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1 **Ambient-Noise Wave-Equation Tomography of the Alps**
2 **and Ligurian-Provence Basin**

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6 ²See Appendix A

7 **Key Points:**

- 8 • A new 3-D velocity model of the crust and upper mantle beneath the Alps and
9 Northwestern Mediterranean, based on wave-equation tomography
- 10 • 3-D acoustic-elastic waveform modeling of Rayleigh-Scholte waves from ambient noise
- 11 • 3-D high-resolution Moho depth map of the Alps and Ligurian-Provence basin

Abstract

Taking benefit of the AlpArray temporary network and permanent networks in W-Europe, we construct a 3-D onshore-offshore velocity model of the crust and upper mantle using ambient-noise wave-equation tomography (WET). We use a frequency-dependent phase traveltime misfit function in an iterative procedure to refine a recent 3-D V_s model computed from a Bayesian two-step ambient noise tomography (ANT). Observed waveforms consist in vertical-component noise correlations from 600 broadband stations in the Alps and surroundings, including ocean-bottom seismometers (OBS) in the Ligurian sea. We perform 3-D inversion in the 5-85 s period range. In the long-period band (20-85 s), an elastic approximation is considered, while in the 5-20 s band, we account for the effect of water layer in the Ligurian sea by applying a fluid-solid coupling for acoustic-elastic waveform simulations. The resulting V_s model enhances the shape and contrast of velocity structures, accounting for 3-D and finite-frequency effects. It emphasizes the deep sediments of the Ligurian-Provence basin and focuses the low-velocity anomalies of the crust in the W-Alps. We obtain a high-resolution Moho depth map covering the Alps and Ligurian sea. In the W-Alps, this map confirms the deepening of the European crust following the subduction beneath Adria and the existence of major structures such as the Moho jump beneath the external crystalline massifs and shallow depths associated with the Ivrea Body. It provides further constraints on the deep structure beneath the Ligurian-Provence basin, regarding the lateral and along-strike crustal-thickness variations from the oceanic domain to the conjugate margins.

Plain Language Summary

In the complex tectonic setting of the Alps and Northwestern Mediterranean, a precise seismic mapping of the 3-D structure of Moho is crucial to understand the deep dynamic processes involved in the orogenesis of the western Alps and the opening of the Ligurian-Provence back-arc basin. Using an innovative wave-equation tomography (WET) of ambient noise recordings, we compute a high-resolution velocity model covering the Alps and Ligurian sea. In particular, we account for the influence of water and seabed relief on the 3-D propagation of surface waves in the Ligurian sea. We derive a high-resolution onshore-offshore 3-D map of Moho depth beneath the Alps and Ligurian-Provence basin. Our Moho map confirms the subduction of the European lithosphere beneath Adria. It shows strong

43 depth variations in the Ligurian-Provence basin, from 12 km in the basin axis to \sim 20-25 km
44 in the conjugate margins of Provence and Corsica.

45 **1 Introduction**

46 The geodynamic evolution in the Alps and Northwestern Mediterranean occurs in a
47 context of plate reorganization, dominated since the Late Cretaceous by the convergence
48 between the European and African plates (e.g., Schmid et al., 2004; Faccenna et al., 2014).
49 In this convergence context, the western Alps result from three successive tectonic episodes
50 involving Europe, the Adria micro-plate and the Tethyan oceanic domain (Tethys) in be-
51 tween: (i) closure of the oceanic domain by subduction from Late Cretaceous to Early
52 Eocene (e.g., Handy et al., 2010); (ii) continental subduction of the European margin be-
53 neath Adria during the Late Eocene (e.g., Chopin, 1984; Duchêne et al., 1997; Guillot et
54 al., 2009; Zhao et al., 2015); (iii) continental collision from Oligocene onwards (e.g., Polino
55 et al., 1990; Dumont et al., 2012). This multiphase evolution led to a complex present-day
56 arcuate architecture of Western Alps, where different units are involved (Fig. 1): (i) to the
57 west, the European continental domain (or European foreland) corresponding to the lower
58 plate; (ii) to the east, the Adria continental domain (or Adriatic foreland) corresponding
59 to the upper plate; (iii) the Eocene subduction wedge (or accretionary prism) in between,
60 bounded by two major crustal-scale faults: the Frontal Pennine Fault (PFT) to the west and
61 the dextral strike-slip Insubric Fault (IF) to the east. During Miocene and Pliocene Epochs,
62 the opening of back-arc extensional basins occurred at the back of the Adria northwest-
63 dipping subduction zone (e.g., Gueguen et al., 1998; Jolivet et al., 2020). This extension
64 started along the Provence region leading to the Ligurian-Provence basin, and has further
65 spread from east to west and south, resulting in the opening of the Algerian basin, and
66 later, of the Tyrrhenian basin (e.g., Rollet et al., 2002; Séranne, 1999). The opening of
67 the Ligurian-Provence basin initiated at 35 Ma by a rifting phase between Europe and the
68 Corsica-Sardinia block. The progressive south-eastward roll-back and retreat of the Adria
69 slab below the Corsica-Sardinia domain led to Oligocene stretching of the continental crust
70 followed by continental break-up during the early Miocene, and to the genesis of an oceanic
71 crust between 15 and 11 Ma (e.g., Séranne, 1999; Jolivet & Faccenna, 2000). As a result,
72 three principal domains, mainly identified from active seismic data, describe the present-day
73 geological setting of the Ligurian-Provence basin (Fig. 1): (i) two thinned conjugate con-
74 tinental passive margins corresponding to the Ligurian-Provence margin to the north and

75 Corsican margin to the south; (ii) an oceanic domain in between, described as an 'atypical
76 oceanic crust' by Rollet et al. (2002) because it is thinner than normal oceanic crust; (iii)
77 two transitional domains separating the margins and the oceanic domain, likely made up of
78 a very thin continental crust overlying a thick rift-related corner of magmatic underplating
79 (e.g., Séranne, 1999).

80 The complex geodynamic context associated with the interaction between the opposing
81 vergence subductions of Europe and the Apennines has resulted in a three-dimensional,
82 heterogeneous lithospheric structure that is difficult to image by seismology. Consequently,
83 questions remain regarding the role of the 3-D geometry of European continental subduction
84 in the present-day architecture of Western Alps (e.g., Malusà et al., 2021), and the role
85 of the rapid retreat of Apennines subduction in the present-day petrological-lithological
86 composition of the crust beneath the Ligurian-Provence basin (e.g., Rollet et al., 2002).
87 Thus, accurate 3-D seismic imaging of the crust and upper mantle is critically needed to
88 complement the 2-D seismological observations, especially since the region benefits from a
89 fairly wide coverage of seismological stations, including the newly deployed AlpArray land-
90 sea network (AASN, Hetényi et al., 2018).

91 With the deployment of the AlpArray temporary seismic network, Western Europe
92 has become one of the most densely instrumented areas in the world. More specifically, the
93 region consisting of the Alps and the Ligurian sea includes, in addition to the onshore part of
94 AlpArray, other temporary experiments such as Cifalps-2 in the northwestern Alps (Zhao et
95 al., 2018) and the marine part of AlpArray in the Ligurian-Provence basin. These dense and
96 high-quality surveys provide an ideal set up to use seismic ambient noise records to resolve
97 the crust and upper mantle, as well as major interfaces such as the Moho. Several ambient
98 noise tomography (ANT) were conducted in the region. Following on from the large-scale
99 ANT by Lu et al. (2018), which used a part of the land component of AlpArray, the recent
100 ANT by Nouibat et al. (2022a, 2022b) was the first study to use the entire AlpArray network,
101 resulting in a land-sea model covering the Alps and Northwestern Mediterranean. These
102 large-scale models, and other models focusing on specific regions such as the western Alps
103 (Kästle et al., 2018; Zhao et al., 2015, 2020), the southeastern Alps (Sadeghi-Bagherabadi
104 et al., 2021), the Vienna basin (Schippkus et al., 2020), the Bohemian massif (Kvapil et
105 al., 2021), the Ligurian sea, (Wolf et al., 2021; Magrini et al., 2022), have substantially
106 improved existing knowledge and have provided new insights into the complex deep structure
107 of the region. They highlighted Moho jumps beneath the external crystalline massifs (Lu

108 et al., 2018; Nouibat et al., 2022a) and Moho dipping following the subduction of European
109 continental crust in the southern Western Alps (Zhao et al., 2015, 2020; Nouibat et al.,
110 2022a), and provided clues about the lithological nature of the crust in the Ligurian basin
111 (Wolf et al., 2021; Nouibat et al., 2022b).

112 The above-mentioned models have been derived using conventional ANT, which involves
113 two steps: (1) 2-D traveltimes inversion for group or phase velocity maps, and (2) 1-D depth
114 inversion for V_s based on local dispersion curves. This procedure has two limitations that
115 may bias the geological interpretations: (1) the ray theory assumption when computing 2-D
116 Rayleigh-wave velocity maps, which is only valid in the high-frequency case (e.g., Snieder,
117 1986; Cervený, 2003), and (2) the local 1-D nature of the depth inversion in the second step,
118 which does not account for the 3-D lateral heterogeneity of the medium, thus limiting the
119 velocity model to be pseudo-3-D by construction. The aim of our study is to overcome these
120 limitations in the context of Western Alps and Ligurian-Provence basin by building a self-
121 consistent onshore-offshore 3-D velocity model using a wave-equation tomography of ambient
122 noise. This allows us to take into account the effect of 3-D structures and of the water
123 layer in the Ligurian sea on the propagation of surface waves. Indeed, wave-equation-based
124 tomographic methods are an alternative of choice to overcome such assumptions, as they
125 naturally accommodate for 3-D heterogeneity and finite-frequency effects, thus providing
126 more realistic sensitivity kernel for surface waves. These approaches consist in iteratively
127 updating the velocity model by minimising a misfit function between observed and synthetic
128 waveforms obtained through 3-D numerical modeling of seismic wave propagation.

129 In the following, we refer to ANT when the two-step method is used to invert the
130 correlations, to FWI when the seismic wave propagation is modeled in 3-D by solving the
131 wave equation and the misfit function involves the waveform i.e. phase and amplitude (e.g.,
132 Virieux & Operto, 2009, and references herein), and to wave equation tomography (WET)
133 when the misfit function involves only the phase (e.g., Luo & Schuster, 1991; Tape et al.,
134 2010).

135 With the availability of accurate large-scale seismic wavefield modeling and computing
136 resources, wave-equation-based tomographic methods have been widely applied to the crust
137 and upper mantle (e.g., Tape et al., 2010; H. Zhu et al., 2012; Yuan et al., 2014; Fichtner
138 & Villaseñor, 2015; Beller et al., 2018), often providing higher resolution images than those
139 obtained from ray theory and finite-frequency tomography. However, their application to

140 ambient noise data is challenging. Indeed, noise correlations provide the Green's function
141 of the medium only under certain conditions, for instance when considering a homogeneous
142 distribution of uncorrelated noise sources (Wapenaar, 2004; Roux et al., 2005), or more
143 generally assuming equipartition of the noise wavefield (Lobkis & Weaver, 2001; Snieder et
144 al., 2007; Weaver & Lobkis, 2001; Campillo, 2006). These conditions are rarely fully met in
145 practice. There are two ways to deal with this difficulty.

146 The first way consists in taking into account the distribution of the noise sources when
147 computing synthetic noise correlations and their sensitivity kernels. Indeed, the cross-
148 correlations waveforms can be strongly affected by the distribution of noise sources (Tromp
149 et al., 2010; Fichtner, 2014) and by the processing applied to the noise records (Bensen et
150 al., 2007; Fichtner et al., 2016, 2020). In particular, the amplitude of noise correlations
151 depends mostly on the energy and distribution of the noise sources (e.g., Hanasoge, 2014).
152 As a consequence, spatial variations of noise correlation amplitude cannot be interpreted
153 unambiguously as lateral contrasts of attenuation (Tsai, 2011; Stehly & Boué, 2017) but
154 can be used to image the distribution of noise sources (Stehly & Boué, 2017; Ermert et al.,
155 2016, 2017, 2021; Igel et al., 2021, 2023). Therefore, methodological efforts have been done
156 to apply FWI to noise correlations without assuming that they are similar to the Green's
157 function of the medium, but by treating them as self-consistent observables. This requires
158 to jointly invert for the distribution of noise sources and the Earth structure in order to take
159 into account the distribution of noise sources when computing synthetic noise correlations
160 and their sensitivity kernels (Tromp et al., 2010; Fichtner, 2014; Sager et al., 2018a, 2018b;
161 Hanasoge, 2014). However applying this approach to image the crust and upper mantle at a
162 continental scale remains challenging due to the complexity of the dynamics of microseismic
163 noise sources and the heterogeneity of the medium.

164 An alternative approach is to apply wave equation tomography (Luo & Schuster, 1991)
165 rather than FWI to noise correlations. Indeed, while heterogeneous distribution of noise
166 sources and data processing affect the correlations waveforms, travelttime measurements
167 of the correlations surface-wave fundamental mode are less affected (Tsai, 2009; Froment
168 et al., 2010; Tsai, 2011; Fichtner, 2014; Delaney et al., 2017). Consequently, in crustal
169 environment, the travelttime residuals depend more on the heterogeneity of the medium
170 than on the source distribution (Yang & Ritzwoller, 2008; Froment et al., 2010; Kimman
171 & Trampert, 2010). It is thus possible to assume that at least a subset of the correlations
172 provide accurately the Green's function surface-wave dispersion. Hence to apply WET, noise

173 correlations are modeled assuming that they are similar to the Green's function, i.e. without
174 taking into account the distribution of noise sources. In this case the traveltimes (phase)
175 of surface-wave fundamental mode is inverted rather than the waveform. However, this
176 approach has two main limitations: (1) it does not account for the noise sources distribution
177 which can bias the apparent arrival time of surface waves (Tsai, 2009; Froment et al., 2010;
178 Delaney et al., 2017). This is mitigated in practice by selecting only station pairs for which
179 the arrival time of surface waves is similar in the positive and negative correlation time; (2)
180 noise correlations that contain mostly the fundamental mode of surface waves are compared
181 with synthetic correlations that contains all possible mode of propagation. This may affect
182 misfit measurements if the different modes are not well separated in time. Nevertheless,
183 WET applications to noise correlations has recently demonstrated its potential to provide
184 accurate velocity models of the crust and upper mantle that better explain the observations
185 than conventional ANT (Chen et al., 2014; Liu et al., 2017; Lu et al., 2020).

186 Using the ANT model from Lu et al. (2018) as an initial model, Lu et al. (2020) were
187 the first to apply wave-equation tomography to the alpine region using noise correlations.
188 They demonstrated that WET is relevant to obtain models that better explain the data,
189 and to improve the resolution of ANT models in the Alps. However, their WET model
190 does not cover the Ligurian sea since the offshore part of the AlpArray network was not yet
191 available. We refer the reader to Nouibat et al. (2022a, 2022b) for more details about the
192 comparison of recent ANT methodologies and models in the region.

193 In this study, we improve the WET methodology used by Lu et al. (2020). These
194 improvements consist in: (i) performing an acoustic-elastic coupled wave-equation tomogra-
195 phy to better represent the wave-physics and consider more realistic constraints on velocity
196 structure in the oceanic domain of the Ligurian basin and along the continental margins;
197 (ii) inverting seismic data in a broader band (5-85 s) in order to constrain the shallow part
198 of the crust as well as the upper mantle, employing a hierarchical inversion strategy that
199 avoids possible cycle-skipping issues; (iii) using a random sub-sampling scheme over 185
200 virtual sources rather than a fixed number of virtual sources. Similar to Lu et al. (2020),
201 we minimize the frequency-dependent phase traveltimes differences of Rayleigh waves and
202 tackle the inversion using the SEISCOPE SEM46 code originally developed for exploration
203 scales (Trinh et al., 2019). Taking advantage of the densest seismological coverage in West-
204 ern Europe including the entire AlpArray network, we perform a large-scale wave-equation

205 tomography in order to refine the land-sea model by Nouibat et al. (2022a) obtained from
206 conventional ANT.

207 The availability of Rayleigh-wave traveltime measurements across and within the Lig-
208 urian sea, and the latest developments of full waveform modeling and inversion in the fluid-
209 solid coupled media (Cao et al., 2022), motivated us to go further in the improvement of
210 wave-equation tomography by considering the effect of the water layer on the 3-D propaga-
211 tion of surface waves from ambient noise correlations.

212 In an oceanic domain, two different types of surface waves can be recorded on the
213 vertical component. At large periods, when the wavelength of the surface waves is large
214 compared to the thickness of the water layer, the effect of the water on wave propagation is
215 negligible, and the surface waves propagate as Rayleigh waves. At shorter periods, however,
216 the effect of the water layer can no longer be neglected and the surface wave propagates
217 as a Rayleigh-Scholte wave, which is a fluid-solid interface wave. Nouibat et al. (2022b)
218 have shown that a water layer thicker than 0.5 km can have a significant effect on the
219 Rayleigh-Scholte wave dispersion at period ≤ 20 s and should therefore be taken into account
220 in tomographic studies. Thus, several studies have taken the water column into account
221 when inverting the local dispersion curve of Rayleigh-Scholte waves extracted from noise
222 correlation into a local 1-D V_s model (e.g., Mordret et al., 2014; Guerin et al., 2020; Wolf
223 et al., 2021; Nouibat et al., 2022b; Carvalho et al., 2022). However, this assumes a strictly
224 1-D coupling in a laterally homogeneous medium.

225 In our study, the 3-D effects of the presence of water layer are considered on the surface-
226 wave propagation. For this purpose, we apply the technique of acoustic-elastic coupling in
227 the waveform modeling to deal with the seabed relief. The objective is twofold. Firstly, to
228 investigate the influence of the water layer on the propagation of surface waves between on-
229 shore and sea-bottom stations, more specifically Rayleigh-Scholte waves. To our knowledge,
230 this is the first study documenting the influence of the water layer on the 3-D propagation of
231 ambient noise surface waves. Secondly, to incorporate the fluid-solid coupling in the inver-
232 sion framework, which makes this study the first ambient noise wave-equation tomography
233 to take into account the effect of the water layer.

234 The structure of the manuscript is organized as follows. Section 2 describes the dataset
235 of ambient noise correlations and the initial velocity model used in the wave-equation to-
236 mography. The overall methodology is presented in Sections 3–4. Section 3 introduces the

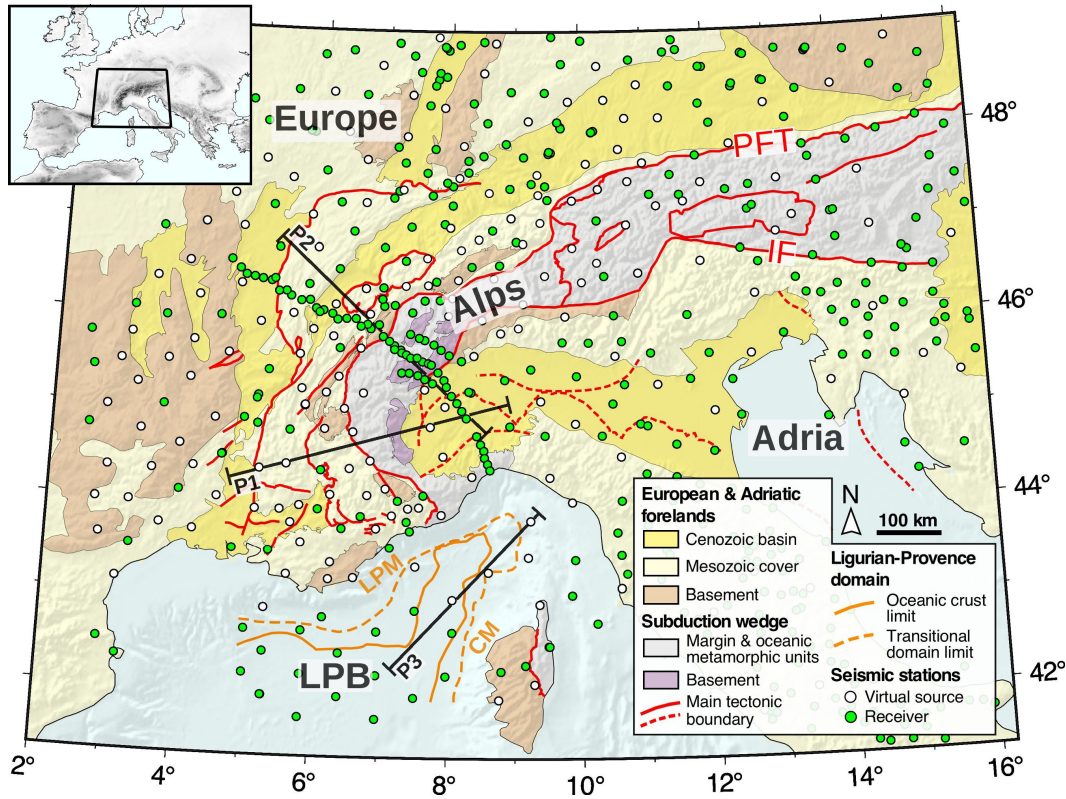


Figure 1. Geological and tectonic setting of the study area, modified from Handy et al. (2010) and Rollet et al. (2002), with locations of seismic stations used in this work (white circles: virtual sources; green circles: receivers). LPB: Ligurian-Provence basin, LPM: Ligurian-Provence margin, CM: Corsican margin. Black lines show locations of the seismic profiles discussed in the text.

237 iterative process of the WET workflow, while Section 4 is dedicated to 3-D modeling of
 238 surface waves, with emphasis on the acoustic-elastic case. In the light of the specific case
 239 of the Ligurian-Provence basin, we document the importance of the fluid-solid coupling for
 240 marine crustal imaging based on ambient noise data. Section 5 presents tomography results
 241 and related discussions. In Section 6, we assess the robustness of the resulting 3-D velocity
 242 model.

243 2 Data

244 2.1 Ambient Noise Data

245 We use a subset of the data used by Nouibat et al. (2022a, 2022b) consisting of vertical
 246 component cross-correlations computed between 600 broadband stations from all available
 247 temporary and permanent networks in the Alpine region (Fig. 1, 2°E-16.2°E; 41.3°N-49.1°N)

248 during the period 2015-2019. The station array includes the entire AlpArray Seismic Net-
 249 work (AASN) and the Cifalps-2 experiment.

250 The offshore component of the dataset consists of high-quality correlations between the
 251 AASN OBSs obtained from an iterative procedure involving onshore stations of the AASN
 252 and permanent networks. Beforehand, the OBSs daily noise records were cleaned from
 253 transients (e.g., glitches) and seafloor-induced noises (compliance, tilt). For more details
 254 about this specific processing, readers can refer to Nouibat et al. (2022b).

255 We keep only reliable cross-correlations by applying the following criteria: (1) signal-to-
 256 noise ratio (SNR) greater than 5, (2) difference in group velocity measured on the positive
 257 and negative parts less than 0.2 km/s, and (3) inter-station distance larger than one wave-
 258 length for the maximum period considered in each period band. The SNR is defined as
 259 the ratio of the maximum amplitude of the surface wave to the standard deviation of the
 260 signal starting after the surface wave window. The final dataset includes $\sim 22 \times 10^3$ to
 261 55×10^3 high-quality cross-correlations depending on the period band (Table. 1). For the
 262 wave-equation tomography, all stations in Figure 1 serve as receivers, out of which 185
 263 stations are used as virtual sources (white circles).

264 **2.2 Initial Model**

265 The initial model consists of the 3-D onshore-offshore V_s model by Nouibat et al. (2022a,
 266 2022b), and a V_p model converted from this V_s by the empirical formula from Brocher (2005):
 267 $V_p = 0.9409 + 2.0947V_s - 0.8206V_s^2 + 0.2683V_s^3 - 0.0251V_s^4$. The V_s model by Nouibat et al.
 268 (2022a, 2022b) was obtained from ambient noise tomography using all available permanent
 269 and temporary seismic networks in Western Europe from 2015 to 2019. This model is derived
 270 from a hybrid data-driven tomography involving a 2-D transdimensional Bayesian inversion
 271 for group-velocity maps and their uncertainties, and a 1-D probabilistic inversion for V_s at
 272 depth. The onshore part of this model has been validated in the western Alps by comparison
 273 with other available geophysical studies, e.g, receiver functions (RFs), controlled-source seis-
 274 mic (CSS) experiments, Bouguer anomaly, along the Cifalps-1 & Cifalps-2 profiles (Nouibat
 275 et al., 2022a; Paul et al., 2022). The offshore part has been validated by comparison with
 276 a high-resolution V_p cross-section derived from refraction and wide-angle seismic profiling
 277 along the axis of the Ligurian basin (Nouibat et al., 2022b). However, it remains limited
 278 by the assumptions made in the two steps of the inversion, in particular the high-frequency

279 assumption for Rayleigh-wave propagation in the 2-D inversion, and the 1-D assumption
 280 in the inversion for V_s . Because of the 1-D assumption, they assumed a 1-D fluid-solid
 281 coupling in the oceanic domain, which is not suitable for dispersion of Rayleigh-Scholte
 282 waves or conversion of Rayleigh waves to Rayleigh-Scholte waves across the margins. The
 283 objective of the present study is to overcome these limitations and improve the resolution
 284 of the V_s model by performing 3-D wave-equation tomography.

285 **3 Iterative Inversion Scheme**

Similar to Lu et al. (2020), we formulate the inverse problem as the minimization of the misfit function (ξ) defined by the differences of frequency-dependent phase traveltimes between observed and synthetic vertical-component waveforms of ambient-noise Rayleigh waves

$$\operatorname{argmin}_{\mathbf{m}} \xi(\mathbf{m}) = \frac{1}{2} \sum_i \sum_{\omega} \Delta T_i(\omega, \mathbf{m})^2, \quad (1)$$

where $\Delta T_i(\omega, \mathbf{m})$ is the phase traveltimes residual for the i th station pair at frequency ω , given the model parameters (\mathbf{m}). We measure ΔT using a multi-taper technique (Tape et al., 2010). We tackle the minimization of (ξ) using a quasi-Newton local optimization method, with the following iterative scheme (Métivier & Brossier, 2016):

$$\mathbf{m}_{k+1} = \mathbf{m}_k + \alpha_k \delta \mathbf{m}_k \quad (2)$$

286 where α_k is the step length estimated by line search strategy and $\delta \mathbf{m}_k = -B_k \nabla \xi(\mathbf{m}_k)$ is
 287 the model update in which B_k is an approximation of the inverse of the Hessian matrix
 288 (i.e. the second-order derivative of ξ with respect to model parameters). We use the limited
 289 memory version of quasi-Newton methods (l-BFGS, Nocedal, 1980) to estimate B_k using
 290 gradients at few previous iterations. We update the velocity model by inverting from low
 291 to high frequencies, in four narrow period bands: 5-10 s, 10-20 s, 20-40 s, and 40-85 s.
 292 Given the dispersive character and depth sensitivity of the Rayleigh-wave phase velocity at
 293 these periods, such a hierarchical approach is useful to prevent cycle skipping issue. This
 294 approach first constrains large-scale anomalies in the upper mantle part of the model using
 295 the long-period content, then it constrains small-scale anomalies in its crustal part using
 296 the short-period content.

297 We reconstruct simultaneously V_s and V_p to account for the sensitivity of the Rayleigh-
 298 wave to V_p , which becomes non-negligible at short periods (e.g., Eddy & Ekström, 2014;
 299 Qiao et al., 2018). Therefore, the relative perturbation of the misfit function (ξ) to per-

300 turbations of the two parameters is given by the linear relation: $\delta\xi(\mathbf{m}) = K_{V_s}(\mathbf{m})\delta\ln V_s +$
 301 $K_{V_p}(\mathbf{m})\delta\ln V_p$, where K_{V_s} and K_{V_p} are the sensitivity kernels for V_s and V_p respectively
 302 (i.e. misfit gradients with respect to V_s and V_p). Due to the computational cost of modeling
 303 for multiple sources, we employ a randomized sub-sampling strategy of the ensemble of virtual
 304 sources over iterations in order to achieve optimal path coverage without compromising
 305 the quality of the inversion results, as usually applied for exploration scales (Warner et al.,
 306 2013). Following the workflow shown in Fig. 2, our wave-equation-based tomography is
 307 summarized in the main steps below:

- 308 (1) Perform the forward modeling for one subset of 64 virtual sources.
- 309 (2) Measure phase traveltimes misfits between observed and simulated waveforms and
 310 compute the adjoint sources.
- 311 (3) Perform the adjoint modeling and extract gradients from cross-correlations of incident
 312 and adjoint fields.
- 313 (4) Gradients are summed and smoothed using a non-stationary Laplace smoothing filter
 314 (approximated by a 2nd-order Bessel filter) to remove high wavenumber artefacts and
 315 constrain the inversion (Trinh et al., 2017). The filter is defined by coherent lengths
 316 that are directly computed from the inverted parameters V_p and V_s : $L_z = L_x = L_y =$
 317 $0.1 \times \lambda_l$ where λ_l is the local wavelength at the maximum frequency considered.
- 318 (5) Estimate the step length and update the current velocity model.
- 319 (6) Change the subset each 3 iterations. The output model after 9 iterations is used
 320 as a starting model for the next shorter period band following the same iterative
 321 procedure. The density model is updated accordingly using the empirical formula
 322 from Brocher (2005): $\rho = 1.6612V_p - 0.4721V_p^2 + 0.0671V_p^3 - 0.0043V_p^4 + 0.000106V_p^5$.

Period band [s]	Nb of elements (nZ,nX,nY)	Element size (dZ, dX, dY) [km]	Nb of data
40 - 85	16, 81, 60	10, 15, 15	31 212
20 - 40	20, 101, 75	8, 12, 12	43 350
10 - 20	27, 152, 112	6, 8, 8	54 870
5 - 10	41, 202, 150	4, 6, 6	22 066

Table 1. Spectral elements number/size and number of measurements used in each period band.

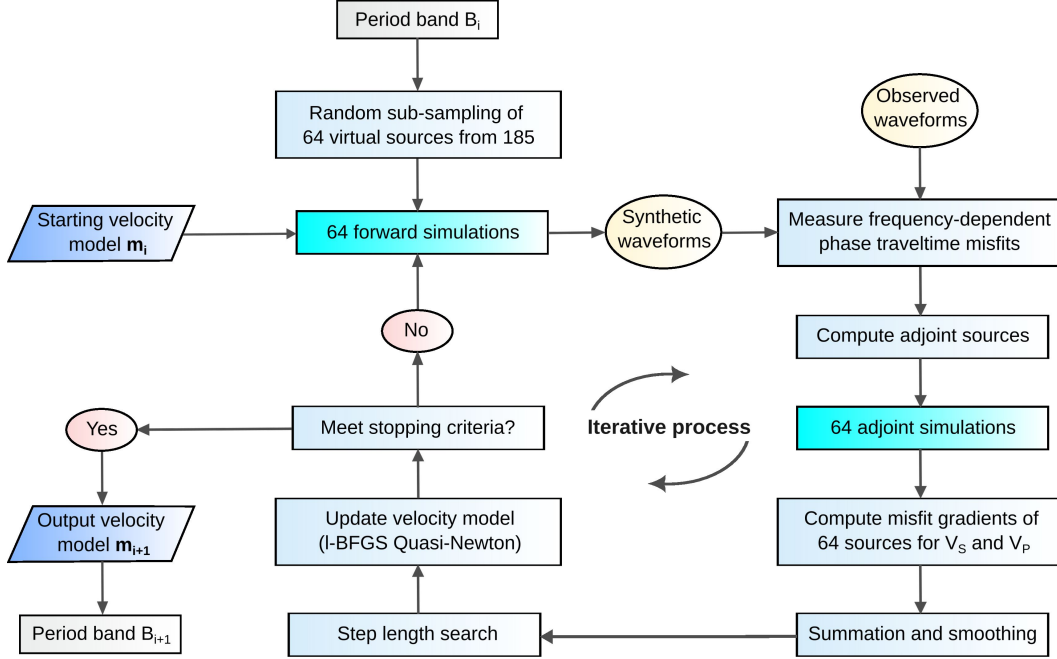


Figure 2. General workflow of the wave-equation tomography (WET).

4 3-D Simulations of Surface-Wave Propagation

We perform forward and adjoint simulations using the time-domain wave equation modeling of the SEM46 spectral-element-based code. We perform simulations for the vertical component of Rayleigh waves by applying a point-force at the location of virtual sources with a filtered Dirac delta as source time function (identical for all virtual sources). Therefore, the synthetic waveform (u_{syn}) at a receiver position for model (\mathbf{m}) is the convolution product between the source time function (s) and the synthetic Green's function (G_{syn}) for the source–receiver pair (e.g., Lu et al., 2020): $u_{syn}(\mathbf{m}, t) = s(t) * G_{syn}(\mathbf{m}, t)$. To compare with synthetic waveforms, we convolve the source time function (s) with the time derivative of the noise cross-correlation (C) for the source–receiver pair, which is a good approximation of the Green's function (G) of the target medium (e.g., Lobkis & Weaver, 2001; Snieder, 2004; Wapenaar, 2004; Roux et al., 2005; Weaver, 2005): $u(t) = s(t) * -\partial_t C(t) \approx s(t) * G(t)$. The time sampling of simulation varies from 0.04 to 0.0015 s, with the Courant–Friedrichs–Lewy (CFL) condition satisfied (Komatitsch & Tromp, 1999). Surface waves are extracted in a time window with $\min[d/5, t_{max} - 1.5T]$ and $\max[d/2, t_{max} + 1.5T]$ as lower and upper limits in the 10-20 s, 20-40 s, and 40-85 s, and a time window with $\min[d/4, t_{max} - 2T]$ and $\max[d/1.5, t_{max} + 2T]$ as lower and upper limits in the 5-10 s, where d is the inter-station

340 distance (in km), t_{max} is the arrival time of the maximum of the envelope of synthetic
 341 waveform, and T is the maximum period of the considered band.

342 4.1 Elastic modeling of Rayleigh Waves

We simulate the incident wavefield in the 20-85 s period band by solving the second-order elastic wave equation:

$$\rho_s \partial_{tt} \mathbf{u} = \nabla \cdot \boldsymbol{\sigma} + \mathbf{f}_s, \quad \boldsymbol{\sigma} = \mathbf{C} \boldsymbol{\varepsilon} \quad (3)$$

343 where \mathbf{u} is the displacement field, \mathbf{f}_s is the force vector, $\boldsymbol{\sigma}$ and $\boldsymbol{\varepsilon}$ are the second-order stress
 344 and strain tensors respectively, ρ_s is the solid density and \mathbf{C} is the fourth-order elastic
 345 stiffness tensor. We employ a flexible Cartesian-based hexahedral mesh with element size
 346 varying from 15 to 6 km in the horizontal direction and from 10 to 4 km in the vertical
 347 direction (see Table. 1). The wavefield is interpolated inside each spectral element using
 348 Lagrange polynomials of order 4. We parameterize Gauss–Lobatto–Legendre (GLL) points
 349 by isotropic S-wave velocity (V_s) starting from the model described in section 2, and (V_p ,
 350 ρ) computed from V_s (Brocher, 2005). Finally, we vertically deform the mesh grid for an
 351 accurate representation of the complex topography(bathymetry) on the free surface and to
 352 handle the Earth curvature.

Misfit elementary gradients with respect to the medium parameters are obtained from the zero-lag correlation of adjoint and incident wavefields, based on the adjoint-state method (e.g., Tromp et al., 2005; Plessix, 2006). Adjoint wavefields are obtained by injecting the adjoint sources at receiver locations in the adjoint wave equations. The adjoint source for the multi-taper method is given by the derivative of synthetic waveforms weighted by the frequency-dependent phase difference measurements (Tape et al., 2010). Gradients of (ξ) with respect to the solid parameters (ρ_s, C_{IJ}) are obtained by the zero-lag cross-correlations

$$\frac{\partial \xi(\mathbf{m})}{\partial \rho_s} = (\bar{\mathbf{u}}, \partial_{tt} \mathbf{u})_{\Omega_s, t}; \quad \frac{\partial \xi(\mathbf{m})}{\partial C_{IJ}} = \left(\bar{\boldsymbol{\varepsilon}}, \frac{\partial \mathbf{C}}{\partial C_{IJ}} \boldsymbol{\varepsilon} \right)_{\Omega_s, t} \quad (4)$$

353 where $\bar{\mathbf{u}}$ is the adjoint wavefield vector associated with \mathbf{u} , and $\bar{\boldsymbol{\varepsilon}}$ is the adjoint of the strain
 354 field $\boldsymbol{\varepsilon}$ for model sets Ω_s .

355 4.2 Acoustic-Elastic modeling of Scholte-Rayleigh Waves

Since we also invert phase dispersion data for S -wave velocities in the Ligurian-Provence basin, we have to consider the influence of the seafloor topography on the propagation of

surface waves in the 5-20 s period band. Indeed, Scholte waves can be excited as a result of the fluid-solid interaction (e.g., Zheng et al., 2013; J. Zhu et al., 2004). Scholte waves can also be generated from Rayleigh-Scholte mode conversion at the continent-ocean transitions. To further investigate this influence, we account for the effect of the water layer by considering fluid-solid coupling in the 3-D wave equation through the second-order equation system (Cao et al., 2022):

$$\begin{aligned}
 \rho_s \partial_{tt} \mathbf{u} &= \nabla \cdot \boldsymbol{\sigma} + \mathbf{f}_s, & \boldsymbol{\sigma} &= \mathbf{C}\boldsymbol{\varepsilon}, & \text{in } \Omega_s, \\
 \frac{1}{\kappa} \partial_{tt} \varphi - \nabla \cdot \left(\frac{1}{\rho_f} \nabla \varphi \right) &= \frac{1}{\kappa} \iint -P_f dt dt, & & & \text{in } \Omega_f, \\
 \mathbf{u} \cdot \mathbf{n} &= \frac{1}{\rho_f} \nabla \varphi \cdot \mathbf{n}, & \boldsymbol{\sigma} \cdot \mathbf{n} &= \partial_{tt} \varphi \mathbf{n}, & \text{on } \Gamma_{fs}.
 \end{aligned} \tag{5}$$

where the first equation describes the elastic wave propagation in the solid domain (Ω_s) in terms of displacement vector \mathbf{u} (same as Eq. 3). The second equation describes the acoustic wave propagation in the fluid domain (Ω_f) in terms of scalar displacement potential φ . P_f is the pressure source associated with a force vector \mathbf{f}_f , ρ_s is the fluid density and κ is the bulk modulus of the fluid. The third system gives the boundary conditions along the fluid-solid interface (Γ_{fs}), describing the interaction between the two domains. In our implementation, we consider to discretize the topography of onshore-offshore transition zones through staircase functions to preserve the configuration of structured mesh used in SEM46. We consider a minimum vertical size of 200 m for elements in the water column, ensuring proper sampling of waves in water. Misfit gradients in the solid domain are given by expressions (4). In the fluid domain, gradient building is performed following the hybrid approach proposed by Cao et al. (2022). Gradients of (ξ) with respect to the fluid parameters (ρ_f, κ) are then obtained by the zero-lag cross-correlations:

$$\frac{\partial \xi(\mathbf{m})}{\partial \rho_f} = \left(\nabla \bar{P}, \frac{1}{\rho_f^2} \nabla P \right)_{\Omega_f, t}; \quad \frac{\partial \xi(\mathbf{m})}{\partial \kappa} = \left(\bar{P}, \frac{1}{\kappa^2} \partial_{tt} P \right)_{\Omega_f, t} \tag{6}$$

356 The gradients on V_p and V_s are estimated using elementary gradients (4) and (6) based on
 357 chain rule.

358 Figure 3 shows snapshots of the simulated vertical displacement recorded at the sur-
 359 face, in the 5-10 s period band for an onshore source station located in southern France. We
 360 performed elastic (Figs. 3a1-c1) and acoustic-elastic modeling (Figs. 3a2-c2) in the initial
 361 model. Snapshots for the simulation in the 10-20 s period band are shown in the supple-
 362 mentary Fig S1. Figure 4 shows synthetic waveform fits to data at an OBS located in
 363 the central Ligurian-Provence basin at a depth of ~ 2700 m (location in Fig. 4a). Fig-

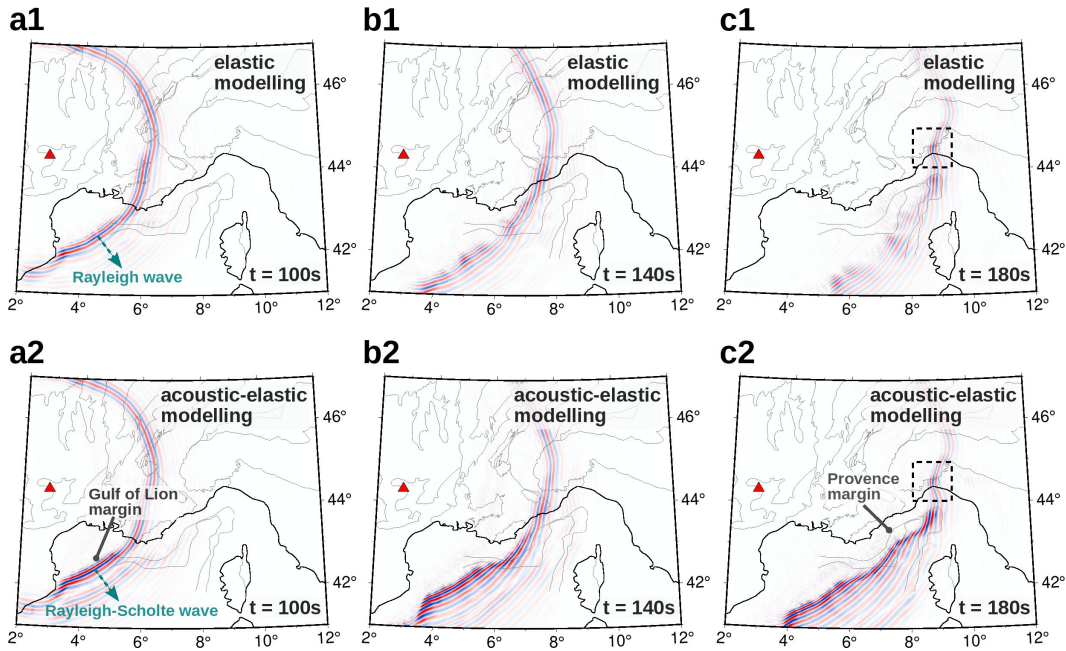


Figure 3. Comparison of snapshots of the vertical displacement wavefield in the 5-10 s period band, simulated using (a1-c1) the elastic wave equation, (a2-c2) the acoustic-elastic coupled wave equation, for the source station indicated by the red triangle. The wave fields are extracted on a 3-D surface corresponding to the topography in the onshore part and to the bathymetry of the seafloor in the marine part. Black dashed frames show scattered Rayleigh-wave packets. The thin black lines are geological and tectonic boundaries from Figure 1.

364 ures 4b1,b2;c1,c2;d1,d2 show simulations in the initial model and Figs. 4b3,c3,d3 show sim-
365 ulations in the final model. In the elastic modeling case, the source–OBS pair is assumed
366 to be located on the free surface. We therefore observe the propagation of a Rayleigh wave
367 everywhere in the oceanic part (Figs. 3a1-c1). In the acoustic-elastic case, we account for
368 the fluid-solid coupling by considering the presence of the water layer in the Ligurian sea. In
369 this case, we observe a mode conversion from Rayleigh to Rayleigh-Scholte wave at the Gulf
370 of Lion margin (Figs. 3a2-c2). The resulting wave packet is characterized by stronger ampli-
371 tude and slower propagation velocity and includes strong wavefront distortions. However,
372 we notice in both cases a dissipation within the Ligurian-Provence basin and almost the
373 same wavefield in the continental part, as expected. For instance, both simulations evidence
374 a scattering of the wave packet to the north of the Ligurian sea (dashed black frame in
375 Figs. 2c1,c2). The discrepancies in amplitude and traveltimes in the oceanic part are signifi-
376 cant. Indeed, the arrival time of the Rayleigh-wave envelope obtained from elastic modeling
377 in the 5-10 s band is ~ 35 s earlier than that of the Rayleigh-Scholte wave obtained from
378 acoustic-elastic modeling (Figs. 4b1-b2). Furthermore, the arrival time of the observed sur-
379 face wave is much closer to the arrival time of the simulated signal with fluid-solid coupling
380 (Figs. 4b2). This fit is further improved after inversion as shown in Fig. 4b3. Differences
381 between the two modeling results are also visible in the 10-20 s period band with a lower
382 but still significant misfit due to the free-surface boundary condition (Figs. 4c1-c3). As
383 illustrated in Figs. 4d1-d2, the differences between elastic and acoustic-elastic simulations
384 are negligible in the 20-40 s period band, showing that the surface wave in these period
385 ranges becomes insensitive to the coupling effect and presence of water. Furthermore, the
386 waveform computed from elastic modeling in the final model fits very well the observed
387 signal (Fig. 4d3), indicating that the elastic assumption is realistic-enough at long periods.

388 Figure 5 shows snapshots extracted at the solid surface (topography and seafloor) and
389 at 10 km depth, in the 5-10 s period band for an OBS source located in the northeastern part
390 of the Ligurian-Provence basin. We observe a significant amplitude attenuation at depth of
391 the wave propagating in the Ligurian-Provence and Tyrrhenian basins. This is consistent
392 with the trapping of the Rayleigh-Scholte wave in the vicinity of the elastic-acoustic interface
393 (e.g., Nayfeh, 1995). Hence, we detect the long-period dispersion of the elastic wave at 10 km
394 depth, characterized by less distorted wavefront. Finally, Figure 5 shows that wave packets
395 are dissipated in the oceanic part (Ligurian-Provence and Tyrrhenian basins) or through
396 the sharp Provence margin towards the continental part. This suggests that considering the

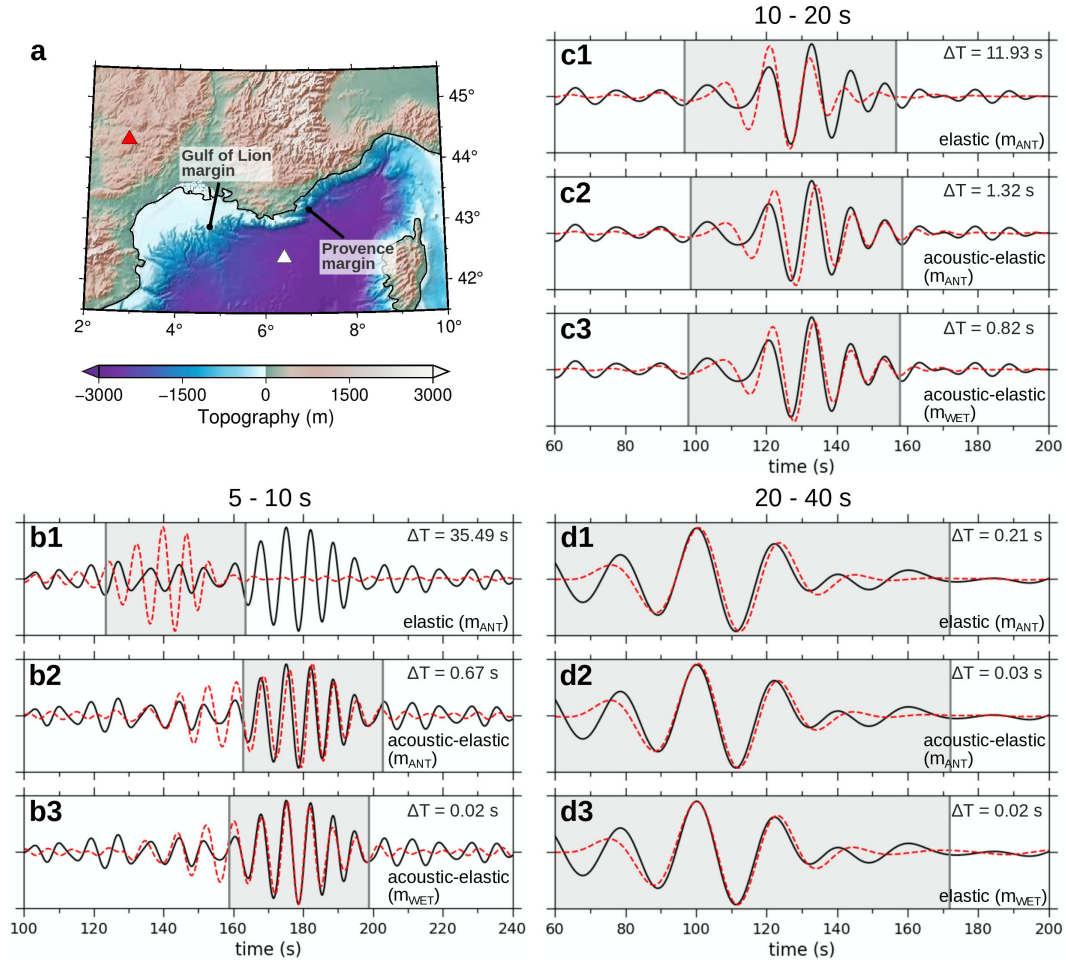


Figure 4. Comparisons between simulated vertical-component seismograms (red) and observed noise correlation waveforms (black), filtered in the 5-10 s, 10-20 s and 20-40 s period bands, for the source-receiver pair in (a) (red: station source, white: receiver). Elastic and acoustic-elastic simulations in the initial model (labeled " m_{ANT} ") are shown in b1, c1, d1 and b2, c2, d2 respectively. Seismograms in b3, c3 are computed from acoustic-elastic modeling in the final model (labeled " m_{WET} ") while d3 is computed from elastic modeling. The gray areas indicate the surface-wave window and ΔT corresponds to the cross-correlation-based time shift between observed and synthetic windowed waveforms.

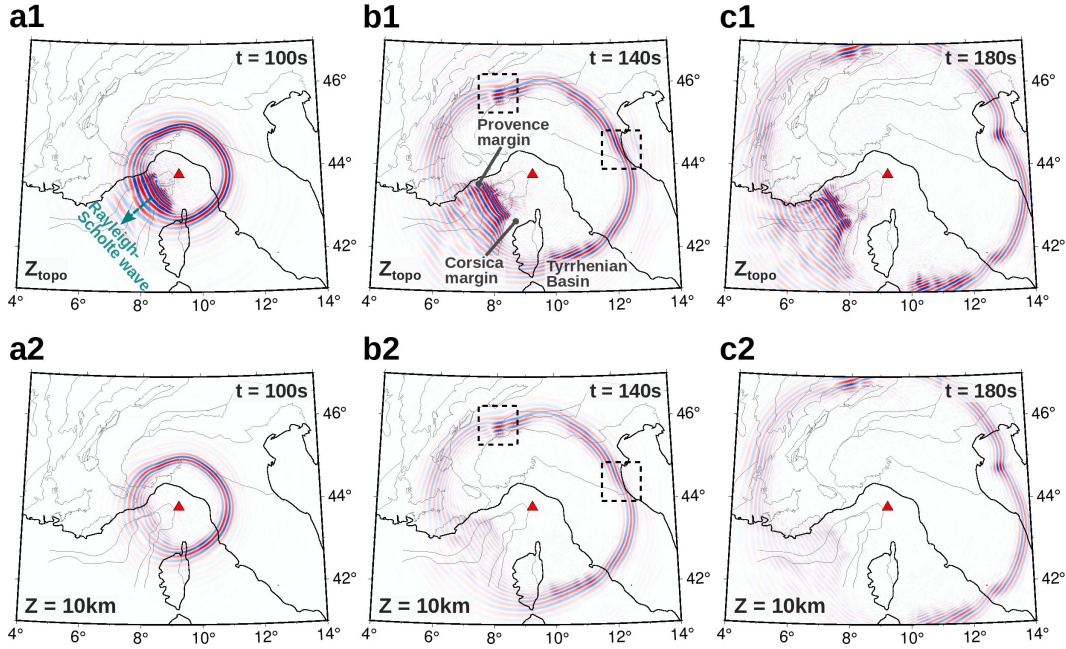


Figure 5. Snapshots of the wavefield computed from the acoustic-elastic coupled wave equation in the 5-10 s period band, showing the depth attenuation of the Rayleigh-Scholte wave generated by an offshore source (red triangle). Wavefields are extracted (a1-c1) at the surface (Z_{topo} , same definition as in Fig. 3) and (a2-c2) at 10 km depth. Black dashed frames show scattered Rayleigh-wave packets.

397 fluid-solid interaction would help to recover the shallow 3-D velocity structures, as it will
 398 be illustrated in the next section.

399 5 Results and Discussion

400 As Rayleigh waves from ambient-noise correlations in the period range 5-85 s are mainly
 401 sensitive to shear-wave velocities at depth, i.e. the most significant variations from the initial
 402 model would occur in V_s , we analyse here depth slices and Moho maps from the obtained
 403 3-D V_s model and discuss their geological implications.

404 5.1 Depth Slices

405 Figure 6 shows the depth slices of the initial and final V_s models at 6 km (Figs. 6a1-a2)
 406 and 26 km (Figs. 6b1-b2) and corresponding relative variations (Figs. 6a3,b3). We observe
 407 a roughly preserved geometry of velocity structures but strong relative variations. At 6 km-

408 depth, the final model exhibits lower velocities (≤ 2.5 km/s) with strong negative velocity
 409 variations ($\leq -8\%$) in the Ligurian-Provence basin. These low velocities probably reflect the
 410 slow propagation of Rayleigh-Scholte waves, induced by the sedimentary cover. Another new
 411 pattern appears in Corsica where the western Variscan Corsica (see Fig. 1) exhibits higher
 412 velocities (≥ 3.5 km/s) than the northeastern Alpine Corsica. Such variations are probably
 413 due to Rayleigh-Scholte mode conversions across the Corsican margin, which are taken into
 414 account by the 3-D fluid-solid coupling. On the onshore domain, relative variations roughly
 415 coincide with geological structures, with a velocity decrease in the western and eastern parts
 416 of the subduction wedge and an increase in the forelands.

417 The 26-km maps (Figs. 8b1-b3) display two striking features: (1) a velocity decrease
 418 (down to -8%) in the subduction wedge with a northeastward focusing of the low-velocity
 419 anomaly (LVZ) beneath the W-Alps, and (2) a velocity increase (up to 8%) in the crust
 420 beneath the Alpine forelands and the Apennines. The transitions between these domains
 421 show small variations to the initial model, highlighting the effectiveness of the ANT model
 422 in dealing with sharp transitions along the main tectonic boundaries (e.g., the subduction
 423 wedge-European foreland boundary, see Fig. 1). We will see in Section 6.1 that such transi-
 424 tions exhibit a low average misfit. Nevertheless, the strong variations in velocity between the
 425 ANT and WET models demonstrate the strengths of WET in addressing the high-frequency
 426 approximation and 1-D assumption issues associated with the ANT procedure. The com-
 427 parison with the WET model by Lu et al. (2020) is shown in Supplementary text S1 and
 428 Figure S2. The initial and final V_p models remain fairly close due to the limited sensitivity
 429 of Rayleigh waves to P-wave velocity (Supplementary Fig. S3).

430 5.2 Vertical Cross-Sections

431 We now focus on the western Alpine and Ligurian-Provence regions, and compare the
 432 initial and final V_s models along 2-D cross-sections of the crust and uppermost mantle
 433 (profiles P1-3 in Fig. 1). These profiles traverse complex geological domains that are well
 434 known at the surface and have already been investigated at depth using other techniques
 435 with independent data. The underlying three-dimensional structure is therefore particularly
 436 interesting for validating the continental and oceanic parts of the model.

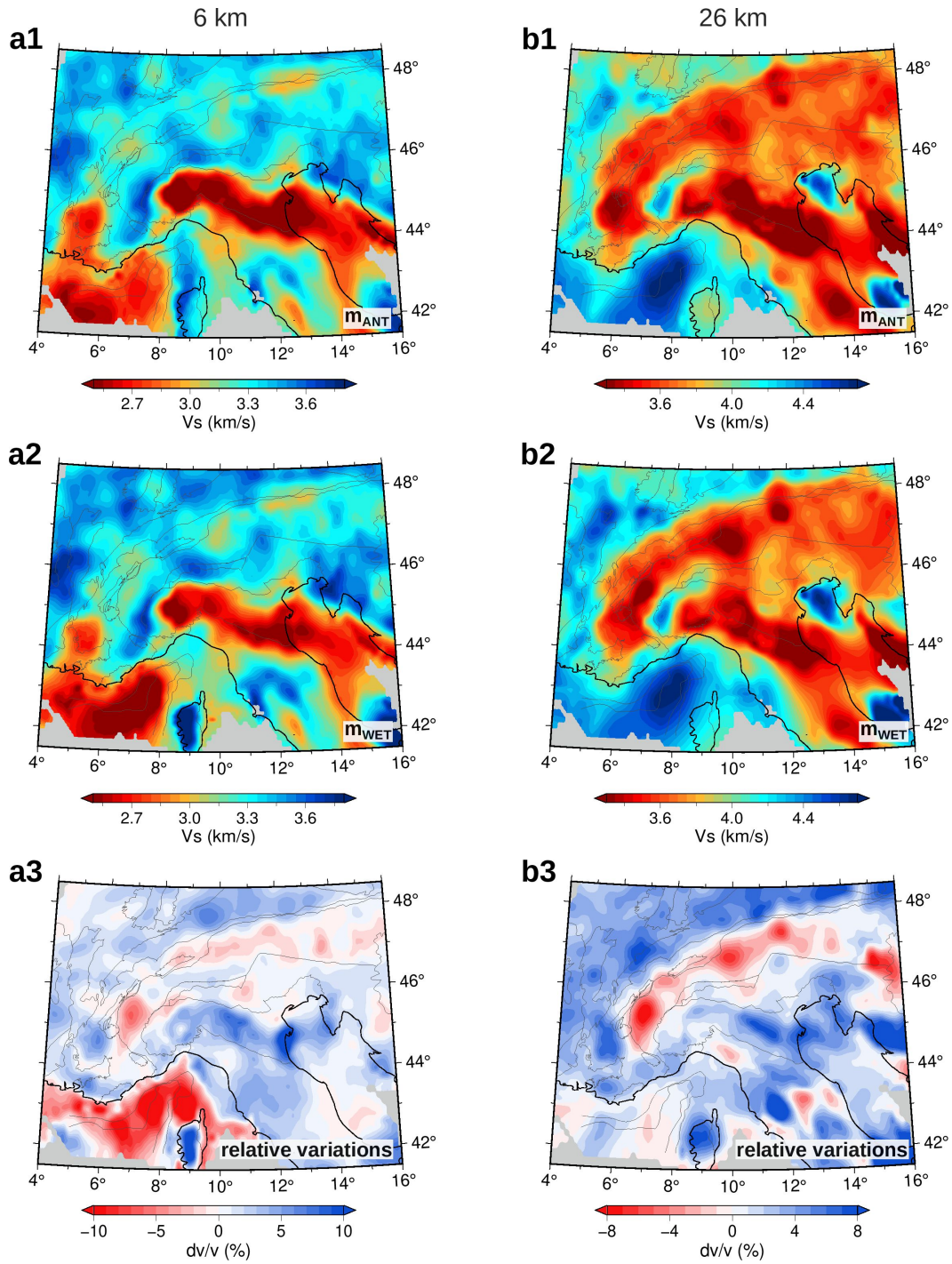


Figure 6. Depth slices in the initial (a1-b1) and final (a2-b2) shear-wave velocity models and the relative variations to the initial model (a3-b3), at 6 km (a1-a3) and 26 km (b1-b3) depths.

5.2.1 Cross-sections in the western Alps

The profile P1 (Fig. 1) coincides with the Cifalps transect, which was extensively investigated in a number of tomographic studies (Malusà et al., 2021, and references herein). The profile P2 is located close to ECORS-CROP wide-angle reflection transect (Nicolas et al., 1990) and Cifalps-2 RFs transect (Paul et al., 2022). The two profiles cross the W-Alps from the European foreland (subducted lower plate) to the Adriatic foreland (upper plate), and the subduction wedge delimited by the Penninic Frontal Thrust (PFT) to the west and the Insubric Fault (IF) to the east (Figs. 7a1-b1).

The cross-sections in initial (ANT) and final (WET) V_s models as well as their relative velocity changes are shown in Figure 7. The final model displays lower velocities in the European mantle and higher velocities in Adria mantle ($\pm 2-3\%$). We note a progressive increasing of velocities beneath the subduction wedge, as highlighted by positive-velocity variations from 40 km depth and 180 km offset. Such variations at large depths indicate a significant contribution to the model update from the long-period component of the dispersion data. The European crust (ECC) is marked by an attenuation ($dv/v \geq +5\%$ in P1) of the lower crust LVZ (label "1") beneath the foreland, and strengthening ($dv/v \leq -3-4\%$) of the LVZ (label "2") beneath PFT trace. These two velocity changes are already visible in Figure 6, indicating a north-eastwards shift of LVZ1. This shift is likely the result of correcting for a bias from the seismic ray deviation that is not accounted for in the ANT model due to the high-frequency approximation. It also reflects the higher resolution of the final model. The resolution tests in Section 6.4 show that such perturbations are fairly reliable. The final model exhibits smaller changes in Adria crust (ACC), which is less structurally complex than ECC, indicating that it is indeed well constrained at depth by the ANT.

Similarly to RFs Moho (dashed yellow line), the 4.2 km/s contour of the final model dips continuously following the subduction of European lithosphere, reaching a maximum depth of ~ 80 km. In the north (profile P2), the European Moho reaches only 55 km maximum depth, which is consistent with CSS and RFs data along ECORS-CROP and Cifalps-2. We therefore define this contour as proxy for Moho in W-Alps. Below the subduction wedge (profile P1), the Moho geometry is more stable than that of the ANT which extends to depths of more than 90 km (location "3"). Similar changes along P2 underline a decrease in the maximum depth (55 km instead of 70 km). We note strong changes in the V_s gradient at Moho depths, particularly along P2 ($dv/v \geq +5\%$). However, no changes are observed at

469 the Moho jump in this profile, indicating that this structure is well constrained by the ANT.
 470 Finally, the crust-sediment interface, which can be roughly approximated by the 2.8 km/s
 471 contour, shows a velocity decrease below the Po plain and below the Rhone valley. Such
 472 variations along the main interfaces are likely the result of correcting for a bias from the 1-D
 473 assumption in the construction of the ANT model. Moreover, their sharpness undoubtedly
 474 demonstrates the resolving capacity of our wave-equation tomography.

475 **5.2.2 Cross-Section in the Ligurian-Provence Basin**

476 We now focus on the offshore part of our study area and discuss the P3 profile along
 477 the axis of the Ligurian-Provence basin (location in Fig. 1). This transect crosses the three
 478 main domains involved in the genesis of oceanic crust. The oceanic southwestern part of the
 479 transect coincides with the seismic refraction and wide-angle reflection line by Dannowski
 480 et al. (2020), while the northeastern part coincides with the marine part of the wide-angle
 481 reflection line by Makris et al. (1999).

482 Figure 8 shows that the final model displays major velocity changes with respect to
 483 the ANT model: (i) strong decrease in sediments and crust of the oceanic domain ($dv/v \geq$
 484 -8%); (ii) strong increase in sediments and decrease in crust of the transitional domain;
 485 (iii) decrease in sediments and crust of the southwestern margin domain, and an increase to
 486 the north-est. Such changes at shallow depths are likely related to the consideration of 3-D
 487 fluid-solid coupling. Conversely, the uppermost mantle displays lower velocity changes with
 488 an overall decrease in the oceanic domain and an increase in the transitional and continental
 489 domains. Similarly to Nouibat et al. (2022b), we define the Moho proxy as the 4.1 km/s
 490 contour which is the more consistent with the V_p Moho proxy defined as the 7.3 km/s V_p
 491 contour by Dannowski et al. (2020) (dashed yellow line). Depth and geometry of the Moho
 492 proxy are almost unchanged in the oceanic part. However, we observe a significant increase
 493 in velocity reflecting a stronger gradient at this interface, similar to Moho in W-Alps. Finally,
 494 we observe a shallower Moho in the final model at the transition to the margin.

495 **5.3 New large-scale map of Moho depth**

496 In a tectonically complex setting as in the Alps and Northwestern Mediterranean, a
 497 precise definition of the three-dimensional Moho structure is fundamental to understand
 498 the link between mantle dynamics and the geological setting.

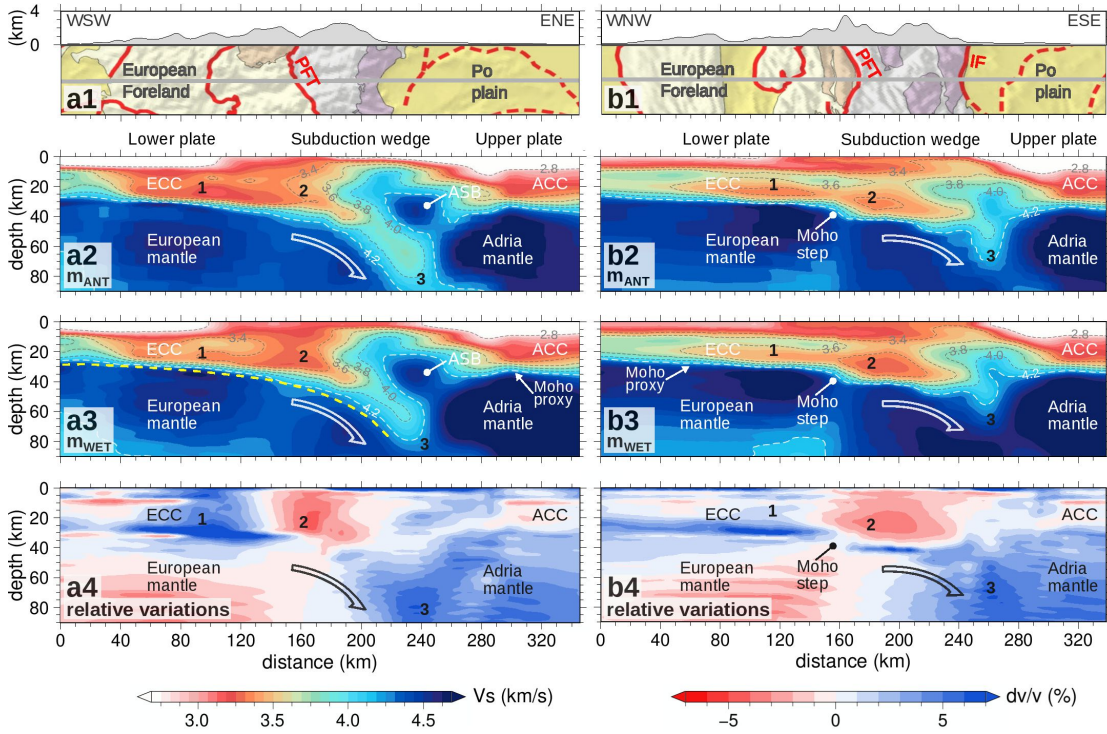


Figure 7. Comparison of depth sections for the initial and final models along the WSW-ENE transect P1 (left-hand side) and the WNW-ESE transect P2 (right-hand side). Locations of P1 and P2 are indicated in Figure 1. (a1-b1) Topographic profiles and geological maps (extracted from Fig. 1). (a2-b2) Shear-wave velocities from the ANT initial model (“ m_{ANT} ”). (a3-b3) Shear-wave velocities from the WET final model (“ m_{WET} ”). (a4-b4) Relative variations to the initial model. The white and black curved arrows highlight the subduction of the European lithosphere beneath Adria. The dashed yellow line in (a3) show the European Moho picked from a receiver functions section (Zhao et al., 2015). Note the remarkable agreement with the V_s Moho proxy in (a3). Acronyms: ECC, European continental crust; ACC, Adria continental crust; ASB, Adria seismic body (Nouibat et al., 2022a). Labels 1, 2 and 3 indicate velocity structures discussed in the text.

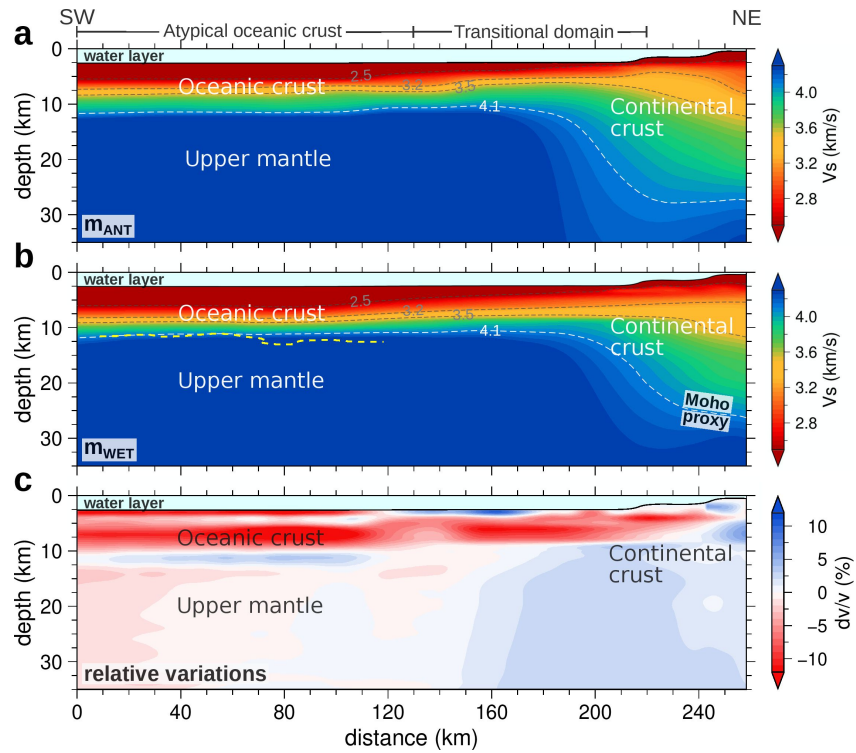


Figure 8. Comparison of depth sections for the initial and final models along the SW-NE offshore transect P3 (see location in Fig. 1). (a) Shear-wave velocities from the ANT initial model. (b) Shear-wave velocities from the WET final model. The yellow dashed line represents the Moho proxy from the active seismic P-wave velocity section of (Dannowski et al., 2020). (c) Relative variations to the initial model.

499 The Moho depth cartography in the Alpine region has been the target of a variety
500 of seismological methods, such as interpolating data from CSS, RFs and LET (Fig. 9a;
501 Waldhauser et al., 1998; Lombardi et al., 2008; Piana Agostinetti & Amato, 2009; Diehl et
502 al., 2009; Di Stefano et al., 2009; Spada et al., 2013), conventional ANT (e.g., Molinari &
503 Morelli, 2011; Kästle et al., 2018; Lu et al., 2018) and more recently from WET (Fig. 9b; Lu
504 et al., 2020). Zhao et al. (2015) showed the first evidence of continental subduction beneath
505 the western Alps obtained by analyzing receiver functions along the Cifalps-1 seismic pro-
506 file. By using simultaneously observations from receiver functions across the Cifalps-1 and
507 Cifalps-2 profiles and the V_s model by Nouibat et al. (2022a), Paul et al. (2022) documented
508 a marked change in the depth of the European Moho from north to south of the western
509 Alps, which occurs over a distance of a few tens of kilometers only.

510 In this section, we present a new 3-D Moho map at the scale of the Alps and Northwest-
511 ern Mediterranean (Fig. 9c). The Moho depth is estimated from the depth of the iso-velocity
512 surface $V_s = 4.2$ km/s. In the western Alps, our 3-D map of the Moho is overall similar to
513 the one obtained by Lu et al. (2020), with some important discrepancies (Figs. 9b-c). In
514 the southwestern Alps, while Lu et al. (2020)'s model exhibits ~ 40 -45 km depths, the Eu-
515 ropean Moho in our model deepens to depth greater than 60 km (see Fig. 7), which is more
516 consistent with the continental subduction from RF observations (Zhao et al., 2015). Using
517 the iso-velocity 4.2 km/s as a proxy, we obtain a shallower Moho (20-25 km) associated with
518 the Ivrea Body (white dashed line on Fig. 9c). The strong along-strike variations in our
519 Moho depth from north to south are in line with RF observations between the northwestern
520 and southern Alps (Paul et al., 2022), indicating a non-cylindricity of the deep structure of
521 the western Alps. Besides these differences, we observe a similar Moho topography than Lu
522 et al. (2020). In particular, we observe a Moho jump below the external crystalline massifs
523 that in our model appear to be linear and aligned with the Variscan Accident (arrows on
524 Fig. 9b-c).

525 **5.4 Moho topography beneath the Ligurian sea**

526 3-D crustal imaging of oceanic domains based on ambient-noise surface waves is a
527 challenging topic due to the difficulty of accounting for effects of the water layer and complex
528 seabed relief in a hybrid medium, as illustrated in this study. One of the main objectives
529 of this study is to produce an homogeneous 3-D map of the Ligurian Moho in line with the
530 physics of surface-wave propagation.

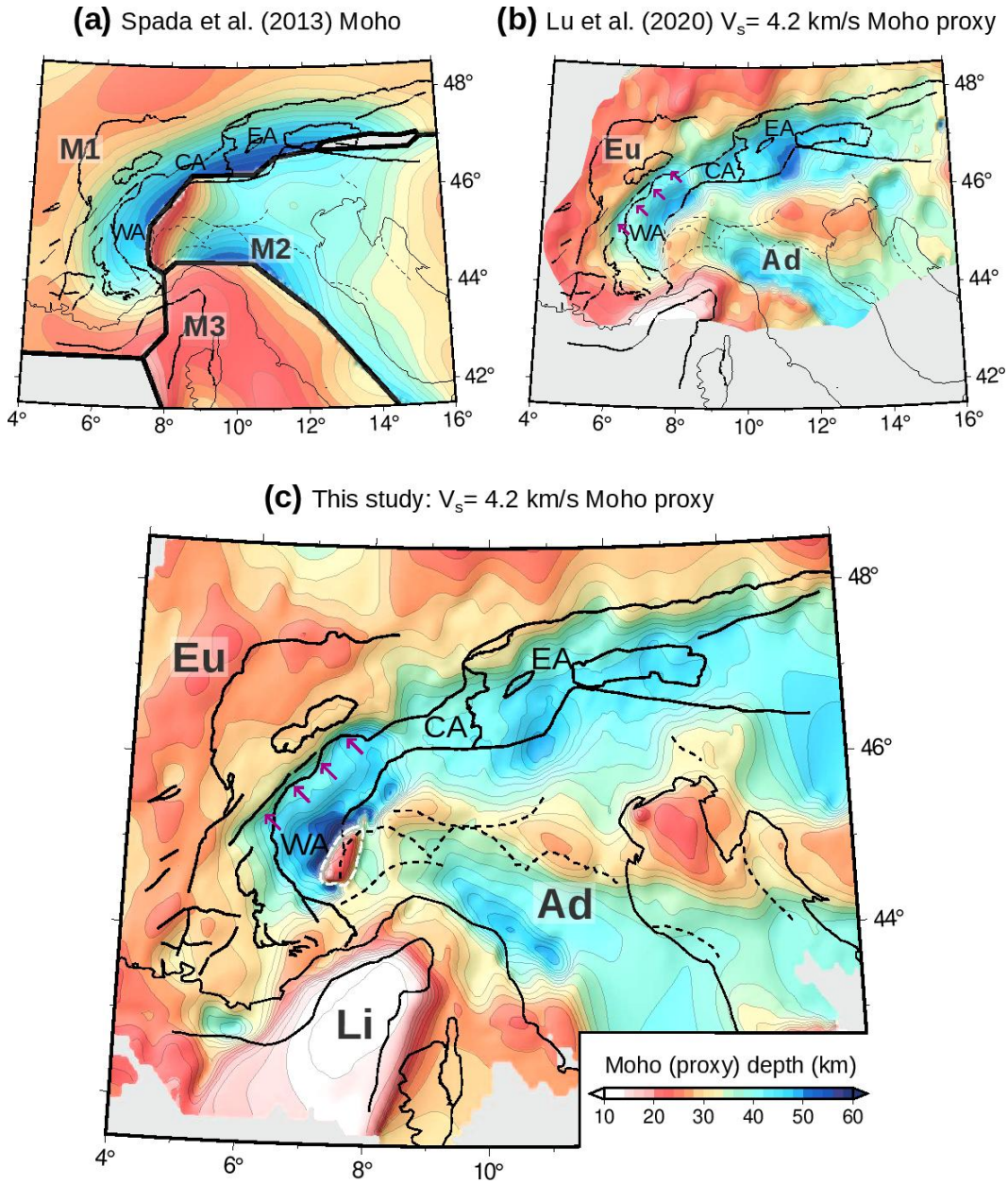


Figure 9. Depth maps of Moho and Moho proxy in the Alps and Northwestern Mediterranean. (a) Spada et al. (2013)’s map composed of three blocks: European Moho (M1), Adriatic Moho (M2) and Ligurian–Sardinia–Corsica–Tyrrhenian Moho (M3). (b) Lu et al. (2020)’s depth map of the $V_s = 4.2$ km/s iso-velocity surface from their WET model. (c) Depth map of the $V_s = 4.2$ km/s iso-velocity surface from our WET model. For a better illustration, the color scale is saturated at 60 km depth; Our model shows 70–75 km depths in the southwestern Alps (continental subduction; see Fig. 7). Acronyms: Eu, European Moho; Ad, Adriatic Moho; Li, Ligurian Moho; WA, Western Alps; CA, Central Alps; EA, Eastern Alps. Arrows in purple: Moho step beneath the external crystalline massifs (Lu et al., 2018). Dashed white line: shallow Moho associated with the Ivrea Body. The gray areas hide regions where Moho maps are unconstrained.

531 Unlike the Alpine region, the 3-D structure of Moho in the Ligurian sea is poorly
 532 constrained by 3-D velocity models. In this region, the few constraints on the Moho depth
 533 are mainly provided by CSS data (e.g., Makris et al., 1999; Rollet et al., 2002; Contrucci
 534 et al., 2001; Dannowski et al., 2020). This gap is partly due to the poor coverage of
 535 seismological stations in this area before the deployment of AlpArray OBSs, making it
 536 difficult to achieve 3-D high-resolution imaging with tomography methods such as ANT or
 537 LET.

538 Figure 10 shows the Moho depth map in the Ligurian-Provence basin estimated from
 539 the iso-velocity surface $V_s=4.1$ km/s. The Moho map shows strong lateral variations with
 540 depths, ranging from 12 km in the axis of the basin to ~ 20 -25 km in the conjugate margins
 541 of Provence and Corsica. The transition to the Ligurian-Provence and Corsica margins
 542 is characterised by a strong gradient in Moho depth while the transition to the Ligurian
 543 margin is smoother. In the oceanic domain, the Moho gradually deepens along the axis of
 544 the basin, from 12 km in the northern and central parts to ~ 14 -18 km in the southern part.
 545 Red dashed show depth contours of a recently published Moho surface from conventional
 546 ANT (Magrini et al., 2022). This map shows smoother depth variations than ours within
 547 the basin and at transitions to the conjugate margins. Along the basin axis, it shows depths
 548 increase from the southern part (15 km) to the northern part (18 km) of the basin. While
 549 our Moho has a depth of about 12 km in the oceanic domain traversed by the active seismic
 550 profile of Dannowski et al. (2020), the Magrini et al. (2022)'s Moho is 15-18 km deep in the
 551 northern portion of the profile and more than 18 km deep in the southern portion. However,
 552 Dannowski et al. (2020) detected a Moho of ~ 12 km depth, which is more consistent with the
 553 depths displayed by our model (see Fig. 8). These comparisons demonstrate that our Moho
 554 model provides a better resolution of the crustal thinning beneath the Ligurian-Provence
 555 basin.

556 **6 Model robustness**

557 In this section, we will document the robustness of our resulting velocity model. First,
 558 we analyse the distribution of phase traveltime misfits of surface waves from the ambient-
 559 noise database used in this study. Then, we validate our model with earthquake seismic
 560 waves in the western Alps, (1) by comparing P-, Rayleigh-wave observed waveforms with
 561 synthetic waveforms computed in our model, and (2) by analysing the distribution of first-

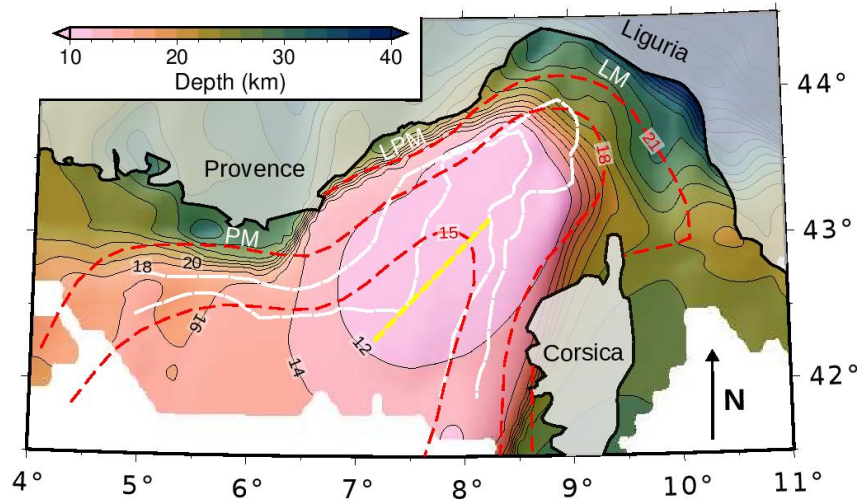


Figure 10. Depth map of the $V_s = 4.1$ km/s iso-velocity surface from our WET model as a proxy of Moho in Northwestern Mediterranean. Black lines are iso-depth contours from our Moho proxy. Red lines: iso-depth contours from Magrini et al. (2022). Dashed yellow line shows the trace of Dannowski et al. (2020)’s CSS profile. Dashed white lines are limits of the main geological domains in the Ligurian-Provençe basin from Rollet et al. (2002). Acronyms: PM, Provençe margin; LPM, Ligurian-Provençe margin; LM, Ligurian margin.

562 arrival traveltimes residuals between picked P, S arrivals and synthetic arrivals predicted in
 563 our model.

564 6.1 Misfit Analysis

565 We first document the model robustness by analyzing histograms of misfit for the initial
 566 and final model and their spatial and azimuthal distributions. The misfit is expressed as
 567 traveltimes delay for waves propagating on 100-km distance (s/100km) in order to avoid
 568 biases related to long inter-station distances. Figure 11 shows histograms of traveltimes
 569 misfits for the initial (in red) and final (in blue) models at 8, 20, 35 and 35 s period. The
 570 mean and standard deviation of the misfit distribution allow a quantitative comparison. The
 571 initial histograms are rather narrow and globally centred around zero misfit value (absolute
 572 misfit ≤ 0.1 s/100km). The initial standard deviation increases with period (0.3 s/100km at
 573 8 s to 0.73 s/100km at 55 s). We do not observe a significant shift of the mean value, which
 574 illustrates the consistency of our ANT model. Since our initial model does not suffer from
 575 biases towards wrong mean-value velocities, our inversion further contributes in refining the

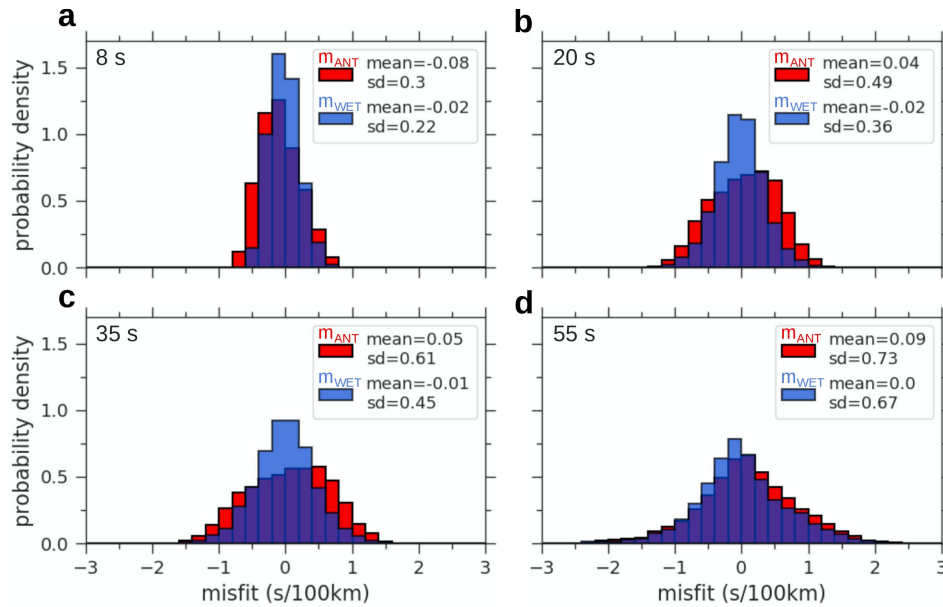


Figure 11. Comparison of histograms of misfit (in s/100km inter-station distance) for the initial (m_{ANT} , red) and final (m_{WET} , blue) velocity models at 8, 20, 35 and 55 s periods. Labels "mean" and "sd" refer respectively to the mean value and standard deviation of the misfit distribution.

576 intrinsic shape and magnitude of velocity structures rather than the overall mean velocity
 577 at depth. The final model fits the data significantly better than the initial model, with
 578 overall smaller average misfits and standard deviations. Histograms of cross-correlation
 579 type traveltimes differences are shown in supplementary Figure S4.

580 Figure 12 shows misfit maps at 8, 20, 35 and 55 s periods, obtained by averaging
 581 misfit values for the initial and final models over $0.2^\circ \times 0.2^\circ$ cells assuming great-circle
 582 ray paths. The spatial distribution of the misfit allows to see whether some part of the
 583 model are better constrained than others. Positive misfit values mean that the velocity is
 584 underestimated, while negative values mean that the velocity is overestimated. In the initial
 585 model (Figs. 12a1-d1), the misfit patterns tend to coincide with geological structures due to
 586 imperfections in the isotropic initial velocity model. The 8-s map (Fig. 12a1) exhibits rather
 587 small misfit values along the central and eastern Alps, west Po basin, southeastern French
 588 foreland and the central Ligurian-Provence basin. Stronger negative values are observed
 589 in the northeast French foreland, north and south Apennines, and in the northern and
 590 western parts of the Ligurian-Provence basin. Besides this, we observe a strong positive
 591 anomaly in the Adriatic sea. The 20-s and 35-s misfit maps (Figs. 12b1-c1) evidence positive

592 misfits all around the Alpine belt, while negative misfits are observed in the Alpine arc
 593 and west Po basin. Note that the western boundary of the Alps has rather low misfit
 594 values (absolute misfit < 0.4 s/100km) which indicates that the 1-D inversion in the ANT
 595 model is sufficiently robust to recover such structural changes at depth. The transition
 596 to 55 s is marked by an inversion of polarity: the European foreland and central-western
 597 Ligurian-Provence basin evidence negative anomalies, while the central and eastern Alps and
 598 Apennines display positive anomalies. Strong negative anomalies are observed in the western
 599 Ligurian-Provence basin and central Adriatic sea. Most of these regionally organized misfit
 600 patterns are corrected after WET inversion (Figs. 12a2-d2). Although the strong misfits in
 601 the south of the study region are significantly smaller, some residuals remain, in particular
 602 along boundaries of the study region where path coverage is poorer (e.g., southern Corsica,
 603 in the 20-s and 35-s maps, or in the western Ligurian-Provence basin at 55 s). The azimuthal
 604 distribution of misfit for the initial and final models is shown in supplementary text S2 and
 605 Figure S5.

606 **6.2 Validation with earthquake waveform modeling**

607 We validate our final (V_p , V_s) model by performing 3-D wave propagation simulations of
 608 earthquakes in Western Alps. The synthetic waveforms are simulated up to 1 Hz using the
 609 same spectral-element-based solver described in section 4. Since the absolute waveform am-
 610 plitude can be affected by the attenuation quality factors (Q_p , Q_s) at such high frequencies,
 611 and we are interested in comparing both amplitudes and phases between observed and simu-
 612 lated waveforms, we consider here a visco-elastic wave propagation. The model is discretized
 613 with an adaptive mesh built in the Cartesian system. We parametrize the GLL points with
 614 isotropic V_p , V_s from our model, and isotropic ρ , Q_p , Q_s computed from these velocities by
 615 empirical formulas (Brocher, 2005). We represent the earthquakes as a moment source, us-
 616 ing moment tensor solutions obtained by Petersen et al. (2021) from centroid moment tensor
 617 (CMT) inversion using the AlpArray network. We perform simulations using a Gaussian
 618 source-time function with half duration determined by the seismic moment (Komatitsch &
 619 Tromp, 2002). The seismograms are filtered in three narrow period bands (2-5 s, 3-7 s and
 620 5-10 s) and the time windows are defined as $[t_{max} - T, t_{max} + T]$ for the P wave-packet and
 621 $[t_{max} - 1.5T, t_{max} + 1.5T]$ for the Rayleigh wave-packet, where T is the maximum period of
 622 measurement and t_{max} is the maximum envelope arrival time. t_{max} is detected in velocity
 623 range for P waves (5.0-8.0 km/s) and for Rayleigh waves (1.5-3.5 km/s). Finally, we mea-

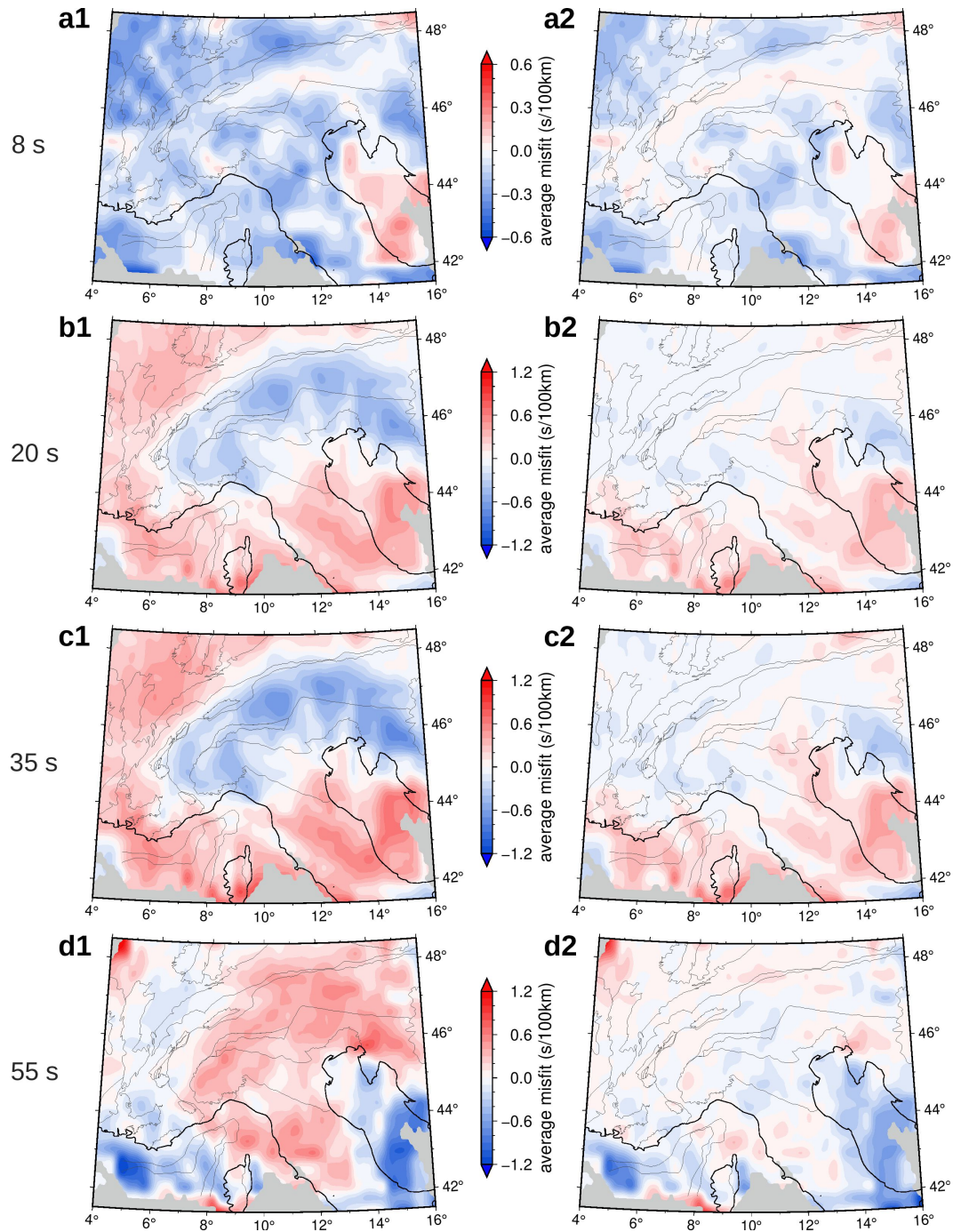


Figure 12. Comparison of the spatial distributions of misfit for the initial (left-hand side: a1-d1) and final (right-hand side: a2-d2) velocity models at 8, 20, 35 and 55 s periods. Misfits are averaged over cells of $0.2^\circ \times 0.2^\circ$ assuming great circle ray paths. The gray area hides regions where the initial model is unconstrained.

624 sure the cross-correlation-based time shifts to quantify the mismatch between the synthetic
625 waveforms and observations.

626 Figure 13 shows an example of a comparison between synthetic and observed waveforms
627 for an earthquake of magnitude $M_w = 4$ and 5-km depth that occurred in North Western
628 Alps in 2017 (Fig. 13a for location). The seismograms are recorded by 6 stations from
629 permanent networks (FR, France; CH, Switzerland; GU, Italy) and the temporary AlpArray
630 network (Z3). We observe a striking fit between synthetic and observed waveforms for both
631 P and Rayleigh waves, as supported by an overall time delay lower than 0.65 s (Fig. 13b-c-d).
632 As we expect, the Rayleigh waves are quite well fitted in the 5-10 s period band (Fig. 13d),
633 which is the shorter period band used in our inversion. Besides that, the striking fit in
634 the 3-7 s period (Fig. 13c) highlights the potential of our model to recover the shallower
635 and lateral small-scale structures. The good coherency in phase and amplitude for the
636 P waves in the 2-5 s and 3-7 s period bands (Fig. 13b-c) highlights the potential of our
637 model to recover the small-scale intra-crustal structures. This example illustrates that our
638 WET model could be a relevant starting model for wave-equation tomography of regional
639 earthquake body and surface waves – a natural extension of the wave-equation tomography
640 of ambient noise correlations. A more comprehensive validation of this model will be the
641 subject of a future publication.

642 **6.3 Validation with first-arrival traveltimes of local earthquakes**

643 We further document the robustness of our final velocity model by comparing observed
644 and synthetic body-wave traveltimes associated with more than 75×10^3 earthquakes in the
645 region of Western Alps, over a period of more than 30 years (Fig. 12a). We use $\sim 1.6 \times 10^6$
646 picked times (93×10^4 *P*-wave picks and 66×10^4 *S*-wave picks) coming from the dataset of
647 Potin (2016) for the period between 1989 and 2014, complemented by picked times available
648 from the European plate observing system (EPOS) for the period 2015-2021.

649 The velocity models are discretized according to a regular Cartesian grid of 5 km cell
650 size in horizontal directions and 2 km in the vertical direction. Hypocenters are relocated
651 in our velocity models using the NLLOC software (Lomax et al., 2000). Initial first-arrival
652 traveltimes are predicted in our 3-D V_s and V_p final models through an efficient eikonal solver
653 (Podvin & Lecomte, 1991) on a finer interpolated square grid of 2 km. Using receivers as
654 eikonal sources saves computer resources, and, therefore, back-tracing rays between each

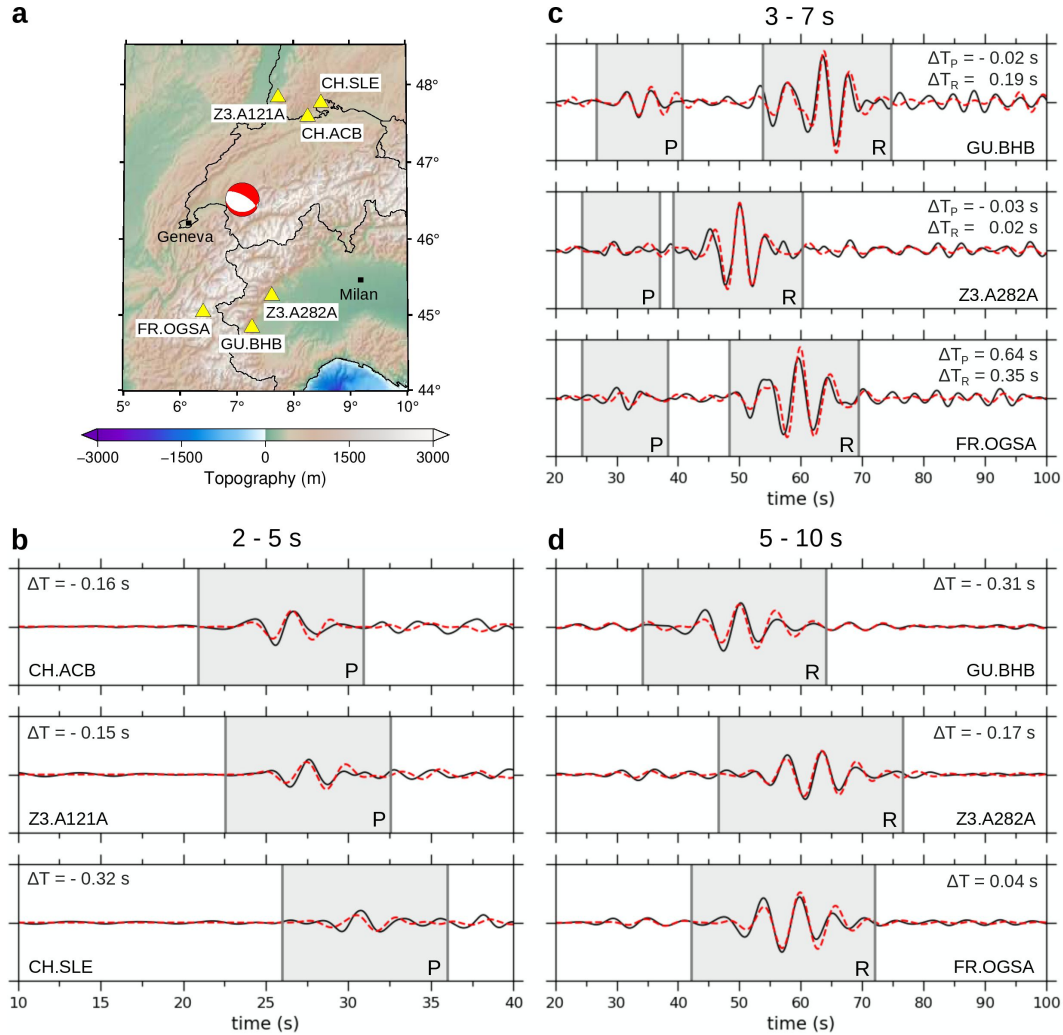


Figure 13. Comparisons between simulated vertical-component seismograms (red) in the WET model and observed waveforms (black), filtered in the 2-5 s, 3-7 s and 5-10 s period bands for the earthquake and receivers in (a) (red: earthquake, yellow: receiver). The gray areas indicate the *P*-wave (*P*) and Rayleigh-wave (*R*) windows and ΔT corresponds to the cross-correlation-based time shift between observed and synthetic windowed waveforms (ΔT_P and ΔT_R respectively for *P*-, Rayleigh -wave time shifts).

655 couple (source, receiver) are obtained. Finally, traveltimes are estimated along these rays
 656 with a ray sampling of 0.5 km in order to achieve higher-computational accuracy. Using the
 657 same procedure, we calculate traveltimes in a 3-D model obtained from a local earthquake
 658 tomography using the same dataset (Virieux et al., 2023).

659 The P and S residuals from our model (blue histograms in Fig. 14b-c) show normal dis-
 660 tribution shapes that are rather narrow (standard deviation < 1 s/100km) and remarkably
 661 well centred around zero misfit (mean < 0.15 s/100km). The standard deviation of the S
 662 residuals is slightly larger, but it should be noted that the total number of P picks is almost
 663 1.4 times larger. However, the mean value of the S distribution is smaller. Such small mean
 664 values indicate that our model does not suffer from biases towards wrong mean-value ve-
 665 locities. These histograms obtained from a completely independent dataset are comparable
 666 to those in Figure 11 based on surface waves from ambient noise. To further assess the
 667 relevance of our model, we compare with residuals from the LET model (red histograms in
 668 Fig. 14b-c). The distribution of residuals from our model is notably narrower than that of
 669 the LET model. The overall good agreement between observed first-arrival traveltimes and
 670 those predicted in our model indicates that it is an accurate enough initial model for 3-D
 671 local earthquake tomography. Since it has been obtained using ambient-noise surface waves,
 672 this model has the advantage to constrain the lower crust that is often poorly sampled by
 673 earthquakes.

674 **6.4 Resolution tests**

675 We perform 3-D spike tests to evaluate the resolution of the final V_s model. Three
 676 spikes of ~ 25 km radius and ~ 20 km thickness are located in the crust, beneath Western
 677 Alps, Central Alps and northern Ligurian-Provence basin (Fig. 15a1). A spike of ~ 50 km
 678 radius and ~ 40 km thickness is located in the uppermost mantle beneath the northwestern
 679 Po basin (Fig. 15a2). These patterns correspond to velocity perturbations of up to ~ 8 %
 680 in the crust and ~ 5 % in the uppermost mantle, with respect to the initial model. Using
 681 the same station-pair coverage as our observations, we apply the methodology described in
 682 Section 3. Inversion results are shown in the right-hand panel of Figure 15.

683 At 20 km depth, the crustal anomalies are well recovered. Overall, the shape of the
 684 anomalies appears to be slightly stretched. At 50 km depth, the perturbation is recovered
 685 but its shape is affected by a strong eastward horizontal smearing. The peak amplitude is

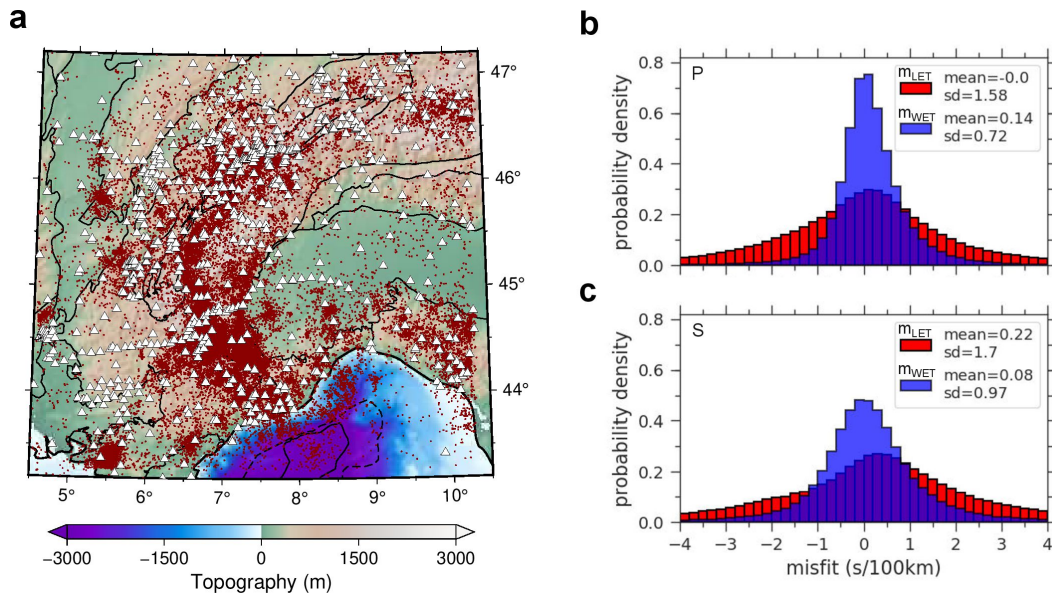


Figure 14. (a) Location map of the seismic events (red dots) and stations (white triangles) constituting the database of first-arrival traveltimes described in section 6.3. (b,c) Histograms of the traveltime residuals for the P waves (b) and the S waves (c) between observations and predicted synthetic arrivals from our velocity model (m_{WET} , blue) and from the LET model (m_{LET} , red). Labels "mean" and "sd" refer respectively to the mean value and standard deviation of the misfit distribution.

686 weaker as for the crustal anomalies. In both cases, the recovered peak is approximately in the
687 right position. The cross-sections (Figs. 15a3-b3) show that most of the smearing occurs in
688 the horizontal direction. The discrepancies between input and recovered patterns are mainly
689 due to path coverage, to the propagation of Rayleigh wave which induces lateral smoothing
690 of heterogeneities in the crust and upper mantle, and to its sensitivity to the medium which
691 decreases with depth. Nevertheless, part of these discrepancies is also attributable to the
692 tomographic method itself and the convergence rate.

693 Though these spike tests do not stand for the entire study area, they indicate that our
694 model is fairly resolved in the crust and can be used for the geological interpretation of
695 the three-dimensional geometry of crustal structures and for their petrophysical character-
696 ization. While the resolution in upper mantle remains tolerable for a first-order geological
697 interpretation, it is obviously not sufficient for precise petrophysical characterization.

698 **7 Conclusion**

699 Using ambient noise data from exhaustive coverage of permanent and temporary arrays
700 in the western European region, we derive a 3-D onshore-offshore velocity model covering the
701 Alps and Ligurian-Provence basin, using wave-equation tomography (WET). We iteratively
702 refine the recent ANT model of Nouibat et al. (2022a) by minimizing the phase traveltime
703 differences between observed and simulated Rayleigh-wave waveforms in the 5-85 s period
704 band. Observed signals are obtained from ambient noise cross-correlations and synthetics are
705 computed from SEM-based 3-D elastic and acoustic-elastic modeling of surface waves. The
706 specificity of this study is to highlight the effect of the water layer on the 3-D propagation of
707 Rayleigh waves by applying a fluid-solid coupling for 3-D acoustic-elastic simulations, taking
708 the Ligurian sea domain as an illustration. We demonstrate that the elastic propagation
709 assumption is no longer valid at short periods (5-20 s), since the surface-wave packet is
710 dominated by a composite Rayleigh-Scholte mode propagating with lower velocities. Finally,
711 we incorporate the fluid-solid coupling in the inversion framework of the WET.

712 In line with the true physics of surface-wave propagation, the WET corrects for the
713 biases in the ANT model related to the high-frequency assumption and the 1-D inversion.
714 The resulting model has a better resolution, with significant intra-crustal changes. In the
715 superficial part, the WET better emphasizes the sediments of the Ligurian-Provence domain
716 and a high-velocity anomaly beneath the Variscan part of Corsica. This improvement is

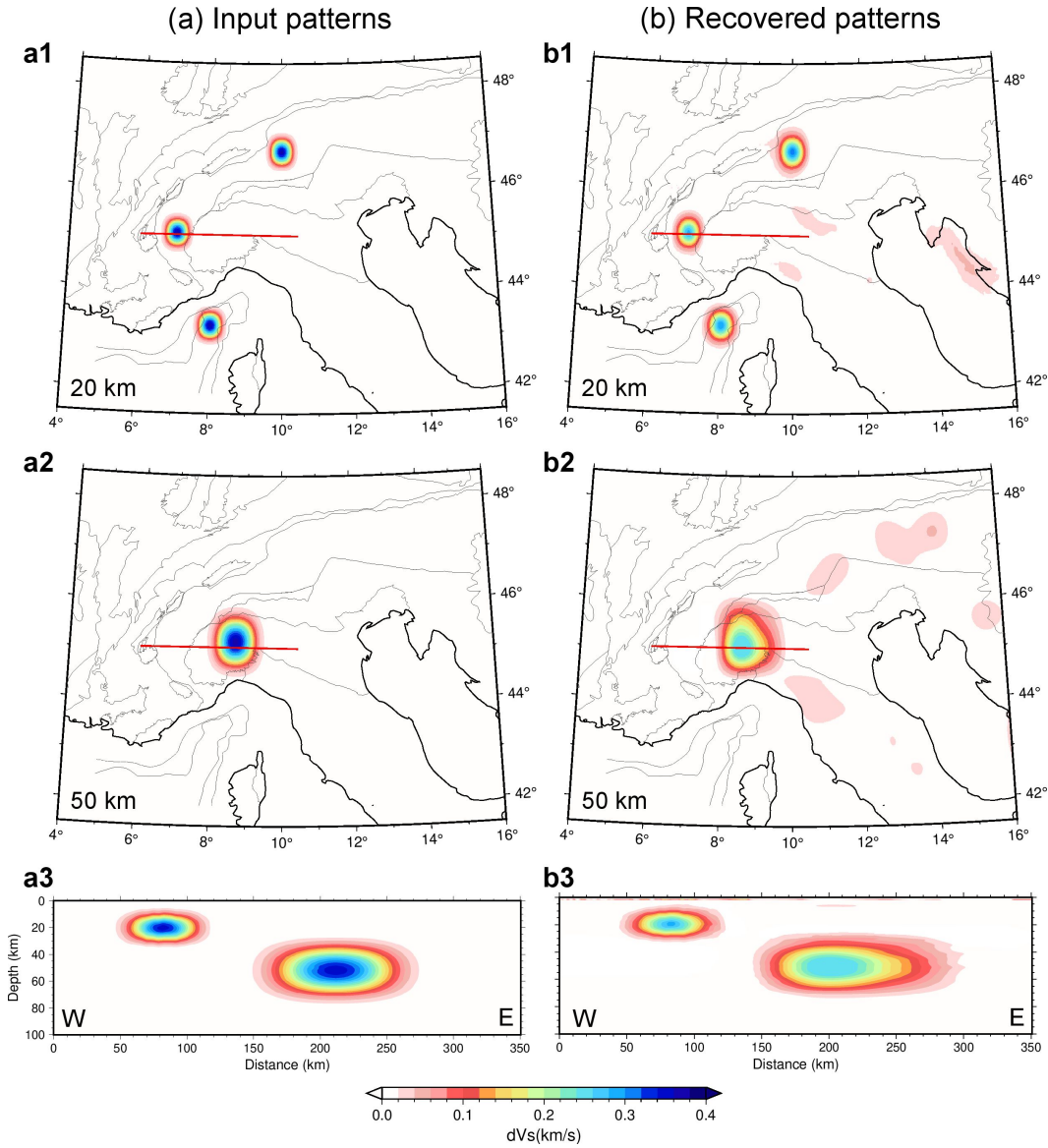


Figure 15. Resolution assessment of the WET V_s model using 3-D spike tests. Left: input perturbations with respect to the initial model (m_{ANT}), right: recovered perturbations. (a1-b1) 20 km depth slices. (a2-b2) 50 km depth slices. (a3-b3) E-W cross-section (red profile in the depth slices).

717 partly due to accounting for Rayleigh-Scholte wave 3-D sensitivity kernels at short periods.
718 In the crust, the WET mainly change the velocity contrast with an overall velocity decrease
719 in the subduction wedge, focusing the LVZ in the W-Alps to the northeast, and an overall
720 increase in the two forelands. In addition, the WET validates major structures already
721 present in the initial model, such as the subduction of the European crust beneath Adria in
722 the southwestern Alps, and the slow anomalies at the base of the crust in the northwestern
723 Alps, thus validating recent first-order interpretations on the deep structure in this area (e.g.,
724 Nouibat et al., 2022a; Paul et al., 2022). Significant changes are observed in the oceanic
725 crust of the Ligurian-Provence domain, where intra-crustal velocities decrease significantly
726 along the axis of the basin.

727 We present a new depth map of the Moho proxy of Western Europe. In the western Alps,
728 this model shows the deepening of the European crust and confirms the Moho jump under
729 the outer crystalline massifs (Lu et al., 2020) and the high-velocity anomalies associated
730 with the Ivrea Body (Nouibat et al., 2022a). This proxy is the first truly 3-D representation
731 of the land-sea Moho in the Alps and Northwestern Mediterranean. We show a strong
732 deepening of the Moho towards the Ligurian-Provence and Corsica conjugate margins (from
733 12 to 20-25 km), and a smoother deepening towards the Ligurian margin and the southern
734 part of the basin (from 12 to 14-18 km).

735 We validated our tomographic model with independent data, through 3-D waveform
736 modeling and first-arrival traveltimes of seismic waves from earthquakes in the region of
737 Western Alps. In a future perspective, this model can be used as a starting model for 3-D
738 imaging based on earthquake data (local earthquake tomography, wave-equation tomogra-
739 phy or full waveform inversion) to derive a higher resolution V_p and V_s models of Western
740 Alps and Ligurian sea.

741 **Open Research**

742 Waveform data used in this paper are available through the EIDA (European Integrated
743 Data Archive) service of ORFEUS (<http://www.orfeus-eu.org/eida/>) and belong to the per-
744 manent networks with codes AC, BE (Royal Observatory Of Belgium, 1985), CA (Institut
745 Cartogràfic I Geològic De Catalunya - Institut D'Estudis Catalans, 1984), CH (Swiss Seis-
746 mological Service (SED) At ETH Zurich, 1983), CR (University Of Zagreb, 2001), CZ
747 (Institute Of Geophysics Of The Academy Of Sciences Of The Czech Republic, 1973), ES

748 (Instituto Geografico Nacional, Spain, 1999), FR (RESIF, 1995), G (Institut de Physique
 749 du Globe de Paris (IPGP) & Ecole et Observatoire des Sciences de la Terre de Strasbourg
 750 (EOST), 1982), GB (GEOFON Data Centre, 1993), GR (Federal Institute for Geosciences &
 751 Natural Resources (BGR), 1976), IV (INGV Seismological Data Centre, 2006), NL (KNMI,
 752 1993), OE (ZAMG-Zentralanstalt Für Meteorologie Und Geodynamik, 1987), SL (Slovenian
 753 Environment Agency, 2001), and UP (SNSN, 1904). We also used data of the temporary Al-
 754 pArray network (network code Z3, AlpArray Seismic Network, 2015), Cifalps-2 experiments
 755 (network code XT, Zhao et al., 2018) and EASI experiments (network code XT, AlpArray
 756 Seismic Network, 2014). Our tomographic 3-D model will be available in the IRIS EMC -
 757 Reference Earth Models repository (<http://ds.iris.edu/ds/products/emc-earthmodels/>).

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786 **Appendix A**

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