Italian and Alpine three-dimensional crustal structure imaged by ambient-noise surface-wave dispersion

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Key Points.

New 3D vs model of central-europe crust, that correlates well with existing vp models

The model is coherent with tectonic features, and has higher resolution than, e.g., EPcrust

Uncertainties are quantified: the model can be seen as a consensus vs model for the region

Abstract. We derive the 3D crustal structure (S wave velocity) under-
neath Italy and the Alpine region, expanding and exploiting the database

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of ambient noise Rayleigh-wave phase- and group-velocity of Verbeke et al. [2012]. We first complement the database of Verbeke et al. [2012] with a dense set of new ambient-noise-based phase-velocity observations. We next conduct a suite of linear least squares inversion of both phase- and group-velocity data, resulting in 2D maps of Rayleigh-wave phase and group velocity at periods between 5 and 37 s. At relatively short periods, these maps clearly reflect the surface geology of the region, e.g. low velocity zones at the Po Plain; at longer periods, deeper structures such as Moho topography under Alps and Apennines, and lower-crust anomalies are revealed. Our phase- and group-velocity models are next inverted via the Neighbourhood Algorithm to determine a set of one-dimensional shear-velocity models (one per phase/group-velocity pixel), resulting in a new three-dimensional model of shear velocity ($v_S$) parameterized in the same way as the European reference crustal model EPcrust [Molinari and Morelli, 2011]. We also show how well $v_S$ is constrained by phase and group dispersion curves. The model shows the low velocity area beneath the Po Plain and the Molasse basin; the contrast between the low-velocity crust of the Adriatic domain and the high-velocity crust of the Tyrrhenian domain is clearly seen, as well as an almost uniform crystalline crust beneath the Alpine belt. Our results are discussed from the geological/geodynamical standpoint, and compared to those of other, interdisciplinary studies.
1. Introduction

Detailed maps of the seismic structure of the crust are valuable because they reflect the effects of past and present tectonic processes, and because crustal structure must be known if seismic data are used to image deeper, mantle structure or to constrain the parameters that define seismic events. Seismic observations are very sensitive to crustal structure, but are often unable to image it unambiguously. In particular, teleseismic observations are affected by seismic velocities and depth of discontinuities within the lithosphere, in particular the Moho, but it is hard to separate crust and mantle effects by analyzing teleseismic data alone. As a consequence, seismic tomography studies often rely on a priori descriptions of crustal structure, not necessarily based on direct observation [e.g. Arlitt, 1999; Auer, 2014, and references therein]. An inaccurate crustal model may significantly reduce the accuracy of mantle imaging [e.g. Bozdag and Trampert, 2008], and our knowledge of the long-wavelength seismic structure of the crust is still, in many instances, unsatisfactory.

The Alpine and Apennines mountain ranges have been the subject of countless geological and geophysical studies for the last two centuries at least [e.g. Doglioni et al., 1999; Handy et al., 2014, and references therein]. Owing to the mentioned difficulties in crustal imaging, a number of fundamental questions are still open concerning orogeny and mantle dynamics in this complex area [Faccenna and Becker, 2010]. Local-scale crustal models have been derived from geological datasets [Molinari et al., 2015], active reflection and refraction seismic experiments [e.g., Waldhauser et al., 1998; Brückl et al., 2007], P-wave travel-time tomography [e.g., Chiarabba and Amato, 1996; Piromallo and Morelli, 2003;
Serretti and Morelli, 2011; Gualtieri et al., 2014], local earthquake tomography (LET) [e.g., Diehl et al., 2009; Di Stefano et al., 2009], receiver-function studies [e.g., Piana Agostinetti and Amato, 2009], and combinations of the latter two data types [Wagner et al., 2012; Spada et al., 2013]. These methods provide important constraints on three-dimensional (3D) P-wave velocity structure and on the depth of sharp discontinuities. However, they all have intrinsic limitations: the scarcity of seismic events in large regions prevents LET to be applied at a regional scale; observations of receiver functions are limited to station locations and thus relatively sparse; active seismic experiments need 3D migration [Waldhauser et al., 1998]. Hence, these techniques cannot be easily applied at large scales. Even more importantly, none of them constrains S-wave velocity (v_S) directly. Some studies have combined information from different methods, to determine larger-scale crustal models [e.g., Tesauro et al., 2008; Grad et al., 2009; Baranov, 2010; Molinari and Morelli, 2011; Molinari et al., 2012]. This approach requires some subjective assumptions: for instance, information from distantly neighbouring refraction profiles could be interpolated on the base for instance of gravity data, and v_S could be derived by scaling P-wave velocity (v_P) using some standard relations. Crustal v_S-sensitive data are much sparser than v_P-sensitive ones, and current crustal models of v_S are relatively unreliable.

The recent introduction of seismic interferometry has allowed to map crustal v_S by observing surface waves in seismic ambient noise [e.g. Boschi and Weemstra, 2015, and references therein]. As a general rule, surface waves are much more sensitive to v_S than to v_P. While teleseismic surface-wave observations are limited to long periods and thus mostly affected by mantle structure [e.g. Boschi and Ekström, 2002], ambient-noise seis-
mology allows to measure surface-wave dispersion at shorter “epicentral” distances, and
at shorter periods: in practice, this means that ambient-noise surface-wave data are sen-
sitive to crustal and lithosphere structure but not to the mantle. Today, a number of
authors apply ambient-noise interferometry to reconstruct the structure of the crust with
increasing resolution, in regions covered by networks of seismographic stations. For in-
stance, Delorey and Vidale [2011] have been able to refine a 3D $v_S$ model of the Seattle
area for seismic hazard assessment purposes; Zulfakriza et al. [2014] retrieved the upper
crustal structure of central Java (Indonesia) from transdimensional seismic noise tomog-
raphy; Saygin and Kennett [2012] imaged the Australian crust; Kao et al. [2013] imaged
the crust and upper mantle $v_S$ structure of Canada and adjacent regions; Lin et al. [2008]
and Ekström [2014] have studied North American crustal structure via phase- and group-
velocity maps; preliminary ambient-noise studies of Alpine crustal structure have been
conducted by Stehly et al. [2009] and Fry et al. [2010].

Verbeke et al. [2012] compiled a large database of surface-wave dispersion measurements
from seismic ambient noise in the Alpine and Italian regions. Their group-velocity maps
at periods between 5 and 40 seconds reveal velocity anomalies with high detail, not ex-
ploited yet to define a 3D crustal model. These data have strong sensitivity to crustal $v_S$
structure, but weak sensitivity to discontinuities and their depth. The geometry of crustal
seismic discontinuities under Italy and the Alps, however, has been determined in a num-
ber of studies using other observation techniques: model EPcrust [Molinari and Morelli,
2011] includes robust interface geometry determinations based on seismic reflection and
refraction studies; [Spada et al., 2013] have constrained Moho depth via controlled-source
seismology, local earthquake tomography and receiver-function analysis; the 3D structure
and depth of the deep Po-plain sedimentary basin has been determined from geological data [Molinari et al., 2015].

In this study, we derived the 3D crustal structure (S wave velocity) underneath Italy and the Alpine region, expanding and exploiting the database of Verbeke et al. [2012]. We first derive new maps of fundamental-mode Rayleigh-wave group- and phase-velocity, making use of phase-velocity observations that were not available to Verbeke et al. [2012], and enhancing lateral resolution significantly. The study of Verbeke et al. [2012] was limited to phase- and group-velocity maps; here we infer 3D $v_S$ structure using a fully nonlinear direct-search inversion algorithm [Wathelet, 2008]. This method allows for a priori information other than seismic dispersion curves to be taken into account in the inversion: we thus combine ambient-noise surface-wave data with the above-mentioned, independent observations of discontinuity depths, which surface waves are only weakly sensitive to; in practice, the resulting 3D map can be seen as a model integrating information on receiver functions and surface wave.

In the following, we first describe the linear tomographic inversion that provides phase- and group-velocity maps; we next illustrate the results of the nonlinear inversion for $v_S$, including an assessment of uncertainties; finally, we discuss our results from the geological/geodynamical standpoint, and compare them to those of other, interdisciplinary studies.

2. Rayleigh-wave phase velocity maps from ambient noise

Verbeke et al. [2012] have built a database of Rayleigh wave group and phase velocity measurements from seismic ambient noise extending across Italy and the Alps towards Central Europe. Following Bensen et al. [2007], they cross-correlated and analysed one
year (2008) of continuous vertical-component seismograms recorded by pairs of stations belonging to the Italian, Swiss and German permanent broadband networks (Fig. 1). For each station pair, surface-wave group velocities have been measured applying the time frequency analysis (FTAN) method \cite{Ritzwoller and Levshin, 1998}, while phase velocities have been obtained following the two-station method as implemented by Meier et al. \cite{2004}. The accuracy of phase-velocity measurements was verified by Boschi et al. \cite{2013}, who validated the method used by Verbeke et al. \cite{2012} against an independent approach based on frequency-domain cross-correlation.

Our database is based on that of Verbeke et al. \cite{2012}. With respect to the work of Verbeke et al. \cite{2012}, group-velocity data are exactly the same, while phase-velocity dispersion curves have been entirely recomputed from the original cross-correlations, to include phase-velocity measurements made at a denser and broader set of frequencies as illustrated in the following. With respect to group velocity, phase-velocity data sample a wider depth range and are less contaminated by interfering phases, because measurements are made within a narrower time window \cite[e.g. Boschi et al., 2013]{2013}. On the other hand, group velocities are more sensitive to relatively shallow depth and are easier to measure.

We take advantage of both data types to constrain 3D crustal structure at the highest possible resolution. After measuring dispersion as described between all available station pairs (Fig. 1), we derive phase-velocity maps between 5 and 37 seconds periods and group-velocity maps from 8 to 35 seconds.

2.1. Phase- and group-velocity imaging method

Assuming, as is done in most seismic ambient-noise literature, that the effects of non-uniformity in noise source distribution can be neglected, an ambient-noise database is
equivalent to a "traditional" one, with earthquake sources replaced by seismic stations acting as "virtual" sources \cite{Boschi and Weemstra, 2015}. We set up a linear system in the ray-theory approximation as described e.g. by Boschi and Dziewonski \cite{1999}. Based on the available ray coverage, we parameterize the region of interest in terms of $0.25\degree \times 0.25\degree$ cells, independent of period and for both phase- and group-velocity. We take the phase- and group-velocities predicted by PREM \cite{Dziewonski and Anderson, 1981}, period-dependent but laterally homogeneous, as reference values, and determine phase- and group-velocities through a suite of least-squares inversions (one per period for both group and phase data) via the iterative LSQR algorithm \cite{Paige and Saunders, 1982}.

LSQR approximates the exact least square solution, converging, according to a rigorous "stopping criterion" \cite{Paige and Saunders, 1982}, after some tens of iterations. In order to obtain a stable solution, we regularize the inverse problem via the roughness minimization scheme of Boschi and Dziewonski \cite{1999}, which is equivalent to requiring the solution model to be smooth. We do not apply norm minimization because, in cases of nonuniform data coverage, this form of regularization can result in solution models with unnaturally large velocity gradients Boschi and Dziewonski \cite[e.g. 1999]{}. We carry out many tentative inversions with different regularization parameter values until a satisfactory solution is achieved, i.e.: (i) variance reduction of the data is relatively high; (ii) using resolution tests (sec. 2.2) as a reference, we verify that there are practically no features in our maps at wavelengths shorter than our target resolution. We verified that solutions found in this way correspond to the corner region of the L-curve as defined by Hansen \cite{1992} and shown in Fig. 2: the roughness damping value is 0.75 for both phase and group velocity and for all periods. As can be expected, variance reduction changes as a function of surface-wave
period. Highest variance reductions (up to ∼ 75%) of both group and phase data are achieved in the 8-25 s period range; at periods <8 s and >25 s variance reduction does not exceed 50%.

2.2. Resolution test

We quantify the horizontal resolution of phase- and group-velocity allowed by our dispersion data sets via synthetic tests [e.g., Boschi and Dziewonski, 1999]. Following Husen et al. [2009] and Verbeke et al. [2012], we choose as ”input” model a random 2D map of velocity anomalies (Fig. 3) whose size and distribution is statistically similar to typical seismic maps at this scale length. This is preferable to unrealistic though widely used ”checkerboard” input models who tend to yield too optimistic estimates.

The input model of Fig. 3 is characterized by anomaly values ranging between −10% and +10% with respect to PREM [Dziewonski and Anderson, 1981], filtered via 2D Fourier transformation to isolate the wavelengths of interest, resulting in anomalies of spatial extension > 50 km and < 200 km. We calculate phase and group delay times predicted by this model for all ray paths and periods in our database (station coverage depends on period and on whether phase or group velocity is measured); we add Gaussian noise to the data (the standard deviation of random noise is the same as that of the data, \( \sigma = 0.2 \text{km/s} \)). Finally, we solve the two inverse problems — for phase and group velocities — using the inversion algorithm and regularization scheme and parameters that we apply to our real data set. The results associated with 7, 12 and 20 s Rayleigh-wave phase velocity data and with 8, 16, 24 s group velocity data are shown in Fig. 3. We only show our solution in cells sampled by more than 5 rays, estimating that less well sampled pixels might not be sufficiently well constrained. As a general rule, the test model is recovered
fairly well, within the resolved area, independently of period; group velocity, for which we have a denser ray coverage (see maps in Figure 3), is recovered more accurately than phase velocity; amplitude of both phase- and group-velocity anomalies is underestimated. At short periods (up to 14 seconds), phase velocity is resolved well in Switzerland, Southern Germany, Western and Eastern Alps and North-Western Apennines; at longer periods phase-velocity resolution becomes acceptable also in North-Eastern, Central and Southern Italy. Group-velocity resolution is fairly good throughout the region of interest, with the exception of Western Tyrrhenian Sea, Sardinia, Corsica and Southern Adriatic sea.

We anticipate that, as a consequence, the resolution of our final 3D model changes as a function of location (and not only in depth, due to differences of depth sensitivity for phase and group velocity at each period, see Section 2.3): in areas where short period surface-wave velocity is not constrained by either group or phase dispersion observation, shallow crustal structure is underconstrained (see Section 3.2).

2.3. Rayleigh-wave phase- and group-velocity maps

We show in Fig. 4 the phase and group velocity maps obtained from our data sets at periods between 5s and 37s, and 8s and 35s, respectively. Long-period surface waves sample deeper into the Earth and are correspondingly characterized by higher velocities than short-period ones, as noted in Fig. 4. As a first order approximation, surface waves of 7s image structure from surface down to 10 km depth, those of 20s period are most sensitive to structure between 10km and 30km, and those of 37s period between 35km and 55km depth [Warren et al., 2013]. At any given period, group velocity is sensitive to structure at shallower depth than phase velocity.
The longer wavelength patterns of the maps in Fig. 4 are in agreement with the lower-resolution maps of Verbeke et al. [2012]. At short periods we clearly recognize the effects of shallower geological features. Low-velocity anomalies (~2.3 km/s for phase and < 1.5 km/s for group velocity) can be associated to sedimentary basins in the Po plain, known to be characterized by a low seismic velocity foredeep sedimentary basin that reaches a depth of 7-8 km and to a lesser degree in the Swiss and German Molasse basin (visible especially well in the 8s group velocity map). As expected for shallow crustal levels, the Alpine belt exhibits higher velocities than the Apennines, indicating the presence of shallow high-velocity crystalline structure in the Alps.

At intermediate periods (12s for phase and 16s for group velocity) the signature of the Po Plain is still clearly visible, suggesting that deep and slow sedimentary basins affect the dispersion curves at longer periods more than estimated by theoretical maximum sensitivity. This means that, in order to get a realistic velocity 3D model in this region, we need to constrain those basins with a good a priori model. Using a finer parameterization grid, we recover the Ivrea body high-velocity zone at lon ~ 8° and lat ~ 44.5° (Figure 3). Lower phase and group velocities are found along the Apennines than along the Alps, presumably owing to the different properties of rocks forming the uppermost part of the crust under those two mountain ranges: the Apennines are predominantly characterized by Adriatic consolidated sediments with a low degree of metamorphism, as opposed to relatively high-velocity crystalline upper-crustal and metamorphic basement rocks under the Alps. High velocities appear mainly over the Tyrrhenian Sea, as a result of the shallow Moho depth in this oceanic region.
Longer periods (20s and 32s for phase; 24s and 35s for group velocity) are most sensitive to lateral variations of lower crust, Moho depth and uppermost mantle, depending on location. The contrast between Tyrrhenian and Adriatic crust is strong (about 0.3-0.4 km/s), but this effect is explained to a large extent in terms of Moho depth, the Tyrrhenian crust being anomalously thin with respect to the rest of the region of interest. At periods > 30s, we can clearly distinguish the deep crustal root along the inner parts of the Alpine orogen and beneath the Apennines, and the low-velocity anomaly proposed by Di Stefano et al. [2009] as the "Central Apennine" window, based on P-wave travel-time tomography.

3. Inversion for 3D structure

At each cell of our phase-/group-velocity maps, we construct fundamental-mode Rayleigh-wave dispersion curves, ranging from 5 to 37 s period in phase velocity, and 8 to 35 s in group. We subsequently invert each combination of phase- and group-velocity dispersion curves, to determine isotropic \( v_S \), \( v_P \) and layer thicknesses. Each inversion consists of a stochastic direct search, implemented via the "Neighbourhood Algorithm" [Sambridge, 1999a; Wathelet, 2008]. In this approach, the solution space is sampled sequentially and non-uniformly, taking into account the data fit achieved by old samples when generating new ones: in practice, while possible solutions are sought randomly, the search is biased towards regions of the solution space were better-fitting solutions have been found [Wathelet, 2008]. The cost function is the L2-norm of the difference between the observed dispersion curves, and those calculated from a solution model.

This approach is suited to the inverse problem under consideration because of its inherent nonlinearity. It has the advantage of widely exploring the admissible solution space, without anchoring the solution model to the reference one as would happen in a linearised
inversion. However, the strong nonlinearity of the problem and the averaging properties of
the dispersion measurements can always result in significantly biased solutions, especially
in combination with deep sedimentary basins or wherever particularly strong velocity dis-
continuities are present. Naturally, where data coverage is poor, nonlinear inversion can
result in strong depth variations clearly incompatible with known geological features [Say-
gin and Kennett, 2012]. As mentioned above, the main weakness of surface-wave inversion
is the lack of sensitivity to the depth of seismic interfaces. Following Wathelet [2008], we
therefore define the boundaries of the solution space to be sampled (i.e. the range of
possible values for all parameters we invert for), on the basis of a priori information from
independent geophysical and geological data. A priori constraints on the model search
are important to speed up the inversion by limiting the volume of model space searched
and to define what we judge to be physically plausible candidate models.

The vertical parametrization of our 3D models is the same as that of EPcrust, including
three crustal layers plus two 25-km thick mantle layers. The three crustal layers are
defined as the following: a top layer which can be either sediment in the sedimentary
basins (Po Plain, Molasse basin and Ligurian Sea basin), a stack of sedimentary thrust
sheets in i.e Apennines or upper-most crustal like material; a middle layer corresponding
to the crystalline upper crust; a lower layer corresponding to the lower crust. $v_S$ in each
layer is described with a linear gradient, obtained by subdividing each layer into five sub-
layers. The water layer is not taken into account: this simplification, inherent limit of our
inversion software, might cause biased results under Tyrrhenian Sea. Since our study is
focused on continental structure, we think it is reasonable to ”neglect” this layer for the
time being. The allowed parameter ranges in each layer are defined based on the most
recent crustal models and Moho maps of the region of interest. Upper- and lower-crustal
parameters (especially $v_P$ and density) should be in the same range as EPcrust values;
velocities in the mantle should be similar to EPmantle values; Moho depths are required
to be close to those given by Spada et al. [2013]; in the Po Plain, sediment-layer thickness
and velocity should resemble those provided by the MAMBo model [Molinari et al., 2015].
Outside the Po Plain, we allow the top-layer thickness to vary between 1-5 km. The exact
variability range of each parameter is given in Table 1.

Each inversion consists of a nested exploration of the solution space, carried out in 140
steps; at each step, the 110 best-fitting solutions are kept. In summary, 35,400 possible
solution models are evaluated at each cell. In Fig. 5, the best-fitting 15,400 models are
plotted for a few sample cells, illustrating the range of possible models sampled by the
inversion, and the goodness of fit they achieve. The similarity between best and mean
models in Fig. 5 reflects the velocity resolution afforded by our database: the solution is
generally robust (similar to the mean) except for the first few kilometers, and for the top
of the mantle.

We run our algorithm once for each of the $0.25^\circ \times 0.25^\circ$ pixels of our dispersion maps.
The root mean square (RMS) of the misfit is shown, as a function of period, in Fig. 6,
where we compare the fit achieved by crustal models EPcrust and CRUST2.0 [Bassin
et al., 2000] and by our final best model. We infer from Fig. 6 that inversion improves
data fit significantly with respect to existing regional models. The improvement of fit is
particularly remarkable at periods $< 15$ s, where both EPcrust and CRUST2.0 show large
deviations from measured data. This could be ascribed to inaccurate EPcrust/CRUST2.0
estimates of uppermost crustal layer, at least in some regions. Importantly, EPcrust and
CRUST2.0 (i) are continental-/global-scale models, not designed to be employed at the scale of this study, and (ii) they were derived from observations independent of surface waves (whether earthquake or noise-generated). The fact that EPcrust and CRUST2.0 reduce the variance of our ambient-noise data at most periods is a nontrivial result, and a significant validation of those models.

Fig. 6c shows the geographical distribution of model misfit. The misfit is highest at sedimentary basins and at North-West border of our region of interest. The overall misfit strongly depends on whether a good fit is achieved at the shortest periods, where poor estimates of top crustal layer thickness and velocity and coarse parameterization can easily deteriorate it, and at the longest periods in consideration, where ambient-noise-based measurements are scarcer.

Our final best-fitting model ultimately accounts for a broad variety of data; it strongly improves the fit of surface-wave dispersion observations and additionally accounts for a-priori knowledge of crustal structure derived by other independent methods; it is overall an important upgrade of EPcrust, in particular as far as \( v_S \) structure is concerned.

### 3.1. Local velocity profiles

In Fig. 5 we select examples of nonlinear \( v_S \) inversions involving a diverse set of geological settings. Our group- and phase-velocity data, combined with the mentioned a priori information on the depth of the main sedimentary basins and, in particular, of the Moho depth, allow us to map \( v_S \) crustal structure with higher resolution than previously possible. The lack or poor quality of data at very short periods (<5s) may lead to near-surface velocities that are poorly correlated with geology, but this effect is marginal,
owing in particular to the careful a-priori constraint introduced in our inversion strategy as explained above.

We noted that both phase- and group-velocity data are generally well explained by our model. In areas where crustal structure is particularly complex (e.g., deep sedimentary basins, Moho offsets at plate boundaries), the data fit is occasionally poor: the problem might be solved by means of a local, ad-hoc, finer parameterization, but this would be extremely time-consuming and is beyond the scope of this study.

Fig. 5b corresponds to the Northern border of the Molasses basin, and is representative of European lithosphere outside the orogenic belts. We find here a crustal thickness of about 25km, with an almost uniform upper crust and a large velocity gradient in the lower crust. The Moho appears to be relatively shallow, but a closer look at Fig. 5b shows that many solution models exist that fit the data about equally well.

In the Alpine orogen cell (Fig. 5c) our model is characterized by a relatively high near-surface velocity of about 3.15km/s reflecting the granitic basement rocks exposed at surface after intensive uplift and erosion. The Moho is more than 50km-deep [Schmid and Kissling, 2000], near the lower limit or resolution by our data. That the Moho is anomalously deep could be directly inferred from the phase-velocity dispersion curve, where the kink typically associated to Moho depth [Lebedev et al., 2013] is not found, presumably because it takes place at longer periods, outside our measurement range.

The south-west Alpine foreland cell (Fig. 5d) is located at the Southern edge of the Ivrea Body [Solarino et al., 1997], and shows very low surface velocity of about 2.3km/s and velocities of about 3.4km/s in the whole crust, in good agreement with known near-
surface structure. Imaged Moho depth is relatively shallow (20km), corresponding to the Ivrea mantle upwelling [Spada et al., 2013].

In the Po Plain, along the Adriatic coast (Fig. 5e), Tertiary and Mesozoic sediments reach 7-km thickness [Molinari et al., 2015]. Their effect is visible in the corresponding dispersion curves up to 25-30s period; group velocity is particularly low, i.e. ~2 km/s at period < 15 s. We emphasize that, in this anomalous location, our a priori definition of the solution space plays a particularly important role: we have only explored solution models that included a thick sediment layer. Preliminary inversions had shown that, in the absence of such constraint, group- and phase-velocity anomalies would have been explained in terms of very low $v_S$ throughout the crust. Despite that, a broad range of plausible solution models are shown in Fig. 5e, suggesting that this remains a relatively poorly constrained location that could benefit from further seismic observations of all kinds. Besides the thick sediment layer, our best profile at this location includes upper crustal basement rocks of cumulative thickness > 15km, overlying rather average Adriatic lower crust.

The last cell is representative of the Tyrrhenian crust (Fig. 5f). At this location the Moho is relatively deep (~ 25 km); the upper crust is characterized by a strong velocity gradient; the middle-lower crust is approximately uniform.

3.2. Model variability

Before discussing 3D $v_S$ variations, we evaluate how robustly they are constrained by our procedure. Non-linear, direct-search inversion techniques — such as the Neighbourhood Algorithm used here — allow the exploration of the model space, guided by explicit a priori informations that designate physically plausible models on the basis of supple-
mentary data. This approach can thus provide an assessment of the range of ‘possible’
models — i.e. those that reach an acceptable fit to observations — in a more general
fashion than linearised standard error analyses. While a rigorous model appraisal, geared
for accurately picturing posterior probability density functions in a Bayesian sense, may
often be an expensive task [Sambridge, 1999b], the suite of models tested during the
search stage provides a fair picture of model uncertainty. Our non-linear search technique
[Sambridge, 1999a; Wathelet, 2008] generates an ensemble of acceptable models that may
be taken to represent the posterior probability distribution function (PDF) of the earth
structure reflected by the observations. We limit our estimation of model variability to
this approximate PDF representation.

The marginal velocity probability density distribution of a single model parameter (i.e.,
shear-wave velocity at some particular depth in vertical $v_S$ profile) appears well behaved,
not far from a Gaussian. This can be verified in Fig. 7, that shows the distribution
of $v_S$ values at three fixed depths for a sample geographical location (cell e) in Fig.
5). The a priori PDFs — in each case, a boxcar representing a range of variability
with uniform probability density — is also plotted, showing that significant improvement
in information is reached by the inversion. Note that we sample wider velocity ranges
at shallower depths, consistently with physical expectations of heterogeneity. Posterior
marginal PDFs taper off near the ends of the sampled ranges, indicating that they are
wide enough and compatible with information contained in the data. We may claim that
the a priori distributions, being flat and non-informative, do not influence results, or the
shapes of posterior PDFs. As posterior PDFs resemble Gaussians (Fig. 7 a), the standard
deviation $\sigma$ is a convenient way to represent their width, hence uncertainty. Therefore, in
Fig. 7 b), c), and d), we plot maps of the standard deviation at three sample depths. We note that standard deviation is generally rather small, typically less than, say, 150 m/s, and only seldom exceed 200 m/s. Larger values of $\sigma$ can be due to weak data constraints, and this is the case of dark pixels in the eastern Po Plain region. Complex crustal structure in the deep Po Plain sedimentary basin may also be a cause of larger spread of velocity values of well fitting models. Yet another case of larger indetermination may occur when the interface between two layers is placed near some specific depth, and the model $v_S$ jumps from values of one layer to those of the next one. These maps picture the uncertainty associated to our best solution, in the inversion of shear-wave velocity profiles to fit local Rayleigh phase and group dispersion curves. We recall that the quality of fit of the dispersion maps to observed data is depicted by Figure 6 showing the geographical variation of global RMS misfit.

3.3. S-wave velocity structure

The maps of $v_S$ anomalies associated with our best-fitting model are shown in Fig. 8. At each depth, the solution model is only shown in cells that are sampled by at least 5 ray paths for which both phase and group velocity measurements have been successfully obtained at 16s period. We conservatively estimate that less well sampled cells are not sufficiently robustly constrained. The well resolved cells at each depth depends on the surface wave velocity coverage (Figure 3), at periods whose sensitivity is maximum at that depth (Warren et al. [2013], Fig. 7b). As discussed in the previous section, we do not have sufficient resolution to resolve the structure of shallow sedimentary layers, which are therefore controlled, in our inversions, by a priori constraints provided by geological models.
The 3-km deep section in Fig. 8 shows the uppermost part of the crust including the sediment layer. The Po basin and the Ligurian-Sea basin are the most prominent features of this image. The a-priori $v_S$ variability range is from 1.2 to 3.2 km/s and the inversion results show a $v_S \approx 2$ km/s, in good agreement with Molinari et al. [2015]. The deeper part of the Molasse basin is also well marked and shows a higher velocity (2.5 km/s) than Po basin highlighting the different evolution of the two Alpine foreland basins. The Alps and Apennines shallow velocities appear to be consistent with the current surface geology knowledge: high $v_S$ in the Alps indicating the presence of crystalline and metamorphic rocks and consolidated sediment. The Tyrrhenian back-arc basins, that has started forming in the late Miocene, is partially imaged by our tomography and shows $v_S$ around 2.4 km/s.

The 10-km deep section in Fig. 8 shows a strong variability in the upper crust: distinct difference between the Alp and its Northern foreland with a uniform $v_S$ of about 3.2 km/s while South of Alps a the same depth we find the various lithologies from the deep Po Basin sediments ($v_S \approx 2.8$ km/s) to uppermost crystalline crust in central Apennines ($\approx 3.0$ km/s). The low $v_S$ values corresponding to the Po Plain and the North and Central Apennines can be caused by ”smearing” in depth of the low surface wave velocity anomaly from 5 to 20 seconds (Fig. 4). Within each domain, $v_S$ does not show strong 3D variations but highlights the difference in composition (and in origin) of Alpine and Apennines upper crust. In the southern Tyrrhenian, the crust is oceanic, the Moho depth is known to be about 10-15 km depth and low S-wave velocity values found at 10 km depth is consistent with a large percentage of partial melting and high temperature.
The 20-km deep section shows the lower crust, both in the Tyrrhenian and Adriatic domain, and the upper mantle in the Tyrrhenian sea ($v_S \sim 4.0$ km/s). Alps and Apennines show up very clearly in our imaged structure. Beneath southern Germany and northeastern Italy, $v_S$ is as high as 3.6-3.9 km/s. Moreover the contrast between relatively high $v_S$ under the Tuscan Apennines and low $v_S$ under the Marche region is in agreement with new interpretations based on the CROP03 profile [Pauselli et al., 2006], which samples that area. Ligurian upper mantle is visible as high shear-speed beneath the South-Western Alps and western-most part of the Po Plain.

At 35km depth, the geographic pattern of $v_S$ heterogeneity is correlated with that of Moho topography. $v_S$ values typical of the lower crust (3.6 -3.9 km/s) are still found along the Alpine belt and the Northern and Central Apennines; besides that, $v_S$ is within the range of typical upper mantle values. However, with respect to their surrounding areas, Central Apennines have relatively low $v_S$, i.e. between 3.2-3.5 km/s; this confirms published observations of anomalously low lower-crust $v_P \approx 5.8$ km/s in this region, and thus the requirement of anomalously high lower-crust temperature, given the granitic composition [Chiarabba and Amato, 1996; Chiarabba et al., 2009; Di Stefano et al., 2009].

Structure at larger depth (> 50 km) is not resolved well by our dataset. Our dataset does not allow to resolve the complex geometry of subduction zones, which should be discussed in future work involving longer-period surface-wave and/or body-wave travel-time data.

Our inversion provides also an updated Moho map since we allow crustal layer thickness to vary up to 10 % with respect to the total crustal depth from the Moho of Spada et al. [2013]. The allowed variation is roughly within the error bar specified in the a-priori Moho. Although surface waves are only marginally sensitive to sharp discontinuities,
some adjustments with respect to the a-priori Moho are expected. We show in Fig. 9 the comparison between the map of Moho depth resulting from our inversion, the EPcrust one, and the one obtained by Spada et al. [2013] via a combination of CSS, LET and RF. Our inversion and our dataset result in Moho depths from 10 to nearly 60 km, with the highest values found along the entire Alpine arc and locally beneath the Northern Apennines. These depth estimates are generally higher than those of EPcrust (Fig. 9-b), and in better agreement with those of Gualtieri et al. [2014]. The boundary between the Adriatic and Tyrrhenian tectonic domains appears to be defined well, with an average Moho depth of about 35-40 km in the former and about 20-30 km in the latter. Crustal thickness grows from East to West across the Western Alps. The inversion of ambient-noise data has "sharpened" the Alpine Moho with respect to EPcrust, shifting it to an overall larger depth; in the Po Plain area and throughout Adria, our final Moho appears closer to that of EPcrust. Finally, the imaged thin Ligurian crust is similar to the results of Spada et al. [2013].

Vertical sections (Fig. 10) of our final model reveal several prominent tectonic and geological features. These sections provide an overall view on the crustal structure and highlight the different crustal domains present in this complex region. Our model sections are not smoothed in order to well identify parameterization and to not over-interpret our results. To facilitate the discussion, on top of each section, we plot the a-priori Moho depth and, for the AA' profile (NFP-20 West section), we also show the correspondent section from Diehl et al. [2009] and Gualtieri et al. [2014] P-models. This profile is representative of the Western-Central Alps and cross the northern part of the Ivrea body [Hunziker and Zingg, 1980; Rivalenti et al., 1975]. This feature is present in all the three works at...
about 200-240 km along the section and at 20-40 km depth. The similarity of models
in Fig. 10 is particularly significant if one considers that they have been derived using
completely different data and modelling techniques, and suggests that our overall picture
of this region's structure is robust. The AA' profile documents the strong variation in
crustal thickness between the European crust (25-30 km), the Alpine crustal root (55
km), the strong intrusion of Ivrea body (16-20 km), the Adriatic domain (beneath the
Po Plain) with 40 km of total crustal thickness and the Tyrrhenian one (25-30 km).
Within the crust, the sediments and upper and lower crustal domain are clearly marked.
The Po plain is a prominent feature; as explained above, the a priori model in this area
is characterized a thicker-than-average sediment-layer thickness, which compensates the
lack of short-period data; \(v_S\) is, however, a free parameter of the inversion (see Table 1),
and, importantly, \(v_S\) values consistent with geological expectations are found within the
layer. The BB' section is a N-S profile across the (western part of the) Eastern Alps, east
of the Giudicaria line and goes from the Pannonian basin to the Adria-Tyrrhenia Moho
suture zone, crossing the Alps, the deepest part of the Quaternary and Pliocene sediment
in the Po Plain and the northern Apennines. The contrast between upper and lower crust
is less pronounced in the Alpine region, indicating an almost uniform crust. The CC' and
DD' sections (the latter correspond to the EE' section by [Di Stefano et al., 2009]) show
the Adria-Tyrrhenian Moho suture zone at an angle almost perpendicular to the strike
of the plate boundary and illustrate the very large Moho offset that is an expression of
strong ongoing tectonics.
4. Conclusions

We have derived a new 3D \( v_S \) model that fits surface wave group and phase dispersion curves between 5 and 37 s. Virtually all other relevant data on the Italian and Alpine crust at the scale length of interest have been accounted for, in the form of an a-priori model. Especially at high frequency, our model leads to a significant reduction of surface-wave data variance, with respect to some recent crustal models. We found that, at periods between ~20 and ~40s, both EPcrust and Crust2.0 fit our ambient-noise data fairly well, with the higher-resolution model EPcrust achieving a systematically better fit. The fact that EPcrust fits a data set that was not employed in its derivation is an important proof of its reliability.

To obtain our model, we have first inverted the database of Verbeke et al. [2012], to find Rayleigh-wave fundamental-mode phase- and group-velocity maps at periods from 5 to 37 s, and from 8 to 35 s, respectively. We have then applied a direct-search method (Neighbourhood Algorithm) to obtain the \( v_S \) structure under the region of interest with a nominal resolution of \( 0.25^\circ \times 0.25^\circ \). Our final model (i) is a complete and consistent model, including all the most relevant elastic parameters \((v_P, v_S, \rho)\), (ii) can be considered an improvement of existing reference models like EPcrust in the Alpine and Northern Italian areas, (iii) has the same vertical parametrization of EPcrust and (iv) incorporates, as a priori information, the most recent progresses in terms of crustal structure from independent methods (LET tomography, receiver function studies and sedimentary basin modeling).

Our visual comparison with published results indicates that our model of \( v_S \) as a function of depth is physically consistent with current knowledge of \( v_P \) structure in the same region,
independently obtained by other seismic methods. Tomographic maps presented here reflect the complexity of the crustal structure of the region and are highly correlated with the known surface geology features of this portion of the European crust: our 7-10s maps are reminiscent of the geographic distribution of deep sedimentary basins, such as the Po Plan and the Molasse Basin; deeper features, including the topography of the Moho and the three-dimensional structure of the crust under the Apennines, the Alps, Adria and the Thyrrenian sea are resolved well by our data.

Our final preferred model consists of three layers (sediments, upper crystalline and lower crust); within each layer, the seismic parameters are described with a linear gradient. This model is suited to many applications, such as wave propagation modelling at regional scale, crustal correction in tomography, gravity studies, dynamic topography inference.

Relatively poor data fit at periods below 10s suggests that the shallowest part of the crust, i.e. the sediment layer, is not as robustly constrained by the data as the deeper structure. Lack of resolution in shallow layers could result in trade-off and thus loss of model quality for deeper structure. In a future study, the European Plate reference crustal model may be further improved by (i) replacing its current sediment information with more detailed sediment models, and (ii) replacing its current Moho map with one of the new generation of Moho maps. In particular the sediment layers could be obtained by the integration of other sources of information, such as seismic reflection/refraction profiles, geological maps, borehole data will lead to a better definition of the shallowest properties and discontinuities within the crust.

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available at http://www.bo.ingv.it/eurorem/EPcrust; CRUST2.0 model is available at http://igppweb.ucsd.edu/gabi/crust2.html; MAMBo model of the Po basin is available at http://www.bo.ingv.it/molinari/MAMBo.

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References


Table 1. Setting of the variability range of the parameters in the inversion with respect to the a priori informations: MAMBo model [Molinari et al., 2015] in the Po sediment, EPcrust velocity range in the crust and EPmantle in the mantle.

<table>
<thead>
<tr>
<th>Thickness</th>
<th>$v_S$ (km/s)</th>
<th>$v_P$ (km/s)</th>
<th>Density (g/cm$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Top layer – sedimentary basins</td>
<td>±5%</td>
<td>1.2 – 3.4</td>
<td>3.3 – 5.5</td>
</tr>
<tr>
<td>Upper crystalline crust</td>
<td>±10%</td>
<td>2.8 – 3.7</td>
<td>6.2 – 7.0</td>
</tr>
<tr>
<td>Lower crust</td>
<td>±10%</td>
<td>3.1 – 4.0</td>
<td>6.8 – 7.5</td>
</tr>
<tr>
<td>1$^{st}$ mantle layer</td>
<td>fixed</td>
<td>±10%</td>
<td>fixed</td>
</tr>
<tr>
<td>2$^{nd}$ mantle layer</td>
<td>fixed</td>
<td>±6%</td>
<td>fixed</td>
</tr>
</tbody>
</table>

Figure 1. Stations where the ambient noise analyzed in this study was recorded.
Figure 2. L-curve resulting from phase (red crosses) and group (blue circle) inversions for 16s period. Selected models are marked by arrows and labelled by the corresponding value of the roughness-damping parameter [e.g. Boschi and Dziewonski, 1999]. The image roughness is defined as the squared modulus of the dot-product of roughness damping matrix times vector of model coefficients. It is normalized dividing by the total sum of squared model coefficients. Following e.g. Boschi [2006], we show here misfit defined as one minus variance reduction.
Figure 3. Results of the reconstruction test with randomly distributed velocity anomalies as input. The left panels show the input models used while the other columns show retrieved models at different periods as indicated for phase (top row) and group (bottom row) velocity. We highlight the well resolved area with the grey line for each shown period.
Figure 4. Maps of phase velocity (top) and of group velocity (bottom) resulting from tomographic inversion of ambient-noise dispersion data. Period (s) is specified on each map, and grows from left to right. The shortest periods are sensitive e.g. to the thickness of the sediment layer, while the longest ones sample the crust down to depths close to that of the Moho and below [e.g. Fry et al., 2010, Fig. 5]. We also show the ray coverage for each period (insets).
Figure 5. 1D profiles of the $v_S$ resulting from inversion. Parameterization is the same as EPcrust (i.e., three crustal layers, and two mantle layers). The panels refer to a different sample cell shown in panel (a): Northern Alpine Foreland (b), Alpine Orogen crust (c), South-West Alpine foreland (d), Po Plain (e) and Tyrrhenian crust (f). In each main panel the suite of resulting $v_S$ profiles, as a function of depth, and corresponding Rayleigh phase and group dispersion curves, as a function of period, are shown. Both types of curves are plotted with different colour depending on data misfit (colour scales vary slightly). The red line represents the best model, while the blue line represents the preferred model (average of the best 500 models) in both depth profiles and dispersion curves plots. We reduce the colour intensity in all regions where we have no observations (i.e., periods shorter than 5s and at depth above the 5-6 km and below the 50km ).
Figure 6. The root mean square (RMS) in km/s of the misfit to ambient-noise-based group (a) and phase (b) velocity at various periods over all the cells, achieved by EPcrust (squares), by CRUST2.0 (diamonds) and by the model we obtain after 3-D inversion (circles). (c) Map of the RMS in each cell for group and phase velocity together.
Figure 7. (a) Variability range for the $v_S$ parameter at 3 different arbitrarily chosen depths representative of the three layers for the cell in Fig. 5e, defined as the standard deviation of the gaussian distribution. The red lines represent the best value and the dashed lines (black and grey) the a priori range of variability allowed during the inversion for that layer. (b) Map of the sigma calculated in each cell at 3 km, (c) 20 km and (d) 35 km depth.
Figure 8. Maps of the shear velocity model obtained by averaging the 500 models that best-fit the data in the local neighborhood algorithm inversions. Depth in km is specified on each panel and grows left to right, top to bottom.
Figure 9. Comparison of the Moho depth values obtained in a) this study, b) Spada et al. [2013] and EPcrust.
Figure 10. Vertical sections of the shear velocity model along the profiles showed in the map. The sections are not smoothed in order to well identify parameterization. In each section the Moho depth from Spada et al. [2013] is overimposed (grey line). The AA’ profile (NFP-20 West section) is compared with the correspondent section from Diehl et al. [2009] and Gualtieri et al. [2014] where the Ivrea Body is identified. The BB’ section is a N- S profile across the (western part of the) Eastern Alps; CC’ and DD’ sections document Adria-Tyrrhenian Moho suture zone.