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Ambient-noise tomography of the Ligurian-Provence Basin using the AlpArray onshore-offshore network: Insights for the oceanic domain structure

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Key Points:

- Efficient processing scheme to remove transients and reduce tilt and compliance from continuous ambient noise recorded by OBSs
- Computation of iterative correlations between OBSs based on a virtual reconstruction of the Rayleigh waves
- Thin, anomalous oceanic crust with gabbroic intrusions evidenced in the basin axis from a joint interpretation of V_s and V_p models

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Abstract

We derive a three-dimensional shear-wave velocity model of the Ligurian-Provence back-arc basin (Northwestern Mediterranean Sea) using ocean-bottom seismometers (AlpArray OBSs) and land stations from permanent and temporary seismic networks. The quality of OBS continuous records is enhanced by a specific processing that reduces instrumental and seabed-induced noises (transients, tilt, compliance). To further improve the resolution of ambient-noise tomography in the offshore area, we compute the Rayleigh-wave part of the Green functions for OBS-OBS pairs by using onshore stations as virtual sources. 2-D group-velocity maps and their uncertainties are computed in the 4 - 150 s period range by a transdimensional inversion of Rayleigh-wave travel times. The dispersion data and their uncertainties are inverted for a probabilistic 3-D shear-wave velocity model that includes probability densities for V_s and for the depth of layer interfaces. The probabilistic model is refined by a linearized inversion that accounts for the water layer in the Ligurian Sea. Our S -wave velocity and layer boundary probability models correspond well to a recent, high-resolution P -wave velocity cross-section derived from controlled-source seismic profiling along the Ligurian-Provence basin axis. A joint interpretation of the P - and S -wave velocity sections along this profile reveals a thin, anomalous oceanic crust of low P -wave velocities but high S -wave velocities, intruded by a gabbroic body. The illuminated part of the upper mantle appears to be devoid of serpentinization.

Plain Language Summary

The Ligurian-Provence basin (Northwestern Mediterranean Sea) is one of the Miocene-Pliocene back-arc basins that resulted from the retreat of the Adria subduction in the plate reorganization due to Africa-Europe convergence. The crustal structure of the basin is still debated, even though it has been probed by active seismic profiling. We compute a high-resolution shear-wave velocity model of the Ligurian-Provence basin and its margins by making optimal use of ambient-noise recordings of seafloor broadband seismometers. In particular, we improve the usually low quality of surface-wave signals in noise correlations between seafloor stations by involving correlations with land stations. The joint interpretation of our S -wave velocity model with a P -wave velocity section obtained in the basin axis by controlled-source seismic profiling provides compelling evidence for the presence of a thick sediment pile above a thin, ~ 4.5 km-thick oceanic crust, intruded by gabbroic bodies emplaced at the crust-mantle transition. These results show the potential of a joint

48 interpretation of P - and S -wave velocity models since they provide reliable answers to a
49 number of debated questions on the petrological nature of the crust and uppermost mantle,
50 in particular in the Ligurian-Provence basin.

51 1 Introduction

52 In the last two decades, the deployment of extensive and dense seismic networks of tem-
53 porary broadband sensors (e.g., USArray, IberArray) has provided a better understanding
54 of the deep structures of the crust and upper mantle, in particular through the emergence
55 of increasingly more precise 3-D seismic imaging (e.g., Moschetti et al., 2010; Levander et
56 al., 2011). However, many of these dense arrays included exclusively onshore sensors, thus
57 preventing 3-D imaging of continent-ocean transitions and oceanic domains. The AlpArray
58 seismic network (AASN) that covers the European Alps and their foreland, is one of the few
59 dense seismic networks consisting of both onshore and offshore stations (Fig. 1; Hetényi et al.,
60 2018). The onshore AASN has been used in ambient-noise imaging studies at the scale of the
61 Alps and Apennines to construct 3-D models of shear-wave velocity (Lu et al., 2018, 2020),
62 attenuation (coda-Q, Soergel et al., 2020), and radial anisotropy (Alder et al., 2021), as well
63 as at the regional scale, e.g., Vienna basin (Schippkus et al., 2018), Western Alps (Zhao
64 et al., 2020), Southeastern Alps (Sadeghi-Bagherabadi et al., 2021) and Bohemian Massif
65 (Kvapil et al., 2021). The German-French ocean-bottom seismometers (OBS) of the AlpAr-
66 ray network have been deployed to gain insights into the 3-D structure of the lithosphere
67 beneath the Ligurian-Provence basin and its margins, where major geological/geodynamical
68 issues remain to be clarified. The OBS recordings have been used together with data of on-
69 shore permanent and temporary stations in the transdimensional ambient-noise tomography
70 of Nouibat et al. (2022) that covers a large part of Western Europe. In Nouibat et al. (2022),
71 we described the probabilistic inversion strategy for Rayleigh-wave group-velocity maps and
72 their uncertainties, the injection of these uncertainties in the inversion for the V_s model,
73 and the validity of the model in the southwestern Alps. The present paper is a complement
74 to Nouibat et al. (2022) that focuses on the Ligurian-Provence basin. We describe here
75 the specific processing of OBS records and the use of iterative noise correlations that are
76 required to improve ray coverage in the Ligurian Sea, hence improving the model resolution
77 in the basin. We further explain how the water layer and its thickness changes are taken
78 into account in the inversion for S -wave velocity. Finally, we compare our V_s model to a V_p
79 section derived from controlled-source seismic profiling along the basin axis.

80 The geodynamic context of the western Mediterranean region is controlled by the
81 northward motion of Africa (1 cm/yr) with respect to Europe since Late Cretaceous times
82 (Fig. 1). This global convergence was accommodated by several collision episodes involving
83 Europe with continental micro-plates (Iberia and Adria), leading to the formation of peri-
84 mediterranean mountain belts (Alps, Apennines, Pyrenees, Dinarides and Betics; e.g., van
85 Hinsbergen et al., 2020). During Miocene and Pliocene times, part of this convergence was
86 accommodated by development of back-arc extensional basins behind the Adria northwest-
87 dipping subduction zone (e.g., Gueguen et al., 1998; Jolivet et al., 2020). This extension
88 started in the Ligurian-Provence basin, and has further spread from west to east, resulting
89 in the opening of the Algerian basin, and later, of the Tyrrhenian basin (e.g., Séranne,
90 1999; Rollet et al., 2002). Crustal thinning in the central Ligurian basin resulted in the
91 formation of a narrow oceanic domain mainly identified from geophysical data, including
92 seismic reflection, refraction and wide-angle profiling (e.g., Egger et al., 1988; Déverchère &
93 Beslier, 1995; Dannowski et al., 2020), altimetry data and magnetic data from aeromagnetic
94 surveys and reduction to the pole (Sandwell et al., 1995). Although the crust of the central
95 Ligurian basin is considered "atypical" because it is thinner than normal oceanic crust and
96 highlights non-linear magnetic anomalies and a concomitant low gravity anomaly (Bayer
97 et al., 1973; Sandwell & Smith, 1997; Rollet et al., 2002), its petrological and lithological
98 nature is poorly constrained and still debated. Several hypotheses have been considered: (1)
99 thin oceanic crust with tholeiitic volcanism overlying mantle rocks, similar to the Tyrrhenian
100 sea (e.g., Mascle & Rehault, 1990; Bonatti et al., 1990); (2) partly serpentinized peridotites
101 of exhumed upper mantle mostly devoid of volcanic crust (e.g., Boillot et al., 1989; Beslier
102 et al., 1993; Jolivet et al., 2020); (3) thinned and stretched continental crust related to an
103 hyper-extended margin (e.g., McKenzie, 1978; Pascal et al., 1993; Dannowski et al., 2020).

104 We combine the OBS records of the AlpArray network with those of 890 onshore stations
105 from the AlpArray temporary network, the Cifalps-2 temporary experiment and European
106 permanent networks (Fig. 1). Ambient-noise data recorded by OBSs have already been
107 used to build shear-wave velocity and anisotropy models in different regions, e.g., South-
108 central Pacific (Harmon et al., 2007), Southeast of Tahiti Island (Takeo et al., 2016) and
109 Western Indian Ocean (Hable et al., 2019). The originality of our approach lies in the
110 specific processing of OBS records and their use within an innovative tomographic framework
111 based on data-driven Bayesian inversions, which has recently been successfully applied to
112 image the lithosphere at the scale of Western Europe (Nouibat et al., 2022). Specific and

113 careful pre-processing of OBS records is compulsory because they are affected by noise
114 sources at the seabed such as compliance and tilting, and they may also be impacted by
115 intrinsic instrumental noise (Crawford et al., 1998; Crawford & Webb, 2000; Deen et al.,
116 2017). Furthermore, OBSs are sensitive to local noise sources such as tides and currents,
117 boat traffic, or marine animals, which are not recorded coherently over long distances (e.g.,
118 Batsi et al., 2019). Such noises are therefore unsuitable for ambient-noise tomography, and
119 they even alter the signal-to-noise ratio of surface waves reconstructed by noise correlation.
120 Finally, the water column above the seismometers (water depth of 1100 - 2800 m in our case)
121 may have a significant impact on the quality of the Rayleigh-wave dispersion measurements
122 between distant OBSs, particularly at periods shorter than 15 s. Wolf et al. (2021) have
123 highlighted the difficulties in using the AASN OBSs to measure Rayleigh waves dispersion
124 curves from noise correlations. These difficulties are partly related to the high level of sea-
125 floor noises. To overcome these difficulties, we propose an innovative way of computing
126 seismic noise correlations using OBS data that consist (1) in a pre-processing of the OBSs
127 noise records that decrease efficiently the seabed noise, and (2) in computing iteratively
128 noise correlations between OBS stations using onshore stations as virtual sources.

129 The overall methodology and its results are presented in sections 2 to 4. Section 2 is
130 dedicated to the description of a specific pre-processing that aims at cleaning OBS daily
131 noise records from instrumental transient glitches and seafloor noises. Section 3 presents
132 how iterative noise correlations are computed between OBSs using onshore stations as vir-
133 tual sources. In section 4, we show how noise correlations are used to build a 3-D S -wave
134 velocity model of the Ligurian-Provence basin by computing 2-D probabilistic transdimen-
135 sional Rayleigh wave group velocity maps and their uncertainties at different periods, and
136 by inverting local dispersion curves to derive a probabilistic V_s model. The fifth and final
137 section is dedicated firstly to a validation of the resulting 3-D V_s model through a compar-
138 ison with a recent V_p model obtained along a linear refraction, wide-angle seismic profile in
139 the center of the basin by Dannowski et al. (2020). Finally, we show how the combination
140 of the V_p and V_s models along the same profile provides insightful clues to the structure and
141 nature of the crust in the central Ligurian-Provence basin.

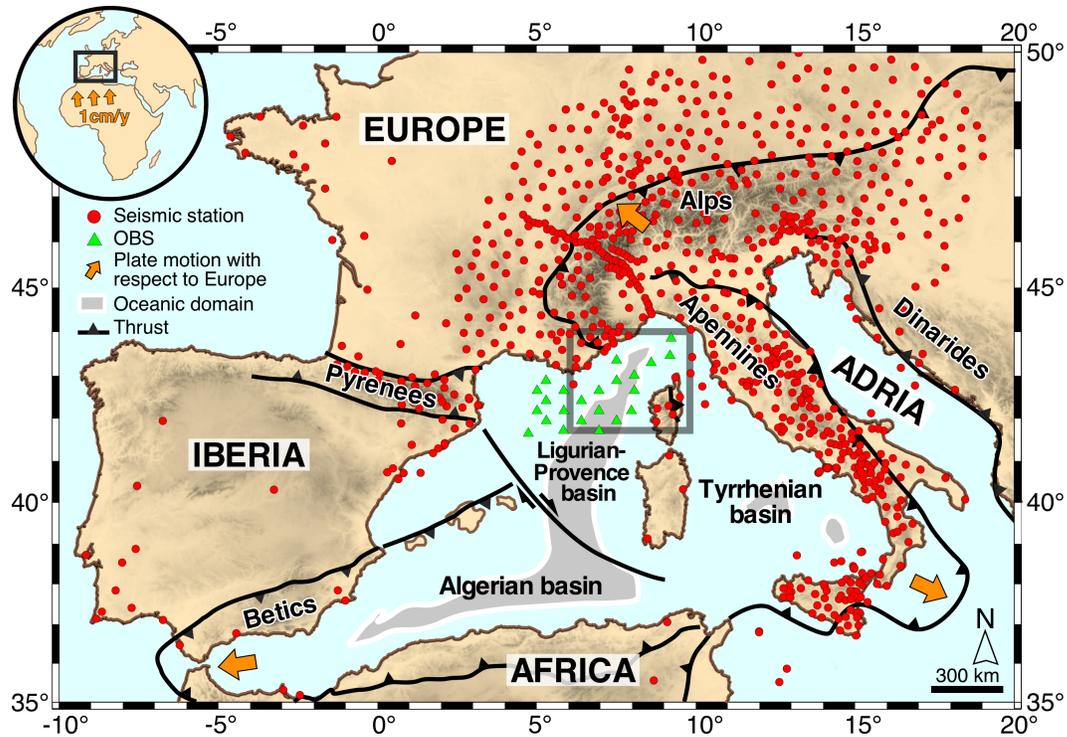


Figure 1. Tectonic map of the western Mediterranean region (modified from Faccenna et al., 2014; Jolivet et al., 2020) with locations of seismic stations used in this work (red circles: onshore sensors; green triangles: AASN OBS). The oceanic domains of the Ligurian-Provence, Algerian and Tyrrhenian back-arc basins are filled with gray color. The gray frame shows the location of the map in Figure 8a.

2 Data processing

2.1 Description of the AASN sea-bottom instruments

We processed ambient-noise records from 23 ocean-bottom-seismometers (OBS in Fig. 1) that were deployed for eight months in the northwestern Mediterranean Sea as the offshore part of the AlpArray temporary seismic network (AASN; network code Z3). All OBSs were deployed in June 2017 by the AlpArray-OBS cruise of the French R/V "Pourquoi-Pas?" (Crawford, 2017), and recovered in February 2018 by the MSM71 cruise of the German R/V "Maria S. Merian" (Kopp et al., 2018).

Sixteen LOBSTER instruments (Long-term Ocean Bottom Seismometer for Tsunami and Earthquake Research) belong to the GEOMAR and DEPAS pools (German instrument pool for amphibious seismology). Designed by K.U.M. Environmental- and Marine Technology GmbH, these instruments were equipped with HTI-01-PCA hydrophones from High Tech Inc., Trillium compact velocimeters from Nanometrics and K.U.M., and with CMG-40T velocimeters from Guralp for four of them. Seven broadband OBS (BBOBS) belong to the French OBS pool of INSU-IPGP. Designed by the Scripps Institution of Oceanography, these instruments were equipped with deep-sea differential pressure gauges and Nanometrics Trillium-240 very broadband velocimeters.

The drift of the sensor clocks was measured before the deployment and after the recovery. Recordings were then corrected for clock drift in the RESIF and GEOFON datacenters that archive and distribute the dataset. Hable et al. (2018) has demonstrated that the assumption of a linear clock drift is adequate for the ocean-bottom instruments used here.

2.2 Glitch removal

The first step of the processing of OBS data is the removal of instrumental transient nearly-periodic impulsive noises (glitches) from continuous records of the 7 French broadband OBSs (BBOBS in section 2.1). Indeed, the vertical-component signals exhibit glitches of 1-hr period caused by the activation of the hourly-check of the internal mass centering of the sensor. The pressure component exhibits glitches with a period of 2.65 hrs, related to data writing on the hard disk. No glitch was detected on the horizontal components. Similar glitches have been observed in other datasets recorded with the same instruments (e.g., Deen et al., 2017).

172 Similarly to Deen et al. (2017), we remove glitches from the data using an average
173 glitch waveform matching algorithm. Detailed explanation can be found in Text S1 of the
174 supporting information. Supplementary Fig. S1 shows a vertical-component daily record of
175 OBS A416A before and after removing the hourly glitches. It documents the efficiency of the
176 processing by comparing the original signal (in blue), the synthetic glitch signal time series
177 (in red), and the final, glitch-free signal (in black). Similarly, Supplementary Fig. S2 shows
178 a daily record of the pressure component before and after removing the 2.65-hrs-period
179 glitches. The power spectral density curves of an example of 1-day raw and pre-processed
180 vertical component record displayed in Fig. 2a, show that removing the glitches reduces the
181 noise level by up to 15 dB at frequencies lower than 10^{-1} Hz.

182 **2.3 Seafloor-noise reduction**

183 At frequencies below 5×10^{-2} Hz, the power spectral densities (PSD) of the three-
184 component records of all OBSs are dominated by noise due to compliance and tilt (Fig. 2).
185 The compliance is a long-period pressure signal generated by infra-gravity waves induced
186 by pressure variations in the water column (Crawford et al., 1998). Tilting corresponds to
187 displacements and rotations of the sensor induced by seafloor currents that also generate
188 long-period noise. As documented by Fig. 2a, long-period noise is stronger on the horizontal
189 components than on the vertical-component. Crawford and Webb (2000) have shown that
190 noise on the vertical-component is lower when the instrument is better levelled. However,
191 long-period noise on the vertical component strongly increases if the instrument is tilted,
192 even slightly, since acceleration induced by seafloor currents on the horizontal components
193 is projected onto the vertical component.

194 Tilt and compliance noises on the vertical-component records are reduced by using a
195 frequency-dependent response function method (Crawford et al., 1998; Crawford & Webb,
196 2000). The horizontal components are firstly corrected for compliance noise by subtracting
197 coherent signals derived from the pressure component. In a second step, tilt noise is reduced
198 in the vertical-component by subtracting coherent signals derived from the compliance-
199 corrected horizontal components (black to green in Fig. 2a). Finally, the resulting vertical-
200 component signal is corrected for compliance by subtracting the coherent signal derived from
201 the pressure-component signal (green to purple). Text S2 of the supporting information
202 provides a detailed explanation of this procedure.

Figure 2a shows the power spectral density of a daily, vertical-component record of OBS A416A before any correction is applied (blue curve), after correcting from the hourly glitches (black), from the tilt (green) and from the compliance (purple). As shown in Fig. 2b, the coherence of the raw vertical-component with the horizontal components (black and blue curves) increases below 5×10^{-2} Hz (see also Supplementary Fig. S3a) while coherence with the pressure component increases around 1.2×10^{-2} Hz. Figure 2a shows that the reduction of the tilt noise (green curve) is maximum (10 dB) at 5×10^{-3} Hz, which is the frequency with the maximum coherence between the vertical component and the 2 horizontal components. The correction for the compliance noise is almost negligible for this record except around 1.2×10^{-2} Hz, where the coherence between the vertical and the pressure components increases (green area in Fig. 2b). Supplementary Fig. S3b confirms that the correction for compliance is maximized for this station around 1.2×10^{-2} Hz. In this specific case, the tilt noise is stronger than the compliance noise, which indicates that pressure variations generate less noise than sea-current induced tilt. This strong tilt noise may result from strong currents at the seabed and/or the presence of poorly consolidated sediments directly under the OBS. The lower compliance noise may be due to the large water depth of OBS A416A (~ 2630 m), damping the effect of pressure-induced infra-gravity waves.

3 Computation of noise correlations and group-velocity measurements

Once OBS records have been corrected for glitches and seafloor noises, we apply the same pre-processing scheme described below to the records of all stations, onshore and offshore, in order to prepare the calculation of inter-station cross-correlations. As explained in Soergel et al. (2020) and Nouibat et al. (2022), we first down-weight the contribution of earthquakes and other high-amplitude transients by removing all 4-hr segments with a peak amplitude 4 times greater than the standard deviation of the current daily record, and with a RMS greater than 1.5 times the daily mean RMS. Each daily record is then filtered into six period bands (3 - 5, 5 - 10, 10 - 20, 20 - 40, 40 - 80 and 80 - 200 s) and amplitudes are normalized by their envelope. Finally, the 6 filtered and normalized signals are stacked to obtain the 4-hr pre-processed broadband signal.

3.1 First-order correlations

As in Nouibat et al. (2022), we compute seismic noise cross-correlations for all station pairs by segments of 4 hrs. The 4-hr correlations are normalized and stacked to obtain a

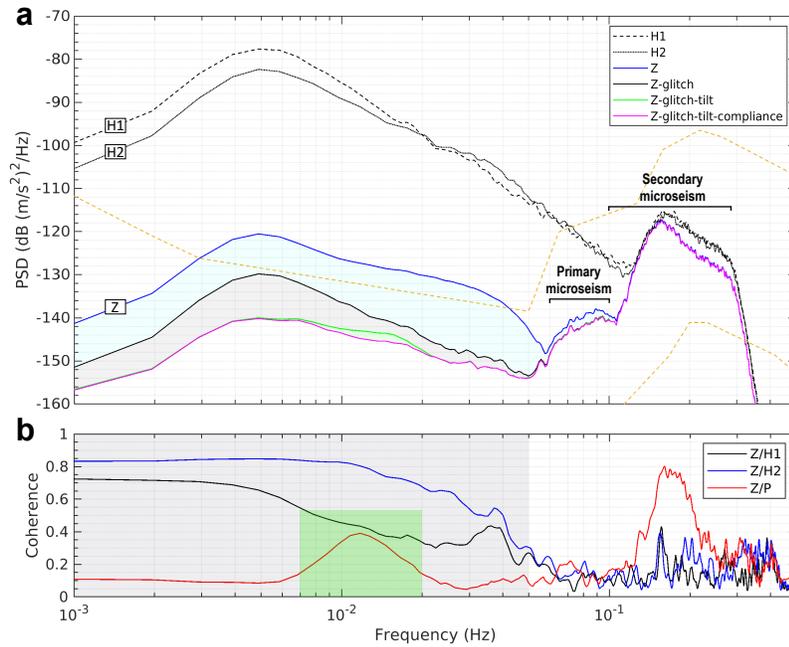


Figure 2. Effect of the different corrections for glitch and seafloor noises on the PSD of a 1-day vertical-component record of OBS A416A. Raw signals are band-pass filtered between 2.5 and 250 s, corrected from the instrumental response and decimated to 1 Hz sampling frequency. (a) PSDs before and after corrections; Orange dashed lines indicate the Peterson (1993) New Low and High Noise Model (NLNM & NHNM); Blue: PSD of the raw vertical-component record (blue signal in supplementary Fig. S1a); At frequencies $> 5 \times 10^{-2}$ Hz, the PSD is dominated by the primary (~ -140 dB) and secondary (~ -120 dB) microseisms. At longer periods, the PSD is dominated by the effect of the periodic glitches and the seafloor noises with levels similar to the secondary microseism and higher than the NHNM (in the 3×10^{-3} Hz - 4×10^{-2} Hz band) with a peak amplitude at 5×10^{-3} Hz; Black: PSD of the Z-component after correction for the 1-hr glitches (black signal in supplementary Fig. S1a); The -130 dB peak at 5×10^{-3} Hz remains, pointing to its oceanic origin; Its amplitude is weaker than the secondary microseism but still stronger than the primary microseism; Green: PSD of the Z-component after correction for the glitches and the tilt noise; Purple: PSD of the Z-component after correction for the glitches and seafloor noises; The maximum amplitude of the residual signal is now well below the primary microseism and the NHNM. The dotted and dashed black curves are the PSDs of the horizontal components. Colored areas show the reduction of the noise level after correcting from glitches (blue), from the tilt (gray), and from the compliance (green). (b) Coherence between the vertical channel and the horizontal (blue and black curves) and pressure (red curve) channels. The gray area shows the frequency domain where the Z-component is coherent with the horizontal components due to tilt noise. The green area shows the domain where the Z-component is coherent with the pressure component due to compliance noise. At frequencies higher than 5×10^{-2} Hz, the coherence is due to primary and secondary microseisms.

234 single reference correlation per station pair. We use up to 4 years of continuous vertical
235 records for on-land stations pairs, and up to 8 months data to compute correlations for
236 OBS-OBS and OBS-land-station pairs.

237 As shown by Nouibat et al. (2022) and supplementary Fig. S4a, Rayleigh waves are
238 clearly visible in the correlations for on-land station pairs in a wide period band (5 - 150 s).
239 The Rayleigh waves have an average signal-to-noise ratio (SNR) greater than 3.5 (Supple-
240 mentary Table S1). Although correlations for OBS-land station pairs are computed from
241 only 8 months of data, Rayleigh waves have a $\text{SNR} > 3$ in the 5 - 70 s period band, except in
242 the 40 - 70 s band where the SNR is slightly lower ($\text{SNR} = 2.87$, see supplementary Fig. S4b
243 and Table S1). These noise correlations can therefore be used for Rayleigh-wave tomography.
244 Supplementary Figures S4b-c demonstrate the effectiveness of corrections for glitches and
245 seafloor noise in enhancing the SNR of correlations between OBS and land-station records.

246 First-order correlations between raw OBS records are displayed in the Fig. 3a. They
247 can be compared to correlations for land-land station pairs and OBS-land station pairs
248 shown in Supplementary Fig. S4. The SNR of correlation signals for OBS pairs is poorer
249 than for other types of pairs in all period bands (average $\text{SNR} < 2.6$, Table 1). Therefore,
250 Rayleigh waves are hardly detectable at periods shorter than 40 s in OBS-OBS correlations,
251 and undetectable at longer periods ($\text{SNR} < 1.6$, Table 1). This may be explained by several
252 factors such as local noises generated around the sensors by seafloor currents, or seismic
253 noises generated between the stations that induce signals around time 0 s of the correlations,
254 masking the Rayleigh waves. Therefore, these first-order OBS-OBS correlations of raw
255 records cannot be used for Rayleigh-wave tomography.

256 Figure 3b and Table 1 document once again the effectiveness of the corrections for
257 glitches and seafloor noises applied to OBS records, which improve the inter-OBS correlation
258 signals. The correlations have a better SNR and are more symmetrical than those obtained
259 from uncorrected signals, particularly at periods shorter than 40 s (columns a-b of Fig. 3).
260 However, correlation signals are still noisy, particularly at short lag times, and accurate
261 measurements of Rayleigh-wave group velocities remain challenging. To further improve
262 the quality of correlations for OBS pairs, we chose to virtually reconstruct the Rayleigh
263 waves by computing iterative correlations.

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3.2 Iterative correlations for OBS-OBS paths

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We have seen in the previous section that correlations for OBS pairs can hardly be used for Rayleigh-wave tomography. Since correlations computed between onshore stations and OBSs exhibit clear Rayleigh waves, we will use onshore stations as virtual sources in order to measure the travel time of Rayleigh waves between OBSs. Indeed, first-order correlations computed between onshore stations and OBSs contain the Rayleigh-wave part of the Green's function. It is thus possible to use them to mimic the case where Rayleigh waves are emitted on the continent and recorded by OBSs. By computing correlations of Rayleigh waves emitted by each virtual source and recorded by two OBSs (i.e., by computing a second-order correlation), it is possible, thanks to the stationary phase theorem, to isolate the Rayleigh-wave propagating between the two OBSs, and therefore to measure its travel time. We explain this with more detail in the following.

Let us consider any medium with a distribution of sources f . The wavefield recorded at a station A can be expressed using the Green's function G of the medium:

$$u(\vec{r}_A, t) = \int_{\Omega} \int_0^{\infty} G(\vec{r}_s, \vec{r}_A, t') f(\vec{r}_s, t - t') dt' d\vec{r}_s \quad (1)$$

It has been shown that the time-derivative of the first-order cross-correlation $C_{\vec{r}_A, \vec{r}_B}^1(\tau)$ computed between wavefields recorded at two stations A and B is the Green's function $G_{\vec{r}_A, \vec{r}_B}(t)$ of the medium, assuming for instance a perfectly homogeneous distribution of white noise everywhere in the medium (e.g., Lobkis & Weaver, 2001; Snieder, 2004; Wapenaar, 2004; Roux et al., 2005; Weaver, 2005; de Verdière, 2006):

$$\frac{d}{d\tau} C_{\vec{r}_A, \vec{r}_B}^1(\tau) = G_{\vec{r}_A, \vec{r}_B}(\tau) \quad (2)$$

In the case where all white noise sources are spatially uncorrelated, the correlation of wavefields recorded at A and B can be rewritten as the integral of correlations between $G(\vec{r}, \vec{r}_A)$ and $G(\vec{r}, \vec{r}_B)$:

$$\begin{aligned} C^1(\vec{r}_A, \vec{r}_B, \tau) &= \int_0^{\infty} u(\vec{r}_A, t) u(\vec{r}_B, t + \tau) dt \\ &= \int_{\Omega} \int_0^{\infty} G(\vec{r}_s, \vec{r}_A, t') f(\vec{r}_s, t - t') dt' d\vec{r}_s \otimes \int_{\Omega} \int_0^{\infty} G(\vec{r}_s, \vec{r}_B, t') f(\vec{r}_s, t - t') dt' d\vec{r}_s \\ &= \int_{\Omega} G(\vec{r}_s, \vec{r}_A, t) \otimes G(\vec{r}_s, \vec{r}_B, t) d\vec{r}_s \end{aligned} \quad (3)$$

where \otimes denotes the cross-correlation operation. Since the time-derivative of correlations is similar to the Green's function of the medium, it follows immediately by substituting dC^1/dt

to G in Eq. 3 that the first-order correlation is equivalent to a second-order correlation that we will note C^2 :

$$C^1(r_A^{\vec{}}, r_B^{\vec{}}, \tau) = \int_{\Omega} \frac{d}{dt} C^1(r_s^{\vec{}}, r_A^{\vec{}}, t) \otimes \frac{d}{dt} C^1(r_s^{\vec{}}, r_B^{\vec{}}, t) dr_s^{\vec{}} = C^2(r_A^{\vec{}}, r_B^{\vec{}}, \tau) \quad (4)$$

Equation 4 indicates that it is possible to reconstruct the Green's function of the medium between A and B by re-correlating the noise correlations computed between each point of the medium and stations A and B . However, this demonstration assumes that the time-derivative of correlations are the exact and complete Green's function of the medium. The assumption would be correct if stations that could be used as virtual sources would exist everywhere in the medium.

In practice, the seismic noise recorded at the Earth surface in the period band considered in this work (5 - 150 s) is dominated by Rayleigh waves. As a consequence, noise correlations computed between onshore stations and OBSs do not provide the full Green's function of the medium including all propagating modes, but they do provide robust estimates of the travel time of Rayleigh waves.

Rather than attempting to reconstruct the full Green's function, we will measure the travel time of the Rayleigh waves between OBSs A and B by using a simplified approach inspired by Eq. 4. It consists in re-correlating only the Rayleigh-wave parts of the correlations computed between onshore stations and OBSs A and B , that we consider as the Rayleigh-wave part of the Green's function:

$$\begin{aligned} \frac{d}{d\tau} C^{2, Ray}(r_A^{\vec{}}, r_B^{\vec{}}, \tau) &= \frac{d}{d\tau} \int_{\Omega} \frac{d}{dt} C^{1, Ray}(r_s^{\vec{}}, r_A^{\vec{}}, t) \otimes \frac{d}{dt} C^{1, Ray}(r_s^{\vec{}}, r_B^{\vec{}}, t) dr_s^{\vec{}} \\ &\approx G^{Ray}(r_A^{\vec{}}, r_B^{\vec{}}, \tau) \end{aligned} \quad (5)$$

where superscript "Ray" indicates that we only correlate the fundamental mode of Rayleigh waves of first-order correlations. In that way, we only retrieve the fundamental mode of the Rayleigh-wave part of the Green's function.

In practice, the distribution of virtual sources is never homogeneous. Instead, we use land stations deployed all over Western Europe, while OBSs are located in the Ligurian Sea (Fig. 1). Therefore, the condition of the stationary phase theorem are not completely met. To circumvent this difficulty, we select virtual sources that are expected to contribute constructively to the correlations (Fig. 4). To that end, we design a virtual-source azimuthal-filter that only retains sources located in the end-fire lobe of the OBS couple, that is in azimuths at $\pm 20^\circ$ with respect to the azimuth of the OBS pair. Moreover, we enhance the

297 virtual source coverage by using separately the causal and anticausal parts of the first-order
298 correlations to compute the second-order correlation.

299 Because onshore stations are well distributed and OBS-onshore correlations exhibit
300 clear Rayleigh waves, the use of virtual sources in iterative correlations for OBS-OBS paths
301 leads to a higher quality of Rayleigh waveforms (columns b-c of Fig. 3 and Table 1). This is
302 achieved through: (1) separately recovering the causal and anticausal parts of the Green's
303 function by using separately the causal and anticausal parts of the first-order correlations,
304 thus avoiding interferences at long periods, and (2) controlling the distribution of virtual
305 sources, thus guaranteeing a higher quality of the C^2 s by contrast to OBS-OBS C^1 s that
306 exhibit low signal-to-noise ratio probably due to local noise sources. These local noise
307 sources do not contribute significantly to OBS-onshore stations paths, and therefore neither
308 to second-order correlations. Supplementary Fig. S5 shows that phases of the Rayleigh
309 waveforms reconstructed from the C^1 and C^2 processes match. In the 5 - 10 s band,
310 iterative correlations do not systematically improve the signal quality as compared to first-
311 order correlations. Therefore, we select for each path the correlation of highest quality after
312 checking that the C^1 and C^2 are coherent.

313 The strengths of the iterative correlations make it possible to substantially improve
314 the path coverage in the Ligurian-Provence domain (Table 1). Our results demonstrate
315 the efficiency of this method in providing robust group-velocity measurements. Further
316 illustration and validation will be the subject of a future paper.

317 3.3 Group-velocity measurements

318 Once first-order and iterative correlations have been computed for onshore and offshore
319 stations respectively, we derive group-velocity dispersion curves of the causal and acausal
320 parts of the correlations by using multiple filter analysis (MFA, Dziewonski et al., 1969;
321 Herrmann, 1973). As in Nouibat et al. (2022), we adapt the width of the gaussian filter
322 to the inter-station distance to accommodate the trade-off in resolution between the time
323 and frequency domains (Levshin et al., 1989). We correct our group-velocity measurements
324 from the biases that occur when the MFA method is applied on signals having a non-flat
325 spectrum (Shapiro & Singh, 1999). This is especially important when measuring Rayleigh-
326 wave velocities using noise correlations around the first and second microseismic peak, i.e.,
327 around 7 s and 14 s.

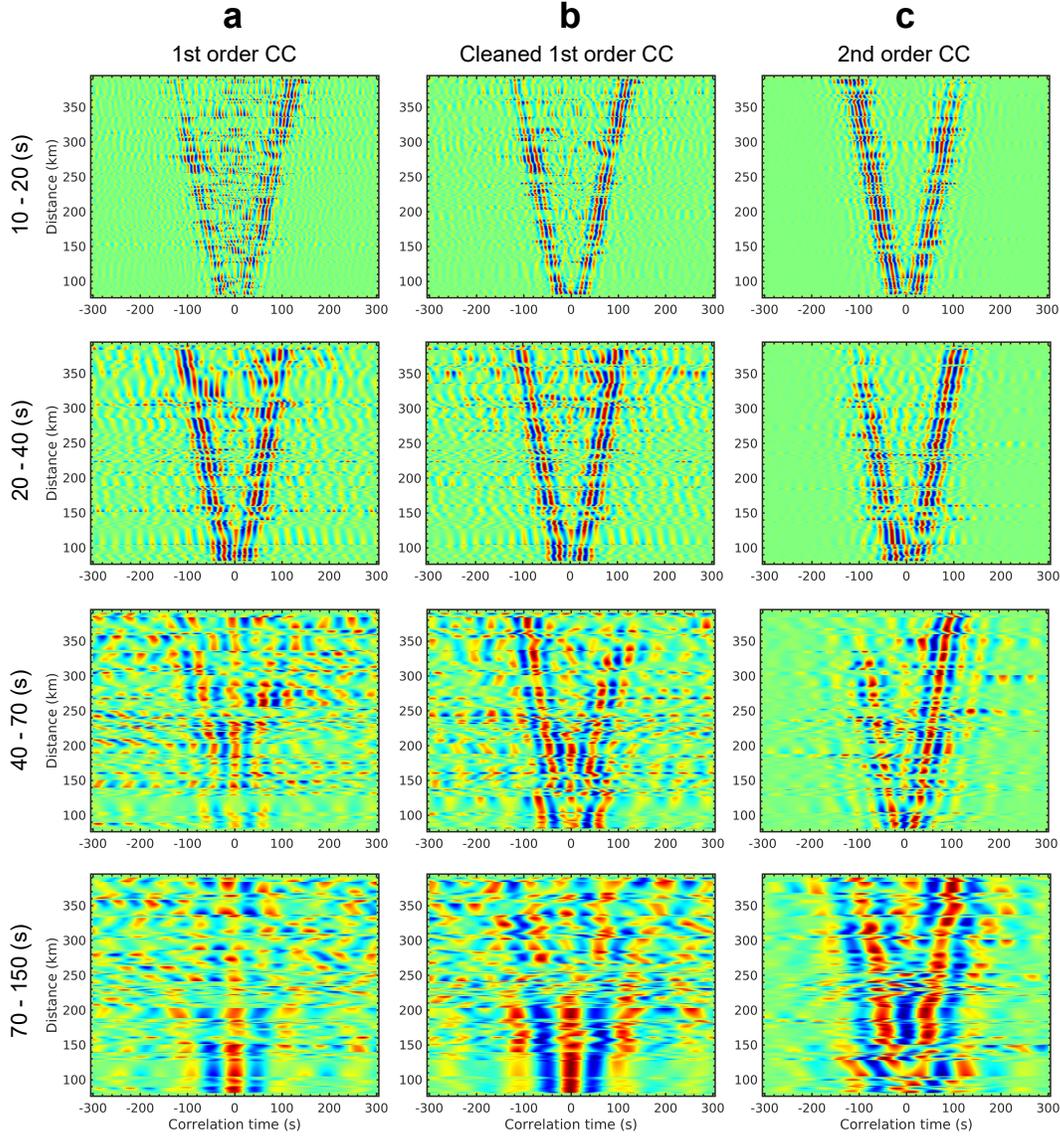


Figure 3. Time-distance plots of correlation signals for OBS-OBS pairs, obtained in different period bands at different steps of the processing. (a) First-order cross-correlations (C^1s) of raw signals. (b) C^1s of pre-processed signals (glitch removal & seafloor-noise reduction). (c) Second-order correlations (C^2s). The C^1s in (a) are generally of poor quality and poor symmetry, with strong signals at short lag times due to interferences even at short periods. These interferences become stronger with increasing period, and they progressively overshadow the Rayleigh wave-trains. In (b), the C^1 signals are strongly improved due to corrections for glitches and seafloor noises, in particular in the three short- and medium-period bands. The 70 - 150 s band is still affected by interferences that hide the Rayleigh wave-trains. In (c), the quality and the time-symmetry of correlation signals are significantly improved in all period bands by the calculation of C^2s . In particular, the Rayleigh wave-train emerges from the noise in the long-period band.

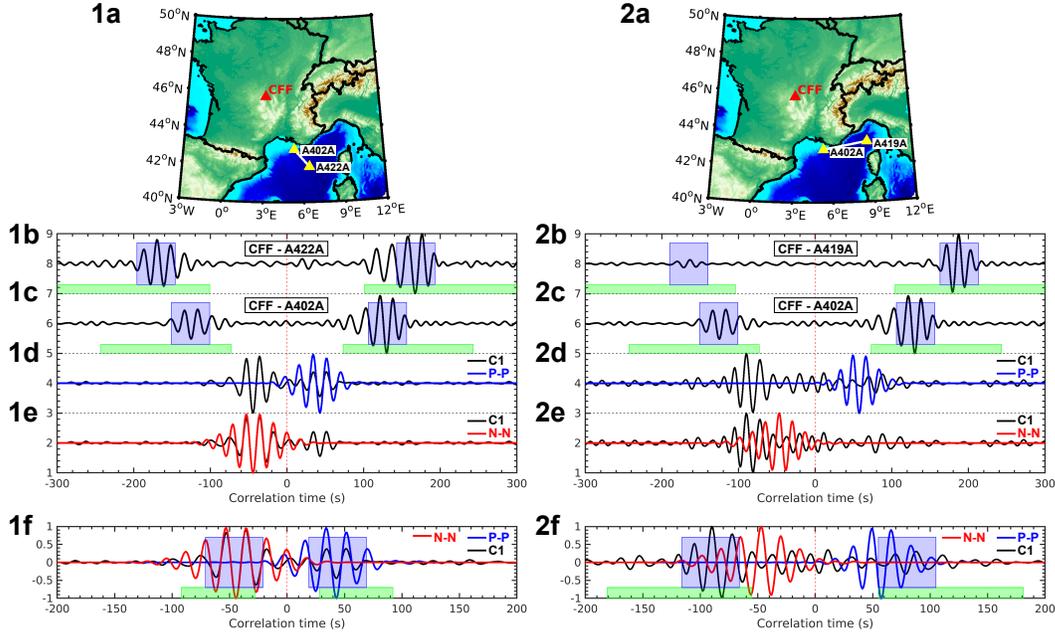


Figure 4. Details of the computation of second-order correlations for 2 examples of OBS-OBS pairs, using station CFF (FR network) as virtual source. (1-2a): Station location maps. In example (1), the virtual source is roughly aligned with the OBS pair, while in example (2), the azimuth of the virtual source is almost perpendicular to the pair. (1-2b, 1-2c): First-order correlation signals between CFF and each OBS. The green areas show the Rayleigh-wave search windows (wave propagation of 1 to 5 km/s over the inter-station distances). The blue areas show the Rayleigh-wave detection windows. (1-2d): First-order correlation of the OBS signals (C^1 , in black), and correlation of the positive-time (causal) parts of the C^1 s between CFF and the OBS couple (P-P, in blue). (1-2e): First-order correlation of the OBS signals (C^1 , black), and correlation of the negative-time (acausal) parts of the C^1 s between CFF and the OBS couple (N-N, red). (1-2f): Comparison of the causal (P-P, blue) and acausal (N-N, red) C^2 s (that will be summed to obtain the final C^2), with the first-order correlation of the OBS signals (C^1 , black). The C^2 and C^1 signals have similar phases in (1) where the virtual source is aligned with the OBS pair. They have incoherent phases in (2), since CFF is not located in the end-fire lobe of this OBS pair (see text).

SNR — % of retained paths			
Period band	1st order CC	Cleaned 1st order CC	2nd order CC
10s – 20s	2.52 — 2.45%	3.29 — 8.11%	5.77 — 39.6%
20s – 40s	2.32 — 1.52%	3.61 — 11.5%	5.62 — 23.4%
40s – 70s	1.57 — 0.03%	3.12 — 5.22%	4.52 — 14.3%
70s – 150s	1.13 — 0.01%	1.42 — 2.61%	4.37 — 7.31%

Table 1. Signal-to-noise ratio and percentage of selected paths for group-velocity tomography in different period bands using: (1) first-order cross-correlations of raw OBS vertical-component records, (2) first-order cross-correlations of pre-processed signals (corrected for glitches, and seafloor noises), and (3) second-order cross-correlations of pre-processed signals.

328 In order to build the group-velocity maps in the Ligurian-Provence domain, we maxi-
 329 mize the path coverage over the Ligurian Sea by using simultaneously OBS-OBS, land-land,
 330 and land-OBS station pairs. A careful selection of group-velocity measurements is achieved
 331 to keep the most reliable ones and discard those that are biased by an unfavorable distribu-
 332 tion of noise sources, or by interferences of causal and acausal Rayleigh waves for instance.

333 For first-order correlations (C^1s) computed between land-land and land-OBS stations,
 334 at each period, we keep measurements if: (1) the SNR defined as the ratio of the Raleigh-
 335 wave peak amplitude and the standard deviation of the following signal, is greater than 3
 336 on the positive and negative correlation times, (2) group velocities measured in positive and
 337 negative correlation times differ by less than 0.2 km/s, and (3) the inter-station distance
 338 is greater than 2 wavelengths. For iterative correlations (C^2s) computed between OBS
 339 stations, we do not use the SNR criteria, since iterative correlations only exhibit the fun-
 340 damental mode of the Rayleigh waves, owing to their construction. Nevertheless, for each
 341 C^2 satisfying the distance and symmetry criteria, we only keep the group-velocity measure-
 342 ment of the side of the correlation (positive or negative time) where the amplitude of the
 343 Rayleigh-wave is maximum (i.e., where we have more virtual sources contributing).

344 Table 1 shows that for OBS-OBS pairs, the selection procedure would reject more than
 345 97% of Rayleigh waves velocity measurements performed on first-order correlations com-
 346 puted using OBS data that were not corrected from the compliance and tilt noises. This

347 illustrates that these signals are obviously not usable for ambient-noise tomography. By
348 contrast, we selected between 2.6% to 11.5% of the measurements performed on first-order
349 correlations done using OBS data corrected from sea floor noises. This highlights the im-
350 portance of the pre-processing scheme described in section 2. Moreover, we kept between
351 7.3% and 39.6% of group-velocity measurements performed on iterative correlations (C^2s)
352 depending on the period-band considered. Second-order correlations provide a substantial
353 gain over first-order correlations, leading to a significant improvement of the raypath cov-
354 erage in the Ligurian-Provence basin with respect to the ANT of Wolf et al. (2021) based
355 on first-order correlations.

356 4 3-D shear-wave velocity model

357 4.1 Inversion for 2-D group-velocity maps

358 We compute 2-D group-velocity maps and associated uncertainties using a "data-
359 driven" transdimensional approach at discrete periods from 4 s to 150 s. At each period,
360 probabilistic group-velocity maps are derived by exploring millions of 2-D models using the
361 reversible-jump Markov-chain Monte-Carlo method (rj-McMC, Bodin et al., 2012). The
362 method used for the inversion and the spatial resolution of the resulting group-velocity
363 maps are discussed in detail in Nouibat et al. (2022). Uncertainty and path density maps
364 at different periods are presented in Supplementary Figure S6.

365 Resulting group-velocity maps of the Ligurian-Provence basin and its margins are shown
366 in Fig. 5 for periods from 6 s to 35 s. The 6 s and 8 s maps (sensitive to $\sim 4 - 8$ km depth)
367 highlight low-velocity anomalies ($U < 1.6$ km/s) in the central and southwestern parts of
368 the basin (C-SLPB in Fig. 5), that are probably associated with thick sediment sequences.
369 These velocities are lower than those of the southeast-France basin and the Gulf of Lion
370 (respectively SFB and GL in Fig. 5). The northeastern Ligurian-Provence basin (NLPB),
371 its northern, Provence coast margin and its southern margin in Corsica have larger velocities
372 ($U \geq 2.4$ km/s). From simulations of Rayleigh-wave dispersion in synthetic 1-D models, we
373 show that such periods are also highly sensitive to the presence and thickness of the water
374 column (Supplementary Figure S7). The 10 s map still shows velocities lower than 2.5 km/s
375 in the central and southwestern Ligurian-Provence basin. The 12 s and 15 s maps (sensitive
376 to $\sim 10 - 15$ km depths) highlight velocities larger than 3.5 km/s associated with the thin
377 crust of the Ligurian basin. However, the northeastern basin has lower velocities indicative

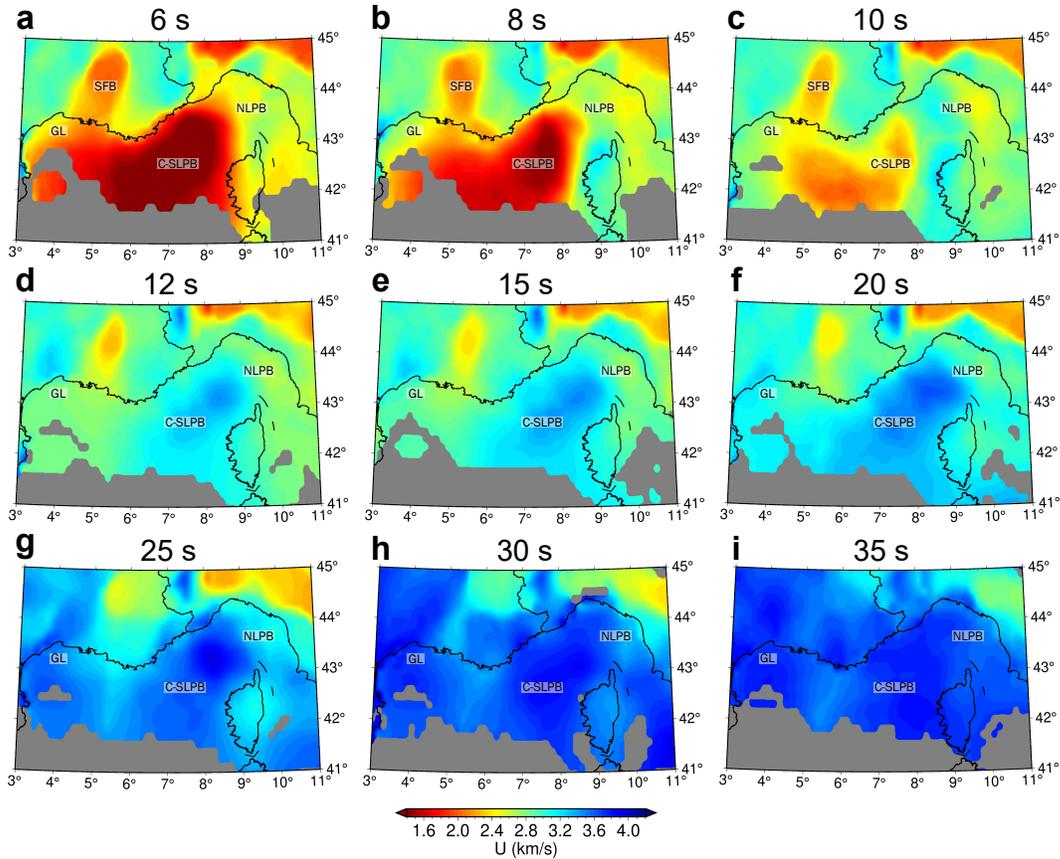


Figure 5. Group-velocity maps (average solutions) at 6 to 35 s periods, obtained with the Hierarchical Bayes reversible-jump algorithm. Only areas with uncertainty lower than 0.5 km/s are shown. C-SLPB: central and southwestern parts of the basin, GL: Gulf of Lion, NLPB: northeastern Ligurian-Provence basin, SFB: southeast-France basin.

378 of a deeper Moho in the Gulf of Genova. Velocities lower than 3.2 km/s correspond to thick
 379 crust under Corsica and the Provence coast. At 25 s period (sensitive to $\sim 15 - 30 \text{ km}$
 380 depth), velocities are still lower along the western coast of the Gulf of Genova than in the
 381 basin, indicative of a thicker crust, as in Corsica. At 30 s and 35 s periods, group velocities
 382 are homogeneous and exhibit large velocities ($U > 3.5 \text{ km/s}$) in most of the study region.

383 4.2 Inversion for shear-wave velocity

384 The group-velocity maps and their uncertainties are used to derive a 3-D V_s model. For
 385 this, we perform a two-step data-driven inversion to tackle the non-unicity of the inverse
 386 problem. The main part of this process is described in detail in Nouibat et al. (2022), so

387 we will only summarize it here. We will rather focus on the specificity of the inversion for
388 the offshore region, which is the consideration of the water layer. A result of the inversion
389 for V_s at an offshore location in the Ligurian Sea is shown in Supplementary Fig. S8.

390 Firstly, a 3-D probabilistic solution is computed that gives at each location the prob-
391 ability distribution of V_s and the probability of having an interface as a function of depth.
392 This is achieved using an exhaustive grid search on a set of ~ 130 million synthetic four-layer
393 models, which include a sedimentary layer, the upper crust, the lower crust and a half-space
394 representing the upper mantle. The strength of this first-step Bayesian framework lies in
395 constraining the structural complexity of the crust (i.e., of the short-period part of the
396 dispersion curve) by means of an ensemble of models, fitting the dispersion curve to the
397 degree required by its uncertainties. However, due to the four-layer model assumption, this
398 procedure is not sufficient to fully describe the complexity of the model structure, particu-
399 larly in the mantle part, that is in the long-period part of the dispersion curve. Hence, we
400 use an iterative linear least-square inversion (Herrmann, 2013) as a complement to update
401 the mantle part of the model and further refine the fit to the local dispersion curve for the
402 crustal part (Supplementary Fig. S8).

403 The initial model for the linear inversion (second step) at a given location is the average
404 of all selected probabilistic solutions of the first inversion step, weighted by their likelihood
405 values. We discretize the crustal and mantle parts at intervals of 1 and 5 km, respectively,
406 and assume a gradual increase of V_s below Moho according to the global model PREM
407 (Dziewonski & Anderson, 1981). For offshore locations, we incorporate on top of this initial
408 model an additional layer of thickness equal to water depth at the given location. The pa-
409 rameters of this water layer are kept fixed during the linear least-square inversion (thickness,
410 $V_s = 0$ km/s, $V_p = 1.5$ km/s, $\rho = 1 \times 10^3$ kg/m³). Since we invert for short periods as well,
411 an appropriate parameterisation of the water column is crucial. Indeed, we highlight its
412 influence by computing group-velocity dispersion curves for synthetic 3-layer crustal models
413 representative of the oceanic crust of the Ligurian-Provence basin, and different water levels
414 from 0 to 3.1 km (Supplementary Fig. S7). This shows that the effect of the water layer on
415 the dispersion curve is substantial at periods shorter than 15 - 20 s, where the water depth
416 changes impact: (1) the absolute group velocities, and (2) the shape of the dispersion curve,
417 particularly in the vicinity of the Airy phase.

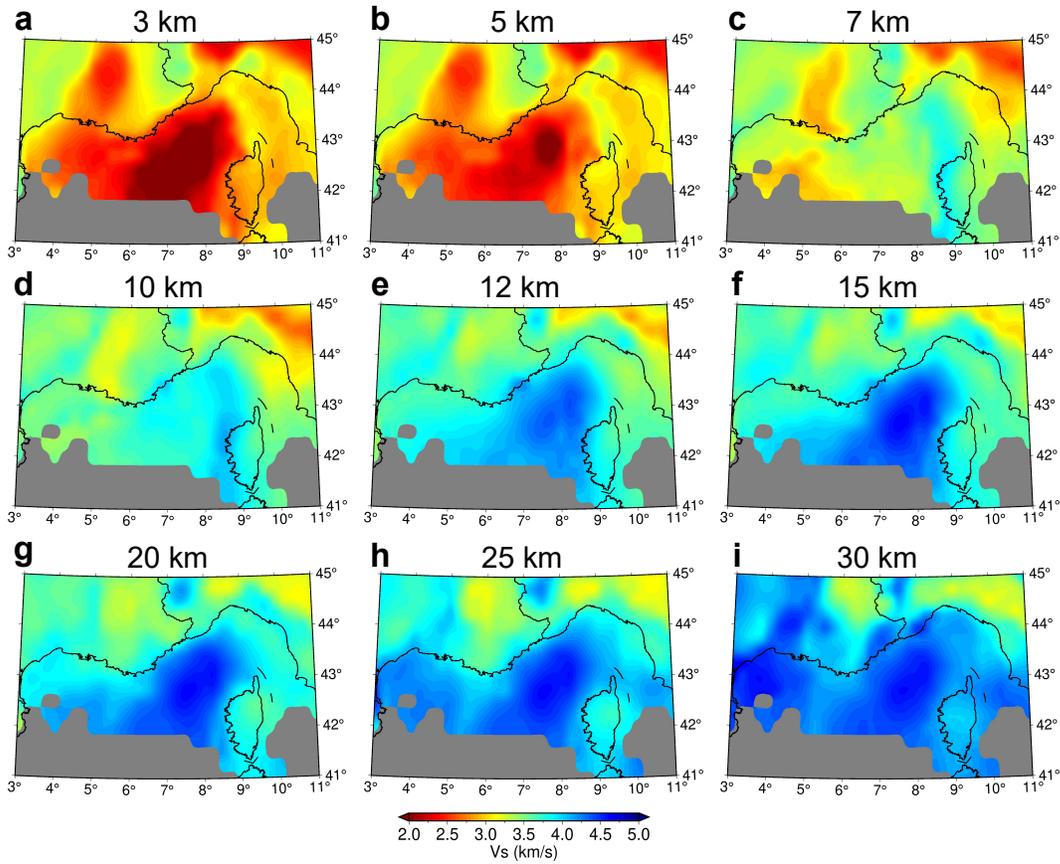


Figure 6. Depth slices in the final V_s model at 3 to 30 km depths. Only regions with 1σ error $< 8\%$ are shown.

418 Figure 6 shows depth slices in the 3-D shear-wave velocity model at 3 to 30 km depth.
 419 The sediment layer is clearly visible at 3 - 7 km depth in the central and southwestern
 420 Ligurian-Provence basin, with low velocities of 2 - 3.2 km/s. The transition from crustal
 421 (3.5 - 4.1 km/s) to mantle velocities (4.1 - 4.5 km/s) is located between 10 km and 12 -
 422 15 km in the parts of the basin with thinnest crust. Between 20 and 30 km, areas of strong
 423 velocities corresponding to the mantle extend westward from the Ligurian-Provence basin
 424 to the Gulf of Lion across a nearly N-S transition of slower velocities at $\sim 5.5^\circ$ E.

4.3 Comparison with the V_s model by Wolf et al. (2021)

425
 426 Figure 7 shows a comparison of our V_s model and the model by Wolf et al. (2021),
 427 which was the first published ambient-noise tomography using data of AlpArray OBSs in

428 the Ligurian basin. Our V_s model covers a wider area as part of the large-scale model by
429 Nouibat et al. (2022) that uses all available broadband stations in Western Europe.

430 At 3-km depth, our model (Fig. 7-1a) highlights: (1) the sedimentary cover in the
431 central and southwestern basin (1.8 - 2.2 km/s), where Wolf et al's model (Fig. 7-1b) exhibits
432 patches of very low velocities ($V_s \leq 1.4$ km/s), and (2) gradual increasing of velocity from the
433 central basin towards the conjugate margins, which is not clearly visible in Wolf et al's model
434 where velocities fluctuate from very low ($V_s \leq 1.8$ km/s) to high values (2.4 - 2.5 km/s). At
435 5-km depth, we still observe typical sediment velocities (Fig. 7-2a), while Wolf et al's model
436 (Fig. 7-2b) exhibits higher velocities ($V_s > 3.5$ km/s). At 13-km depth, our model (Fig. 7-
437 3a) shows almost homogeneous mantle velocities in the basin ($V_s \geq 4.1$ km/s) while Wolf et
438 al's model (Fig. 7-3b) exhibits numerous small-size velocity anomalies, fluctuating between
439 mantle-like ($V_s \geq 4.1$ km/s) and crust-like velocities ($V_s \leq 3.7$ km/s). These heterogeneities
440 suggest a much more irregular Moho surface than in our model. At 20-km depth, our model
441 (Fig. 7-4a) documents crustal thinning towards the basin axis with V_s increasing across the
442 conjugate margins. Again, velocities in the central basin are more heterogeneous in Wolf et
443 al's model (Fig. 7-4b), with localized small-size anomalies. Although the transition domain
444 has an irregular shape in Wolf et al's model, it remains to first order similar to that shown
445 in our model.

446 We think that such strong differences between the two V_s models are mostly due to
447 differences in coverage and quality of Rayleigh-wave dispersion data. Indeed, we greatly
448 enhanced the path coverage in the Ligurian basin by using iterative correlations for OBS-
449 OBS paths. In addition, the use of all broadband stations in Western Europe provides
450 long ray paths across the Ligurian sea, allowing us to improve path coverage to the west
451 and southeast and make use of Rayleigh-wave dispersion measurements up to 150 s. On
452 the other hand, Wolf et al. (2021) used only 23 onshore stations nearby the basin, thus a
453 limited aperture. Therefore, they used earthquake records for periods greater than 20 s.
454 To a lesser extent, the differences between the two models may also be explained by the
455 different strategies used to invert Rayleigh-wave dispersion measurements. Wolf et al. (2021)
456 computed their 2-D dispersion maps using a linear inversion that depends on an explicit
457 regularization, while our transdimensional approach does not. Moreover, our 1-D Bayesian
458 inversion for V_s take uncertainties on dispersion measurements into account, which is key
459 for controlling model complexity.

460 As we will see in the following section, our V_s model is more coherent with current
461 knowledge on the crustal structure of the Ligurian-Provence basin than the one by Wolf
462 et al. (2021). In particular, the thickness of the sedimentary cover and Moho depth and
463 geometry estimated from our V_s model are coherent with the V_p model by Dannowski et al.
464 (2020) along the basin axis and with the stratigraphic log after Leprêtre et al. (2013).

465 5 Discussion

466 We now focus on the oceanic domain of the Ligurian-Provence basin (Fig. 8a), in
467 particular on the comparison of our S -wave velocity model with a recent P -wave velocity
468 model derived by Dannowski et al. (2020) from a controlled-source seismic profile recorded
469 along the basin axis (thick black line in Fig. 8a). The availability of this high-resolution
470 V_p section provides a unique opportunity to assess the accuracy and validate our V_s model
471 against an independent dataset. Moreover, the existence of V_p and V_s models along the same
472 profile may provide clues on the petrological structure of the crust in the oceanic domain of
473 the basin, which is still debated. Indeed, it has been proposed that the oceanic domain is
474 made of an oceanic crust with a thin basaltic layer (e.g., Mascle & Rehault, 1990; Bonatti
475 et al., 1990), or an exhumed and serpentinized mantle devoid of any volcanic upper layer
476 (e.g., Boillot et al., 1989; Beslier et al., 1993; Jolivet et al., 2020), or even an hyper-extended
477 continental crust (e.g., McKenzie, 1978; Pascal et al., 1993; Dannowski et al., 2020).

478 5.1 Geological setting of the Ligurian-Provence basin

479 The basin opening initiated at 30 Ma by a rifting phase between Europe and the
480 Corsica-Sardinia block, as a result of back-arc extension above the Adria oceanic micro-
481 plate, initially subducting north-westward (e.g., Faccenna et al., 1997). The progressive
482 south-eastward roll-back and retreat of the Adria slab below the Corsica-Sardinia domain
483 led to stretching of the continental crust followed by continental break-up during the early
484 Miocene, and to the genesis of an oceanic crust between 20 and 15 Ma (Séranne, 1999).
485 As a result, the Ligurian-Provence basin includes two thinned conjugate continental passive
486 margins separated by an oceanic domain (Fig. 8a). According to Rollet et al. (2002), the
487 entire region is characterized by magnetic anomalies, and by the presence of magmatic bodies
488 identified from acoustic facies in seismic reflection profiles. The area between the margins
489 and the oceanic domain is described as a transitional domain, likely made up of a very
490 thin continental crust overlying a thick rift-related corner of magmatic underplating (e.g.,

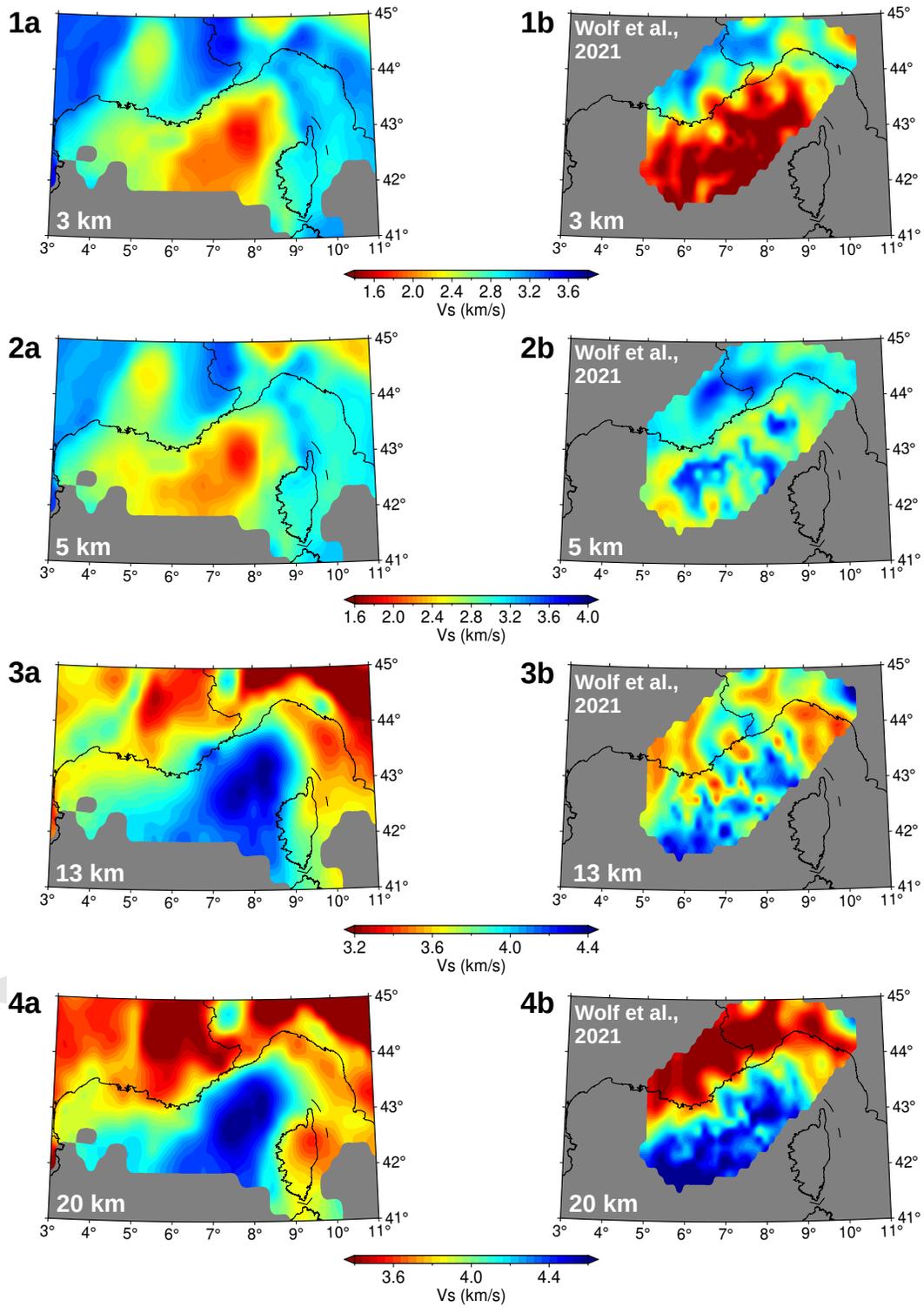


Figure 7. Comparison between our V_s model (left panel) and the V_s model by Wolf et al. (2021) (right panel) at four depths.

491 Séranne, 1999). This limit is marked by an abrupt change in the amplitude of magnetic
492 anomalies with a transition from mostly positive values in the deep basin (i.e., oceanic
493 domain) to negative values at the continent-ocean transition, and by a change in acoustic
494 facies on seismic reflection profiles (e.g., Réhault et al., 1984; Déverchère & Beslier, 1995;
495 Rollet et al., 2002). While the magmatism occurring in the margins has been associated to
496 back-arc magmatic activity strongly influenced by subduction (e.g., Coulon, 1977; Bellon,
497 1981; Réhault et al., 2012), the nature of magmatism observed in the oceanic domain remains
498 unknown. It could not be investigated by direct geochemical analysis due to the presence
499 of a sedimentary cover several kilometres thick.

500 Figure 8b shows a stratigraphic log representative of Western Mediterranean oceanic
501 basins that includes P -wave velocity estimates. It is derived from the results of joint
502 seismic wide-angle and reflection profiling in the Algerian and Western Sardinia basins
503 (Klingelhoefer et al., 2008; Gailler et al., 2009; Leprêtre et al., 2013). The sedimentary layer
504 of 5-*km* average thickness and 1.9 - 5 km/s P -wave velocities is made up of Plio-quaternary
505 sediments, Messinian and pre-salt units (Fig. 8b). The transition from Plio-quaternary to
506 Messinian units occurs at $V_p \approx 2.5$ km/s. The Messinian sequence exhibits strong thick-
507 ness variations ascribed to salt diapirism. It is separated from the pre-salt unit by the
508 4.2 km/s velocity boundary. The deepest sediments overlay an oceanic basement that starts
509 at $V_p > 5$ km/s. The oceanic crust is relatively thin with an average thickness of 4.5 km. Its
510 P -wave velocities range from 5 to 7.2 - 7.3 km/s at ~ 12 km depth, which corresponds to the
511 Moho depth ($V_p > 7.3$ km/s in the upper mantle). Controlled-source seismic data are useful
512 to constrain the layer thicknesses and the depths of major interfaces (intra-sedimentary,
513 sediment-crust and Moho), but their interpretation in terms of petrology only rely on P -
514 wave velocity estimates. The interpretation is ambiguous as two lithologies of different
515 petrological natures may have similar V_p (or V_s) signatures. However, P - and S -wave ve-
516 locities can be used jointly to yield information on lithologies and their hydration degree
517 (e.g., Grevemeyer et al., 2018; Malusà et al., 2021). For instance, the V_p/V_s ratio is com-
518 monly used to assess the degree of serpentinization in oceanic domains and margins (e.g.,
519 Bullock & Minshull, 2005; Reynard, 2013; Grevemeyer et al., 2018). We will take advantage
520 of the availability of the V_p cross-section by Dannowski et al. (2020) and our V_s section
521 along the same profile to further constrain the petrological nature of the crust in the central
522 Ligurian-Provence basin.

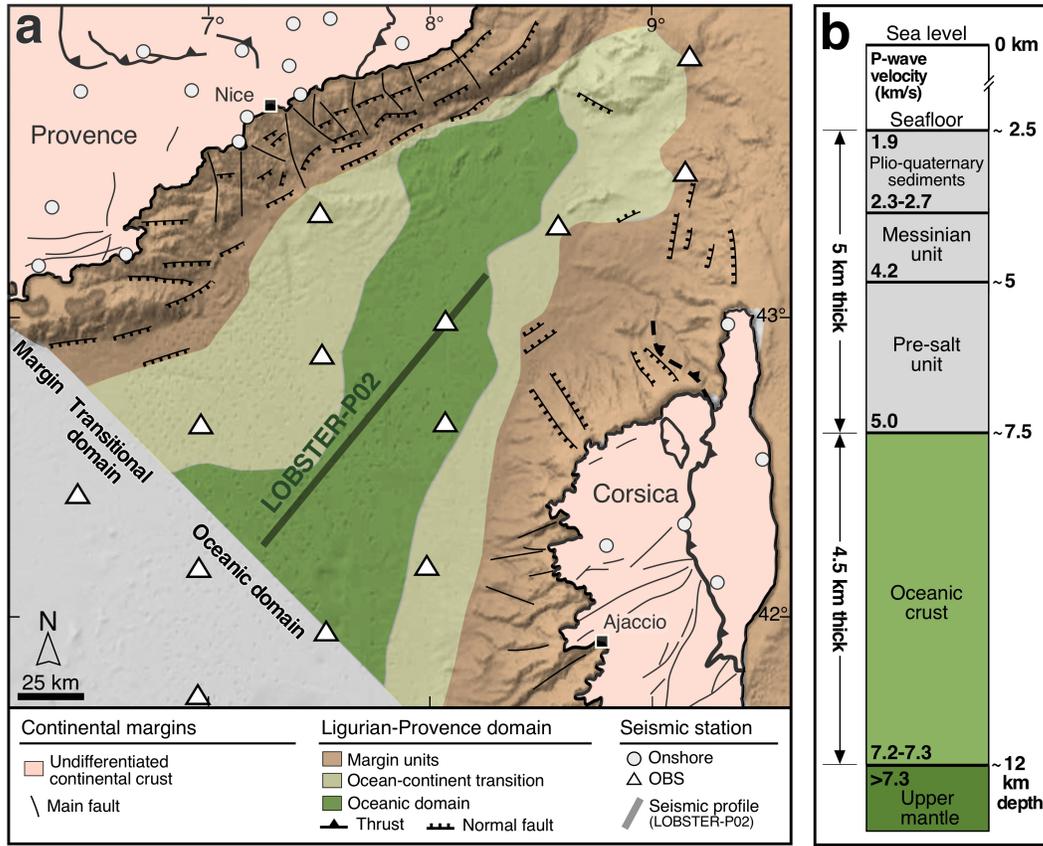


Figure 8. (a) Geological and tectonic setting of the Ligurian-Provence basin and European domains of southeast France and Corsica, showing: (1) continental margins, (2) transitional domains, and (3) the central oceanic domain. Gray line: trace of the seismic profile used in the discussion (Dannowski et al., 2020). Seismic stations are indicated by white triangles (AASN OBSs) and white circles (onshore stations). (b) stratigraphic log showing *P*-wave velocity and thickness of geological units observed in Western Mediterranean oceanic basins (after Leprêtre et al., 2013).

5.2 Seismic velocity cross-sections in the central oceanic domain

Figures 9a-b show vertical sections through our 3-D V_s model and probability of presence of interfaces along the SW-NE transect investigated by Dannowski et al. (2020). The P -wave velocity section of Dannowski et al. (2020) is shown in Fig. 9c.

Figure 9b shows two major layer boundaries at ~ 5 and ~ 12 km depth with rather high probabilities of presence, and a third one of weaker probability at ~ 7.5 km depth. The good correspondence of the shallowest boundary with the velocity contour $V_s = 2.5$ km/s suggests that it is probably an intra-sedimentary interface. This boundary also coincides with the 4.2 km/s P -wave velocity contour (Fig. 9c), which corresponds to the base of the Messinian salt unit according to Fig. 8b. This interpretation is consistent with S - and P -wave velocities, typical of salt (e.g., Yan et al., 2016). The intermediate interface at ~ 7.5 km depth, which is slightly less pronounced than the Moho boundary (Fig. 9b) coincides with the velocity contours $V_s = 3.5$ km/s (Fig. 9b) and $V_p = 5$ km/s (Fig. 9c), which may support its interpretation as the sediment-crust boundary, in agreement with Fig. 8b. However, this interface does not correspond to a marked change in seismic velocity in the final model of Fig. 9a. Therefore, we cannot detect unambiguously the depth of the sediment-crust transition. We will use the proxy $V_s = 3.5$ km/s (or $V_p = 5$ km/s) for the sediment-crust boundary.

In the northeastern part of the transect ($x > 70$ km), the lower crust has higher, but still crustal P -wave velocities (6.0 - 7.2 km/s, in green in Fig. 9c), hence a weaker P -wave velocity gradient at the Moho than in the southwestern part. By contrast, S -wave velocities are high and almost mantle-like (4.0 - 4.4 km/s), with a low S -wave velocity gradient at Moho depth, in particular at the northeastern end of the profile ($x > 90$ km in Fig. 9a). These P - and S -wave velocities are typical of gabbro (Grevemeyer et al., 2018), which suggests a gabbro intrusive body within the oceanic crust. Our interpretation is at odds with Dannowski et al. (2020) who interpreted the high V_p part of the deep crust as hyper-extended continental crust based on gravity modeling. The observed high S -wave velocity rules out the continental crust hypothesis.

As outlined by Dannowski et al. (2020), the $V_p = 7.2$ km/s is a good Moho proxy because it coincides with a very strong velocity gradient. In the southwestern part of the profile ($x < 70$ km), the $V_p = 7.2$ km/s contour closely corresponds to the $V_s = 4.1$ km/s contour while it corresponds to the $V_s = 4.4$ km/s contour for $x > 70$ km, that is beneath the

555 gabbroic intrusion (Fig. 9c). Owing to the presence of the gabbro intrusion, a single S -wave
556 velocity contour cannot be used as proxy for the petrological Moho in the Ligurian-Provence
557 basin, unlike in continental areas (Nouibat et al., 2022).

558 In the southwestern part of the profile, the depths of the V_s (4.1 km/s) and V_p (7.2 km/s)
559 Moho proxies differ by less than 1 km. Such a small discrepancy is remarkable, given that the
560 two models are totally independent. The depth profiles of the two shallower layer boundaries
561 are also remarkably similar to those of the V_p contours that define lithological layering in
562 the western Mediterranean basins (Fig. 8b). Such similarity to the P -wave velocity model
563 of Dannowski et al. (2020) validates the offshore part of our shear-wave velocity model, as
564 similarity to the receiver function section of the Cifalps profile validated its onshore part
565 (Nouibat et al., 2022).

566 In the few locations where information on P -wave velocity is available, the uppermost
567 mantle has the seismic signature of a dry peridotite, with $V_p > 7.2$ km/s and $V_s > 4.0$ km/s
568 (Grevemeyer et al., 2018). We find no evidence of serpentinized mantle, which would show
569 much lower P - and S -wave velocities. Dannowski et al. (2020) and Wolf et al. (2021) also
570 concluded on low mantle serpentinization in this part of the basin axis.

571 6 Conclusion

572 Using data of 23 ocean-bottom seismometers of the AlpArray network with those of 890
573 temporary and permanent onshore stations, we have derived a 3-D high-resolution shear-
574 wave velocity model encompassing the Ligurian-Provence basin and its conjugate margins.
575 The OBS continuous records could be fully exploited after a careful, specific pre-processing
576 scheme including removal of instrument noises (glitches) and reduction of seabed-induced
577 compliance and tilt noises. We enhanced the quality of correlations between OBSs and
578 maximized the path coverage in the Ligurian-Provence basin by involving correlations with
579 onshore stations to virtually reconstruct Rayleigh waves propagating between OBSs. As in
580 Nouibat et al. (2022), we computed 2-D group-velocity maps and their uncertainties using a
581 data-driven transdimensional inversion of Rayleigh-wave group-velocity measurements. The
582 dispersion data and their uncertainties have then been used jointly in a Bayesian probabilis-
583 tic approach to derive a 3-D probabilistic shear-wave velocity model. The output average
584 model was further refined using a linear inversion that accounts for the presence of the water
585 column.

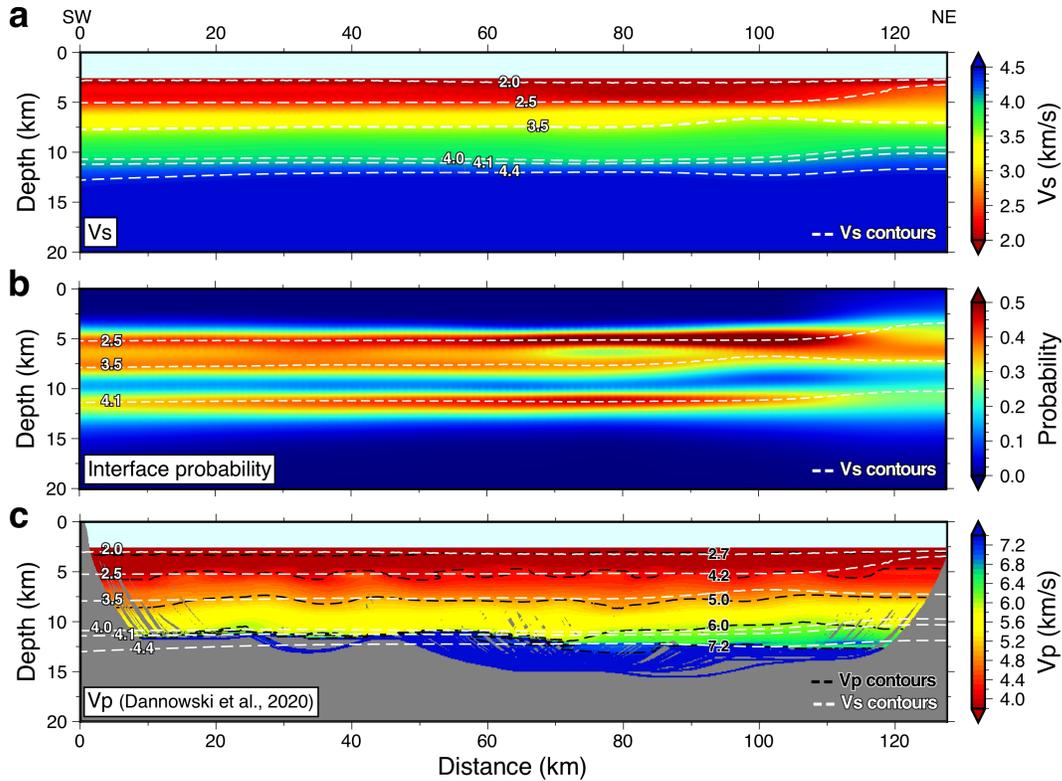


Figure 9. Depth sections along the LOBSTER-P02 transect (location shown in Fig. 8a). (a) Shear-wave velocities from our final model. The 2, 2.5, 3.5, 4, 4.1, and 4.4 km/s V_s contours are shown as white dashed lines. The water column is in sky blue. (b) Posterior probability densities of presence of a layer boundary obtained from the Bayesian inversion. White dashed lines indicate the 2.5, 3.5 and 4.1 km/s V_s contours. (c) P -wave velocities from Dannowski et al. (2020). Black dashed lines indicate the 2.7, 4.2, 5, 6, and 7.2 km/s V_p contours; white dashed lines as in (a).

586 The comparison with the high-resolution P -wave velocity section derived by Dannowski
587 et al. (2020) from travelt ime inversion of controlled-source seismic data along the basin
588 axis validates our 3-D ANT model. Layer boundaries revealed by high probabilities of
589 presence of an interface and V_s contours are remarkably consistent with V_p contours. The
590 joint interpretation of the V_p and V_s models highlights a relatively thin anomalous oceanic
591 crust of low P -wave velocities but rather high S -wave velocities. In the NE part of the
592 profile, the lower part of the crust exhibits a gabbroic intrusive body. The underlying
593 mantle is anhydrous and shows no evidence of serpentinisation. These results show the
594 potential of a joint interpretation of V_p and V_s models since they provide reliable answers
595 to a number of debated questions on the petrological nature of the crust and uppermost
596 mantle of the Ligurian-Provence basin, at least along the LOBSTER-P02 seismic profile.
597 They also warrant the same type of study on the SEFASILS controlled-source seismic profile
598 that crosses the northern margin of the basin (Dessa et al., 2020).

599 The use of OBS recordings in ambient-noise tomography is more challenging than with
600 onshore stations due to shorter recording times, a higher potential of technical problems,
601 sea-floor noises and generally a poorer signal-to-noise ratio in the frequency bands of micro-
602 seismic noise that are key for ambient-noise tomography. Parts of these problems have been
603 solved here by a specific pre-processing of OBS records that reduces instrument and sea-floor
604 generated noises such as tilt and compliance. When available, land stations can be used in
605 combination with OBSs to provide higher quality surface-wave signals between offshore and
606 onshore stations than for OBS pairs, therefore improving interstation path coverage of ANT
607 in particular at the ocean-continent transition. We went a step further by showing how to
608 take full benefit of onshore stations to enhance the quality of correlations for OBS pairs.
609 Our Rayleigh-wave reconstruction scheme for OBS pairs based on second-order correlations
610 between OBSs and land stations has proven to be effective in improving the coverage of the
611 offshore domain. Finally, the transdimensional inversion of the enhanced set of Rayleigh-
612 wave group-velocity observations for group-velocity maps and their uncertainties and the
613 following hybrid inversion for V_s that accounts for the water layer have lead to a high-quality
614 3-D V_s model of the study region. We have therefore set up a complete, efficient and reliable
615 ambient-noise imaging methodology of oceanic domains and their margins using OBSs and
616 land stations that opens new perspectives for the processing of similar datasets.

617 **Data availability statement**

618 Our tomographic 3-D model will be available in the Résif products repository (Réseau
619 sismologique et géodésique Français; <https://www.resif.fr/en/data-and-products/products-repository/>). The inversion code is available on request to the first author.

621 Waveform data used in this paper are available through the EIDA (European Integrated
622 Data Archive) service of ORFEUS (<http://www.orfeus-eu.org/eida/>) and belong to the per-
623 manent networks with codes AC, BE (Royal Observatory Of Belgium, 1985), CA (Institut
624 Cartogràfic I Geològic De Catalunya - Institut D'Estudis Catalans, 1984), CH (Swiss Seis-
625 mological Service (SED) At ETH Zurich, 1983), CR (University Of Zagreb, 2001), CZ
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634 (network code XT, Zhao et al., 2018) and EASI experiments (network code XT, AlpArray
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996 Supporting information

997 Additional Figures and information are presented as the Supplementary Material. Text S1
 998 provides details on the processing applied to remove glitches of instrumental origin from ver-
 999 tical and pressure component records of seven AlpArray seismic network broadband ocean-
 1000 bottom seismometers (AASN BBOBSs). Text S2 describes how coherent signals are detected
 1001 and used to reduce tilt- and compliance-induced noises in vertical-component records of the
 1002 23 AASN OBSs used in this work. Table S1 and Figures S1 to S8 complement sections 2 to
 1003 4 of the main text.