HIGH RESOLUTION ANISOTROPIC IMAGING OF THE CENTRAL EUROPEAN CRUST FROM PHASE AND GROUP VELOCITIES USING AMBIENT NOISE SURFACES WAVES

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Abstract

The morphology of Europe results from the convergence of the African plate towards the European plate via micro-plates such as the Adriatic micro-plate. The direct result of the convergence are numerous subductions in the Mediterranean basin as well as many mountain ranges like the Alps, or the Apennines.

During the last decade several efforts have been directed to image the structure of the European lithosphere using many methods such as, controlled source seismology (CSS), receiver functions, or local earthquake tomography (LET). Many tomographic models using teleseismic events have illuminated the structure of the European mantle. Despite these significant efforts, there is no consensus regarding an European lithosphere model. Hence it remains necessary, first to successfully image the European crust via independent techniques to access the \( P \) and \( S \) structure, and then merge the information and compile a comprehensive and coherent picture of the lithosphere/asthenosphere European system.

The work presented here is a first step towards these objectives. The continued increase in the usage of ambient noise method, combined with the significant increase in the number of broadband seismic networks in Europe allow to access the \( S \) structure of the European’s lithosphere at high-resolution. The \( S \) structure is key to understanding the structure and evolution of the European lithosphere and presently remains poorly resolved.

Using stations from the Swiss Seismic Network (SED), the German Regional Seismological Network (GRSN) and the Italian broadband stations managed by Italian Instituto Nazionale di Geofisicae Vulcanologia (INGV) and the European Orfeus network a database of continuous signals from 196 stations was assembled with inter-station distances from 40km to 2000 km. The cross-correlations between each pair of stations are processed to provide coherent surface waves signals. Group and phase velocities dispersion curves are then measured to infer the structure of the European crust.

Using a linear tomographic inversion algorithm, a set of 2D maps of isotropic velocities were obtained for periods between 5s and 35s from the group and phase velocity measurements. Comparison of our group velocity tomographic results with those obtained from previous studies using ambient noise technique or using combined information from CSS and LET to image the Moho topography, validates our database and methodology, including the pseudo-Green’s functions obtained by cross-correlation of continuous signals. Phase velocities, of the same period, are sensitive to greater depths than the group ve-
velocities. Measurement of phase velocities were also validated by comparing results by the classical two stations method and by the analysis of the cross-correlation frequency.

With the phase velocities measurements, the 2D azimuthal anisotropic structure is inferred for the Alpine region of southern Germany and northern Italy where our dataset provides a high quality coverage along all azimuths. At greater depths, asthenospheric flow induced around the subducting slabs beneath the western Alpine regions. We attribute the observed crustal and mantle lithosphere anisotropy to Hercynian and Alpine orogenies. Anisotropy results of other studies are in general agreement with our 2D anisotropic models. The anisotropy caused by both Hercynian and Alpine orogens are observed. At deeper depths, asthenospheric flow induced around the subducted plate is clearly imaged. The anisotropy results were compared with other studies and the agreement with our 2D anisotropic models.

The compilation of group velocities and phase measurements on the cross-correlation using ambient noise continuous records allowed us to obtain a database of high quality with reliable information of the 2D isotopic and anisotropic structure of the central European lithosphere. This information denotes an important step towards a 3D model of the lithosphere of Europe. First results of a 3D tomographic inversion of our ambient noise data set document a Moho topography that compares well with the recently proposed EPCRustal model. This suggests other well-resolved parts of our preliminary 3D model contains reliable informations.
Résultats

La morphologie de l’Europe résulte de la convergence de la plaque africaine vers la plaque européenne par le biais de micro-plates telle que la micro-plate adriatique. L’expression de cette convergence s’effectue par des nombreuses subductions dans le bassin méditerranéen et aussi de nombreuses chaînes de montagnes telles que les Alpes, ou les Apennins.

Lors de la dernière décennie de nombreux efforts ont mis en place pour imager la structure de la lithosphère européenne grâce à de nombreuses et différentes méthodes telles que la sismique grand angle, les fonctions récepteurs, la tomographie locale. De nombreux modèles dérivés de la tomographie des ondes télé-sismiques imaginent le manteau. Malgré ces efforts significatifs, il n’existe pas de consensus entre tous les modèles. Ainsi il subsiste un réel besoin de, tout d’abord réussir à imager la croute européenne via des techniques indépendantes pour accéder à la structure en ondes P et S ou encore la densité et ensuite réussir à accéder à une image globale et cohérente du système lithosphère/asthénosphère européenne qui incorpore tout types d’informations.

Les travaux présentés ici s’inscrivent dans cette logique. L’explosion de l’utilisation de la technique de bruit de fond combinée à l’augmentation significative du nombres de réseaux de sismomètres en Europe permet aujourd’hui d’accéder à la structure en ondes S de la lithosphère européenne à haute résolution. La structure en onde S est une des clefs pour comprendre la structure de la lithosphère européenne et reste, peu connue.

Grâce aux réseaux sismiques Swiss (SED), German regional seismological network (GRSN), et Italian dirigé par Instituto nazionale di geofisica e vulcanologia (INGV) ainsi que le réseau européen Orfeus, une base de données de signaux continus de 196 stations est assemblée avec des distances inter stations allant de 40 km à 2000 km. Les cross-corrélations entre chaque couple de stations sont effectuées et fournissent des signaux cohérents d’ondes de surface. Ainsi les vitesses de phase et de groupe sont mesurés dans le but d’obtenir la structure de la croute européenne.

Grâce à une inversion utilisant un algorithme d’inversion tomographique linéaire, un ensemble de cartes 2D des vitesses isotopiques est obtenu pour des périodes comprises entre 5 et 35 s provenant de l’inversion des vitesses de groupes et de phases. Le succès de la comparaison des résultats obtenus avec les vitesses de groupes avec des études antérieures utilisant le bruit de fond et une carte de topographie du Moho fournie grâce à la combinaison d’ondes de volumes validant la base de données et plus précisément les pseudo-fonctions
de Green obtenus par la cross-corrélation des signaux continus. Un effort est alors fourni pour l’utilisation des vitesses de phase qui, à la même période, sont sensibles à des profondeurs plus larges que les vitesses de groupe. Les vitesses de phase sont, de plus, validées par une étude de leur mesure grâce à la comparaison de la méthode classique 2 stations et l’analyse de la cross-corrélation en fréquence. La cohérence des résultats obtenus par deux méthodes indépendantes valide d’autant plus nos vitesses de phase.

Ainsi la structure 2D anisotropique est analysée pour la région des Alpes, du sud de l’Allemagne et de du Nord de l’Italie la couverture des données est haute résolution. L’anisotropie causée par l’orogène hercynienne mais aussi alpine est observée. À plus grande profondeur, le flux asthénosphérique induit autour de la subduction est clairement imagée. Les résultats d’anisotropie ont également été mis en comparaison avec d’autres études et la cohérence des résultats permet de valider la structure 2D anisotropique des résultats.

La compilation de vitesses de groupes et de phases mesurées sur les cross-corrélations d’enregistrement de bruits de fond continues permet d’obtenir une base de données haute résolution de la structure 2D isotopique et anisotrope de la lithosphère du centre de l’Europe. Ces informations sont une étape vers un modèle 3D de la lithosphère européenne. Des résultats préliminaires de la topographie 3D du Moho en comparaison avec EPcrust permettent la construction d’un tel modèle avec des données de bruits de fond.
One of the most exciting advances in seismology over the last decade is passive imaging or in other words imaging of the earth based on seismic ambient noise. The source of seismic oscillation could be artificial: man-made explosions, or natural earthquakes. The main interest of seismology is to understand the mechanisms of earthquakes. The strongest earthquakes represent one of the most significant destroying natural hazards and they could occur all over the planet. (Fig. 1.1). Understanding earthquakes means understanding their occurrence in time, to predict them (in a statistical sense) and prevent and educate the population of earthquake-prone areas. The study comprehension of earthquakes involves the study of their source mechanism, the energy released, their location and the geometry but it’s also crucial to know the structure of the Earth to model wave propagation, as seismic waves are possibly the most informative direct measure of all mentioned properties of an earthquake. This is closely related to another branch of the seismology: seismic tomography. In fact, the more precisely we know the structure of the earth, the more accurately we can understand the earthquake. Naturally there is a trade-off between mapping the properties of the earthquake and those of the Earth, since we use the earthquake itself to image the Earth. Imaging the earth is not only useful to study seismic events; tomographic models are a snapshot of mantle convection and therefore a fundamental constraint for geodynamical models. In a more applied perspective, seismic imaging is a fundamental scale for reservoir-scale imaging and the exploration of natural resources like oil or gas.

Seismic tomography is based on the analysis of the different phases of a seismogram (e.g. Fig. 1.2). If one can locate accurately an earthquake or an explosion and describe the propagation of the different waves, thus generated the structure through which the waves
propagate can be inferred. Seismic tomography can be applied at any scale, from reservoir at shallow surface to the whole Earth, since waves propagate through the whole earth (e.g. Fig. 1.3). Since earthquake are the source of essentially all seismological data, the quality of tomographic images depends on (i) our capability to understand and describe the earthquake; (ii) the spatial and temporal occurrence distributions of earthquakes. As we noticed earlier, our knowledge of earthquakes is still limited and we are not able to describe perfectly the seismic source or even to located earthquake with an extreme precision. Moreover, the seismicity on Earth, as shown by Fig. 1.1 is very uneven and several regions are image including Europe, can only be imaged with limited resolution owing to their relatively low seismic activity.

1.0.1 Ambient noise method

Precisely because of the non-uniformity in the geographic distribution of earthquakes, being able to extract information about the shallow earth from seismic noise, which is
Figure 1.2: 3-comp. records of a large event in Peru, at increasing epicentral distances. To the left are the names and epicentral distances of each station. From http://wwwrses.anu.edu.au/seismology/SHon2002/Brian Kennett’s site

present everywhere, represents a important advance in seismologic imaging techniques. The ambient noise technique was first applied by Duvall et al. (1993), who measured the fluctuations of the Sun surface and used them to infer the internal structure of the Sun. The concept behind this technique resides in the theoretical fact that the cross-correlation of ambient signal recorded at two locations contains the Green’s function associated with those locations, one being treated as the source, the other as the receiver (Snieder, 2004;
Weaver and Lobkis, 2001). This method has been used widely, first in acoustic (Weaver and Lobkis, 2001), and later applied in seismology by Shapiro et al. (2005) in California, Europe (Yang et al., 2007; Stehly et al., 2009; Yao et al., 2006).

1.0.2 Europe

The European continental region and especially the Alps are two of the most studied regions in geosciences. Enormous quantity of literature is available about Europe from the surface to the mantle. Our aim in this section is to give a general overview of the European tectonic that are imaged by various seismic methods in the following chapters.
The African plate is converging towards the European plate since the last 80 Myr (Faccenna et al., 2001). The general motion of Africa towards Europe is locally accommodated by micro-plates of continental and oceanic lithosphere (Schmid et al., 2004; Boschi et al., 2010). Indeed with the convergence of those two majors plate, the Tethys ocean was closed and the oceanic lithosphere subducted. As a result of this convergence, numerous subductions took and take place today in the Mediterranean basin (Jolivet et al., 2009) mainly under the Calabrian and the Hellenic arc (e.g Fig. 1.4). Several orogenic belts also result from the convergence like the Alps, the Pyrenees, the Apennines, the Dinarides, and the Carpathians (Dercourt et al., 1986; Platt et al., 1989; Dejonge et al., 1994).

![Figure 1.4: Topographic and bathymetric map of the Mediterranean region with kinematics reconstruction of the position of the trenches Jolivet et al. (2009)](image)

The convergence of the African and European plate is also maps by GPS measurements by (Kreemer et al., 2003). Fig. 1.5 shows the velocity of the different plates and clearly documents the convergence of the African plate towards European.
In the Alps collision occurred in mid-Eocene times after relatively slow subduction of Alpine Tethys oceanic lithosphere attached to Europe during early Tertiary. Subsequently, the oceanic slab break off and the European lithosphere slab delaminates and rolls back further (Handy et al., 2010). Several plate tectonics plate reconstructions have been conducted to explain evolution of the Alpine orogen and their forlands (Faccenna et al., 2001; Handy et al., 2010).

The tomographic image of the European mantle from Piromallo and Morelli (2003) down to 1000km show the presence of several positive P anomalies interpreted as subducted remnants of Alpine Tethys (c.f Fig. 1.6). Lombardi et al. (2009) using receiver functions to image the 440 and 660 km discontinuities interpret these high velocity anomalies resting on top of the 660 discontinuity as remnant oceanic lithosphere (c.f Fig. 1.7).
More recently the high resolution tomographic image from Lippitsch et al. (2003) imaged the European subducted slab under the Alps as well as Po plain and the European lower lithosphere (c.f Fig. 1.8).
The Moho topography was imaged at high resolution by CSS method and CSS associated with receiver functions data Waldhauser et al. (1998); Di Stefano et al. (2009) and their results shown in Fig. 1.9 nicely document the dipping of the European Moho, the Adriatic Moho with a bump in the North-West of Italy with Moho depth around 30 km while in the rest of the Alpine arc Moho is around 50 km caused by the Ivrea Body.
After the slab-break-off (Blanckenburg and Davies, 1995), the episode of Alpine collision convergence results in the Alpine orogen. Mid to lower crust material was exhumed and stacked together to form the Alpine arc with its deep crustal root. Schmid et al. (2004) proposed a schematic interpretation of Alps and the arrangement of the material inside the Arc (c.f Fig. 1.10). Different types of collisions are present. In the Western Alps, there is indentation of the uplifted Ivrea body which acted as a backstop and the exhumation and stacking of the Penninic nappes and the Brianconnais terranes. In the eastern Alps, the lower crust of Adria indents the European lower crust resulting in exhumation of Penninic Austraalphine nappes system on the top.

Alpine dynamics is driven by buoyancy forces meaning surface erosion and subduction/mantle lithosphere delamination processes (Kissling, 2008). The subduction process concern one slab attached to the European plate and 2 to the Adria. The different pending of each slab provokes the difference between Western, center and eastern Alps.

Most of the seismic proof of those phenomenon are based on Moho topography and only few tomographic models have been proposed about the European crust since the seismicity in Europe in low.

Figure 1.9: Moho topography from Waldhauser et al. (1998)
Recently, Molinari and Morelli (2011) proposed a crustal model of Europe assembling all different models available from the literature (c.f Fig. 1.11). They provided a 3 layers model: sediments, upper crust, lower crust for the whole European region: from North Africa to the North Pole (20N - 90N) and from the Mid-Atlantic Ridge to the Urals (40W - 70E). This model is a compilation of several models and there is still the need to provides S velocity properties at crustal of Europe which might give high resolution information could be add to a model like EPcrust.
Figure 1.11: EPcrust: crustal model for the European plate, derived from several models and information of Europe selected from the literature from Molinari and Morelli (2011)

1.0.3 Outline of the thesis

The aim of this thesis is to develop a 3D European model of the lithosphere derived from seismic ambient noise continuous records. The challenges here, are first to apply ambient
noise method to a large dataset and second to provide a S velocity structure model of Europe at crustal scale. We hence present here the database of group and phase velocities measurement we compute from averaged cross-correlation. The velocities measurements are then inverted via a linear tomographic inversion algorithm to infer the isotropic structure though 2D velocity maps from 8s to 35s periods. Main features as the Alps, the Po plain, the Ivrea body or the molasse basin in Germany are well-resolved and our models fit nicely previous studies of the region using also ambient noise method. We then focus on phase velocity measurements and access robust measurements by comparing two way of measurements them (classical 2 stations method and cross-spectrum analysis). The phase velocities are then used to infer 2D azimuthal anisotropy of central Europe. By inferring for both isotropic and anisotropic 2D structure we complied a high resolution and robust database that is a first step to provide a 3D structure of Europe.
**Bibliography**


Abstract

We present a new database of surface-wave group and phase-velocity dispersion curves derived from seismic ambient noise, cross-correlating continuous seismic recordings from the Swiss Network, the German Regional Seismological Network (GRSN), and the Italian national broadband network operated by the Italian Instituto Nazionale di Geofisica e Vulcanologia (INGV). In order to increase the aperture of the station array, additional measurements from the Mediterranean Very Broadband Seismographic Network (MedNet), the Austrian Central Institute for Meteorology and Geodynamics (ZAMG), the French stations (obtained through Orfeus) are also included. The ambient noise we are using to assemble our database was recorded at the above mentioned stations between January 2006 and December 2006. Correlating continuous signals recorded at pairs of stations, allows us to extract coherent surface-wave signal traveling between the two stations. Usually the ambient noise cross-correlation technique provides information at periods of 30s or shorter. By expanding the database of noise correlations, we seek to increase the resolution of the central Europe crustal model.
We invert the resulting datasets of group and phase velocities associated with 8s-35s Rayleigh waves, to determine 2-dimensional (2D) group and phase-velocity maps of the European region. Inversions are conducted by means of a 2D linearized tomographic inversion algorithm. The overall good agreement of our models with previous studies and good correlation with well-resolved velocity anomalies and geological features such as sedimentary basins, crustal roots and mountain ranges, demonstrates the effectiveness of our approach.

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2.1 Introduction

The European lithosphere is shaped by the convergence of the African and European plates involving between them a mosaic of microplates of oceanic and continental lithosphere (Schmid et al., 2004; Boschi et al., 2010). The resulting strong 3-dimensional (3D) heterogeneities in crust and upper mantle are naturally difficult to image seismically. Yet, reliable seismic 3D lithosphere models are necessary for accurate earthquake locations, and as constraints for geodynamical modeling. Most published tomographic models are based on observations of $P$-wave travel times (e.g. Bijwaard and Spakman (2000); Lippitsch et al. (2003)) or surface-wave dispersion recorded from teleseismic events (e.g. Boschi et al., 2009, 2010; Chang et al., 2010). Teleseismic body waves are appropriate for imaging mantle structure, but they are only partially sensitive to the crustal-lithospheric depth range (Schivardi and Morelli, 2009). High-frequency signals associated with teleseismic surface waves are generally weak, and high-quality measurements are only available at periods equal to or higher than 30s. An alternative method is Local Earthquake Tomography (LET), which is well-suited to image strong lithosphere heterogeneities in 3D (Diehl et al. (2009); Di Stefano et al. (2009)) but the relative scarcity of seismic events in large regions of Europe prevents LET to be consistently applied on a regional scale. Moreover, high quality local earthquake $S$ data are difficult to pick (Diehl et al., 2009) and have not yet been used. Accurate maps of Moho depth and local crustal structure can be obtained by Controlled Source Seismology (CSS) (Waldhauser et al., 1998, 2002), but while good at identifying crustal geometries like the Moho discontinuity, CSS yields relatively little information on the lateral variation of the $S$ velocity structure; moreover CSS is a 2D method and needs 3D migration because sources and receivers are on the same side of the target structure. In summary, “traditional” imaging techniques, applied individually have been insufficient to provide a high-resolution image of crustal and lithospheric $P$-wave and $S$-wave velocity structure at the scale of Europe.

A promising complementary seismic approach to enhance resolution of the shallow Earth wave velocity field is the so called ambient noise technique. The ambient noise method is based on the theoretical result that the cross-correlation of ambient seismic signal observed at two locations is generally very close, if not exactly coincident, with the Green’s function associated with those locations (one being treated as the source, the other as the receiver) (Snieder, 2004; Weaver and Lobkis, 2001).
The technique was first used in helioseismology (Duvall et al., 1993) to interpret oscillations observed at the surface of the sun in terms of propagating waves. Weaver and Lobkis (2001) base their acoustic-wave treatment on the assumption of equipartition between all the modes of the propagation medium, which yields the equality between the derivative of the displacement Green’s function and ambient noise cross-correlation. Sanchez-Sesma and Campillo (2006) extended this result to the case of seismic waves. Snieder (2004) came to similar conclusions following the stationary-phase approach: if the station pair is surrounded by seismic sources at all azimuths (representative of a diffuse, or equipartitionned seismic wavefield), the cross-correlation of cumulative recorded noise implicitly cancels out the contribution of sources that are not aligned with the station-station azimuth, and the surviving signal corresponds, again, to surface-wave propagation along the station-station azimuth. Later, Wapenaar (2004) proved the connection between the Green’s function and ambient noise correlation through an application of the reciprocity theorem. All theoretical studies are based on either of the following assumptions: (i) that (from a standing-wave viewpoint) noise is equipartitioned over all modes; (ii) that (from a traveling-wave viewpoint) the wavefield is diffuse, as a result of strong scattering and/or a geographically uniform distribution of noise sources.

Useful analyses of the performance of ambient noise cross-correlation techniques in the real world, i.e. in the absence of noise equipartition, are provided for example by Weaver et al. (2009), Cupillard and Capdeville (2010), Froment et al. (2010) and Tsai (2010). It has been noted that ambient-noise measurements are sensitive not only to the azimuthal distribution of the sources, but also to their distance from the station array (Harmon et al., 2008; Cupillard and Capdeville, 2010). The importance of scattering has been verified at least at the local scale and relatively high frequency (Gouédard et al., 2008; Froment et al., 2010). Tromp et al. (2010) have proposed an adjoint, numerical approach to quantify the effects of nonuniformity in the noise-source distribution, and to compute sensitivity kernels that account for such effects; the database presented in our study is currently being used by (Basini et al., 2011) in one of the first practical applications of this method.

In general, the high correlation between ambient-noise-based tomography and independent results in various, densely instrumented regions of the world suggest that real-world conditions are often sufficient: successful examples are California (Shapiro et al., 2005) and Europe (Yang et al., 2007; Stehly et al., 2009), where group velocities were measured, or Tibet (Yao et al., 2006), where phase velocities were measured.
The distribution of ambient noise sources averaged on several months is sufficiently homogeneous to apply this method to image at crustal scale. We now know that observed surface-wave ambient noise is only generated at the Earth’s surface, and essentially over the oceans (storms, and the coupling of oceans with the solid Earth) (Stehly et al., 2006), with most released energy roughly between 5-20s. In the case of Europe, ambient noise comes mostly from the Atlantic, and only marginally from the Mediterranean sea (Stehly et al., 2006; Chevrot et al., 2007; Kedar et al., 2008; Yang and Ritzwoller, 2008).

With this study, we build on the earlier works of Stehly et al. (2009) and Li et al. (2010), compiling a larger database of surface-wave dispersion measured by noise cross-correlation of European stations. The size of our region of interest is double that of either of those previous studies as it includes Germany, Switzerland and the Alpine region, Italy and the Tyrrhenian Sea. The earlier study of Yang et al. (2007) is similar to ours in that it covers a wider region but includes fewer stations; it is limited to lower frequencies and group-, rather than phase-velocity observations.

Since 2006, Italy is covered by a very dense network of at least 125 broadband instruments. Combined with the central-Europe stations already used by Stehly et al. (2009), a cumulative array of 196 receivers provides a coverage of crust and lithosphere unprecedented in Europe (Fig. 2.1). With this station array the Alps and the Po plain, where the regions of interest of Li et al. (2010) and Stehly et al. (2009) overlap, can be imaged with significantly better resolution than in either earlier study. Importantly, the large number of available high-quality observations allows us to observe not only group-, but also phase-velocity dispersion, and to extend our observations to relatively long periods of up to 35s. The ultimate goal of this latter effort is to fill the gap between teleseismic and ambient noise techniques of surface-wave observation. Improving the reliability of ambient noise-based dispersion curve at periods equal or higher than 30s, somewhat too short for teleseismic observations, is equivalent to significantly improving the seismic coverage of the lithosphere-asthenosphere boundary.

In the following, we first discuss the set of stations and associated seismic records that we used. We next describe the processing algorithm (similar to Bensen et al. (2007), Stehly et al. (2009)) that we followed to extract, from such seismic records, estimates of the station-station Green’s functions, and, subsequently, of the station-station group and phase velocities. We proceed with several synthetic tests to assess quantitatively the resolution of our dataset. We apply phase-velocity tomography (e.g. Boschi (2006)) to measure group- and phase-velocity dispersion data at periods between 8-35s, to derive from our data a
set of 2D maps of crustal and lithospheric structure of the region of interest. Our results are generally consistent with those of Li et al. (2010) and Stehly et al. (2009), which were both limited to smaller areas and/or to lower resolution. Our short period group- and phase-velocity maps are characterized by a few small scale features, which appear to be in good agreement with the geology of the region; our longer-period group and phase velocity maps correlate very well with a Moho map recently established from CSS and LET results from Wagner et al. (2011).

2.2 From continuous records to dispersion curves

As described by Stehly et al. (2006), one year of continuous recording is needed for the successful application of the ambient-noise method to regional-scale seismology. Over one year, seasonal effects associated with the geography of ocean storms cancel out, and the cumulative source distribution of stacked data is closer to being uniform: a condition for our cross correlations to approximate the Green’s functions well. To improve on our current knowledge of the European lithosphere, we combine and cross-correlate recordings from several broadband European networks. We developed a new database of surface-wave group- and phase-velocity dispersion curves, which we obtained by cross-correlating continuous seismic recordings mostly from the Swiss Network, the German Regional Seismological Network (GRSN), the Italian national broadband network operated by the Instituto Nazionale di Geofisica e Vulcanologia (INGV). In order to increase the aperture of the station array, we included additional measurements from the Mediterranean Very Broadband Seismographic Network (MedNet), the Austrian Central Institute for Meteorology and Geodynamics (ZAMG), the French Broadband Seismological Network and the Slovenia Seismic Network. The resulting station distribution is illustrated in the Fig. 2.1. We aimed at collecting continuous recordings for all these stations starting in January 2006 and until December 2006, though of course not all stations were constantly operational in this time interval.

2.2.1 Sampling, whitening and emergence of coherent surface wave signal

We compute our correlations in the same way as Bensen et al. (2007) and Stehly et al. (2009). Bensen et al. (2007) extract Rayleigh wave velocities from the cross-correlation of
Figure 2.1: Location of the broadband stations used in this study, from the combination of regional European networks detailed in Section 2.2.

vertical with vertical, and radial with radial components recorded at the two stations; while Stehly et al. (2009) used the four possible combinations of cross correlation between the vertical and radial component of the two stations ($Z - Z$, $R - R$, $Z - R$ and $R - Z$) and then extract 8 velocities measurements and averaged them to obtain the velocity between the two stations. In our study we used only the vertical component of the record. The signal on the vertical component is more energetic than the radial one; the radial component could additionally be affected by azimuthal anisotropy or by bending of Love-wave paths caused by lateral heterogeneity. Limiting our analysis to the vertical signal is a way to avoid all these potential issues.
The continuous data that are analyzed here were recorded by broadband stations, deployed with the primary goal of recording earthquakes and all the associated information. Since we are interested only in the diffuse, “background” signal, our processing of the data is aimed at emphasizing what traditional seismology normally neglects as “noise” (Bensen et al., 2007). We first remove trend, mean and instrumental response from the signal. We know through numerous previous studies that the spectrum of seismic ambient noise is not flat in the frequency domain (e.g., Bensen et al., 2007) but characterized by several peaks. All complexities in the spectrum of noise sources should be somehow corrected for, to better satisfy the requirement of equipartition (Section 2.1). We achieve this by systematically whitening the noise signal. Even though the resulting spectrum is not completely flat, the amplitude of the mentioned peaks and the bias towards longer periods are reduced.

Our cross-correlation algorithm consists of the following steps: (i) we divide the continuous records in large sets of day-long files; (ii) we cross-correlate the vertical component for all possible pairs of stations and for each day; (iii) we stack together the resulting daily station-station cross-correlations over the whole year. A separate stack is calculated for each available station pair. The result of this exercise, for stations AIGLE and ARBF, is shown in Fig. 2.2 as an example. As noise arrives, at different times, from different predominant azimuths, stacking is equivalent to combining the effects of sources located at different azimuths: the stacked signal is thus closer to the required assumption of source uniformity/equipartition. At the same time, the effect of “ballistic” waves coming from a single direction (e.g., an earthquake) will naturally tend to cancel out: we don’t need to artificially remove days of important seismic activity before processing the data. Our choice of sampling rate for the cross-correlation is determined by the type of structure that we are looking for. Since we are ultimately aiming at crustal/lithosphere-scale seismic imaging, our target is roughly the 0.025-0.5Hz frequency range. Based on Shannon’s theorem, we need a sampling rate that is approximately twice the frequency of interest: we choose a sampling rate of 1 Hz. In practice, our analysis is limited to somewhat lower frequencies (∼0.2Hz) as a consequence of a fairly large average inter-station distance of ∼100km.

As seen in Fig. 2.2 noise cross-correlations tend to be characterized by two symmetric maxima, one at positive and the other at negative time. These two portions of the cross-correlated signals are respectively referred as causal and anticausal, as illustrated e.g. in Fig. 1 of Stehly et al. (2006). Essentially, the causal part corresponds to energy propagating from AIGLE to ARBF and the anticausal one to energy propagating from ARBF to AIGLE. Notice that, regardless of the frequency band at which the cross-correlation was filtered,
a systematic difference in amplitude is evident between the causal and anticausal parts of the traces in Fig. 2.2, with causal parts showing larger amplitude than the anticausal ones. We infer that most ambient noise energy propagates from the north to the south. This observation is consistent with what we found at other European station pairs, and confirms that most seismic ambient noise recorded in Europe is generated in the Atlantic ocean (e.g., Stehly et al., 2006, 2009).

We see in Fig. 2.2 a clear asymmetry in the amplitude of the causal and anticausal parts of the cross-correlations, at all period bands except 15-20s. The surface waves on the causal and anticausal part have a similar travel time for all period bands. We are primarily interested with the phase and not the amplitude. It is thus possible to obtain a good-quality measurement of dispersion, since dispersion is essentially related to phase and not amplitude. For the station pair of Fig. 2.2, as well as for most other station pairs in our study region, the asymmetry is minor compared to the overall cross-correlation signal, indicating that the source distribution is sufficiently close to uniform for the ambient noise to hold.

Figure 2.2: (a) Locations of stations AIGLE (western Switzerland) and ARBF (southern France). (b) Cross-correlation of stacked continuous signal (Section 2.2.1) recorded at AIGLE and ARBF, filtered over different frequency bands as indicated.
2.2.2 Analysis of dispersion

In the ambient noise formalism assumed here, cross-correlations of stacked signal at a station pair are approximately coincident with the surface-wave Green’s function between the two stations (i.e., one station can be thought of as the source, and the other as the receiver), and can be treated as such. It is thus legitimate to apply the frequency-time analysis (FTAN) method (Levshin et al., 1989; Ritzwoller and Levshin, 1998; Bensen et al., 2007) to our cross-correlation, and measure the surface-wave dispersion between all cross-correlated station pairs. Before applying the FTAN, we fold the causal and anticausal part on top of each other. We first apply a phase-matched filter to remove possible contamination of energy by higher modes and to increase the signal-to-noise ratio (SNR) (Herrin and Goforth, 1977; Bensen et al., 2007). Similar to Fig. 2.2, the FTAN then consists of band-pass filtering the signal around the different frequencies we want to measure (vertical axis in the bottom panel of Fig. 2.3), and plotting it as a function of time and frequency as shown in Fig. 2.3. The FTAN is used to measure group velocity. We identify, at each frequency, the maximum of the resulting envelope of signal amplitude, and find the corresponding group velocity as the ratio of time to the known interstation distance: the group velocity as a function of frequency, i.e. the dispersion curve, is found. We pick manually the amplitude maximum. In certain cases it is not possible to identify it at any frequency and we only make measurements in the frequency range where the maximum is sufficiently well defined. In practice, we measure SNR by comparing the peak of the cross-correlation waveform (corresponding to the peak of the Green’s function) and divide by the pick of a time-window of the same waveform well away from the Green’s function. We then only measure group velocity on cross-correlation with SNR of five or larger. We repeat our measurement procedure for all station pairs in our database, and show in Table 2.1 the number of group-velocity observations that we finally keep for each analyzed period. As a general rule, higher quality observations correspond to station couples with a longer available time window for cross-correlation (typically larger than 300 days for the large majority of the measurements included in our database).

Though the main focus of this paper is the development of a new group-velocity database, it is also useful to measure phase velocity to better constrain the S velocity of the crust, since the two types of measurements are well known to be sensitive at different depth (e.g., Ritzwoller et al., 2001). We measure phase velocity via the 2-station method as implemented by Meier et al. (2004). This method was originally designed to resolve local structure based
### Table 2.1: Number of station-station group-velocity observations included in our database. The total number of possible station-station combinations is 19,110.

<table>
<thead>
<tr>
<th>period</th>
<th>number of data</th>
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</thead>
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<tr>
<td>8s</td>
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</tr>
<tr>
<td>12s</td>
<td>5822</td>
</tr>
<tr>
<td>16s</td>
<td>6917</td>
</tr>
<tr>
<td>24s</td>
<td>5512</td>
</tr>
<tr>
<td>30s</td>
<td>3311</td>
</tr>
<tr>
<td>35s</td>
<td>2890</td>
</tr>
</tbody>
</table>

**Figure 2.3:** Frequency-time analysis applied to the cross-correlation of continuous records at AIGLE and ARBF (station locations are shown in Fig. 2.2).

on surface-wave recordings of distant earthquakes. Instead of cross-correlation of earth-
quake signals, we apply it here to our cross-correlations of background noise. According to Meier et al. (2004), given the observed surface-wave phase \( \phi \) as a function of frequency \( \omega \), we obtain a phase-velocity dispersion curve via the formula

\[
c(\omega) = \frac{\omega \Delta}{\arctan(\Im(\phi(\omega))/\Re(\phi(\omega))) + 2n\pi},
\]

where \( c \) denotes phase velocity, \( \Delta \) the interstation distance, \( \Re \) and \( \Im \) the real and imaginary parts of the cross-correlation. The integer number \( n \) accounts for the ambiguity of the arctangent function, whose associated error on phase is a multiple of a full cycle \( 2\pi \) (Meier et al., 2004). In practice, as illustrated in Fig. 2.4, we need to implement Eq. (2.1) for a set of possible values of \( n = 0, 1, 2, ... \), and then, following Fry et al. (2010), we pick the dispersion curve closest to that predicted by PREM. Again, we only measure phase velocity on cross-correlations with a signal-to-noise ratio of five or higher.

### 2.3 Surface-wave tomography

We derive group- and phase-velocity maps from the databases described above, applying the ray-theory formulation of, e.g., Boschi and Dziewonski (1999). As long as effects caused by nonuniformity in the noise source distribution (e.g., Tromp et al., 2010) are neglected, our group- and phase-velocity databases can be treated as traditional ones, with our station pairs corresponding to the source-station pairs of earthquake-based tomography. The region of interest is subdivided into approximately equal-area pixels whose size would be \( 0.3^\circ \) by \( 0.3^\circ \) at the equator; their longitudinal extent is corrected to keep the area approximately constant. Only pixels in the region of interest are sampled by the data and contribute to the inverse problem. A linear system is set up as described by Boschi and Dziewonski (1999) and solved in least-squares sense via LSQR (Paige and Saunders, 1982). The inverse problem is non-unique and regularized via roughness minimization.

The use of ray theory implicitly limits the resolution of our surface-wave group- and phase-velocity maps to heterogeneities of wavelength comparable to, or larger than that of the inverted data. In two-dimensional surface-wave tomography, the limits of ray-theory and the improvement to be expected from the application of finite-frequency methods are analyzed in detail (though at longer wavelength) e.g. by Peter et al. (2009). We choose here to use a simple ray-theory algorithm and derive approximate maps to evaluate the quality and information content of the data.
Figure 2.4: Phase-velocity dispersion curve from cross-correlation of the continuous recordings made at AIGLE and ARBF. Different colored thin curves are the different dispersion curves corresponding to different values of $n$. The thick red curve is the theoretical dispersion curve derived from PREM model. The thin red curve corresponds to $n = 0$ and turns out to be our preferred one. The black portion of this curve denotes the frequency range where we trust the measurement.

2.3.1 Resolution

We use ray theory to invert group and phase velocity measurements made at periods between 8 and 35s. This poses the theoretical limit of resolution correspondingly between 20 and 130 km. In practice, resolution depends on the geographic coverage of our database. We quantify the resolution through a set of synthetic data inversions (e.g., Kissling, 1988; Boschi and Dziewonski, 1999). We first define a “checkerboard” input group-velocity map coinciding with the spherical harmonic function of degree 60 and order 30, corresponding to anomalies extending a few hundred km laterally and compute the corresponding synthetic phase anomalies by a simple matrix multiplication. No noise is added to the data. We invert the resulting synthetic data through an application of the same tomography algorithm that we use on real observations, including the regularization scheme and weight. The results of
this exercise are illustrated in Fig. 2.5. At this stage, resolution limits associated with the approximations inherent to our theoretical formulation (ray theory, which is only strictly valid in the infinite-frequency limit) are neglected, and resolving power is independent of frequency. Since, for our dataset, different surface-wave frequencies have approximately the same geographic coverage, it is then unnecessary to repeat this exercise at all considered surface-wave modes, and we only show in Fig. 2.5 results for 8s and 35s group velocity. The input (Fig. 2.5a and d) and output (Fig. 2.5b and e) models are shown together with the density of raypaths at each period. Generally, the relatively long-wavelength pattern of group-velocity heterogeneity is reproduced well throughout the region that is most densely covered by stations (Fig. 2.1), from northern Germany all the way down to the Tyrrhenian Sea. There is, however, a clear loss in recovered amplitude, which is a systematic problem in damped seismic tomography. The locations marked 1, 2 and 3 on the map are areas of high sensitivity of our data where anomalies are well recovered. Locations 5 and 6 are clearly affected by strong smearing artifacts due to very low ray coverage (Fig. 2.5c and f). In area 4, the results from the two frequencies show different level of coverage even if the region is covered by stations. Looking at the density of raypaths at that specific location, 35s shows more raypaths (Fig. 2.5f) than 8s, and indeed the anomaly in location 4 is better resolved.
Figure 2.5: “Checkerboard” test (a) and (d) Input model coinciding with the spherical harmonic function of degree 60 and order 30, and 1% velocity anomalies (b) and (e) Output model obtained inverting a synthetic database associated with our 8s and 35s Rayleigh-wave group velocity data set, and the input model at (a) (c) and (f) Density of raypaths associated with our measurements at 8s and 35s Rayleigh-wave group velocity.
To evaluate our algorithms power to resolve structures like we might observe in the real Earth, we performed a characteristic model test (Husen et al., 2009), replacing the checkerboard input model of Fig. 2.5a with a model containing randomly distributed (in 2D) velocity anomalies of various sizes. Namely we generate a $128 \times 128$-pixel image, with random values of velocity anomaly ranging between $-1\%$ and $1\%$ with respect to velocity as predicted by the Preliminary Reference Earth Model (PREM) (Dziewonski and Anderson, 1981), apply a 2D Fourier transform to the image, filter it in the Fourier domain so that heterogeneities of scalelength similar to those actually observed become dominant, and inverse-Fourier-transform it back into the spatial domain. The resulting synthetic model is shown in Fig 2.6 and contains anomalies of 100 km minimal length and up to 600 km length. We compute synthetic data and invert them as described above. The resulting models associated with 8s, 16s and 35s Rayleigh-wave group velocity data are shown in Fig. 2.6b. Four images per period are show, one with the input model, one with the raw results, one with the results annotated for specific anomalies that are of interest and finally the result and the outlines of the well-resolved area as derived from this test. Fig. 2.6 shows that velocity heterogeneities of relative short length (100 km) can be resolved by our data coverage in those areas densely covered by stations namely Switzerland and northern Italy. In Germany, most longer-wavelength heterogeneities of 200 km or more are reproduced well, but most of the smaller scale features are lost. High-amplitude features of 150 km length are fairly well-resolved in south-eastern France, Austria and southern Italy. The systematic underestimation of heterogeneity amplitude results from the choice of regularization used in the inversion. More specifically, anomalies 1, 2 and 4 are fairly well resolved within the station array but the amplitudes are underestimated due to the lack of stations in that area. Anomaly marked 3 is an example of 100 km structure well-resolved with only limited reduction in amplitudes thanks to locally high station density. The big square in Italy is well resolved while in Corsica the shape of the anomaly is distorted and the amplitude is significantly reduced. Analyzing the recovery of the velocity anomalies in all regions one can draw the limitations of the well-resolved area and for each period we invert data from this information is incorporated in our interpretation. As an example the limits we draw for 8s period and the one for 16s period are globally similar in general but vary in the west of Germany where at 8s the ray coverage is not sufficient.
Figure 2.6: Synthetic test with randomly distributed velocity anomalies of various size as input: (1a), (2a), (3a) associated with our 8s, 16s and 35s Rayleigh-wave group velocity dataset. Panels. (1b), (2b) and (3b): Raw output model associated with 8, 16 and 35s respectively. (1c), (2c) and (3c) Output models with outline of specific anomalies located inside the correct resolved part of the model. (1d), (2d) and (3d) Output model with the boundaries of the well-resolved area.
To validate our results and our resolution estimates we also compare them with those of previous studies from Stehly et al. (2009) and Li et al. (2010) as shown in Fig. 2.7. Fig. 2.7(a) shows the Rayleigh-group velocity map at 16s derived from our dataset in absolute velocity. The figure pixels coincide with pixels of our parametrization, so that the unsmoothed model is plotted. Fig. 2.7(b) shows the same result but after removing the mean and smoothing as we smoothed the maps of Stehly et al. (2009) and Li et al. (2010) shown in Fig. 2.7(c) and Fig. 2.7(d) respectively. The study by Stehly et al. (2009) includes data from 2004 and 2005 with station coverage limited to the northern portion of the region of interest. The study of Li et al. (2010) is based on the same data and station distribution as ours, but limited to the Italian region. Even though the inversion process and data coverage differ, all studies show similar patterns of velocity anomalies. The Po Plain is visible and imaged at the same location by all studies. Our study and Li et al. (2010) show the same fast feature in the Tyrrhenian sea caused by its thin crust resulting in surface-wave energy propagating through the faster upper mantle. The results by Stehly et al. (2009) and ours show similar features in southern Germany even though amplitudes are different. The Alpine region, from France to Slovenia also shows similar patterns in Fig. 2.7(b) and 2.7(c), while the amplitudes differ. In general, the size of well imaged features in all these studies is superior to 200 km, which is within the resolution limit of ray theory and that is about 100 km based on our synthetic tests.

### 2.3.2 Phase- and group-velocity tomography

We perform a least-squares mean for our group and phase velocity data to derive the group- and phase-velocity maps shown in Fig. 2.8 and 2.10, respectively. Data are selected before inversion, leaving out group- and phase anomalies associated with a signal-to-noise ratio higher than five on cross-correlations. The inverse problem is non-unique and regularization must be applied to counter the effects of noise and of non-uniformities in data coverage. We select the weight of our regularization parameter (roughness damping only) small enough to provide a good recovery of the input model pattern (section 2.3.1), but large enough to eliminate single cell anomalies and sharp, small-scale heterogeneities that our data would not be able to resolve.
Rayleigh-wave group or phase velocity can be thought of as the weighted average of heterogeneities in shear and compressional velocities and density, over a depth range that becomes wider with increasing surface-wave period (e.g., Boschi and Ekström, 2002). The kernel functions that relate structure at depth with group and phase velocity are referred to as sensitivity functions and are discussed and illustrated e.g. by Ritzwoller et al. (2001) and Fry et al. (2010). In the following we interpret the maps of Figs. 2.8, 2.10 based on their sensitivity to structure at depth, and expected geophysical features in different depth ranges.
Figure 2.8: Group-velocity in km/s at (top, left to right) 8s, 12s, 16s, (bottom, left to right) 24s, 30s and 35s periods, superimposed on our actual tomography parameterization grid. Velocity values are only plotted at pixels where there is at least one ray crossing the pixel.
2.4.1 Rayleigh-wave group velocity at 8s and 12s periods

The propagation of 8s and 12s surface waves is strongly affected by shallow crustal structure, with the maximum peak of sensitivity at 5km depth and 8km depth respectively. One of the most prominent features of the 8s and 12s group-velocity map of Fig. 2.8 is a low-velocity anomaly spanning the sedimentary basin associated with the Po plain, a WNW-ESE basin of sediments on average about 7km thick. The Po Plain is filled by sediments originating mainly from the Alps and to a lesser degree from the Apennines reaching a maximum thickness beneath the latter. Minimal group velocities here are as low as 1.4km/s. Low velocities in sedimentary basins are expected based on the elastic properties of sediments. Another prominent feature are the Alps, which are imaged as a relative high velocity anomaly in both the 8s and 12s maps. This observation is consistent with the high shear velocities typically found at shallow depths in orogenic massifs. Lateral structure within Switzerland (the best covered area by our database) also confirms our resolution expectations: directly to the north of the Alps lies the Molasse sedimentary basin. Group velocities are quite high (up to 2.9 km/s) in Switzerland and over most of southern Germany with the exception of the Molasse basin running from Geneva to southern Bavaria (Munich). Compared to the Po Plain or other basin, sediments in the Molasse basin are more compacted, resulting in relatively fast wave propagation.

2.4.2 Rayleigh-wave group velocity at 16s

Group-velocity maps at 16s period are characterized by a number of different features with respect to shorter periods, which reflect the sensitivity of this surface-wave mode to deeper structure, while 8s and 12s waves do not sample the mantle, 16s ones do. Sensitivity of 16s Rayleigh-wave group velocity is highest around 20 km depth, i.e. in the mid/lower crust and, in some areas, the Moho. The main feature of the 16s map is the high velocity mapped throughout the Tyrrhenian sea, clearly associated with the thin oceanic crust of the area (e.g., Marone and Romanowicz, 2007; Tesauro et al., 2008; Grad and Tiira, 2009). In areas of thin crust, most surface-wave energy is focused in the mantle rather than in the crust, and the higher shear velocity in the mantle defines the speed of surface-wave propagation. In Germany and Switzerland, the geology around 20 km depth is analogous to that at shallow depths, and the 16s map accordingly shows a similar pattern to the maps of 8s and 12s. The Molasse basin in Germany is still visible which indicate that the upper crust is deep enough to affect 16s waves. Further south, the Po plain is still prominent,
consistent with the low shallow crust velocities observed in that area by Di Stefano et al. (2009). Other slow features, associated with the Apennines, are comparably important; low group velocities along the Apennines mountain range suggests that the underlying crust might be, at least locally, deeper than previously suggested. As a general rule, we find that features observed at 16s are globally a mixture between upper crust and upper mantle influence, while at shorter periods only the crust is relevant.

2.4.3 Rayleigh-wave group velocity at 24s, 30s, 35s

Longer-period group-velocity maps are overall characterized, as is to be expected, by higher velocities than their shorter-period counterparts: with growing surface-wave period, sensitivity is highest at larger depths where velocity beneath Moho exhibits a significant increase. Hence, maps in this period range are largely correlated to Moho topography depth, with anomalously high group velocity in areas of thin crust (particularly the Tyrrhenian sea), and anomalously low velocities in areas of thick crust (the Alps and the Apennines). The Apennines show lower velocity and present a more prominent signature than the Alps. The Alps are narrower and run mostly E-W than the wider Apennines that run NW-SE. With sources mostly to the north, the waves are more affected by lower velocity structure while they travel through the Apennines than through the Alps. We also compare our group velocity maps with the Moho map of the Alpine region by Wagner et al. (2011) combining LET and CSS migrated information (Waldhauser et al., 1998, 2002) displayed in Fig. 2.9. We plot only the area in common to the two studies and we adapt our colorscale with low velocity in purple and fast velocity in yellow to allow better comparison. On the Moho map the purple color indicates deep Moho and the yellow color indicates shallow Moho. We clearly see that the two maps are in good agreement. First in the Alps, where the Moho is deep, we observe a very slow velocity, with two peaks where the Moho is the deepest. Further south in Italy, beneath Emilia-Romagna and Marche, the same conclusion can be drawn though the positions of the slow velocity anomalies in our map are shifted compared to the locations of deepest Moho in the Apennines according to Wagner et al. (2011). The Adriatic Moho topography, very steep toward the west of the Apennines, is hard to image by surface waves and could create this offset. Overall the two studies obtained by two different methods and approaches are in very good agreement, which confirms the reliability of our dataset and results. The better performance of this study with respect to Stehly et al. (2009) in resolving Alpine structure at 30s period is
expected because of our better station coverage south of the Alps and more in general, the wider aperture of our array: pairs of relatively far-away stations contain the most long-period signal and will be particularly effective at providing observations of 30s waves.
2.4.4 Rayleigh-wave phase-velocity maps

Our phase-velocity maps at 16s and 30s are compared with the group-velocity map at the same period (Fig. 2.10). Based e.g. on Fig. 4 of Ritzwoller et al. (2001), group and phase velocities have different sensitivity to structure at depth. At any given period, group velocity samples a thinner and shallower layer than phase velocity and has an overall higher sensitivity to heterogeneities. This explains the difference in both amplitude and pattern between the left and right panels of Fig. 2.10. At 16s, the Alps are characterized by higher group velocity than the Apennines while phase velocity is approximately the same. To explain this difference, we show in Fig. 2.11 a synthetic cross-section along the European GeoTraverse (EGT) from southern Germany to northern Italy. In the Molasse and Po basin, group velocity is low because its sensitivity is significant within the sediment layer. On the other hand, phase velocity only ”sees” the deepest part of the thicker Po Plain. Our 30s phase-velocity map, also shown in Fig. 2.10, is clearly affected by Moho topography and by the seismic structure of mantle lithosphere (Lucente and Speranza, 2001; Lippitsch et al., 2003; Panza and Raykova, 2008), as is to be expected given the larger depth-range of sensitivity.
Figure 2.10: Group- (left) and phase- (right) velocity maps at 16s (top) and 30s (bottom) in km/s


2.5 Conclusions

We assembled a high-quality database of group and phase measurements in central Europe derived from ambient noise cross-correlation and investigated the resolution power by synthetic tests using a characteristic model. In the region of dense station coverage (< 40 km station spacing) the dataset allows to resolve features as small as 100 km or 200 km in areas of relatively poor station coverage (> 100 km station spacing). We inverted our database to obtain maps of lateral variations in group- and phase-velocity heterogeneity throughout the region covered by the data. Comparing our results with those of earlier studies, we find, in general very good correlation in region of higher station density, but our data set is able to resolve some structures in more detail. Based on sensitivity testing we define the boundaries of regions with high sensitivity by our dataset. Within this region we identify in our maps a number of robust features that can be interpreted geophysically: namely, the Po-plain sedimentary basin, the thin crust of the Tyrrhenian sea, the roots of the Alps and of the Apennines, the Molasse basin. Similar features were found, with somewhat lower resolution, in the earlier ambient noise studies of Li et al. (2010) and Stehly et al. (2009). We also compare our 30s Rayleigh wave group velocity map with the new Moho map from Wagner et al. (2011) based on CSS and LET and find excellent correlation.
between the results obtained by the two different methods. Our results at longer periods document strong sensitivity to Moho topography, with low velocity in our map corresponding to thicker than average crust. In the near future, we will be able to infer from these data the three-dimensional shear-velocity structure of the region of interest, and its pattern of azimuthal anisotropy as a function of depth, expanding the models of Stehly et al. (2009) and Fry et al. (2010). Surface-wave ambient noise is the only seismically recorded signal to sample uniformly the crust-lithosphere depth range: collecting and interpreting these observations in terms of shear-velocity structure is an important step towards the identification of a consensus tomographic model of the European crust and upper mantle.

**Acknowledgements**

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Bibliography


Abstract

We apply two independent algorithms to identify surface-wave phase velocity, as a function of frequency, from seismic ambient noise recorded at pairs of stations from a large European network. One method is based on time-domain cross correlation after filtering the signal around individual frequencies (frequency-time analysis or FTAN), the other on frequency-domain cross-correlation. In both cases, cross correlations are then ensemble-averaged over a relatively long period of time (one year). We verify that the two algorithms give essentially consistent results, and infer that phase velocity can be successfully measured through ensemble-averaging of seismic ambient noise, further validating earlier studies that had followed either approach. The description of our experiment and its results is accompanied by a detailed though simplified derivation of ambient-noise theory, writing out explicitly the relationships between surface-wave Green’s function, ambient-noise cross-correlation, and phase and group velocities.

co-authored by Lapo Boschi, Kees Weemstra, Julie Verbeke, Göran Ekström, Domenico Giardini.
3.1 Introduction

The ability to observe coherent surface-wave signal from the ensemble-averaged cross-correlation of background noise recorded at different stations is crucial to improve our resolution of Earth structure via seismic imaging. Surface waves generated by earthquakes are best observed at teleseismic distances, where the body- and surface-wave packets are well separated, and, owing to different geometrical spreading, surface waves are much more energetic than body waves; teleseismic surface waves, however, are dominated by intermediate to long periods ($\gtrsim 30s$), and their speed of propagation is therefore related to mantle, rather than crustal or lithospheric structure (e.g., Boschi and Ekström, 2002). The ensemble-averaged, cross-correlated ambient-noise signal is instead dominated by periods roughly between 5 and 30 s (e.g., Stehly et al., 2006, 2009), complementary to the period range of teleseismic surface waves, and allowing to extend imaging resolution upwards into the lithosphere-asthenosphere boundary region and the crust.

As first noted by Shapiro and Campillo (2004), the cross-correlation of seismic ambient signal recorded at two different stations approximates the Green’s function associated with a point source acting at one of the stations’ location, and a receiver deployed at the other’s. Such empirical Green’s function can then be analyzed in different ways, with the ultimate goal of obtaining information about Earth’s structure at various depths between the two stations. Most authors either extract group velocity $v_g$ from its envelope (e.g., Shapiro et al., 2005; Stehly et al., 2006, 2009), or isolate the velocity $v$ of individual phases (e.g., Lin et al., 2008; Yao et al., 2006; Ekström et al., 2009). Only the approach of Tromp et al. (2010) allows, at least in principle, to explain the entire waveform in a finite-frequency, adjoint scheme.

Both $v$ and $v_g$ are useful expressions of shallow Earth properties between seismic source and receiver, or, in the present case, between two receivers. To measure $v_g$, one must be able to identify the peak of the surface-wave envelope. This, as a general rule, is easier than isolating the carrying sinusoidal wave (i.e. measuring $v$) at a given frequency. There are, however, several issues with measuring $v_g$ that make phase-velocity observations useful and possibly preferable: (i) the envelope peak is less precisely defined than that of the carrying sinusoidal wave; (ii) at least so far as the surface-wave fundamental mode is concerned, $v_g$ depends on, and is in turn used to image, structure over a narrower and shallower depth range than $v$ (e.g., Ritzwoller et al., 2001); (iii) a $v_g$ measurement needs to be made over
a wider time window than a \( v \) measurement, and contamination by interfering phases is accordingly more likely.

While the validity of group-velocity estimates based on seismic ambient noise is widely recognized, phase velocity is more elusive. For instance, Yao et al. (2006) have noted an important discrepancy between two-station observations of phase velocity obtained from teleseismic vs. ambient signal; the need for a phase correction of \( \pi/4 \) to be applied to empirical Green’s functions before measuring their phase dispersion also has caused some confusion, and e.g. Tsai (2009) claims that its actual origin has been long misunderstood. We apply here two different approaches to measure inter-station surface-wave phase velocity from one year of continuous recording at a dense, large array of European stations, first compiled by Verbeke et al. (2012). Only one of the two approaches requires the mentioned \( \pi/4 \) phase shift. We use the consistency between the two methods’ results as a metric of the accuracy of both.

### 3.2 Theory

We study the properties of the cross-correlation \( C_{xy}(t, \omega) \), function of time \( t \) and frequency \( \omega \), of ambient surface-wave signal \( u \) recorded at two seismic instruments, located at positions \( x \) and \( y \). By definition

\[
C_{xy}(t, \omega) = \frac{1}{T} \int_{-T}^{T} u(x, \tau, \omega)u(y, t + \tau, \omega)d\tau,
\]

(3.1)

with the parameter \( T \) defining the size of the window over which the cross-correlation is computed in practice. We limit our analysis to sources sufficiently far from both receivers for the source-receiver azimuth \( \theta \) to be approximately the same. Said \( \Delta x \) the distance separating the two receivers, it then follows, as illustrated in Fig. 3.1, that the surface wave of frequency \( \omega \) and phase velocity \( v(\omega) \) generated by a single source at azimuth \( \theta \) hits the second receiver with an approximate delay

\[
t_d = \Delta x \cos(\theta)/v(\omega)
\]

(3.2)

with respect to the first.

Our treatment is similar to that of Tsai (2009), that we extend to find complete expressions for cross-correlation, and group, as well as phase-velocity of the ambient signal. Like Tsai
(2009), we assume that sources of ambient noise are far enough from our station pair for the source-receiver azimuth to be approximately the same at the two stations, and that the wavelength of ambient signal is much shorter than the distance between the stations.

A third important assumption of our and most other formulations of ambient-noise theory is that the geographic distribution of noise sources be approximately uniform with respect to azimuth, and, as a result, that the ambient signal be “diffuse”. In practice, this is not true at any moment in time, but can be at least partially achieved if the ambient signal recorded over a very long time (e.g., one year) is subdivided into shorter (e.g., one-day-long) intervals, which are then whitened and (after station-station cross-correlation) stacked. This procedure, that we shall refer to as “ensemble-averaging”, is described in detail by Bensen et al. (2007). Over time, an array of seismic stations will record ambient signal generated at all possible azimuths and distances, and the process of stacking simulates the superposition of simultaneously acting sources. Stehly et al. (2006) show that, at least in the period range \(\sim 5\text{-}15\text{s}\), most ambient-noise signal is likely to be generated by the interaction between oceans and the solid Earth (i.e., ocean storms), and the source distribution of even the stacked ambient signal is accordingly nonuniform. Yet, there are both empirical (Derode et al., 2003) and theoretical (Snieder, 2004) indications that as long as a significant fraction of ambient signal hits a receiver couple along the receiver-receiver
azimuth, ensemble-averaging will provide a sufficiently good approximation of a diffuse wavefield.

### 3.2.1 Monochromatic signal from a single source

In the absence of strong lateral heterogeneity in elastic structure, the momentum equation for a Love or Rayleigh wave can be decoupled into a differential equation in the vertical, and another in the horizontal Cartesian coordinates. The latter coincides with the Helmholtz equation and is solved by sinusoidal functions (e.g., Peter et al., 2007).

Seismic ambient noise is normally the effect of a combination of sources more-or-less randomly distributed in space and time. It is however convenient to start our treatment from the simple case of a single source generating a monochromatic signal of frequency $\omega$. The first receiver then records a signal

$$u(x, t) = S(x, \omega) \cos(\omega t + \phi),$$

where the constant phase delay $\phi$ is proportional to source-receiver distance, and the amplitude term $S(x, \omega)$ is inversely proportional, in the first approximation, to squared source-receiver distance (geometrical spreading). The signal (3.3) is observed at $y$ with a delay $t_d$, i.e.

$$u(y, t) = S(y, \omega) \cos [\omega(t - t_d) + \phi].$$

Expressions (3.3) and (3.4) are essentially the surface-wave Green’s function evaluated at $x$ and $y$.

Let us substitute (3.3) and (3.4) into (3.1), so that

$$C_{xy} = \frac{S(x)S(y)}{T} \int_{-T}^{T} \cos(\omega \tau + \phi) \cos [\omega(\tau + t - t_d) + \phi] \, d\tau.$$  \hspace{1cm} (3.5)

It is convenient to substitute $z = \omega \tau$, to find

$$C_{xy} = \frac{S(x)S(y)}{\omega T} \int_{-\omega T}^{\omega T} \cos(z + \phi) \cos [z + \phi + \omega(t - t_d)] \, dz.$$  \hspace{1cm} (3.6)
We next make use of the general trigonometric identity \( \cos(A + B) = \cos A \cos B - \sin A \sin B \), valid for any \( A, B \), and

\[
C_{xy} = \frac{S(x)S(y)}{\omega T} \int_{-\omega T}^{\omega T} \left\{ \cos^2(z) \cos [\phi + \omega(t - t_d)] \cos(\phi) \\
+ \sin^2(z) \sin [\phi + \omega(t - t_d)] \sin(\phi) \\
- \sin(z) \cos(z) \cos(\phi) \sin [\phi + \omega(t - t_d)] \\
- \sin(z) \cos(z) \sin(\phi) \cos [\phi + \omega(t - t_d)] \right\} \mathrm{d}z,
\]

which can be simplified if one notices that

\[
\int_{-\omega T}^{\omega T} \cos^2(z) \mathrm{d}z = \int_{-\omega T}^{\omega T} \frac{1 + \cos(2z)}{2} \mathrm{d}z = \omega T + \frac{1}{2} \sin(2\omega T), \tag{3.8}
\]

\[
\int_{-\omega T}^{\omega T} \sin^2(z) \mathrm{d}z = \int_{-\omega T}^{\omega T} \frac{1 - \cos(2z)}{2} \mathrm{d}z = \omega T - \frac{1}{2} \sin(2\omega T), \tag{3.9}
\]

and finally

\[
\int_{-\omega T}^{\omega T} \sin(z) \cos(z) \mathrm{d}z = \left[ \frac{\sin^2(z)}{2} \right]_{-\omega T}^{\omega T} = 0, \tag{3.10}
\]

where the notation \( [f(z)]_A^B = f(B) - f(A) \). After substituting the expressions (3.8), (3.9) and (3.10) into eq. (3.7),

\[
C_{xy} = S(x)S(y) \left\{ \left[ 1 + \frac{\sin(2\omega T)}{2\omega T} \right] \cos(\phi) \cos [\phi + \omega(t - t_d)] \\
+ \left[ 1 - \frac{\sin(2\omega T)}{2\omega T} \right] \sin(\phi) \sin [\phi + \omega(t - t_d)] \right\}. \tag{3.11}
\]

It then follows from simple trigonometric identities (cosine of the sum, sine of the sum) that

\[
C_{xy} = S(x)S(y) \left\{ \cos [\omega(t - t_d)] + \frac{\sin(2\omega T)}{2\omega T} \cos [2\phi + \omega(t - t_d)] \right\}. \tag{3.12}
\]

Following Tsai (2009), or all other authors conducting ambient-noise analysis in the time domain, we next assume that the wavelength of the surface waves in question is much shorter than interstation distance, or in other words that the time it takes the wave to
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travel between the two stations is much larger than its period, i.e. \( T \gg 1/\omega \), so that 
\[ 2\omega T \gg 1 \]
Eq. (3.12) then reduces to
\[
C_{xy} \approx S(x)S(y) \cos[\omega(t - t_d)]
\]  
(3.13)
(compare with eq. (1) of Tsai (2009)). From eq. (3.13) we infer that the station-station 
cross-correlation of a “ballistic” signal, i.e. generated by a single source localized in space, 
and not scattered, is only useful if the location of the source is known. It coincides (once 
amplitude is normalized) with the Green’s function between the two stations if and only 
if the two stations are aligned with the source, i.e. azimuth \( \theta = 0 \) or \( \theta = \pi \), so that 
\( t_d = \Delta x/v \).

3.2.2 Monochromatic signal from a discrete set sources

Recorded seismic ambient noise is believed to be the cumulative effect of numerous localized sources, distributed almost randomly all around our couple of recording instruments. The signal generated by a discrete set of monochromatic sources can be written as a superposition of single-source signals, eqs. (3.3) and (3.4), resulting in
\[
u(x,t) = \sum_i S_i(x,\omega) \cos(\omega t + \phi_i) \]  
(3.14)
and
\[
u(y,t) = \sum_i S_i(y,\omega) \cos[\omega(t - t_d^i) + \phi_i], \]  
(3.15)
where the summation is over the sources, \( \phi_i \) is the phase delay associated with source \( i \), 
and the time delay \( t_d^i \) between stations \( x \) and \( y \) also changes with source azimuth, hence 
the superscript \( i \). In analogy with sec. 3.2.1, we next substitute (3.14) and (3.15) into 
(3.1), and
\[
C_{xy} = \frac{1}{T} \sum_{i,k} \left\{ S_i(x)S_k(y) \int_{-T}^{T} \cos(\omega \tau + \phi_i) \cos[\omega(\tau + t - t_d^i) + \phi_k] \, d\tau \right\}.
\]  
(3.16)
At this point we make the important assumption, common to all derivations of noise-correlation properties, that the cross-correlation is performed over a time series sufficiently long for all “cross-terms” (products \( \cos(\omega \tau + \phi_i) \cos[\omega(\tau + t - t_d^i) + \phi_k] \) with \( i \neq k \)) in eq. (3.16) to cancel out (e.g., Snieder, 2004; Tsai, 2009). We are then left with a sum
of integrals of the form (3.5), which we have proved in sec. 3.2.1 to be approximated by (3.13), so that

\[ C_{xy} \approx \sum_i S_i(x)S_i(y) \cos \left[ \omega(t - t^i_d) \right]. \]  

(3.17)

3.2.3 Continuous distribution of sources

Eq. (3.17) can be further generalized to the case of a continuous distribution of sources,

\[ C_{xy} \approx \int \frac{\Delta x}{v} \rho(t_d, \omega) \cos \left[ \omega (t - t_d) \right] dt_d, \]

(3.18)

where we have introduced the function \( \rho(t_d, \omega) \), describing the density of sources as a function of inter-station delay \( t_d \), or, which is the same (recall eq. (3.2)), azimuth \( \theta \).

Integration is accordingly over \( t_d \), and the integration limits correspond, through eq. (3.2), to the interval of possible azimuths, from 0 to \( \pi \). \( \rho \) is also a function of \( \omega \), as signal generated by differently located sources generally has a different frequency content. We have incorporated the continuous version of the source term \( S_i(x, \omega)S_i(y, \omega) \) from eq. (3.17) in the source density function \( \rho(t_d, \omega) \).

In analogy with all earlier formulations of ambient-noise theory, we require the source distribution to be uniform with respect to azimuth \( \theta \). This implies that

\[ \frac{1}{\pi} g(\omega)f(\theta)d\theta = \rho[t_d(\theta), \omega]f[\theta(t_d)] dt_d, \]

(3.19)

for an arbitrary function \( f \), with \( 1/\pi \) is the normalized value of uniform azimuthal density. The factor \( g(\omega) \) serves to remind us that source amplitude generally changes with frequency. After dividing both sides by \( f(\theta) \) and replacing \( dt_d = -\Delta x \sin(\theta)dv/v \),

\[ \frac{g(\omega)}{\pi} d\theta = -\rho(t_d) \frac{\Delta x \sin(\theta)}{v} d\theta, \]

(3.20)

or

\[ \rho(t_d, \omega) = -\frac{v(\omega)g(\omega)}{\pi \Delta x \sin(\theta)}, \]

(3.21)

which is the expression of \( \rho = \rho(t_d, \omega) \) corresponding to azimuthally uniform source density.
3.2.4 Cross-correlation and Green’s function

It is convenient to separate the integral in eq. (3.18) into two integrals, one over positive, and the other over negative $t_d$,

$$C_{xy} \approx \int_{-\Delta x/v}^{0} \rho(t_d, \omega) \cos[\omega(t - t_d)] dt_d + \int_{0}^{\Delta x/v} \rho(t_d, \omega) \cos[\omega(t - t_d)] dt_d. \quad (3.22)$$

Let us first consider the second term ($t_d \geq 0$) at the right-hand side of (3.22), which can be rewritten

$$C_{t_d>0} \approx \Re \left[ e^{-i\omega t} \int_{0}^{\Delta x/v} \rho(t_d, \omega) e^{i\omega t_d} dt_d \right], \quad (3.23)$$

where $\Re(\ldots)$ equals the real part of its argument. It is convenient to replace $\rho(t_d, \omega)$ with its expression (3.21), and the integration variable $t_d$ with $\theta$. By differentiating eq. (3.2), $dt_d = -\Delta x \sin(\theta)/v$, while the limits of integration $0, \Delta x/v$ correspond to azimuth $\theta = \pi/2, 0$, respectively, hence

$$C_{t_d>0} \approx \Re \left[ -\frac{g(\omega)}{\pi} e^{-i\omega t} \int_{0}^{\pi/2} e^{i\omega \Delta x \cos(\theta)/v} d\theta \right]. \quad (3.24)$$

Clearly, positive $t_d$ corresponds to azimuth $0 < \theta < \pi/2$, i.e. to energy propagating from $x$ to $y$; the opposite holds for the $t_d \leq 0$ term ($\pi/2 < \theta < \pi$). We next rewrite the integral in terms of Bessel and Struve functions. Let us first consider the 0-order Bessel function of the first kind in its integral form

$$J_0(z) = \frac{1}{\pi} \int_{0}^{\pi} \cos(z \sin(\theta)) d\theta \quad (3.25)$$
(eq. (9.1.18) of Abramowitz and Stegun (1964)). The integral from 0 to $\pi$ in (3.25) can be transformed into an integral from 0 to $\pi/2$:

$$J_0(z) = \frac{1}{\pi} \int_0^\pi \cos(z \sin(\theta)) d\theta$$

$$= \frac{1}{\pi} \left[ \int_0^{\pi/2} \cos(z \sin(\theta)) d\theta + \int_{\pi/2}^\pi \cos(z \sin(\theta)) d\theta \right]$$

$$= \frac{1}{\pi} \left[ \int_0^{\pi/2} \cos(z \sin(\theta)) d\theta - \int_0^{0} \cos(z \sin(\pi - \theta')) d\theta' \right]$$

$$= \frac{1}{\pi} \left[ \int_0^{\pi/2} \cos(z \sin(\theta)) d\theta + \int_{\pi/2}^{\pi} \cos(z \sin(\theta')) d\theta' \right]$$

$$= \frac{2}{\pi} \int_0^{\pi/2} \cos(z \sin(\theta)) d\theta. \quad (3.26)$$

We then replace $\sin(\theta)$ with $\cos(\theta + \pi/2)$ and change the integration variable $\theta = \pi/2 - \theta'$,

$$J_0(z) = \frac{2}{\pi} \int_0^{\pi/2} \cos \left( z \cos \left( \theta + \frac{\pi}{2} \right) \right) d\theta$$

$$= -\frac{2}{\pi} \int_0^{\pi} \cos \left( z \cos \left( \pi - \theta' \right) \right) d\theta' \quad (3.27)$$

$$= \frac{2}{\pi} \int_0^{\pi/2} \cos \left( z \cos \left( \theta' \right) \right) d\theta',$$

and after substituting $z$ with $\omega \Delta x / v$,

$$J_0 \left( \frac{\omega \Delta x}{v} \right) = \frac{2}{\pi} \int_0^{\pi/2} \cos \left( \frac{\omega \Delta x}{v} \cos (\theta) \right) d\theta$$

$$= \frac{2}{\pi} \Re \left[ \int_0^{\pi/2} e^{i\omega \Delta x \cos(\theta)/v} d\theta \right]. \quad (3.28)$$

The 0-order Struve function also has an integral form

$$H_0(z) = \frac{2}{\pi} \int_0^{\pi/2} \sin(z \cos(\theta)) d\theta, \quad (3.29)$$
which coincides with eq. (12.1.7) of Abramowitz and Stegun (1964) at order 0 and substituting \( \Gamma(1/2) = \sqrt{\pi} \), with \( \Gamma \) denoting the Gamma function. We replace, again, \( z \) with \( \omega \Delta x/v \), and

\[
H_0 \left( \frac{\omega \Delta x}{v} \right) = \frac{2}{\pi} \Im \left[ \int_0^{\pi/2} e^{i \omega \Delta x \cos(\theta)/v} d\theta \right],
\]

(3.30)

with the operator \( \Im \) mapping complex numbers to their imaginary part. It follows from (3.28) and (3.30) that

\[
\int_0^{\pi/2} e^{i \omega \Delta x \cos(\theta)/v} d\theta = \frac{\pi}{2} \left[ J_0 \left( \frac{\omega \Delta x}{v} \right) + i H_0 \left( \frac{\omega \Delta x}{v} \right) \right],
\]

(3.31)

and substituting into (3.24):

\[
C_{t_d>0} \approx \Re \left\{ -\frac{g(\omega)e^{-i\omega t}}{2} \left[ J_0 \left( \frac{\omega \Delta x}{v} \right) + i H_0 \left( \frac{\omega \Delta x}{v} \right) \right] \right\}.
\]

(3.32)

This expression is further simplified by requiring, again (see section 3.2.1), that interstation distance be much larger than the wavelength of the signal under consideration, i.e. \( \omega \Delta x/v \gg 1 \). It then follows from eq. (9.2.1) of Abramowitz and Stegun (1964) that

\[
J_0 \left( \frac{\omega \Delta x}{v} \right) \approx \sqrt{\frac{2v}{\omega \pi \Delta x}} \cos \left( \frac{\omega \Delta x}{v} - \frac{\pi}{4} \right),
\]

(3.33)

and from eqs. (12.1.34) and (9.2.2) of Abramowitz and Stegun (1964),

\[
H_0 \left( \frac{\omega \Delta x}{v} \right) \approx Y_0 \left( \frac{\omega \Delta x}{v} \right) \approx \sqrt{\frac{2v}{\omega \pi \Delta x}} \sin \left( \frac{\omega \Delta x}{v} - \frac{\pi}{4} \right),
\]

(3.34)

with \( Y_0 \) denoting the 0-order Bessel function of the second kind. Substituting equations (3.33) and (3.34) into (3.32),

\[
C_{t_d>0} \approx \Re \left\{ -g(\omega)e^{-i\omega t} \sqrt{\frac{v}{2\pi \omega \Delta x}} \left[ e^{i(\omega \Delta x/v-\pi/4)} \right] \right\},
\]

(3.35)
and it follows that the cross-correlation $C_{xy}^{t_d>0}$ is proportional to the Green’s function between the two stations, separated by a time delay $\Delta x/v$, provided that a phase correction of $\pi/4$ is applied, i.e.,

$$C_{xy}^{t_d>0} \approx -g(\omega)\sqrt{\frac{v}{2\pi\omega\Delta x}} \cos [\omega (\Delta x/v - t) - \pi/4]$$  \ \ \ (3.36)$$

(the cross-correlation of noise is delayed of $\pi/(4\omega)$ with the respect to the Green’s function).

An analogous treatment applies to the negative-time cross-correlation $C_{xy}^{t_d<0}$, i.e. the first term at the right hand side of eq. (3.22), which after after the variable change from $t_d$ to $\theta$ becomes

$$C_{xy}^{t_d<0} \approx \Re \left[ -\frac{g(\omega)e^{-i\omega t}}{\pi} \int_{\pi/2}^{\pi/2} e^{i\omega \Delta x \cos(\theta)/v} d\theta \right].$$  \ \ \ (3.37)$$

To express also this integral in terms of Bessel and Struve functions, we first notice that

$$\int_{\pi/2}^{\pi/2} f(\cos(\theta))d\theta = -\int_{\pi/2}^{\pi/2} f(\cos(\theta))d\theta$$

$$= -\int_{0}^{\pi} f\left( \cos \left( \psi + \frac{\pi}{2} \right) \right) d\psi$$

$$= -\int_{0}^{\pi} f\left( \cos(\psi) \cos \left( \frac{\pi}{2} \right) - \sin(\psi) \sin \left( \frac{\pi}{2} \right) \right) d\psi$$

$$= -\int_{0}^{\pi} f(\sin(\psi)) d\psi,$$

for an arbitrary function $f$. From eq. (3.37) it follows that

$$C_{xy}^{t_d<0} \approx -\Re \left[ -\frac{g(\omega)e^{-i\omega t}}{\pi} \int_{0}^{\pi} e^{-i\omega \Delta x \sin(\theta)/v} d\theta \right].$$  \ \ \ (3.39)$$

Making use of eq. (3.26) and some simple algebra applied to the definition of Struve function (3.29), the above, positive-time treatment can essentially be repeated, to give

$$C_{xy}^{t_d<0} \approx \Re \left\{ -\frac{g(\omega)e^{-i\omega t}}{2} \left[ J_0 \left( \frac{\omega \Delta x}{v} \right) - iH_0 \left( \frac{\omega \Delta x}{v} \right) \right] \right\},$$  \ \ \ (3.40)$$
where only the sign of $H_0$ at the right-hand side has changed with respect to eq. (3.32). We conclude that

$$C_{xy}^{t<0} \approx -g(\omega) \sqrt{\frac{v}{2\pi \omega \Delta x}} \cos \left[ \omega \left( -\Delta x/v - t \right) + \pi/4 \right],$$

(3.41)

i.e. the negative-time phase-shift is symmetric to the positive-time one, in agreement with Tsai (2009).

### 3.2.5 Group and phase velocity

We next consider the more general case of a seismogram formed by the superposition of surface waves with different frequencies. Let us start with our expression (3.36) for the cross-correlated signal, grouping the amplitude terms in a generic factor $S(\omega)$. We then find the mathematical expression of a surface-wave packet by (i) discretizing the frequency band of interest into a set of closely-spaced frequencies $\omega_i$ identified by the subscript $i$, and (ii) combining different-frequency contributions by integration around each frequency $\omega_i$ and summation over $i$, so that

$$u(x, t) = \sum_{i=1}^{\infty} \int_{\omega_i - \varepsilon}^{\omega_i + \varepsilon} S(x, \omega) \cos \left[ \omega \left( \frac{\Delta x}{v(\omega)} - t \right) - \frac{\pi}{4} \right] d\omega,$$

(3.42)

where $\varepsilon \ll \omega_i$. It is convenient to introduce the notation $\psi = \omega(\Delta x/v - t)$, and, since $\varepsilon$ is small, replace it with its Taylor expansion around $\omega_i$, i.e.

$$\psi(\omega) \approx \psi(\omega_i) + (\omega - \omega_i) \left[ \frac{d\psi}{d\omega} \right]_{\omega_i},$$

(3.43)

where $[f(\omega)]_{\omega_i}$ denotes the value of any function $f$ evaluated at $\omega = \omega_i$. We rewrite eq. (3.42) accordingly, and find after some algebra that the integral at its right hand side

$$\int_{\omega_i - \varepsilon}^{\omega_i + \varepsilon} S(\omega) \cos \left[ \omega \left( \frac{\Delta x}{v(\omega)} - t \right) \right] d\omega \approx S(\omega_i) \cos \left[ \psi(\omega_i) \right] \frac{2}{v(\omega_i)} \left\{ \varepsilon \left[ \frac{d\psi}{d\omega} \right]_{\omega_i} \right\}.$$  

(3.44)

If one introduces a function

$$v_g(\omega) = \frac{v(\omega)}{1 - \frac{\omega}{v(\omega)} \frac{dv}{d\omega}},$$

(3.45)
it follows that \[ \frac{d\psi}{d\omega} \] takes the compact form

\[
\left[ \frac{d\psi}{d\omega} \right]_{\omega_i} = \frac{\Delta x}{v_g(\omega_i)} - t; \quad (3.46)
\]

we finally substitute it into (3.44) and substitute the resulting expression into (3.42), to find

\[
u(x, t) = \sum_{i=1}^{\infty} S(\omega_i) \cos \left[ \omega_i \left( \frac{\Delta x}{v(\omega_i)} - t \right) - \frac{\pi}{4} \right] \frac{2 \sin \left[ \varepsilon \left( \frac{\Delta x}{v_g(\omega_i)} - t \right) \right]}{\Delta x \frac{v_g(\omega_i)}{\omega_i} - t}. \quad (3.47)
\]

Each term at the right-hand side of eq. (3.47) is the product of a wave of frequency \( \omega \) and speed \( v(\omega) \) with one of frequency \( \varepsilon \ll \omega \) and speed \( v_g(\omega_i) \). The latter factor, with much lower frequency, modulates the signal, and we call “group velocity” its speed \( v_g \), which coincides with the speed of the envelope of the signal. Eq. (3.45) shows that, in the absence of dispersion (i.e. \( \frac{dc}{d\omega} = 0 \)) phase and group velocities coincide. In practice, the values of \( v \) and \( v_g \) are always comparable, and the large difference in frequency results in a large difference in the wavelength of the phase and group terms.

Comparing eq. (3.47) to (3.42), it is important to notice that when phase velocity is measured from the station-station cross-correlation of ambient signal, a phase correction of \( \pi/4 \) must first be applied; the same is not true for group-velocity measurements.

### 3.3 How to measure phase velocity

To evaluate whether phase velocity can be accurately observed in the ensemble-averaged cross correlation of ambient noise, we use two completely independent approaches to measure it from the same data. Consistency of the results is then a strong indication of their validity. Frequency-time analysis (FTAN) (section 3.3.1) consists essentially of cross-correlating the surface-wave signal (\( \Delta t \)-long records of ambient signal in our case) to then compute phase velocity via a simple relationship involving real and imaginary part of the phase spectrum (Sato, 1955); this requires that a \( \pi/4 \) correction be applied to the data as explained in section 3.2.5. The other approach we consider is based on the result of Aki (1957), confirmed by Ekström et al. (2009) for the frequency range of interest, that the spectrum of the two-station cross-correlation of seismic ambient noise should approxi-
mately coincide with a 0-order Bessel function of the first kind (section 3.3.2); in this case, no $\pi/4$ correction needs to be applied.

### 3.3.1 Frequency-time analysis

The procedure of ensemble-averaging of ambient signal is described in detail, e.g., by Bensen et al. (2007); essentially, a long (e.g., one year) continuous seismic record is subdivided into shorter $\Delta t$ intervals. The records are whitened so that the effects of possible ballistic signal (i.e., large earthquakes) present in the data are minimized. The cross-correlation between simultaneous $\Delta t$-long records from different stations is then computed for all available $\Delta t$ intervals, and the results for each station couple are stacked over the entire year. As explained in detail above, the result is proportional, within a certain approximation, to the combination of causal and anti-causal Green’s functions associated with the station couple, except for a phase shift of $\pi/4$.

Bensen et al. (2007) measure group velocity from noise cross-correlations, and suggest that phase dispersion can be obtained by integration of group dispersion curves. This approach however is not sufficient identify phase velocity uniquely. Meier et al. (2004) provide an algorithm to derive phase velocity from the cross-correlation of teleseismic signals recorded by stations aligned with the earthquake azimuth (as explained in section 3.2.1, for this source-station geometry the cross-correlation of single-source, or “ballistic” records is also proportional to the Green’s function). Fry et al. (2010) and Verbeke et al. (2011) show that the algorithm of Meier et al. (2004) can be successfully applied to the ambient signal recorded at a regional-scale array of broadband stations. This approach is generally referred to as FTAN. In the following we shall analyze a subset of the phase-dispersion database compiled by Verbeke et al. (2011) via their own automated implementation of FTAN.

The phase-velocity measurements of Verbeke et al. (2011) are limited to the 0.02-0.1 Hz frequency range, where seismic ambient noise is known to be strong (Stehly et al., 2009), most likely as an effect of ocean storms and the coupling between oceans and the solid Earth (Stehly et al., 2006). Frequency is discretized with increments whose length increases with increasing frequency (from 0.02 to 0.05 Hz). For each discrete frequency value, ensemble-averaged cross-correlations are (i) band-pass filtered around the frequency in question and (ii) windowed in the time-domain via a Gaussian window centered around the time of maximum amplitude of (filtered) cross-correlation. Causal and anticausal parts are folded together (i.e. stacked after reversing the time-dependence of the anticausal one).
resulting time series is Fourier-transformed, and a simple analytical relationship, eq. (3) of Meier et al. (2004), is implemented to find the phase velocity at that frequency from the phase spectrum. Phase velocity is only known up to a $2\pi n$ “multiple cycle ambiguity”, with $n = 0, \pm 1, \pm 2, \ldots$. After iterating over the entire frequency band, an array of dispersion curves is found, each corresponding to a value of $n$. Verbeke et al. (2011) compare each curve (for all integer values of $n$ between -5 and 5) with phase velocity as predicted by PREM (Dziewonski and Anderson, 1981), and pick the one closest to PREM.

### 3.3.2 Frequency-domain cross-correlations and Bessel-function fitting

A completely different method (hereafter referred to as “AKI”) to extrapolate phase velocity from the ambient signal recorded at two stations is proposed by Ekström et al. (2009), based on much earlier work by Aki (1957). The theoretical basis of this method has been recently rederived by Tsai (2010). In the approach of Aki (1957), ambient signal
recorded over a long time (e.g., one year) is, again, subdivided into shorter $\Delta t$ intervals. Let us call $p_i(\omega)$ the frequency spectrum associated with a $\Delta t$-long record at station $i$ (Fig. 3.3a, with $\Delta t = 2$ hours). After whitening, this is multiplied with the simultaneous $\Delta t$-long recording made at another station $j$ (Fig. 3.3b), resulting in the cross-spectrum, or spectrum of the cross-correlation between the two $\Delta t$-long records (Fig. 3.3c). This procedure is repeated for all available $\Delta t$-intervals in the year, which are then stacked together, i.e. “ensemble-averaged” (Fig. 3.3d). The resulting quantity is usually referred to as “coherence”. According to Aki (1957),

$$\left\langle \frac{p_i p_j^*}{|p_i p_j^*|} \right\rangle = J_0 \left( \frac{\omega \Delta x}{v(\omega)} \right),$$

(3.48)

where $< ... >$ denotes ensemble averaging, the left-hand side is precisely what we call coherence, and the superscript $*$ marks the complex conjugate of a complex number. The quantities at the right-hand side of (3.48) are defined as in section ?? above, with $\Delta x$ distance between stations $i$ and $j$.

Eq. (3.48) can be used to determine phase dispersion. In practice, observed coherence is first of all plotted as a function of frequency (i.e., the ensemble-averaged cross-spectrum is plotted). Values $\omega_i$ ($i = 1, 2, 3, ...$) of frequency for which coherence is zero are identified. If $\omega = \omega_i$ for some $i$, the argument of (3.48) must coincide with one of the known zeros $z_n$ ($n = 1, 2, ...$) of the Bessel function $J_0$,

$$\frac{\omega_i \Delta x}{v(\omega)} = z_n.$$  

(3.49)

Eq. (3.49) can be solved for $v$,

$$v(\omega_i) = \frac{\omega_i \Delta x}{z_n},$$

(3.50)

and we now have an array of possible measures of phase velocity at the frequency $\omega_i$, each corresponding to a different value of $n$. Implementing (3.50) at all observed values of $\omega_i$, an array of dispersion curves is found. Much like in the case of FTAN (section 3.3.1), a criterion must then be established to select a unique curve.

Importantly, the observation of $\omega_i$ on ensemble-averaged cross-spectra like the one of Fig. (3.3d) is complicated by small oscillations that can be attributed to instrumental noise or inaccuracies related to anisotropy in the source distribution. Before identifying $\omega_i$, we determine the linear combination of cubic splines that best fits (in least-squares
sense, via the LSQR algorithm of Paige and Saunders (1982)) observed coherence. Splines are equally spaced, and spacing must be selected so that “splined” coherence is sufficiently smooth (Fig. 3.3d).

It is practical to focus the analysis on zero crossings, rather than measuring the overall fit between $J_0$ and measured coherence, because the latter depends on the power spectrum of the noise sources, of which we know very little, and can be affected importantly by data processing (Ekström et al., 2009).

3.4 Application to central European data and cross-validation of the two methods

Fig. 3.4 shows the set of randomly selected ∼1000 station couples from Verbeke et al. (2011) that we shall analyze here. The corresponding phase-velocity dispersion curves were measured by Verbeke et al. (2011) following the procedure of section 3.3.1, after subdividing the entire year 2006 into day-long intervals and ensemble-averaging the resulting day-long cross-correlations.

We apply the AKI method of section 3.3.2 to continuous records associated with the station couples of Fig. 3.4. Our implementation was originally designed for reservoir-scale application (Weemstra et al., 2011), but could be applied to our continent-scale array of data after only minor modifications. For each station, continuous recording for the entire year 2006 is subdivided into intervals of $\Delta t = 2$ hours, with a very conservative 75% overlap between neighboring intervals to make sure that no coherent signal traveling from station to station is neglected (Weemstra et al., 2011). This results in as many as 45 spectra per day.

In Fig. 3.5 we compare our new phase-velocity measurements with those of Verbeke et al. (2011) for one sample station couples. A visual analysis (which we repeated on a number of different station couples) suggests that the two methods provide compatible results.

To evaluate quantitatively their level of consistency, we first expand FTAN dispersion curves over a set of cubic splines, and apply spline interpolation to estimate FTAN-based phase-velocity values at the exact frequencies (associated with zero-crossings of the Bessel function) where AKI measurements are available; we subtract the AKI phase velocities from the FTAN ones interpolated at the same frequency; we count the number of discrepancy
Figure 3.3: Illustration of the AKI approach. (a) Power spectrum obtained Fourier-transforming two hours of ambient recording at station TORNY. (b) Spectrum from the very same two hours, station VDL. (c) Product of the two spectra (coinciding with the spectrum of the cross-correlation of the two time-domain signals) obtained after whitening both. (d) Results of ensemble-averaging an entire year of spectra like the one at (c), for the same two stations: blue squares and red circles identify values of real and imaginary parts found at various frequencies; the black solid line is the linear combination of cubic splines that best-fits the observed real part of the spectrum.
Figure 3.4: (A) Subset of European stations (circles) from Verbeke et al. (2011) that are also included in our analysis. We only compare phase-velocity measurements associated with ~1000 station couples connected by solid lines. (B) Distribution of epicentral-distance values sampled by the data set at A.
Figure 3.5: For stations TORNY and VDL, the selected FTAN phase-velocity dispersion measurements (black circles, connected by a black line) compared with analogous frequency-domain (AKI) measurements. We have not yet implemented an algorithm for automatic selection of a preferred AKI dispersion curve, but the FTAN curve clearly fits a single branch of AKI datapoints.
Figure 3.6: Frequency of observed phase-velocity misfit (AKI values subtracted from FTAN ones) for the total set of \( \sim \)1000 analyzed station couples. The mean is 13 m/s and the standard deviation is 151 m/s.

observations, independent of frequency, falling in each of a set of 50 m/s intervals, and plot the associated histogram in Fig. 3.6. Both mean and standard deviation of the FTAN-AKI discrepancy are small (13 m/s and 151 m/s, respectively), and we infer that the two approaches can be considered consistent. Outliers exist with misfit stronger than \( \pm 1000 \) m/s, but they would not be visible in Fig. 3.6 even after extending the horizontal-axis range.

We next analyze the dependence of FTAN-AKI discrepancy on interstation distance, through a second histogram (Fig. 3.7a) where the misfit is averaged within \( \sim 0.3^\circ \) interstation-distance bins. In Fig. 3.7b the misfit is likewise averaged within 0.01-Hz increments spanning the whole frequency range of interest. Fig. 3.7a shows that FTAN has a tendency to give slightly higher velocity estimates with respect to AKI; this effect is reversed at very small and very large interstation distances; at very short epicentral distances,
where the causal and anticausal parts of time-domain cross-correlation tend to overlapped, AKI is indeed expected to outperform FTAN, at least at low frequencies (Ekström et al., 2009); at both short and long epicentral distances, relatively few observations are available (Fig. 3.4b), so that the corresponding averages in Fig. 3.7a are presumably less robust. More interestingly, Fig. 3.7b shows that misfit is systematically smaller ($\lesssim 20$ m/s) at relatively high frequencies ($\gtrsim 0.04$ Hz) than it is at low frequencies of $\sim 0.02-0.03$ Hz. This is consistent with the observation of e.g. Stehly et al. (2006) that most seismic-ambient-noise energy is found at frequencies around 0.1 Hz, where ambient-noise-based observations of surface waves are consequently more reliable. We infer that the growth in discrepancy with decreasing frequency in Fig. 3.7b most likely does not reflect inaccuracies in either the AKI or FTAN methods, but rather the difficulty of finding coherent signal in the absence of a sufficiently strong ambient wavefield.

Overall, averaged discrepancies in Fig. 3.7 remain $\lesssim 50$ m/s, with the exception of the lowest considered frequencies, where averaged values can be as high as $\sim 100$ m/s. We take this as an indication that the AKI and FTAN methods provide essentially consistent results, and we infer that such results can be considered reliable.

### 3.5 Conclusions

With this study we have conducted a detailed derivation of the mathematical relationship between ensemble-average cross-correlation of seismic ambient noise recorded at two seismic stations, and the surface-wave Green’s function associated with the locations of those stations. Our treatment is similar to that of, e.g., Tsai (2009), but we additionally provide explicit expressions for both empirical Green’s function (not just its phase) and its group velocity. Our formulation is approximate, and only strictly valid in the assumptions that sources of ambient noise be far from the receiver array and uniformly distributed at all source-receiver azimuths, and that the wavelength of seismic signal be much larger than that of Earth heterogeneity (ray-theory approximation).

Based on these theoretical results, we have applied to different algorithms, FTAN (e.g., Meier et al., 2004; Bensen et al., 2007) and the frequency-domain method of Aki (1957) and Ekström et al. (2009), to measure Rayleigh-wave phase velocity, as a function of frequency, from a year of seismic noise recorded at a dense array of European stations (Verbeke et al., 2012, 2011). A $\pi/4$ phase shift, known empirically and whose physical meaning
Figure 3.7: FTAN-AKI phase-velocity misfit, for the total set of \(~1000\) analyzed station couples, averaged within (a) \(~0.3^\circ\) interstation-distance bins, and (b) \(0.01\)-Hz frequency bins.
is discussed by Tsai (2009), must be applied to the data only if phase velocity is to be extracted in frequency-time analysis (FTAN) approach. The two approaches, albeit very different, provide consistent results, and we infer that Rayleigh-wave phase velocity can be successfully extracted, via ensemble averaging, from continuous recordings of seismic ambient noise, at least within the frequency range $\sim 0.03$-0.1 Hz analyzed here.
Bibliography


Abstract

Measurements of relatively short-period surface-wave dispersion, extracted from seismic ambient noise, are the best tool available to map shear-velocity structure of the Earth’s crust and lithosphere, asthenosphere system at the regional scale, particularly in regions of moderate seismic activity like central Europe. Previous work on data limited to Switzerland yielded tomographic images of layered azimuthal anisotropy in the western Alpine crustal root. We repeat this experiment after extending the available set of ambient-noise cross-correlations to stations from the Italian broadband network and in general from the Orfeus database.
The addition of new data, and the subsequent increase in the aperture of the resolved region, allow us to resolve azimuthal anisotropy more robustly and over a wider area, in particular, also outside the Alps. Based on mineral physics understanding of the dynamical origin of lower crust and mantle anisotropy, our results shed further light on tectonic processes taking place in the area: e.g., Alpine crustal root resulting from stacking of lower continental crustal detached from mantle lithosphere, toroidal asthenosphere flow associated with ongoing subduction and slab retreat.

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4.1 Introduction

In recent years, a number of methodological advances in seismology have made it possible to derive increasingly high-resolution tomographic images of the radial (e.g., Moschetti et al., 2010; Schaefer et al., 2011) and azimuthal (e.g., Becker et al., 2007; Lin et al., 2011; Fry et al., 2010) seismic anisotropy of the Earth’s crust and upper mantle. The seismic anisotropy we observe is most likely an expression of the alignment of mineral fabrics, which in turn form as a consequence of deformation and strain (e.g., Christensen, 1984; Nicolas and Christensen, 1987). Measures of anisotropy can therefore be used to infer deformation within the mantle and crust. Determining current and past deformation within the lithosphere and upper mantle enables one to decipher tectonic evolution of plates at both local and global scale. Using anisotropy tomography the detailed features of plate boundaries, exhibiting a complex pattern of deformation, can be mapped more accurately than by means of simple isotropic imaging. In the mantle, anisotropy may be connected to mantle flow (e.g., Montagner, 2007; Long and Becker, 2010; Schaefer et al., 2011). Seismic anisotropy only provides information at a single instant in time, but in combination with rheological and geological information, one can decouple present and past anisotropy (Deschamps et al., 2008; Long and Becker, 2010; Schaefer et al., 2011).

As alternatives to the seismic techniques discussed above, deformation can also be measured via GPS (McClusky et al., 2000; Sella et al., 2002; Calais et al., 2003; Kreemer et al., 2003; Serpelloni et al., 2007; Altamimi et al., 2011, e.g.,) and classical structural-geology studies. Both approaches provide robust measurements of surface deformation (i.e. upper crust), but with the important limitation of not allowing any estimate of the deformation field at depth. This can only be achieved via seismic-based techniques, and in particular either SKS shear-wave splitting (e.g., Silver, 1996; Silver and Holt, 2002), or isotropic seismic imaging (via controlled source seismology (CSS) (e.g., Waldhauser et al., 2002, 1998) or local earthquake tomography (LET) (e.g., Diehl et al., 2009; Di Stefano et al., 2009)). SKS observations provide a global measure of crust and/or mantle anisotropy, whose specific contributions, however, cannot be isolated. CSS is sensitive to strong velocity contrasts, and thus can be used to robustly detect material interfaces, such as the Moho discontinuity, which are naturally related to strain within their depth range. The geometry of interfaces as mapped by CSS is however only indirectly linked to deformation within different crustal or lithospheric layers (Mckenzie and Jackson, 1986; Schmalholz et al., 2009).
Compensating some limitations and complementing results of the imaging methods described above, tomographic mapping of anisotropy (both azimuthal and radial) makes it possible to (i) establish a complete volumetric representation of the deformation field provided the prescription of appropriate depth-sensitivity kernels, and (ii) to distinguish between deformation within crust and upper mantle.

Whilst imaging anisotropy is not sufficient to determine the origin of the deformation, or constitutive behavior of the material, it does provide the overall kinematics of a region. Furthermore, in combination with our understanding of the rheology and composition of the mantle and crust, it helps to infer long-term mechanisms explaining the deformation of the lithosphere, and connect them with the severe deformation within the mantle.

With this study, we contribute to the understanding of Alpine and central European tectonics through seismic imaging of the region anisotropy. The dynamics of central Europe is controlled by the convergence of the European and African plates, resulting in a multitude of micro-plates (Schmid et al., 2004; Faccenna and Becker, 2010; Boschi et al., 2010), and in large-scale subduction (confirmed by tomography) both at European scale (e.g., Piro- mallo and Morelli, 2003; Spakman et al., 1993; Wortel and Spakman, 2000) and more local (Lippitsch et al., 2003) scales. Few authors have attempted to map seismic anisotropy in Europe: Jolivet et al. (2009); Barruol et al. (2011) estimated it on the basis SKS shear-wave splitting observations, while Fry et al. (2010) used ambient-noise data to mapped azimuthal anisotropy beneath the central Alps. The latter study imaged both lateral and depth variations in crustal anisotropy under the Alps, and interpreted it as caused by differences in the mineral fabrics of European lower crust and lithosphere, and the Helvetic and Penninnic nappes. It should be noted that, SKS-related anisotropy, associated with approximately vertically propagating body waves, cannot be directly compared with the fast azimuth of surface-wave propagation. Mainly since the source of anisotropy imaged with SKS could take his origin at any depth along ray path which travelled through the mantle and the crust (Montagner et al., 2000; Wuestefeld et al., 2009).

We conduct tomographic imaging of anisotropy in central Europe using the high-resolution database of ensemble-averaged European ambient-noise cross-correlations compiled by Verbeke et al. (2012). With respect to the similarly minded work of Fry et al. (2010), we thus have access to a much larger database, including ambient recordings from the entire Italian broadband network; the area of analysis is accordingly extended, to include, besides Switzerland, Northern Italy (the Po plain and part of the northern Apennines) and Southern Germany. The larger aperture of the station array assembled by Verbeke et al. (2012)
allows more robust cross-correlations at relatively long ($\gtrsim 30$ s) period, and hence more reliable images at accordingly larger, asthenosphere and mantle depths. Our work substantiates earlier results on the geographic pattern of seismic anisotropy in the region of interest, and sheds some light on the nature of the relationship between mapped anisotropy and central European geology and tectonics.

4.2 Methodology

Smith and Dahlen (1973, 1975) describe a theoretical relation, valid in a half-space medium, between slight perturbations $\delta C$ in surface-wave phase velocity (with respect to its reference value $C$) and the azimuth $\psi$ of their direction of propagation as,

$$
\frac{\delta C(\mathbf{r}, \psi)}{C} = \gamma_0(\mathbf{r}) + \gamma_1(\mathbf{r}) \cos(2\psi) + \gamma_2(\mathbf{r}) \sin(2\psi) + \gamma_3(\mathbf{r}) \cos(4\psi) + \gamma_4(\mathbf{r}) \sin(4\psi),
$$

(4.1)

where $\mathbf{r}$ is a 2-vector denoting position on the half-space surface, and the values of $\gamma_i$ ($i = 0, \ldots, 4$) naturally change also as functions of surface wave frequency $\omega$.

A set of measurements of $\delta C$ made at the same frequency, each for a different source-station (or, as in this case, station-station) couple, can then be translated via a linear inverse problem, as originally shown by (Tanimoto and Anderson, 1984), into a map of lateral variations in the “$2\psi$” and/or “$4\psi$” terms of Eq. (4.1). In the following, we shall systematically neglect the notoriously elusive $4\psi$ terms ($\gamma_3$ and $\gamma_4$) and focus instead on the $2\psi$ terms.

4.2.1 Automated measurements of phase dispersion

The starting point of this study is the database of ensemble-averaged ambient-noise cross correlations compiled by (Verbeke et al., 2012) and based on continuous recordings made at $\sim 200$ European stations depicted in Fig. 4.2 and covering the entire year of 2006. While Verbeke et al. (2012) had focused on group velocity, we here systematically measure the dispersion of surface-wave phase on the empirical Green functions is provided by that study. As anticipated by Verbeke et al. (2012), this is done via the station-station method
of Meier et al. (2004). In practice, given the observed surface-wave phase $\phi$ as a function of $\omega$, we obtain a phase-velocity dispersion curve via

$$C(\omega) = \frac{\omega \Delta}{\arctan(\Im(\phi(\omega))/\Re(\phi(\omega))) + 2n\pi}$$

(Meier et al., 2004), where $\Delta$ denotes the inter-station distance, and $\Re$ and $\Im$ the real and imaginary parts of the phase $\phi(\omega)$ found by cross-correlation. The integer number $n$ accounts for the ambiguity of the arctan function, whose associated error on phase is a multiple of $2\pi$. A set of possible integer values of $n$ are evaluated and the dispersion curve closest to that predicted by PREM (Dziewonski and Anderson, 1981), which we denote $C_{\text{PREM}}$, is selected. The choice of PREM as a reference model is consistent with the work of Fry et al. (2010). The validity of this approach is confirmed by in Chap.3 after thoroughly comparing its results with those of the independent method of Ekström et al. (2009).

Importantly, the large size of the database requires that we employ an automated algorithm to systematically identify phase-dispersion curves for all possible station couples. The automation process consists of identifying and quantifying the criteria which we had implicitly applied when visually “picking” dispersion curves for each single station couple.

First of all, we limit the analysis to frequencies between 0.02 and 0.1 Hz. Stehly et al. (2006), show that most coherent ambient signal emerging from ensemble averaging at the regional scale is found within this frequency range, and most likely generated by storms and the interaction between oceans and the solid Earth.

We apply Eq. (4.2) to identify a set of dispersion curves corresponding to all integer values of $n$ between -5 and 5, as illustrated in Fig. 4.5 for one particular station couple. Starting at the lowest frequency (0.02 Hz), we next compute the difference $\epsilon_1$ between $C_{\text{PREM}}$ and the eleven dispersion curves associated with all considered values of $n$,

$$\epsilon_1 = |C_{\text{PREM}}(\omega) - C(\omega)|.$$  \hspace{1cm} (4.3)

The curve which minimizes $\epsilon_1$ is selected. For most cases examined, this corresponded to when $n = 0$ case.

While phase can generally be measured from our empirical Green function is at low frequency $\sim0.02$ Hz, there are a number of features a dispersion curve may possess which force us to select an upper frequency limit $\omega_c < 0.1$ Hz. In the final stage of the analysis, we apply two rules to determine if a frequency cutoff is required. First, the dispersion
curve is given a restricted frequency range if $\epsilon_1 > 0.85$ km/s. In addition, we also consider truncating the frequency range if the dispersion curve exhibits a sharp change in gradient. We evaluate this variation in slope, via the measure $\epsilon_2$, which is computed as

$$\epsilon_2 = |C(\omega) + C(\omega + 3\Delta(\omega))|.$$  

(4.4)

where $\Delta(\omega)$ is the spacing used to sample the frequencies. If $\epsilon_2 > 0.85$ km/s, the frequency range is truncated. This measure allows us to identify jumps in the dispersion curve (see Fig. 4.5 for an example), but keep the small fluctuations in the low frequency range. If both the deviation from PREM ($\epsilon_1$) and a measure of the curvature ($\epsilon_2$) are small, we keep all the points of the dispersion curve until 0.1 Hz, past which we expect to have no useful information.

### 4.2.2 Azimuthally anisotropic tomography

We follow the inversion procedure described by Fry et al. (2010), conducting a regularized least squares inversion where phase velocities observed at a certain frequency at each station couple form the data vector and a longitude- and latitude-dependent phase velocity map, including isotropic, $2\psi$ and (in principle) $4\psi$ terms is the unknown. The software we used was originally developed by Lebedev and van der Hilst (2008) following the inversion procedure described by Deschamps et al. (2008).

The parameterization grid is triangular, with 50 km-spacing between nodes. In Fig. 4.3 we show how raypaths are projected onto our grid. The raypaths and azimuthal coverages associated with our parameterization are also shown in Fig. 4.3.

Norm and roughness damping factors are applied to each parameter: isotropic velocity and $2\psi$ and $4\psi$ components of azimuthal anisotropy. Values for all six damping parameters are selected after a series of preliminary tests. As is usually the case in azimuthal anisotropic tomography, the $4\psi$ term cannot be uniquely determined, and we shall neglect it hereafter: we verified that changing the associated damping parameters does not perturb significantly the isotropic and $2\psi$ solutions. A small value of approximately 0.05 is selected for the norm damping parameter of both isotropic and $2\psi$ terms. Our preferred regularization scheme is roughness damping; we select the roughness damping parameter for the isotropic term based on consistency of our results with those of Verbeke et al. (2012), and finally determine the roughness damping parameter for the $2\psi$ term via the L-curve analysis Hansen (1992):
values between 1 and 30 are used and the resulting misfit is plotted in Fig. 4.1 as a function of such damping parameter values.

Figure 4.1: Cumulative data-misfit as a function of roughness damping parameter value for 20s example period

4.2.3 Azimuthal coverage of the dataset

For anisotropy to be mapped, it is crucial that the region of interest be sampled as uniformly as possible by seismic waves traveling in all directions. In Fig. 4.3 we visualize the azimuthal coverage for each node of the grid used for tomographic imaging at 20s. Each circle represents one node and the center of the circle coincides with the location of the node. Each circle is subdivided into eight segments. It is naturally preferable to have fewer raypaths concentrated in several locations than a lot in only one part of the circle. Based on this, we then evaluate which region of our models will be well-resolved. It is important to note that none of the data are removed prior to the inversion. The evaluation of the
azimuthal coverage is nevertheless necessary to define the well-resolved part of our inverted models. We define four classes of circle depending of the quality of the coverage (Table 4.1), ranging from very good, to unacceptable. In Fig. 4.4 we display several examples for each class. Based on this criteria, we then identify the well-resolved part of our models for each period.

### 4.3 Results

#### 4.3.1 Isotropic velocities

We summarize in Table 4.2 the depth of sensitivity for each period, based on the partial derivatives computed by Fry et al. (2010). From numerous previous geophysical studies examining the Moho structure underneath Europe, it is well accepted that the Moho depth in this region is $\sim 40$ km on average, with minima of $\sim 16$ km under Southern Germany and Piedmont (Wagner et al., 2011) and peaks of $\sim 60$ km under the Alps (e.g., Waldhauser et al., 1998). Table 4.2 then indicates that observations at all considered periods are affected by sensitivity to Moho depth; as seismic velocities are higher in the mantle than in the crust, a thicker-than-average Moho will result in lower-than-average surface-wave phase velocity.
Figure 4.2: Geographical distribution of the broadband stations used this study. In blue are the stations used in both the present study and that of Fry et al. (2010). In yellow are the stations used by Fry et al. (2010), but excluded from the present study as these stations only recorded data in 2005. We include the dispersion curve in our inversion. Finally in red are the stations from our database constructed with the recorded data of 2006.

and vice-versa. Visual inspection of Fig. 4.6 indicates that the pattern of mapped phase velocity is correlated with that of Moho depth found e.g. by Waldhauser et al. (1998), Di
Stefano et al. (2009) and Wagner et al. (2011). For example, the strong observed negative phase-velocity anomaly in the Alps can be associated to the thicker-than-average crust found in that area, as opposed to the average-thickness crust of approximately of 30 km in north-western Italy, or in northern Switzerland, where our mapped phase velocities are higher at all periods.

Isotropic maps in Fig. 4.6 are approximately consistent with the large-scale pattern of the lower-resolution phase-velocity images by Verbeke et al. (2012). Perfect agreement cannot be expected since we are using here a different parameterization and inversion algorithm and, most importantly, a different starting model for the inversion.
4.3.2 Azimuthal anisotropy

The direction along which Rayleigh-wave propagation is fastest (or “fast azimuth”), based on the $2\psi$ term only as explained in Section 4.2.2, is shown as a function of location in Fig. 4.6. At relatively short periods $\leq 30$ s, the pattern of lateral variation in the fast azimuth of anisotropy is characterized by three distinct, sharply separated regions: (i) north of the Alpine front, where the fast azimuth is approximately east-west (parallel to the Alpine front); (ii) between Alpine and Adriatic front, where it is approximately north-south (perpendicular to the Alpine front); (iii) south of the Adriatic front, where the fast azimuth is again roughly east-west and perpendicular to the Alpine front. In all these areas, the strength of anisotropy varies between $\sim 2$ and $3\%$.

At periods $>30$ s anisotropy is generally weaker, between $\sim 1 - 2\%$; the fast azimuth south of the Adriatic front is perpendicular to the Alpine front, so that there is no discontinuity in the $2\psi$ anisotropy pattern across the Alpine front. Southern Germany anisotropy partly retains its east-west character, which is however much less pronounced than at shorter periods.
Figure 4.5: Example of the dispersion curve selected by the automatic picker in green, compared to that obtained via manual selection plotted with black markers. The theoretical dispersion curve derived from PREM is shown in red.

4.4 Discussion

4.4.1 Comparison with earlier studies

Our work is essentially an extension of the one from Fry et al. (2010) to a larger database and a geographically much wider region, which presumably results in a more robust model. Within the area covered also by the smaller-aperture station array of Fry et al. (2010) (Fig. 4.2) our results shown in Fig. 4.6 are largely consistent with theirs. The most important features shared by both studies are (i) (isotropically) slow Alpine orogen and fast
Figure 4.6: Inversion at 12s, 16s, 20s (upper row, from left to right), 24s, 28s, 30s (middle row, from left to right), 34s, 40s and 43s (lower row, from left to right). The isotropic part of the velocity is plotted in color. The yellow bars indicate the orientation of the fast azimuth, and the length of each bar represents the intensity of the anisotropy. The grey line represent our high quality coverage part of the model that we defined following the quality criteria we defined in the 4.1.

northern Switzerland at periods between 15 and 30 s; (ii) in the same period range, sharp transition, across the Alpine front, from Alps-parallel fast azimuth to the north, to Alps-perpendicular fast azimuth to the south; (iii) presumed change of anisotropy with depth, as relatively-long-period Rayleigh waves are characterized by a north-south fast azimuth throughout Switzerland, while the pattern of shorter-period anisotropy is more complex, as explained.

One significant difference between our results and those of Fry et al. (2010) is apparent at 12 s period, where Fry et al. (2010) find no systematic change in fast azimuth across the Alpine front. This discrepancy could in principal be caused in principle by differences in data coverage, in the pre-processing of the data, which, as explained in section 4.2.1, we have repeated independently of the earlier work by Stehly et al. (2009) and Fry et al. (2010), or in the choice of inversion parameters (starting model and parameterization and
regularization schemes). This is beyond the scope of this couch to examine the causes for these differences.

### 4.4.2 Geophysical interpretation

Studies of the Earth’s mantle indicate lattice-preferred orientation (LPO) of olivine as the main source of anisotropy (Christensen, 1984; Long and Becker, 2010). In the upper part of our depth range of interest lithosphere and asthenosphere, other causes of anisotropy can be invoked: namely, the shape-preferred orientation (SPO) or LPO of amphiboles and biotite minerals (Nicolas and Christensen, 1987; Barruol and Mainprice, 1993). While Barberini et al. (2007) specifically proposed the LPO of amphiboles to be a major source of anisotropy in the Alpine lower crust, Fry et al. (2010) based on a tectonic interpretation of their tomographic results suggests the SPO of biotite to be the source of peculiar anisotropy in the Alpine lower crustal root zone. We note that in the cases of olivine for the mantle and amphibole for the lower crust, the fast azimuth of anisotropy is expected to be aligned with the direction of deformation, while in the case of biotite, the fast axis is perpendicular to the deformation plane. Our surface-wave anisotropy tomography results are intrinsically coupled with the 3D-isotropic model results obtained by simultaneous inversion. With only general a priori knowledge about the relative size of local anomalies and, therefore, of realistic contributions of isotropic and anisotropic model parts to data variance improvement, the two solution parts may not be regarded as independent even on a regional scale. Reliability of the less well-known anisotropic solution part, therefore, may be tested by comparing the reliably imaged isotropic velocity anomalies (colored back ground maps in 4.6) with previously known seismic structure.

Isotropic results from 12s period surface waves (sensitive to depth range from surface to 30km) document the low velocity associated with the Molasse basin north of the Alps and a pronounced high-velocity anomaly along the inner part of the western Alpine arc that very well matches the Ivrea body (Kissling, 1993). The deep central parts of the Po basin are documented by pronounced low velocity anomalies between 44N to 45N and 8.5E to 10.5E in the maps derived from 16s and 20s waves. In addition, these two maps show a moderately (16s) and pronounced (20s) low velocity anomaly associated with the thickening of the upper and middle crust within the Alpine orogen. Surface waves of 24s, 28s, and 30s are most sensitive to the depth range from 20km to 60km, i.e., lowermost crust and upper mantle lithosphere outside the Alps. Considering the Alpine crustal root
reaches 58km beneath the southeastern front (Kissling et al., 2006) and that the known location of the deep crustal root beneath the northern Apennines (Verbeke et al., 2012; Wagner et al., 2011), we note that the low velocity anomalies in our tomographic results for periods between 20s and 30s precisely correlate with a priori known upper lithosphere structure.

The azimuthally anisotropy maps at periods > 30s in Fig. 6 are consistent with the findings by Barruol et al. (2011) from SKS anisotropy in the western Alpine arc region (recall that the latter cannot be compared directly to surface-wave azimuthal anisotropy) and with the geometries of asthenospheric flow pattern these authors proposed as a source for their results based on previous tectonic studies (Faccenna et al., 2004). The fast azimuths of our anisotropy maps in eastern France are approximately parallel to the Alpine arc, just like the toroidal flow around the sinking slab attached to European lithosphere in Western Alps (Lippitsch et al., 2003), as indicated in Fig. 7 of Barruol et al. (2011).

For periods less than or equal to 30s, the fast azimuths in our anisotropy maps (Fig. 4.6) in the northern Alpine foreland are approximately parallel to the strike of the Hercynian orogenic belt. This moderately strong anisotropy shows a very consistent direction from mid crustal levels down to the uppermost mantle lithosphere. An equally consistent pattern of fast velocity axis but of somewhat stronger anisotropy is observed within the Adriatic lower crust and uppermost mantle lithosphere (Fig. 4.6, periods 16s to 30s). In either case, north and south of the Alps, we interpret these anisotropy patterns as the results of pre-Alpine tectonics that affected the whole continental lithosphere at the time.

Corresponding with the isotropic anomalies associated with the Alpine crustal root (see above), in all maps from periods less than 34s we find the fast azimuth of surface-wave anisotropy to be perpendicular to the strike of the orogen. These findings match the results of Fry et al. (2010) in the central Swiss Alps and document this anomaly to be of more regional scale and representative for much of the Alpine orogen. In correlation with Fry et al. (2010) and based on the above considerations we interpret these findings with orogenic convergence, resulting in compressive stacking of lower crustal slivers detached from subducting mantle lithosphere and accumulated in the Alpine crustal root.
<table>
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<tr>
<th>Period (sec)</th>
<th>Sensitivity (km)</th>
<th>Peak</th>
<th>Range</th>
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Table 4.2: Sensitivity of each period.

4.5 Conclusions

We have collected a large database of station-station Rayleigh-wave phase-velocity observations based on ambient-noise cross-correlation. Phase dispersion is measured on ensemble-averaged cross-correlations for all possible station couples by means of an automated algorithm that we devised. This allows us to derive azimuthally anisotropic Rayleigh-wave phase-velocity maps at a set of periods between \( \sim 10 \) and 40s. With respect to the similarly minded, earlier study of Fry et al. (2010), we have increased enormously the aperture of the array of available stations, so that, besides Switzerland, we could robustly image isotropic and anisotropic phase velocity in southern Germany, northern Italy, and parts of Austria and France. Earlier results are reproduced so far as the smaller, previously studied area is concerned. In particular, we confirm that anisotropy beneath the Swiss Alps is stratified in two layers—a shallower one with an orogen-parallel fast direction, and, at depths larger than 30km, a second one characterized by a strong orogen-perpendicular fast direction. Importantly, northern Italy, which was not covered by the study of Fry et al. (2010), is characterized at most surface-wave periods by a fast azimuth direction perpendicular to that found along the Alpine arc (Fig. 4.6).

Mineral physics studies indicate that for most mechanisms acting in the crustal/asthenospheric depth range, fast seismic azimuth should be aligned with the direction of deformation. From our maps of azimuthal anisotropy in Fig. 4.6 we can then attempt to draw inferences on the geometry of asthenospheric flow/deformation associated with the Alpine arc. The main mechanisms that we suggest to be sampled are: toroidal flow, to the north and west of the Alps, associated with Alpine subduction (consistent with the flow geometry proposed
e.g. by Barruol et al. (2011)); compressive stacking of the Alpine crust which bring lower crustal slivers from subducting mantle lithosphere and Hercynian orogen.

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Bibliography


5.1 Synthesis

We applied the ambient noise method combined with a linear inversion to illuminate European crust with a horizontal resolution of 100 to 200 km from the shallow surface, down to the Moho. Maps of the lateral variations in group- and phase-velocity throughout the region displayed several dominant features: namely, the Po-plain sedimentary basin, anomalously thin crust under the Tyrrhenian sea, the deep crustal roots beneath the Alps and the northern Apennines and, finally, the Molasse sedimentary basin. The accuracy of our S-wave velocity model was validated by a comparison with (i) previous studies using also ambient noise (Stehly et al., 2009; Li et al., 2010) and (ii) with a new Moho map developed using P wave CSS and LET (Diehl et al., 2009; Waldhauser et al., 1998; Wagner et al., 2011).

In both cases, a high degree of correlation was found even if there were (i) differences in ray coverage (for example Stehly et al. (2009) and our study); (ii) differences in the inversion procedure (the tomographic inversion of ambient noise study was conducted here with a different software and different parametrization and regularization strategies); (iii) differences in the seismic method and S-respectively P-wave data used to illuminate the crust (meaning using ambient noise method or CSS for example).

Despite these differences in imaging methodologies, the similarity of the results obtained demonstrates the robustness of the ambient noise technique to produce images at a crustal scale. Moreover, in contrast with Stehly et al. (2009) and Li et al. (2010), with the high density coverage of stations that characterizes our new dataset, smaller scale features have
been resolved. The success of the ambient-noise results constitute the first step towards establishing a complete consensus P and S velocity model of the European lithosphere structure.

From the coherent surface-wave packets (empirical Green’s function) observed from cross-correlations, both group and phase velocities were extracted. Phase velocities are of different interest than group velocities since they are sensitive to structure at larger depth than the group velocities of the same periods. Unfortunately, measuring phase velocities from ambient noise cross-correlations remains less straightforward than measuring group velocities. Yao et al. (2006) observed a difference between phase velocities computed from ambient noise and phase velocities computed from teleseismic events. Tsai (2009) proposes that a $\pi/4$ phase shift must be applied to empirical Green’s function to obtain accurate phase velocity measurements from an ensemble-averaged cross-correlations. In this study (Chap.3) we compare two different ways of measuring phase velocities from ambient noise: (1) the classical two-station method, including the required $\pi/4$ correction and (2) the frequency-domain approach of Aki (1957) and Ekström et al. (2009), which does not require the $\pi/4$ correction. The two techniques are profoundly different but yield consistent results, validating both approaches. This comparison confirms the quality of our phase velocities measurements, which we next use to infer the 2D azimuthal anisotropic structure of central Europe.

To guarantee the robustness of our azimuthal anisotropy maps, the phase velocity dataset is filtered according to a quality evaluation of azimuthal coverage over the tomographic parametrization grid. This evaluation led us to limit our region of interest, spanning from southern Germany, Switzerland and northern Italy. Layered vertical azimuthal anisotropy is computed from mid-crust to the depth of the asthenosphere. To the north and south of the Alpine arc, the fast direction of anisotropy in middle to lower crust is parallel to the strike of the Hercynian orogenic belt and is attributed to the SPO of amphibole.

Along the Alpine arc, following the results of Fry et al. (2010), the anisotropy perpendicular to the arc is interpreted as caused by the stacking of crustal slivers which were compressed and accumulated during Alpine orogen. The main cause of anisotropy in the Alpine lower crustal root is interpreted as the shape-preferred orientation (SPO) of biotite. At periods equal or longer than 34s, the patterns of anisotropy is caused by asthenospheric flow, which also coincides with the results of Barruol et al. (2011).
5.2 Perspectives

Following the validation of the ambient noise derived database and inferring the 2D isotropic and anisotropic structure, a full 3D structure can be obtained. Since the crust is a very heterogeneous medium, applying a linearized type inversion (as was used to obtain the 2D maps in Chap. 2) may not allow us to resolve the 3D structure within the crust. Hence the neighborhood algorithm was used to invert our data for the 3D crustal structure. We specifically used the Dinver module of Geopsy software package (Wathelet et al., 2004; Wathelet, 2008).

The neighborhood inversion is conceptually similar to the Monte Carlo inversion. With the Monte Carlo approach, the solution space is investigated randomly and the misfit with the data is computed for each investigated solution. In contrast, the neighborhood algorithm uses the measured misfit to determine which regions of the parameter space should be sampled. Initially the neighborhood algorithm samples the parameter space randomly. Subsequently, regions within the parameter space with a low misfit are randomly resampled to identify the global minimum of the objective function. The neighborhood algorithm uses misfit information to choose in which part of the parameter space it will search further.

The Dinver module is especially dedicated to the inversion of surface waves and allows the inversion of individual dispersion curves. In practice, from the 2D velocity maps obtained in Chap. 2, a local dispersion curve at each pixel of the grid used in the 2D linear inversion is extracted. The local dispersion curve for each 2D cell is then invested using the Dinver software to derive a local 1D depth profile. Eventually, all local profiles are combined to form a 3D model of the European crust and the lithosphere.

We first select four regions from the grid defined with the 2D inversion: the Po Plain, south Germany, the Alps and the extreme western part of the Alps as shown of the Fig. 5.1.
Figure 5.1: Location of the four regions where we compute the theoretical curve predicted by EP crust using Dinver and the phase velocity dispersion curve obtained from ambient noise cross-correlations for the same location.

Each region represents four pixels according to the grid used for the 2D linear inversion of the isotropic velocity. For each pixel we compare the theoretical dispersion curve predicted by EPcrust (Molinari and Morelli, 2011) model, that we compute using Dinver, and the observed dispersion curve associated with each pixel of our 2D phase velocity models based on ambient noise cross-correlations.

This comparison (illustrated in Figs. 5.2 and 5.3) shows that at low frequency there is a good agreement between our observations and the predictions based on EPcrust model while higher frequency starting at 0.06 or 0.08 Hz, are not matched well, with our observations systematically faster. This might be explained by the relatively poor performance of our measuring technique at relatively high frequency (or, which is the same, low energy of ambient noise) compared to the frequency range 0.02-0.06 Hz. We hence invert our dispersion curve with Dinver focusing on the results at Moho depth. A initial model is needed for inversion of each profile. We choose the same initial model for each of our pixels (c.f. Fig. 5.4).

At each pixel, models of $V_p$, $V_s$ and density are found, together with a estimation of the datafit (c.f Fig. 5.5). For each pixel, we select the best fitting model given by Dinver. Each
Figure 5.2: Comparison of theoretical curve (red) predicted by EP crust using Dinver and the phase velocity dispersion curve obtained from ambient noise cross-correlations (blue) a) in the South Germany and b) in the western Alps (c.f Area 1 and 2 in Fig. 5.1)
Figure 5.3: Comparison of theoretical curve (red) predicted by EP crust using Dinver and the phase velocity dispersion curve obtained from ambient noise cross-correlations (blue) a) Swiss Alps region and b) in Po Plain (c.f Area 3 and 4 in Fig. 5.1)
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Figure 5.4: Initial model used for the inversion with Dinver with 3 layers on half-space, velocities range for $V_p$, $V_s$ and density are set as well as a range of possible surface depth of each layer.

Figure 5.5: Example of results obtained after inversion with $V_p,V_s$ models explored with associated datafit.

vertical profile is then assembled to define the full 3D structure. We here only present the Moho depth we obtained and we compare it with the Moho depth obtained by EPcrust in Fig. 5.8. The Moho depth from EPcrust Molinari and Morelli (2011) is also presented.
together with the Mohodepth obtained by Di Stefano et al. (2009) from CSS and receiver functions methodologies in Fig. 5.6 and Fig. 5.7.
Figure 5.6: Moho topography given by EPcrust (Molinari and Morelli, 2011)
Figure 5.7: Moho topography map obtained by Di Stefano et al. (2009)
Figure 5.8: (left) Moho topography given by EPcrust (Molinari and Morelli, 2011) and (right) the Moho topography obtained from the non-linear inversion of our phase velocity dispersion curves computed from ambient-noise cross-correlations.
While the general pattern agree very well, note that EPcrust exhibits a more smooth Moho topography and Di Stefano et al. (2009) document more locally detailed Moho topography. We have to consider that surface waves measurements are less sensitive to strong velocity contrast and high frequency Moho topography compare to CSS. The resolution capabilities of ambient-noise technique in Alpine region, therefore, is only moderate. The Moho topography map we obtained is similar in the greater area of the Alps and Po plain where the coverage of our data is the best, while in the Tyrrhenian sea, or Austria our data coverage, and hence resolution, are poor. Along the Alpine arc and in Germany we see a similar pattern in both maps: depth around 30 km in Germany and similar maximum topography map in the Alps with values 50 km. Beneath the Po plain we have greater depth compared to EPcrust which can be caused by the vertical smearing of low velocity induced by the sedimentary Po plain to lower crust. The sensitivity of phase velocity are wide and the Po plain has a maximum thickness of 13 km.

We nevertheless obtained a deeper Moho than EPcrust which in general does fit better with the Moho topography of Di Stefano et al. (2009). Some discrepancies might stem from the fact that we made use of the same initial model as the starting point of our neighborhood algorithm inversion over the entire region of interest, regardless of the 3D know, strong lateral heterogeneity. Here more work on finding a good initial 3D model is surely required, as well as more investigating the reason for the discrepancies between the two models at higher frequencies.
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