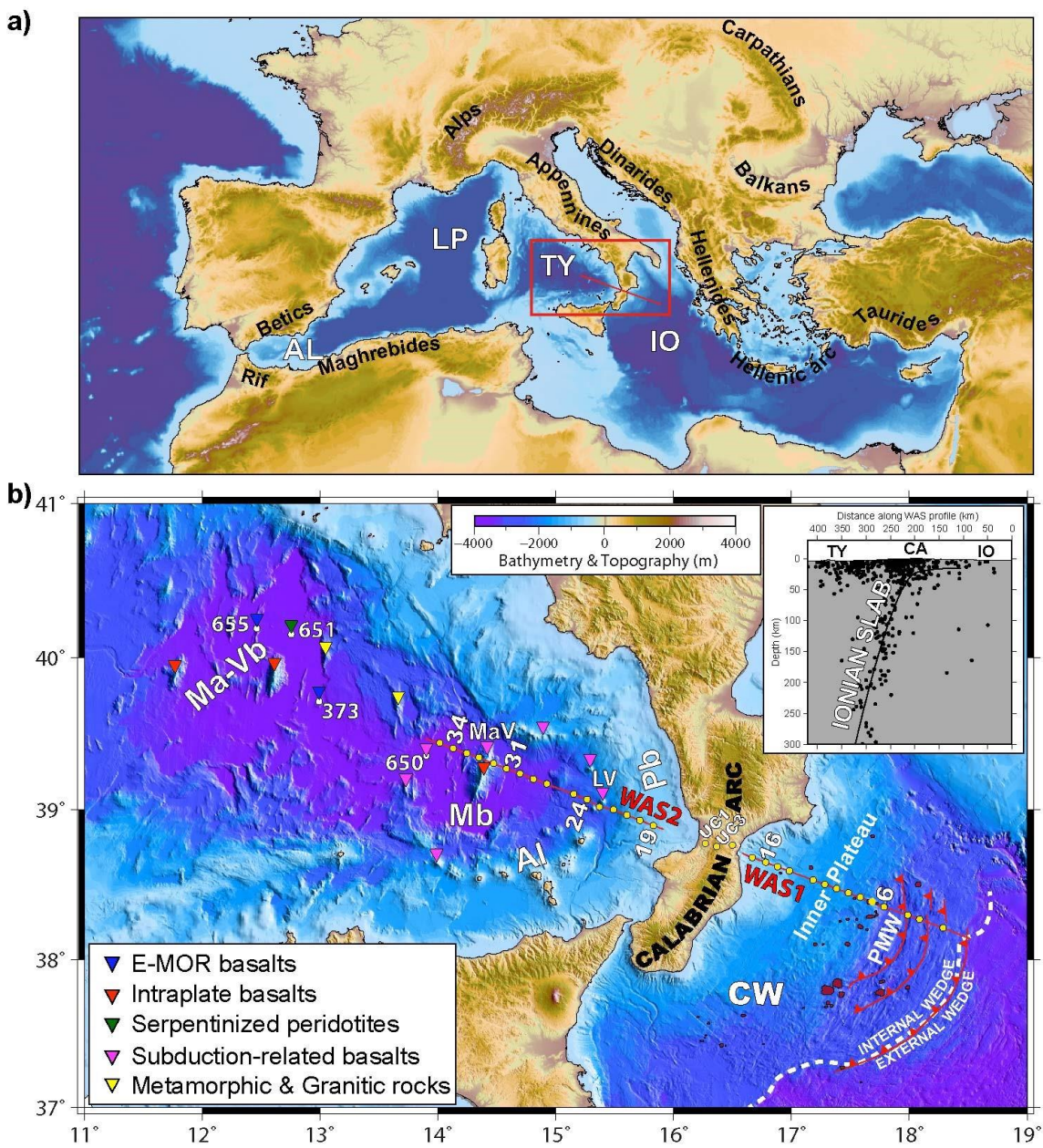
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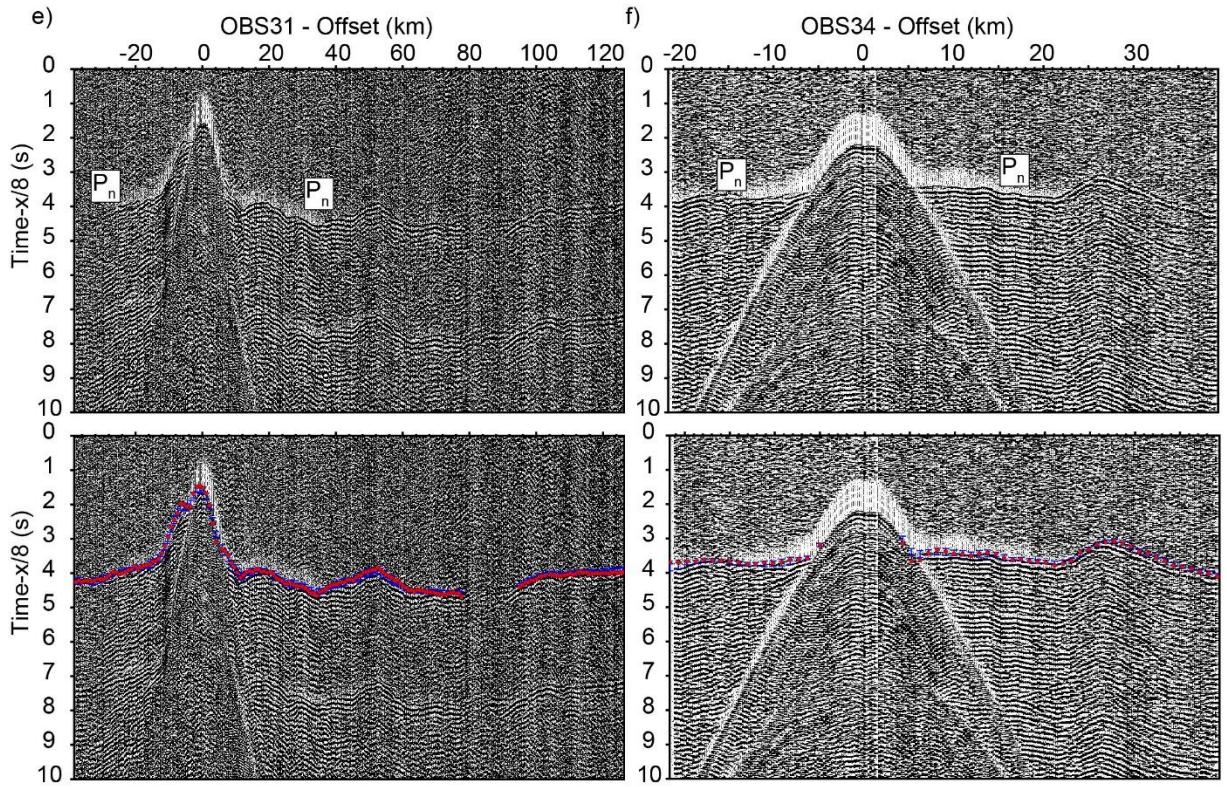
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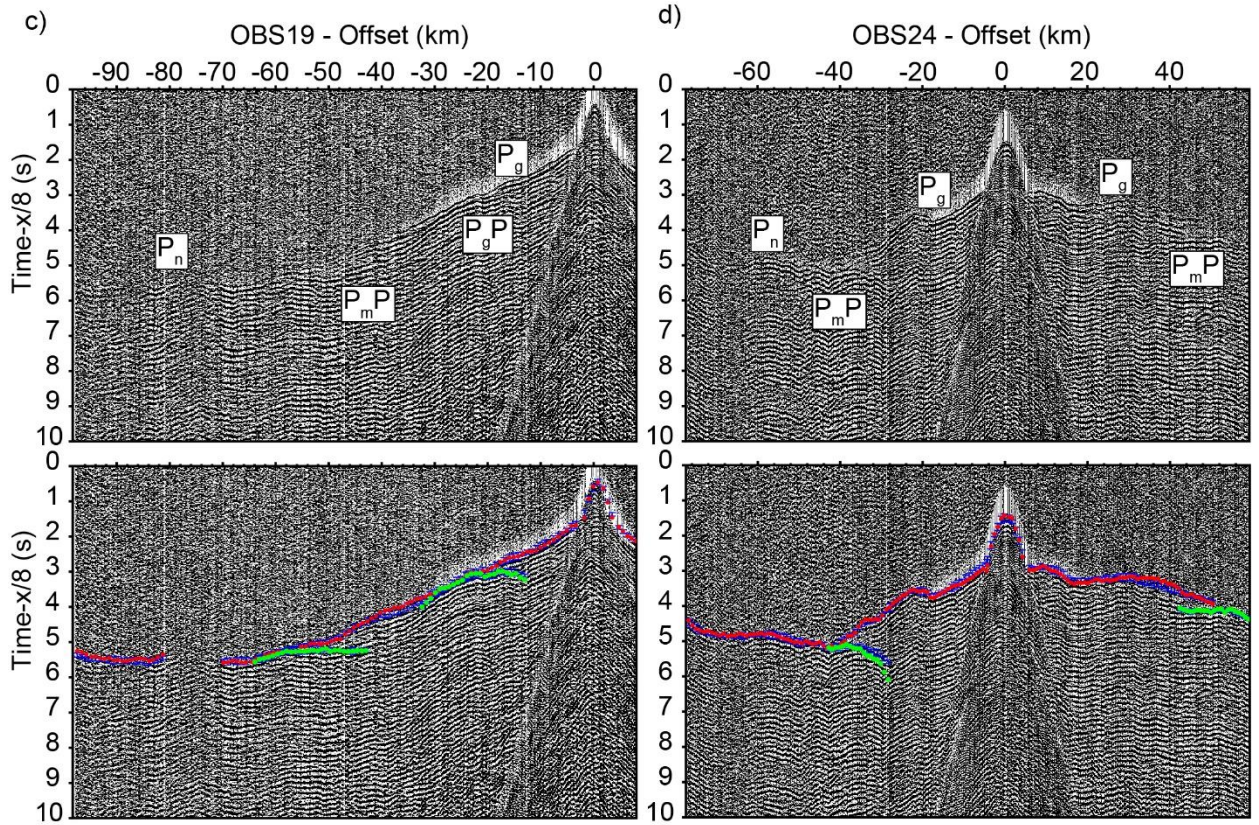
659

660 **Figure 1.** a) Bathymetry and topography map of the Mediterranean region showing the
 661 location of the main Cenozoic orogenic belts, and the location of the seismic cross section
 662 presented in this study. AL: Alboran Sea, LP: Liguro-Provençal, TY: Tyrrhenian, IO:
 663 Ionian. b) Bathymetry map of the Tyrrhenian and Ionian region showing the petrological
 664 nature of basement rocks from ODP sites (Kastens and Mascle, 1990), DSDP sites
 665 (Dietrich et al., 1977), and dredging samples (Colantoni et al., 1981). Yellow circles

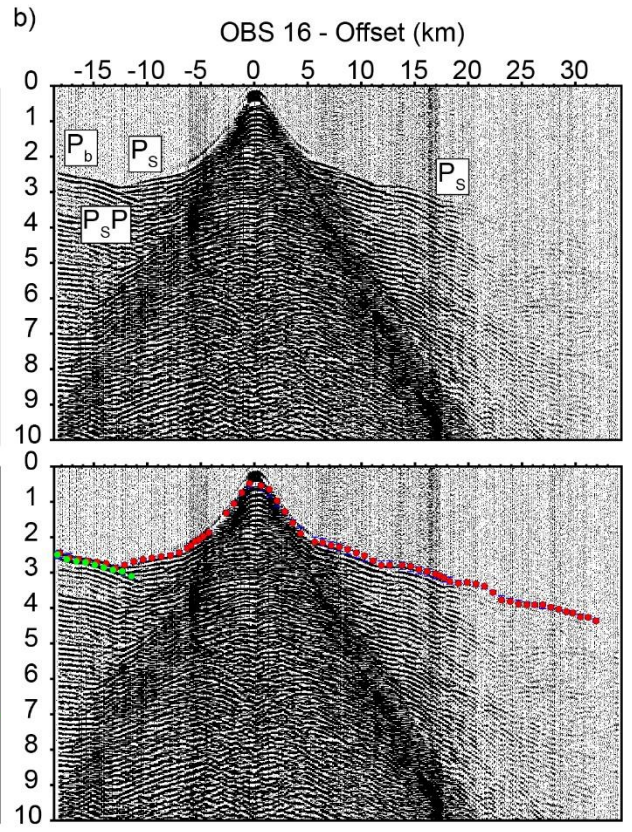
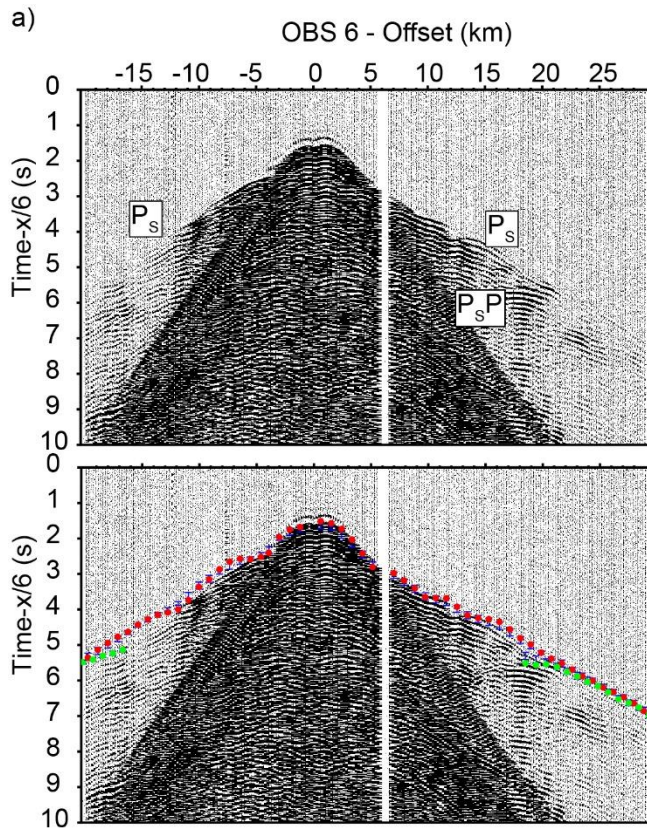
666 depict the location of ocean bottom seismometers and land stations used to record shots
667 along profiles WAS 1 and WAS 2. The record sections of numbered OBSs and land
668 stations UC1 and UC3 are shown in Figure 2. White circles show the location of ODP
669 and DSDP boreholes in the Tyrrhenian, while mud volcanoes in the Contractional wedge
670 (CW) are depicted with dark red circles. The boundary between the internal and external
671 wedge (dashed white line) and the location of mud volcanoes is taken from Gutscher et
672 al. (2017), and Loher et al. (2018), while the location of the morphological domains of
673 the Inner plateau and the Pre-Messinian wedge (PMW) are taken from Polonia et al.
674 (2011). Red lines and triangles in the PMW are splay faults from Polonia et al. (2011).
675 Earthquakes occurring within ± 50 km of distance from the WAS profile are projected
676 along the geophysical cross section and presented in the small inset to show the steep
677 ($\sim 70^\circ$) geometry of the Ionian slab beneath the study area (earthquakes are from
678 Chiarabba et al., 2005). AI: Aeolian Islands, CA: Calabria, LV: Lametini Volcano, Ma-
679 Vb: Magnaghi-Vavilov basins, Mb: Marsili basin, MaV: Marsili Volcano, Pb: Paola
680 basin.

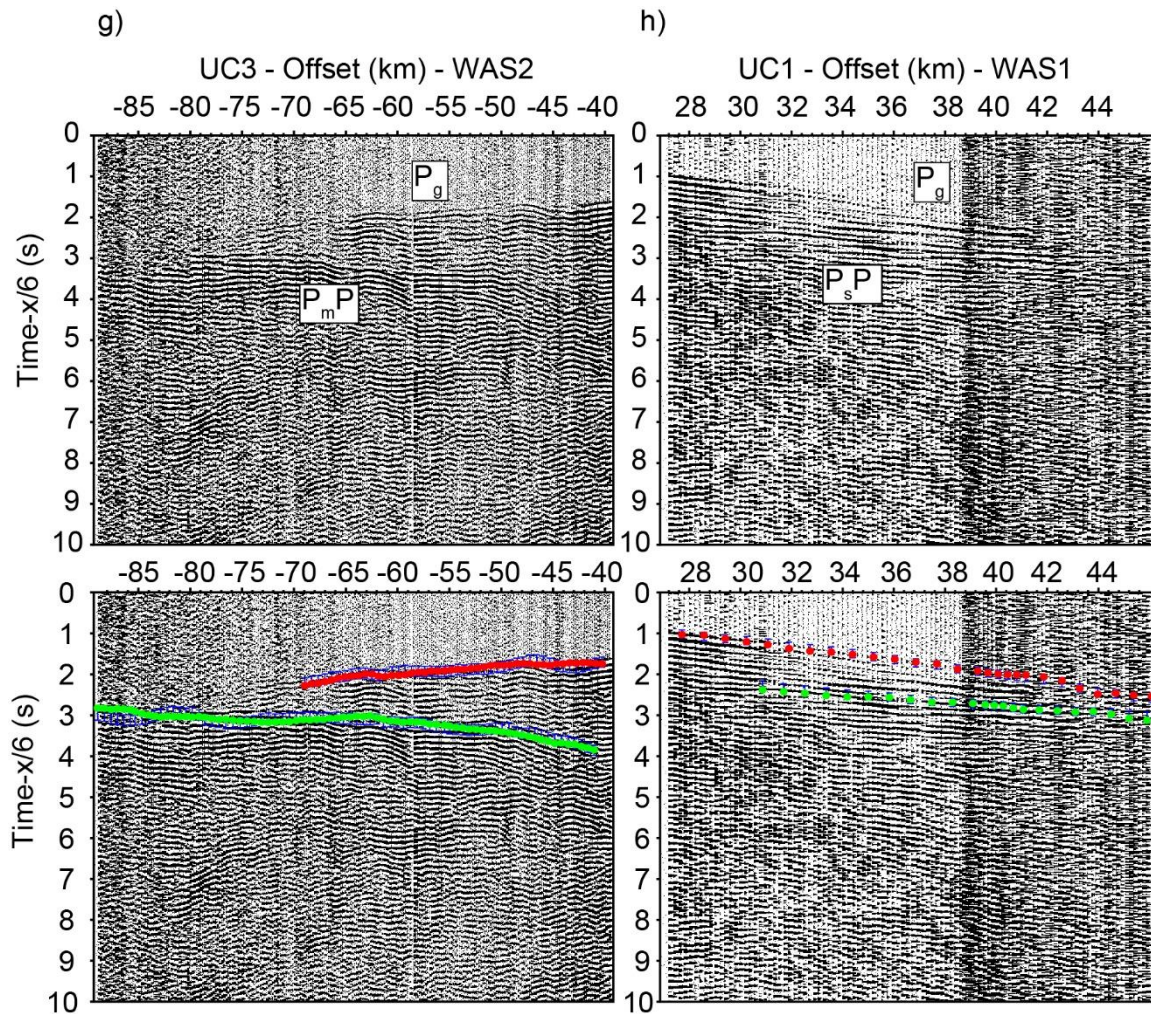


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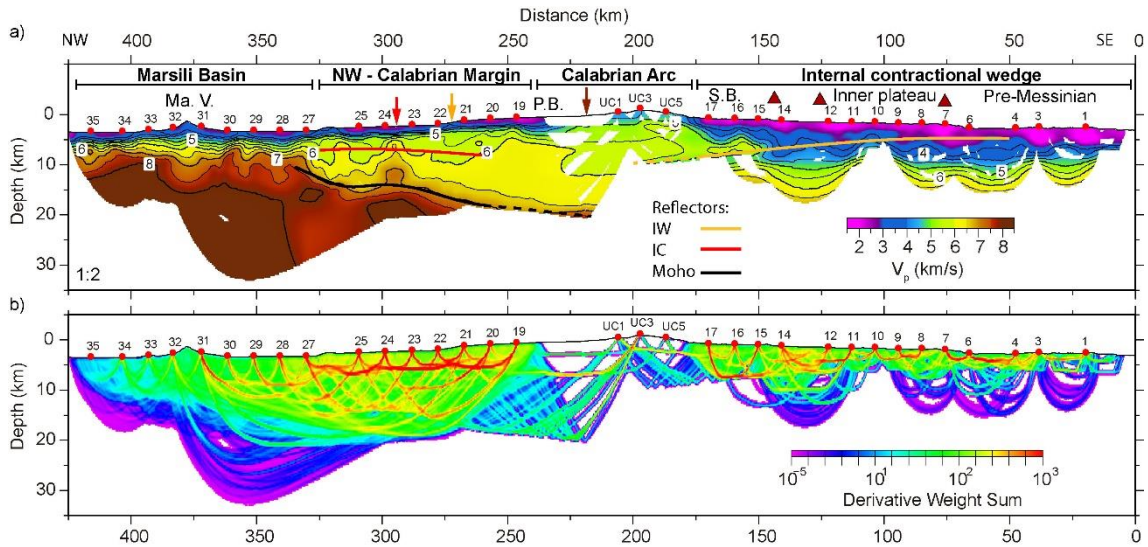
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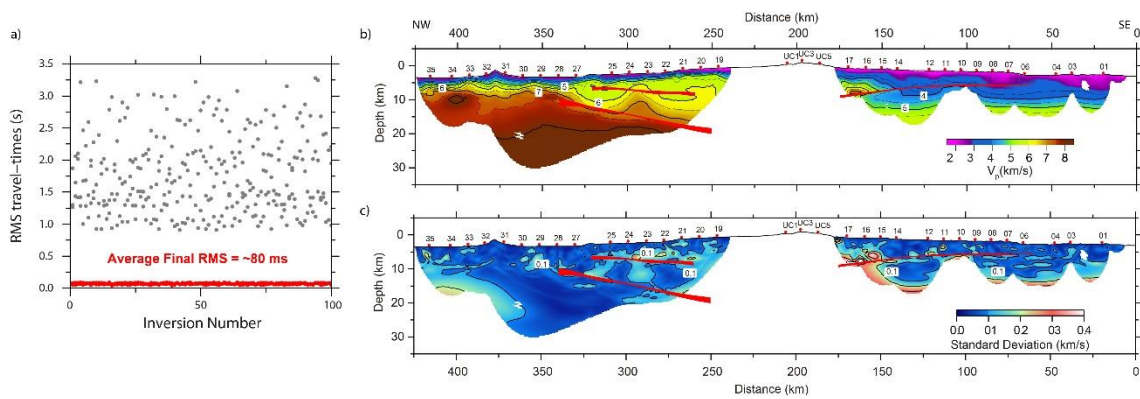
684

685 **Figure 2.** Seismic record sections of OBSs (a)6, (b)16, (c)19, (d)24, (e)31, (f)34, and land
 686 station (g) UC3 record from profile WAS2, and (h) UC1 record from profile WAS1. The
 687 location along the line of each instrument is shown in Fig. 1b. The upper panels show the
 688 interpreted seismic phases, while the lower panels show the records with the picked
 689 refraction and reflection travel-times (blue bars), and the corresponding synthetic
 690 refraction (red dots) and reflection (green dots) travel-times obtained in the final model.



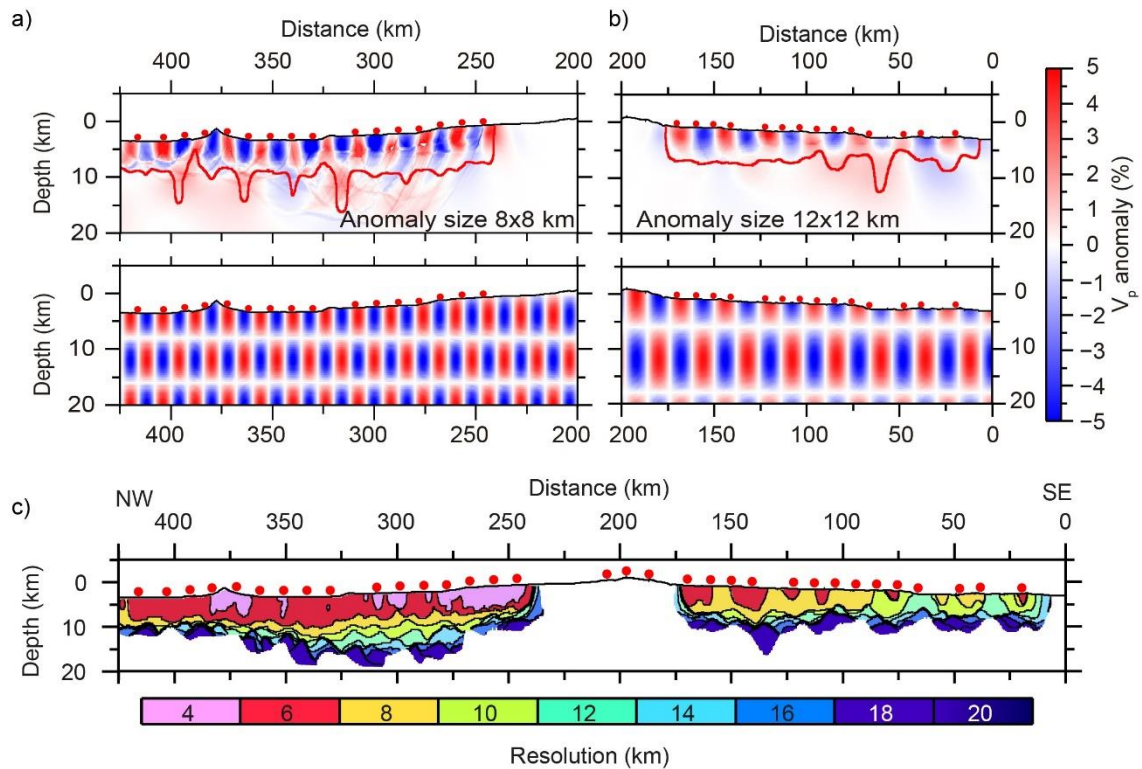
691

692 **Figure 3.** 2D P-wave velocity model of the Calabrian subduction system obtained from
 693 joint inversion of refracted and reflected travel-times. The inverted reflectors are intra-
 694 wedge reflector (IW), the intra-crustal interface (IC) and the Moho along the NW
 695 Calabrian margin. Ma.V: Marsili Volcano, P.B: Paola basin S.B.: Squillace basin. Red
 696 dots are OBS and land stations. Vertical red, orange, and brown arrows indicate the
 697 location of the vertical velocity-depth profiles shown in Figure 7a. Red arrow also shows
 698 the location of the volcanic arc, while the dark red triangle shows the location of mud
 699 volcanism neighboring WAS line 1 in Fig. 1. (b) Derivative weight sum of the
 700 tomographic model. This image can be used as a proxy for the refracted and reflected ray
 701 coverage through the grid during the inversion.



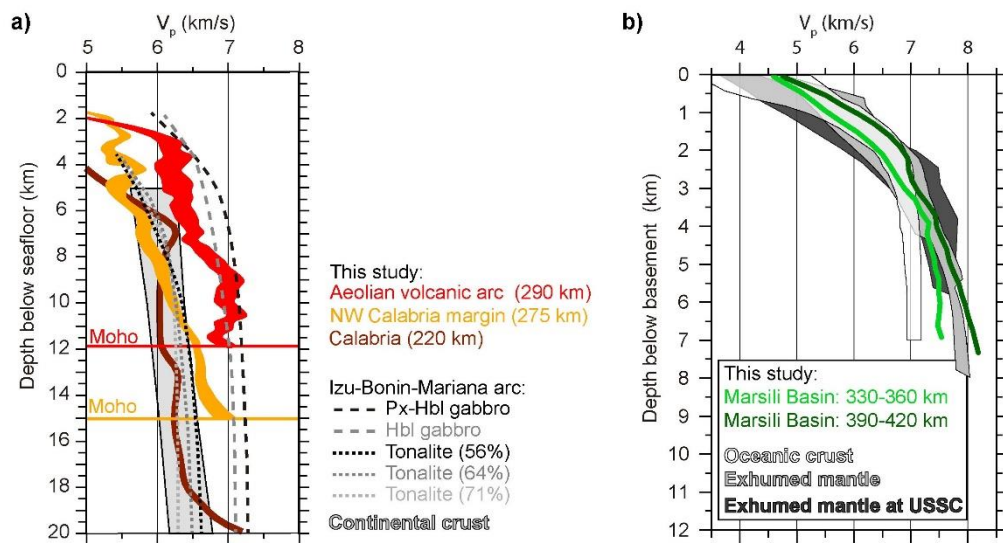
702

703 **Figure 4.** a) Distribution of RMS travel-times of the first (grey dots) and final (red dots)
 704 iteration of each Monte Carlo realization (100 in total) showing how the initial random
 705 distribution converges to a common solution (Final RMS ~ 80 ms). b) Average final
 706 tomographic model of the 100 Monte Carlo final solutions. The red bands depict the error
 707 bar of each reflector. c) Standard deviation of the 100 realizations.



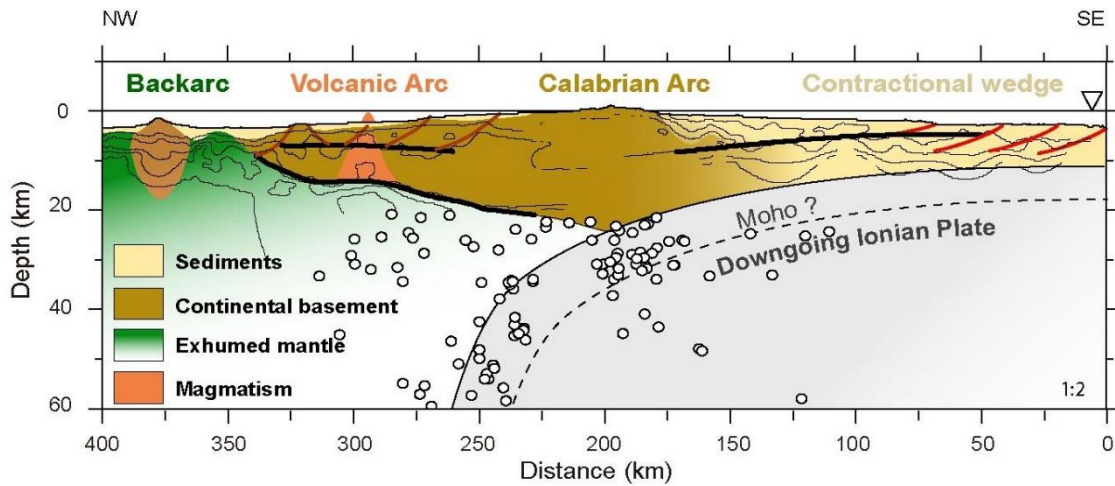
708

709 **Figure 5.** Retrieved (upper) and true (lower) checkerboard pattern of the (a) 8x8 km and
 710 (b) 12x12 km anomaly size. The red contour line indicates where the patter is retrieved
 711 satisfactorily (i.e. where the semblance is ~ 0.7). Red dots are OBSs. c) Resolution map
 712 of the model showing the region at which we are able to retrieve each size of the
 713 checkerboard pattern.



714

715 **Figure 6.** a) Velocity-depth profiles of the region of the volcanic arc (red band), the NW
 716 Calabrian margin (yellow band), and mainland Calabria (dark brown) compared with the
 717 continental crust velocity-depth function of Christensen and Mooney (1995) and the
 718 velocity-depth function for tonalitic and gabbroic rocks of the Izu-Bonin-Mariana arc
 719 (Kitamura et al., 2003). The thickness of each profile from our model corresponds to the
 720 error bars inferred from the Monte Carlo analysis. Note that no error bar is drawn for the
 721 Calabrian profile as the Monte Carlo analysis was focused on the offshore regions. b)
 722 Average velocity-depth profiles of the NW (light green) and SE (dark green) Marsili
 723 volcanic area compared with the velocity-depth reference function for oceanic crust from
 724 Grevemeyer et al. (2018a), and exhumed mantle regions of the Tyrrhenian and the Gulf
 725 of Cadiz from Prada et al. (2015), and at ultra-slow spreading centers (USSC)
 726 Grevemeyer et al. (2018b).



727

728 **Figure 7.** Interpretative cross section of the entire Calabrian subduction system based on
 729 the result of this study and the depth-distribution of earthquakes from 20 to 60 km of
 730 depth (white circles) from Fig. 1b. The geometry of the Ionian subducting slab is taken
 731 from (Chiarabba et al., 2005), while the Ionian oceanic Moho is based on the results from
 732 Cassinis et al. (2003) and Piana-Agostinetti et al. (2009). Brown lines delineate
 733 interpreted normal faults along the NW Calabrian margin, while red lines are
 734 compressional features across the Inner plateau and the Pre-Messinian wedge observed
 735 by Polonia et al. (2011). Thick black lines are wide-angle seismic boundaries constrained
 736 by the tomographic model in this study. Thin black lines are isovelocity contours shown
 737 in Fig. 3.