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EPcrust: a reference crustal model for the European Plate

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SUMMARY

We present a new crustal model for the European Plate, derived from collection and critical integration of information selected from the literature. The model covers the whole European Plate from North Africa to the North Pole (20°N–90°N) and from the Mid-Atlantic Ridge to the Urals (40°W–70°E). The chosen parametrization represents the crust in three layers (sediments, upper crust and lower crust), and describes the 3-D geometry of the interfaces and seismologically relevant parameters—isotropic *P*- and *S*-wave velocity, plus density—with a resolution of 0.5° × 0.5° on a geographical latitude–longitude grid. We selected global and local models, derived from geological assumptions, active seismic experiments, surface wave studies, noise correlation, receiver functions. Model EPcrust presents significant advantages with respect to previous models: it covers the whole European Plate; it is a complete and internally-consistent model (with all the parameters provided, also for the sedimentary layer); it is reproducible; it is easy to update in the future by adding new contributions; and it is available in a convenient digital format. EPcrust could be used to account for crustal structure in seismic wave propagation modelling at continental scale or to compute linearized crustal corrections in continental scale seismic tomography, gravity studies, dynamic topography and other applications that require a reliable crustal structure. Because of its resolution, our model is not suited for local-scale studies, such as the computation of earthquake scenarios, where more detailed knowledge of the structure is required. We plan to update the model as new data will become available, and possibly improve its resolution for selected areas in the future.

Key words: Crustal structure; Europe.

1 INTRODUCTION

Knowledge of crustal structure is crucial for many applications, such as upper-mantle tomographic studies (e.g. Bozdag & Trampert 2008; Ritsema *et al.* 2009), gravity modelling (Yegorova & Starostenko 2002), forward seismic wave propagation (e.g. Capdeville & Marigo 2008), dynamic topography studies (e.g. Faccenna & Becker 2010) and location of seismic events (e.g. Bisio *et al.* 2004).

Seismic waves are very sensitive to crustal structure, but, in the frequency range and ray geometries mostly used in seismic tomography, they are not able to discriminate uniquely between the structure of the crust and that of the mantle below. For instance, the propagation of surface waves, both Rayleigh and Love, with period $T > 35$ s—used, for instance, to calculate seismic moment tensors (e.g. Pondrelli *et al.* 2007) and to image lithospheric structure (e.g. Schivardi & Morelli 2009)—is heavily affected by the strong heterogeneities in the Earth's crust. However, their sensitivity to shear wave speed is spread over both crust and mantle, generating large trade-off between wave speed above and below the Moho. The reduced sensitivity to the velocity jump at the Moho, and to the depth of the discontinuity does not permit the inversion for these parameters in conjunction with seismic wave speed in the mantle.

Focusing the attention on the European continent—where abundant data are available, and the geological structure is quite complex—it becomes clear that current crustal models are not fully adequate for modelling regional data sets with enough detail. The global model CRUST2.0 (Bassin *et al.* 2000) is frequently used for crustal correction, but its resolution (2° × 2°) is too low for continental scale studies (Molinari & Morelli 2009). Some European-scale crustal models are available (Bassin *et al.* 2000; Ritzmann *et al.* 2007; Tesauro *et al.* 2008), but none of them possesses all the desired properties for a complete reference crustal model: either because of a low resolution, a limited geographical extent, or lack of description of some seismic parameter, such as v_S.

Local studies can overcome these limitations. Receiver functions provide point determinations of Moho depth and constraints on the velocity structure in the European crust (e.g. Kumar *et al.* 2007; Geissler *et al.* 2008; Piana Agostinetti & Amato 2009), but such determinations are not dense and extended enough to draw a wide geographical surface. They can however be used to calibrate and improve existing models. Seismic reflection/refraction experiments (e.g. Guterch *et al.* 2003; Grad *et al.* 2003; Wilde-Piorko *et al.* 2008) present the best capability to image crustal structure, and resulting profiles have extensively been used to create local

3-D models (e.g. Guterch *et al.* 2005; Diaz & Gallart 2009; Stratford *et al.* 2009). Recent studies have shown that cross correlation functions of seismic noise computed at two receivers contains the Green's function between these two receivers, that is, the waveform that would be recorded at one of the stations if a point force source was applied at the other station. This provides the possibility of measuring relatively high-frequency surface waves for many short paths. High-frequency surface waves have their sensitivity concentrated in the crust, and their inversion permits the reconstruction the structure (e.g. Stehly *et al.* 2009). However, such studies are for now limited to subregions of the whole European Plate, so by themselves they have not (yet) provided a whole European model.

A lot of detailed information on European crustal structure therefore exists, but at different scales and following different formats. This information needs to be merged into a larger-scale, coherent representation. Our purpose is to sketch the seismological description of the complex crustal structure of the European Plate with a higher resolution and more plausibility than it is offered by existing models. We collected the information currently available from different sources, ranging from active-source seismic profiles, to receiver function studies, to digital maps and models. These large-scale and regional models concur to construct a new comprehensive reference model for the European region: EPcrust. In the next sections we describe the initial models and data sets used, the methodology adopted to create the new model, and we discuss our results in function of the geological setting of the European Plate.

2 DATA SETS

As a first step in our effort, we collected available models and information about the European crust. We selected preferentially more recent studies, and evaluated their significance and confidence level. Our goal is to create a model for the whole European Plate, from North Africa to the North Pole (20°N–90°N) and from the Mid-Atlantic Ridge to the Urals (40°W–70°E). In the following, we provide an overview of the data used. In Table 1 we summarize all the contributions used to assembly EPcrust, and in Table 2 we specify the information included in these contributions. The different regions covered by the original models are shown in Fig. 1.

Perhaps the most commonly used description for the crust at a global scale is CRUST2.0 (Bassin *et al.* 2000). CRUST2.0 is a good global model, but its resolution is too low for regional or continental scale work: in fact, Moho depth defined on $2^\circ \times 2^\circ$ pixels cannot account for important features (e.g. Moho undulations under Alps and Apennines) that are relevant to seismic wave propagation at this scale. Also, CRUST2.0 has sharp tile edges, and a rather complex vertical layering and discontinuous lateral steps, that make it difficult either to represent the model with fidelity in numerical codes, or to smooth it laterally (Molinari & Morelli 2009). Another global model is due to Meier *et al.* (2007), who derived it by inversion of surface wave data. This model has a simpler parametrization, providing thickness and average *S* velocity at the same resolution of $2^\circ \times 2^\circ$. We use these global models as a background for our final model, as they provide information where more detailed studies are unavailable.

Moho depth is certainly the single best known parameter of the crust. Significant efforts have recently been devoted to deriving Moho maps of different subregions of Europe. We collected several models, limited to the description of Moho depth only, for inclusion in our compilation. As a result of a very substantial effort carried on by a Working Group specifically set by the European Seismological Commission, a digital Moho depth map has recently been compiled by assembling more than 250 data sets of individual seismic profiles, 3-D models obtained by body and surface waves, receiver function results, and maps of seismic and/or gravity data compilations (Grad *et al.* 2009). This study represents the first digital, high resolution map of Moho depth for the whole European Plate extending from the Ural Mountains in the east, to mid-Atlantic ridge in the west and from the Mediterranean Sea in the south, to the Barents Sea and Spitsbergen in Arctic in the north (available online at http://www.seismo.helsinki.fi/mohomap/). Because of its recent development, and the broad data compilation that went into it, we consider this study as a reliable reference at a large scale for Moho depth. In a more local framework, Diaz & Gallart (2009) and Stratford *et al.* (2009) compiled Moho maps, respectively of the Iberian Peninsula and southern Norway. Both maps are derived from interpolation of a collection of seismic profiles published in the past. The Iberian Peninsula map has been derived collecting and revising the most relevant seismic experiments carried on in the area in the last three decades, and then interpolating the geo-referred database, using a kriging algorithm, to come up with a continuous Moho depth model. By using receiver function analysis it is also possible to obtain reliable estimates about the Moho depth and the crustal v_P/v_S ratio under the recording seismic station. Piana Agostinetti & Amato (2009) derive a new data set of the Moho depth in the Italian peninsula using more than 270 teleseismic event recorded at the 127 stations of the Italian National Seismic Network. In addition, we also collect a Moho depth data set in the central Europe from receiver functions provided to us by Kind (2009 'Moho depth in central Europe from receiver functions studies', personal communication) from which we derive a map in this region interpolating Moho values.

For the areas where they exist, regional models can provide reliable descriptions of crustal geometry, as well as values of seismic parameters inside. Recently, Baranov (2010) improved the knowledge of the Central and Southern Asia and surrounding regions compiling AsCRUST-09, a $1^\circ \times 1^\circ$ model of the crystalline crust, from an interpolation of a collection of seismic refraction and reflection data. The westernmost part of this model, that overlaps our region of interest (Fig. 1), appears to be well constrained by the presence of many seismic profiles. EuCRUST-07 (Tesauro *et al.* 2008) is instead a digital model for the crust of Western and Central Europe and surroundings (35°N–71°N, 25°W–35°E) based on the assemblage of available results of seismic reflection, refraction and receiver functions studies into an integrated model at a uniform grid ($15' \times 15'$). EuCRUST-07 consists of three layers: sediments and two layers of the crystalline crust. Besides depth to the boundaries, EuCRUST-07 provides average *P*-wave velocities in the upper and lower parts of the crystalline crust but lacks one of the sedimentary layer. This study shows large differences in the Moho depth compared to previous compilations, more than ±10 km in some specific areas (e.g. the Baltic Shield). Furthermore, the velocity structure of the crust is much more heterogeneous than in previous maps.

On a more local scale, BARENTS50 (Ritzmann *et al.* 2007) has been generated by analysis and interpolation of 680 individual seismic profiles in the western Barents Sea to assemble a 3-D crustal model of the region with a resolution of 50 × 50 km. The Authors used a compilation strategy based on the definition of geological provinces to produce a model with a parametrization in sediment layers (soft and hard), upper, middle and lower crust, providing all the seismic parameters.

We also use the $1^\circ \times 1^\circ$ world sediment map by Laske & Masters (1997)—from now on called LM97—and the $5' \times 5'$ thickness

Table 1. All contributions used to assembly EPcrust. For each model is specified: name and reference, covered region, resolution, year of publication, presence of error bar, assigned weight in EPcrust.

Model name	Region	Resolution	Year	Error bars	Weight
CRUST2.0 (Bassin *et al.* 2000)	Globe	2° × 2°	2000	Yes	1
EuCRUST-07 (Tesauro *et al.* 2008)	25°W–35°E/35°N–71°N	0.25° × 0.25°	2007	No	100
ESC Moho (Grad *et al.* 2009)	40°W–70°E/28°N–86°N	0.1° × 0.1°	2009	Yes	100
BARENTS50 (Ritzmann *et al.* 2007)	20°E–155°E/10°N–55°N	50 km× 50 km	2007	No	100
AsCRUST-08 (Baranov 2010)	10°E–70°E/62°N–82°N	1° × 1°	2010	No	10
LM97 sediment map (Laske & Masters 1997)	Globe	1° × 1°	1999	No	1
Sediment NOAA (Divins 2003)	Ocean	5′ × 5′	2003	No	10
Alps model (Stehly *et al.* 2009)	5°E–13°E/44°N–49°N	25 km× 25 km	2009	Yes	100
East Alps model (Molinari *et al.* 2010b)	14°E–22°E/45°N–50°N	0.1° × 0.1°	2010	No	150
Iberian Moho (Diaz & Gallart 2009)	10°E–5°E/35°N–46°N	0.5° × 0.5°	2009	No	150
Italian Moho (Piana Agostinetti & Amato 2009)	Italian Peninsula	Points	2010	Yes	150
Norwegian Moho (Stratford *et al.* 2009)	5°E–13°E/58°N–63°N	Points	2010	No	150
RF (Kind, 2009)	Central Europe	Points	2009	No	50
Meier model (Meier *et al.* 2007)	Globe	2° × 2°	2007	Yes	1
ETOPO1(Amante & Eakins 2009)	Globe	1′ × 1′	2009	No	–

map of the oceanic sediments due to Divins (2003). The Laske & Masters (1997) three-layer sediment model has been obtained, for continental areas, by digitizing the Tectonic Map of the World assembled by the EXXON Production Research Group (1985) and, for oceanic areas, by averaging other published high-resolution maps (e.g. Pacific, Indian and South Atlantic oceans). For regions where digital information was unavailable (e.g. Arctic and North Atlantic ocean), the sediment thickness has been hand-digitized from atlases and maps (http://igppweb.ucsd.edu/∼gabi/sediment.html). Each sediment layer has been assigned a *P*-wave speed, obtained using regional velocity functions for the oceans, and seismic reflection/refraction profiles, complemented by values given by CRUST5.1 (Mooney *et al.* 1998) for the continents.

These data sets provide information within a wide range of resolution and expected accuracy. Moho depth and sediment thickness are however the best constrained parameters (Table 2, a total of 12 and 7 models, respectively), while the Conrad discontinuity between upper and lower crystalline crust is poorly constrained, and is mostly derived from EuCRUST-07 and global models only (with the exceptions of the Barents Sea model and East Alps model). From a geographical point of view, the regions best covered are the Alps, Central Europe, the Italian peninsula, Spain, and Barents Sea, while in North Africa and in East Europe, we only have global models available [CRUST2.0 and Meier *et al.* (2007)]. Where we have more than one crustal measurement we found differences among the data sets: in particular for Moho depth we found differences at times exceeding 15 km. We see considerable differences also in the shape and thickness of sedimentary basins, in some cases of more than 6–8 km (i.e. the Po plain is more than 12 km thick in EuCRUST-07, while in Laske Sediment Map it is about 4 km). Of course, higher resolution models (such as BARENTS50, EuCRUST-07, Italian and Iberian Moho maps) provide presumably better constrained details on the geological structure.

3 MODEL CONSTRUCTION

EPcrust is constructed by joining information from all the global, regional and local models described in the previous section. The approach of combining *a priori* information, rather than fitting data solving an inverse problem, has been adopted for a number of crustal models, with different spatial scales, such as CRUST2.0 (Bassin *et al.* 2000), 3SMAC (Nataf & Ricard 1996), WENA1.0 (Pasyanos *et al.* 2004), AsCRUST-08 (Baranov 2010) and EurID (Du *et al.* 1998). The procedure consists of collecting and amalgamating reliable, although scattered, information about the crust in the region

Table 2. Information on crustal structure included in the contributions used to assembly EPcrust.

	Sediment				Upper crust				Middle crust				Lower crust				Moho	Topo-
Model name	*h*	*Vp*	*Vs*	*ρ*	*h*	*Vp*	*Vs*	*ρ*	*h*	*Vp*	*Vs*	*ρ*	*h*	*Vp*	*Vs*	*ρ*	depth	graphy
CRUST2.0	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
EuCRUST-07	X	–	–	–	X	X	–	–	–	–	–	–	X	X	–	–	X	X
ESC Moho	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	X	–
BARENTS50	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
AsCRUST-08	X	–	–	X	X	X	–	X	X	X	–	X	X	X	–	X	X	X
LM97	X	X	X	X	–	–	–	–	–	–	–	–	–	–	–	–	–	–
Sediment NOAA	X	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–
Alps model	–	–	X	–	–	–	X	–	–	–	–	–	–	–	X	–	X	–
East Alps model	X	–	–	–	X	X	–	–	–	–	–	–	X	X	–	–	X	–
Iberian Moho	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	X	–
Italian Moho	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	X	–
Norwegian Moho	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	X	–
RF	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	X	–
Meier model	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	X	–
ETOPO1	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	X

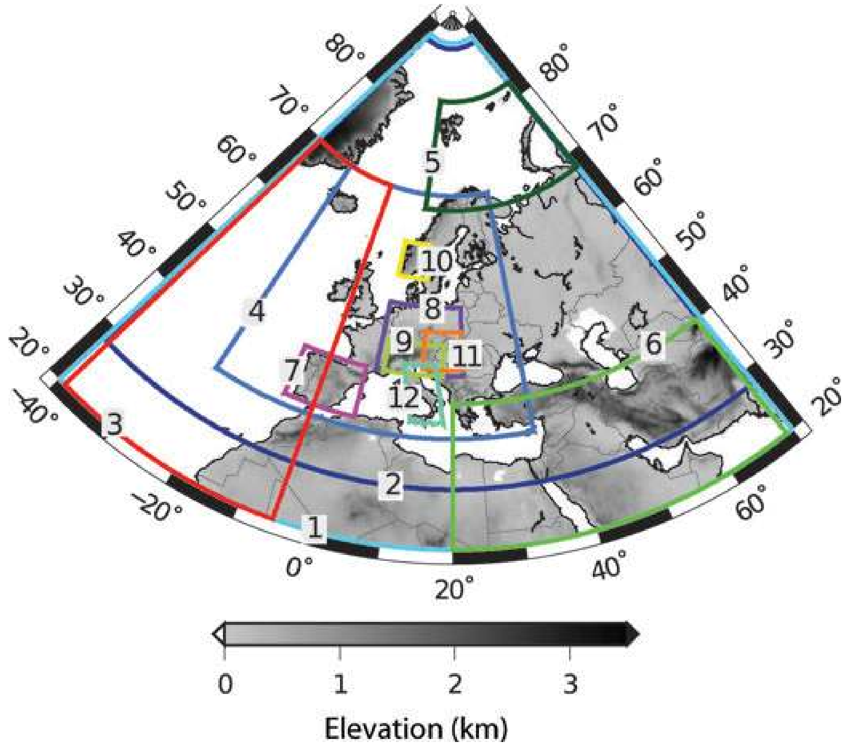

Figure 1. Geographical coverage of the models considered: (1) CRUST2.0 (Bassin *et al.* 2000), sediment map (Laske & Masters 1997) and Meier crustal model (Meier *et al.* 2007); (2) ESC Moho (Grad *et al.* 2009); (3) NOAA sediment thickness; (4) EuCRUST-07 (Tesauro *et al.* 2008); (5) BARENTS50 (Ritzmann *et al.* 2007); (6) AsCRUST-08 (Baranov 2010); (7) Iberian Peninsula Moho (Diaz & Gallart 2009); (8) Moho depth from RF (?); (9) Alps model (Stehly *et al.* 2009); (10) Norwegian Moho map (Stratford *et al.* 2009); (11) Eastern Alps model (Molinari *et al.* 2010b); (12) Italian Moho depth from receiver functions (Piana Agostinetti & Amato 2009).

of interest from a variety of studies, done using diverse data and approaches and characterized by different merits and drawbacks. As such, the resulting model should attempt to retain the best from each constituent and render it with a uniform representation.

The region covered by the new model is the whole European Plate from North Africa to the North Pole (20°N–90°N) and from the Mid-Atlantic Ridge to the Urals (40°W–70°E, Fig. 1). We represent the crust with three layers (sediments, upper crust and lower crust) in each specifying *P* and *S* velocity and density, with a resolution of $0.5^\circ \times 0.5^\circ$ on a geographical latitude–longitude grid. We actually work with a $0.1^\circ \times 0.1^\circ$ pixel size, but then decimate the grid (after anti-aliasing filtering) to the target $0.5^\circ \times 0.5^\circ$ pixels. The longitude–latitude coordinates are defined as geodetic coordinates in WGS84 reference system, the elevation is referred to the reference ellipsoid, and between gridpoints, the model is defined using a bilinear interpolation.

In principle, we could distinguish among soft, intermediate or hard sediments; and upper, middle or lower crystalline crust. However, it is seldom possible to find out information about such interfaces and distinct seismic properties. In fact, just a few of the models we collected (Bassin *et al.* 2000; Ritzmann *et al.* 2007; Baranov 2010) have such high detail in depth, unfortunately coupled to poor resolution horizontally. We thus chose to adopt the simple vertical parametrization, consisting of a sedimentary and two crystalline layers (e.g. Tesauro *et al.* 2008). We did not find any added vertical complexity justified by quality of available information, at the geographical scale of work, or improved ability to model seismic wave propagation. Our goal is a model that can readily be used for the main seismological applications, such as surface wave tomography, *P*-wave tomography, density inversion from gravity data, dynamic topography calculation and waveform simulations. Note that fewer crustal layers are easier to honour in a 3-D mesh, necessary for numerical modelling of seismic wave propagation (Molinari *et al.* 2010a).

We proceed as follows. First of all, global and local models are regridded to our own parametrization. We select the portion of original models that lies in our region of interest and then we obtain the information of interest—such as depth (or thickness) of sedimentary basins; depth of upper, middle and lower crust and velocity structure of these layers. Some models considered have a different layer parametrization than EPcrust. For instance, the sedimentary layer may be divided into soft, middle and hard sediments, such as in the sediment map of Laske & Masters (1997); or the crystalline crust may be divided in more than just an upper and a lower layer, such as in CRUST2.0, AsCRUST-08, BARENTS50. In these cases, the model has to be reduced to our vertical three-layer description. For the crystalline crust, as a general rule, we decide to merge together middle and lower crust into our lower crustal layer. This choice is justified considering that in the models used middle and lower crust have more similar velocity value than upper and middle crust (mean value of *P* speed: upper crust = 5–6.3 km s^{-1}, middle crust = 6.5–6.8 km s^{-1} and lower crust = 6.8–7.3 km s^{-1}). For simplicity, we take the total thickness of the layers and, in order to assign a value of the elastic properties, we made a mean weighted with the sublayer thickness. It may be possible to apply more sophisticated and expensive approaches (Fichtner & Igel 2008; Molinari & Morelli 2009) that ensure strictly equivalent behaviour of a simplified model for instance for surface wave propagation, but such procedures appear specialized for specific seismological applications and do not appear appropriate for this study. Once the model has the same vertical parametrization, each layer is regridded on a finer working mesh of $0.1^\circ \times 0.1^\circ$.

To include point determinations, such as Moho depth from receiver functions, we first need to create a surface (with a grid resolution of $0.1^\circ \times 0.1^\circ$) that honours all the data, and interpolates values between data points. We use the simple *surface* tool of the Generic Mapping Tools (Wessel & Smith 1998) and the ordinary kriging method as implemented by SGeM software (Remy *et al.* 2009).

With all the data reported on grids with the same resolution it is possible to assemble the final model. For each parameter (sediment thickness, upper-lower crustal depth, Moho depth, v_P in sedimentary, upper and lower crustal layers) all the maps are then merged into the wide mesh covering the whole European Plate. In most regions, for each gridpoint more than one estimate of the same parameter is available (Fig. 1). We could then decide and pick for every point the 'best' estimate neglecting the others. However, considering that we have no strict criterion to evaluate the 'goodness' of a model—especially since most of them do not even supply error bars—we proceed differently and average the multiple determinations, with weights chosen to represent a scale of reliability. In principle, this corresponds to the way of combining information with different Gaussian uncertainties multiplying their probability density functions. One could argue that, in the case we have a very detailed result in a region where bad previous knowledge existed, we spoil this high-quality study. However, we note that the weighting scheme can certainly limit this possible pollution to a minimum, and that it would be inappropriate to just hardly cut off pre-existing information that we are not in a position to completely rule out (note that we do not even consider older studies with dubious reliability). Local, recent, high-resolution models overlap with larger scale ones. As we must trust the large-scale model for areas outside the local study, it is undesirable to introduce artificial lineaments along local borders. Assignment of weights remain somehow subjective. A weight is assigned to each model (Table 1, last column) on the basis of the date of publication, the original resolution, the number of data set and the method used in the paper to construct the model. Fig. 2 shows the logarithm of the sum of weights assigned to all models available at each gridpoint, for sedimentary layer thickness and Moho depth. This quantity may be seen as a proxy for information content, useful to identify relatively better-known and less-resolved areas. For each local model, we also use a cosine taper weight at the borders to flatten out the transition between models. Each data grid is then filtered using a Gaussian filter with 60 km half-width.

Several studies concentrate on modelling the depth of crustal discontinuities (Table 2) but specification of the values of seismic parameters is of course essential to characterize the crust for seismological use. Seismic parameters are not as well constrained as the depth of the interfaces. Information derives from laboratories and field experiments, and from refraction studies. Most original information refers to *P*-wave speed so, in each layer, we actually merge the different v_P models. We derive *S*-wave speed and density from scaling relations with respect to v_P (Brocher 2005) derived from a Nafe-Drake curve regression. The so-called 'Brocher's regression fit' is reliable for v_P between 1.5 and 8 km s^{-1}, that is the typical crustal velocity range. Below, we report the formulae we used

$$\rho\ (\mathrm{g\,cm^{-3}}) = 1.6612v_P - 0.4721v_P^2 + 0.0671v_P^3 - 0.0043v_P^4 + 0.000106v_P^5 \tag{1}$$

$$v_S\ (\mathrm{km\,s^{-1}}) = 0.7858 - 1.2344v_P + 0.7949v_P^2 - 0.1238v_P^3 + 0.0064v_P^4. \tag{2}$$

In the upper and lower crust, most of the original information about v_P derives from CRUST2.0, EuCRUST-07, AsCRUST-08 and BARENTS50 (Table 2) whereas in the sedimentary layer we follow a different approach described next. Sediments are characterized by low values of velocity and density, and especially when the thickness of basins becomes large, reliable information about seismic parameters becomes critical. For instance, the sedimentary layer has a strong influence on seismic wave propagation, such as recorded by surface wave dispersion curves. So, a reliable velocity structure of the sedimentary layer is crucial to characterize the seismic behaviour of the final model. Mooney *et al.* (1998), Ritzmann *et al.* (2007) and other authors divide the sedimentary coverage in 'soft' and 'hard' sediments to distinguish between unconsolidated (average v_P of 2.0–3.0 km s^{-1}) and consolidated sediments

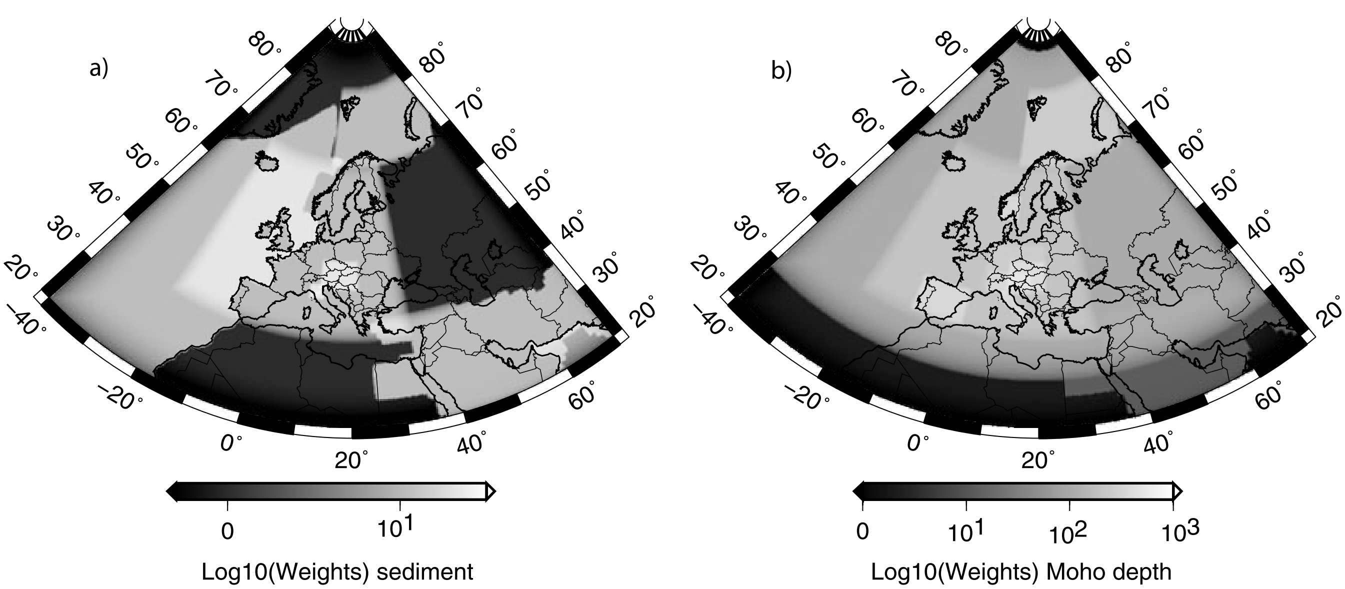

Figure 2. Sum of weights assigned to all the models considered for sedimentary layer thickness (a), and Moho depth (b). This parameter may be seen as a proxy for information content, useful to identify relatively better-known versus less-resolved areas (in brighter or darker shades of gray).

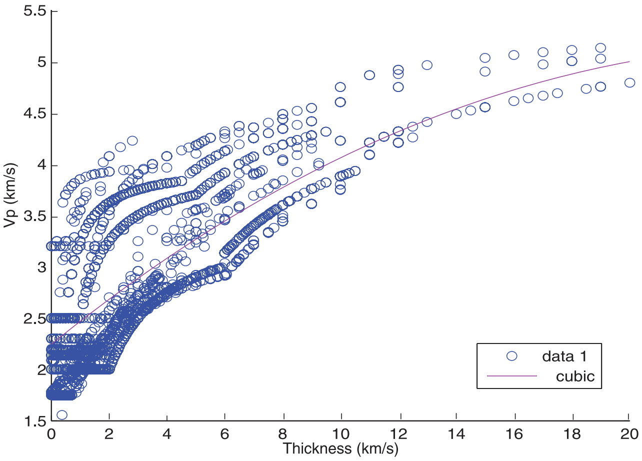

Figure 3. Average *P*-wave velocity in the sedimentary layer as a function of thickness from the LM97 sediment map (Laske & Masters 1997) in the European area. The red line represents the cubic polynomial fit of all the data ($v_P = 2.2 + 0.23\ h - 0.006\ h^2 + 0.000059\ h^3$, v_P in km s^{-1} and h in km).

(average v_P of 4.0–5.3 km s^{-1}). In Europe, the total sediment thickness vary between 0 and 20 km and, looking at Table 2, it may appear that, while basin thickness is rather well constrained, the velocity structure is not. In our data set, information about basin velocity structure can be found in the LM97 sediment maps and in the BARENTS50 model, but in the first, spatial resolution is low and, in the second, the covered region is small. We cannot just use the v_P value of the LM97 sediment map in each gridpoint of our sedimentary layer since we modified the thickness of the layer using other models. Improved information of the sediment properties can be derived from borehole data, seismic profiles (where the determination of v_P still presents difficulties), or laboratory experiments, but retrieving detailed information to locally calibrate the model is beyond the scope of the present study. To overcome this difficulty, that we face in the regions outside BARENTS50 (Fig. 1), we derive an empirical relation between thickness (h) and v_P, based on the data found in the LM97 sediment map. With a polynomial regression fit of all the data points (Fig. 3), we find the relation $v_P(h)$ that could be used to assign a velocity structure in the sedimentary layer. The v_P structure then should be scaled to v_S and ρ using the 'Brocher regression fit' (Brocher 2005).

Our final sediment elastic parameters could be function of the sediment thickness (in km) using the third degree polynomial relation

$$v_P(\mathrm{km\,s^{-1}}) = 2.2 + 0.23h - 0.006h^2 + 0.000059h^3. \quad (3)$$

This empirical relation between thickness and v_P however oversimplifies the assessment of sediment properties. Laske & Masters (1997) and Bassin *et al.* (2000) derived their models using an approach based on geological provinces, consisting of dividing the world crust in crustal types and assigning a 1-D structure to each type. From Fig. 3 it is clear that, for each value of thickness, there is a large variability in velocity due to local geology. In other words, for each geological setting, there is a curve describing average velocity as a function of thickness. As velocity increases with depth in a sedimentary layer, because of pressure and lithogenetic processes, depth-averaged velocity depends on total thickness. We want to keep, in our model, the geological information contained in LM97. We proceed as follows. We analyse v_P as a function of sediment thickness in LM97, and derive the range of v_P values reported in

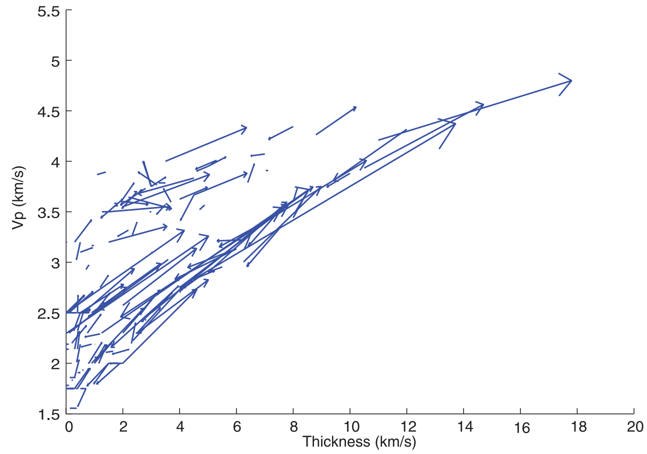

Figure 4. Comparison of average *P*-wave velocities in the sedimentary layer, as a function of thickness, from LM97 (open end of segments) to new values (arrow heads). Only 200 randomly selected points are shown on the graph for clarity. Each arrow represent how we adjust v_P to new layer thickness: start from the original v_P^{LM97} (h) and end up with the new $v_P^{EPcrust}$ ($h_{EPcrust}$).

connection to each value of thickness. Then, for each gridpoint, we find the relative position, within this range, of its v_P as listed by LM97. For the same gridpoint, we finally associate to our new thickness the v_P value corresponding to the same relative position, but within the range related to the new thickness. This transforms the thickness to a new value at some geographical location, but retaining its relative fast/slow nature—presumably connected to geological properties—of LM97. This method is illustrated by Fig. 4, where for each point we plot an arrow starting from the original v_P^{LM97} (h) and ending in the new $v_P^{EPcrust}$ ($h_{EPcrust}$) (only 200 randomly selected points are shown on the graph for clarity).

P-wave velocity at the top of the mantle, representing the velocity of *Pn* waves, is useful for practical purposes, and often associated to crustal models. Only few of the models we collected actually contain such specification (CRUST2.0, BARENTS50). To associate *Pn* velocity to EPcrust, we deem thus more appropriate to include the results of a continental-wide inversion. EPmantle (Schivardi & Morelli 2010) is a tomographic model of the upper mantle obtained by inversion of surface waves, using EPcrust as *a priori* constraint. EPcrust and EPmantle have been conceived as a coherent reference model for European earth structure. Since EPmantle is a v_S model, we calculate v_P from v_S using the scaling relation from Ritsema & Van Heijst (2002).

4 RESULTS AND DISCUSSION

The new Moho depth, sediment thickness and Conrad depth of the new EPcrust crustal reference model are shown in Figs 5(a), 6(a) and 7(a) respectively. We recall that parametrization is based on three layers, representing in turn sediments, upper crust and lower crust. Each layer has laterally varying thickness and seismic parameters (*P*- and *S*-wave speed, density) and is uniform with depth. The working representation is gridline-based on a 0.1° × 0.1° grid, but the distribution format is based on 0.5° × 0.5° grid. Computed *P*-wave velocity for each layer is shown in Fig. 8; v_S and ρ values are derived from the v_P structure using the Brocher relations (Brocher 2005) as explained in the previous section. In the oceanic crust we found *P*-wave speed in the sedimentary layer within the range $1.5 < v_P < 2$ km s^{-1}, with the exception of the deep ocean basin (Barents

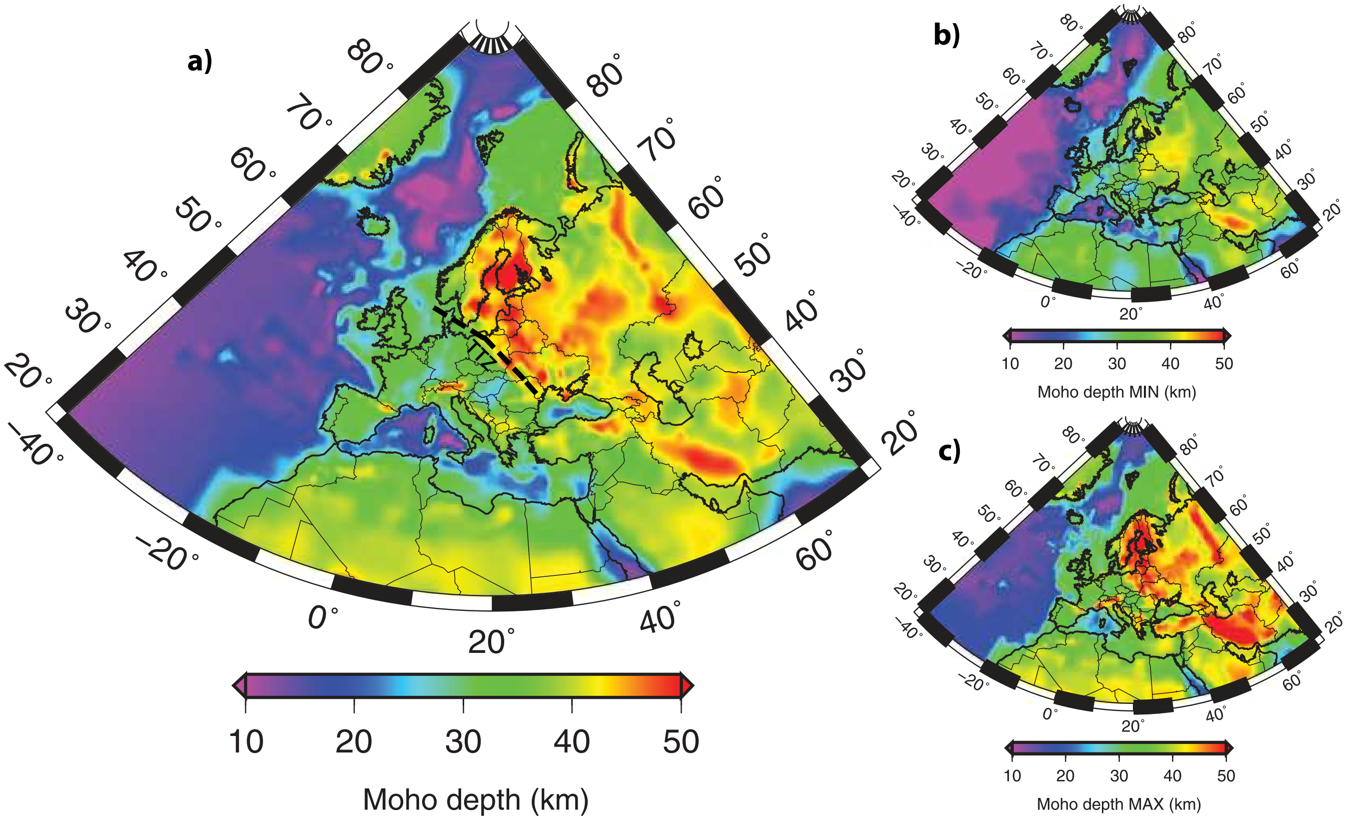

Figure 5. Moho depth (km) in EPcrust (a), compared to absolute minimum (b) and maximum (c) Moho depth in our data set. The dashed black line in (a) represents the Tornquist-Teisseyre Zone (TTZ).

basin, Cadiz, etc.) that have higher v_P (4.5–5 km s^{-1}). In the oceanic upper crust the velocity is between 5 and 6.5 km s^{-1} while in the lower crust we have $6.6 < v_P < 7.3$. In the continental crust we find v_P in the sediments between 2 and 5.3 km s^{-1}; in the upper crust we have $5.6 < v_P < 6.5$ and in the lower crust it is $6.6 < v_P < 7.3$. In the upper mantle, *Pn* velocity determined from Schivardi & Morelli (2010) is in the range from 7.8 to 8.2 km s^{-1} (Fig. 8), typical values for this parameter. These values are in overall agreement with the well known crustal velocities found in the literature (Christensen & Mooney 1995). The lateral resolution of the seismic parameters is limited by the resolution of the original data set.

Our final model is a seismological description of the complex geological structure of the European Plate. Europe is composed by a large variety of tectonic structures ranging from Archean to Cenozoic. The Precambrian part (Artemieva *et al.* 2006) is formed by the East European craton (EEC) ranging from the Urals to the Carpathians, with the unique feature of the presence of a thick sedimentary cover over most of the platform, mainly of 2–4 km, but locally up to 20 km thick. The EEC has in general a flat surface topography (from 0 to 200 m) due to surface erosion since Precambrian (Artemieva 2007) but it shows large undulations of the amplitude of the Moho topography (up to 30 km in variation, from 30 km to more than 60 km) and of the basement thickness (more than 20 km in variation) reflecting its complex tectonic history. The average Moho depth is about 45–50 km. In this region we find the deepest sedimentary basins, that share the common characteristics of a large thickness of the sedimentary cover, uplift of the Moho boundary (up to 36 km), and strong increase of the average velocity in the crystalline crust up to 6.5 km s^{-1}. They are the Dnieper-Donets Basin (a linear rift basin with a thickness of more than 10 km); the Peri-Caspian Basin, with thickness of more than 18 km (a cross-section is shown in Fig. 9); the South Caspian Basin (thickness of 10–16 km) and the Black Sea basin, 18 km thickness (Fig. 10). In Fig. 10(b) we compare a cross-section of EPcrust along the same profile as in Neprochnov & Ross (1978) (Fig. 10a), where the authors review the information about crustal structure in the Black Sea. They found a 25–30 thick crust with a sediment layer of 20 km thickness, consisting of unconsolidated and consolidated sediments with velocity ranging from 3 to 5.5 km s^{-1}. Underneath this layer they put a lower crustal layer with high velocity (6.6–7 km s^{-1}), in good agreement with what we find in EPcrust both for depth and *P* velocity. In Fig. 10c we make a similar comparison, but in a section across the South Caspian Sea (Fig. 10a) taken from Mangino & Priestley (1998). This study was obtained using results from receiver function studies and seismic reflection profiles and, for the consolidated and unconsolidated 16 km-thick sediments, they found $v_P < 4.8$ km s^{-1}; for the granitic crust v_P varies between 4.8 and 6 km s^{-1}, while for the basaltic (lower) crust v_P is between 6.4 and 7.4 km s^{-1}. These values are in good agreement with our final model, while Moho depth under the Caspian Sea is instead 10 km deeper in EPcrust.

In central-southern Europe, between the Tornquist-Teisseyre Zone (TTZ)—Fig. 5—and the Atlantic Ocean, we find a Paleozoic crust where crustal thickness changes very clearly with respect to the deeper East European craton. Moho depth averages at 30 km, with a maximum of 50 km beneath the Alps and the Pyrenees (two of the best constrained zones in our model), and a minimum of 8–10 km under the Tyrrhenian Sea. Some deep basins are also present in this region: the Po Plain basin,

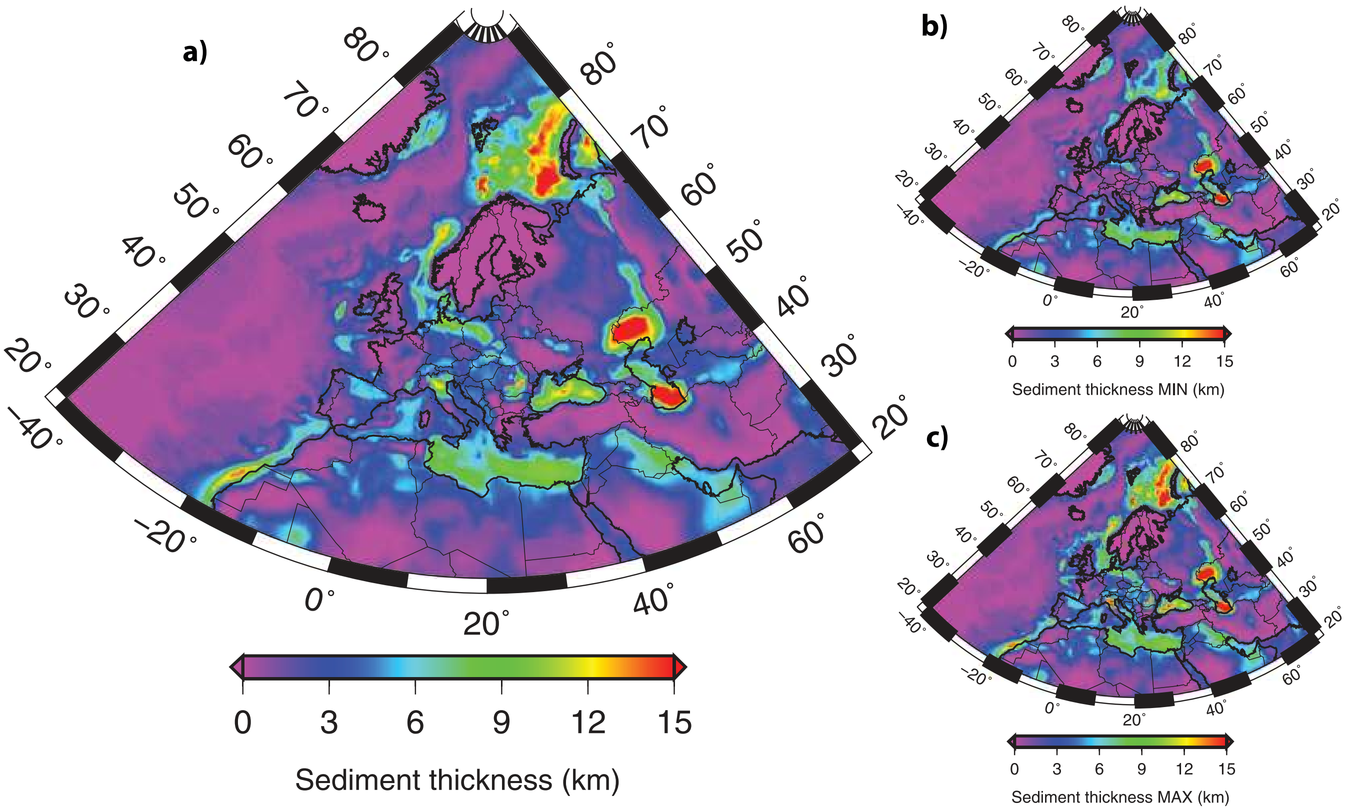

Figure 6. Sediment thickness (km) in EPcrust (a), compared to minimum (b) and maximum (c) sediment thickness in our data set.

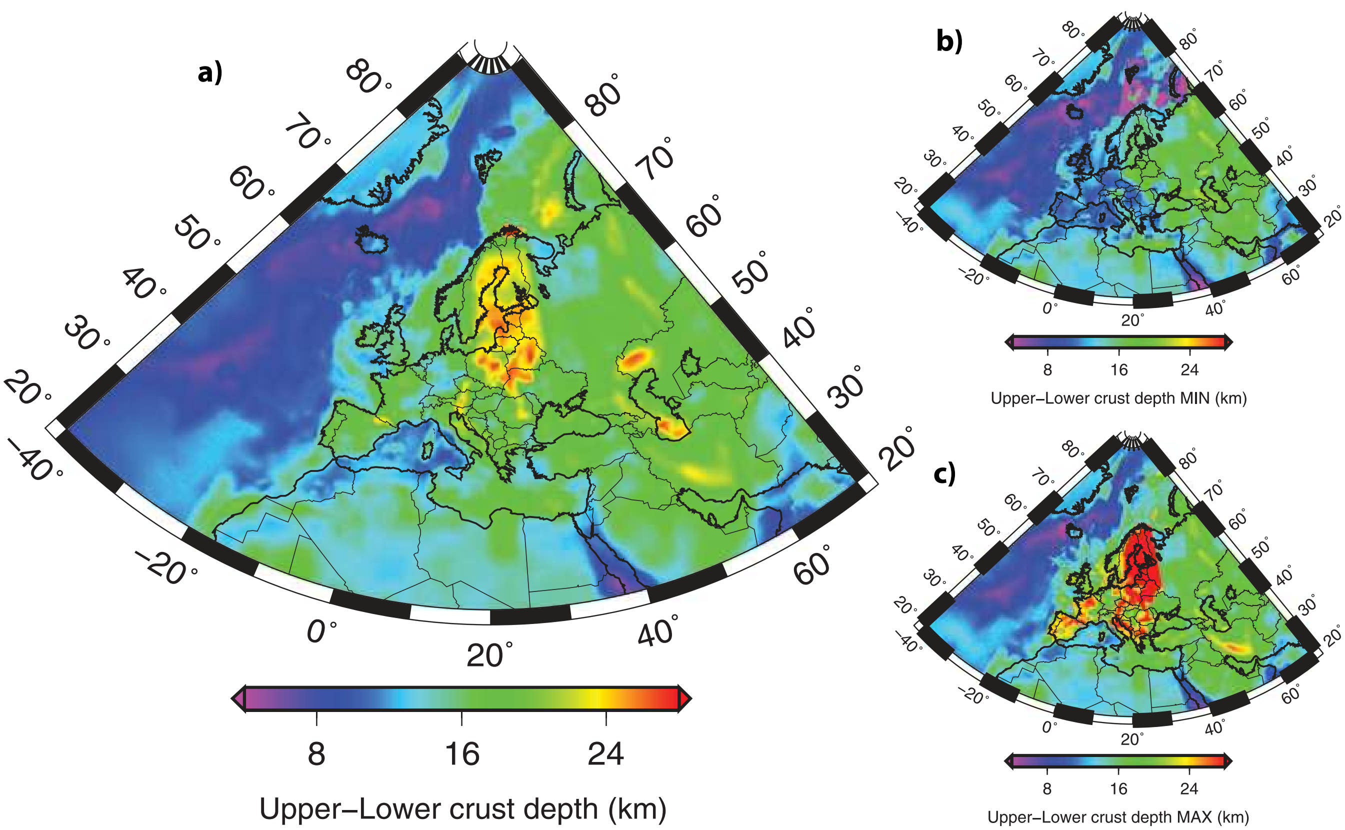

Figure 7. Depth of upper-lower crust discontinuity (km) in EPcrust (a), compared to minimum (b) and maximum (c) discontinuity depth in our data set.

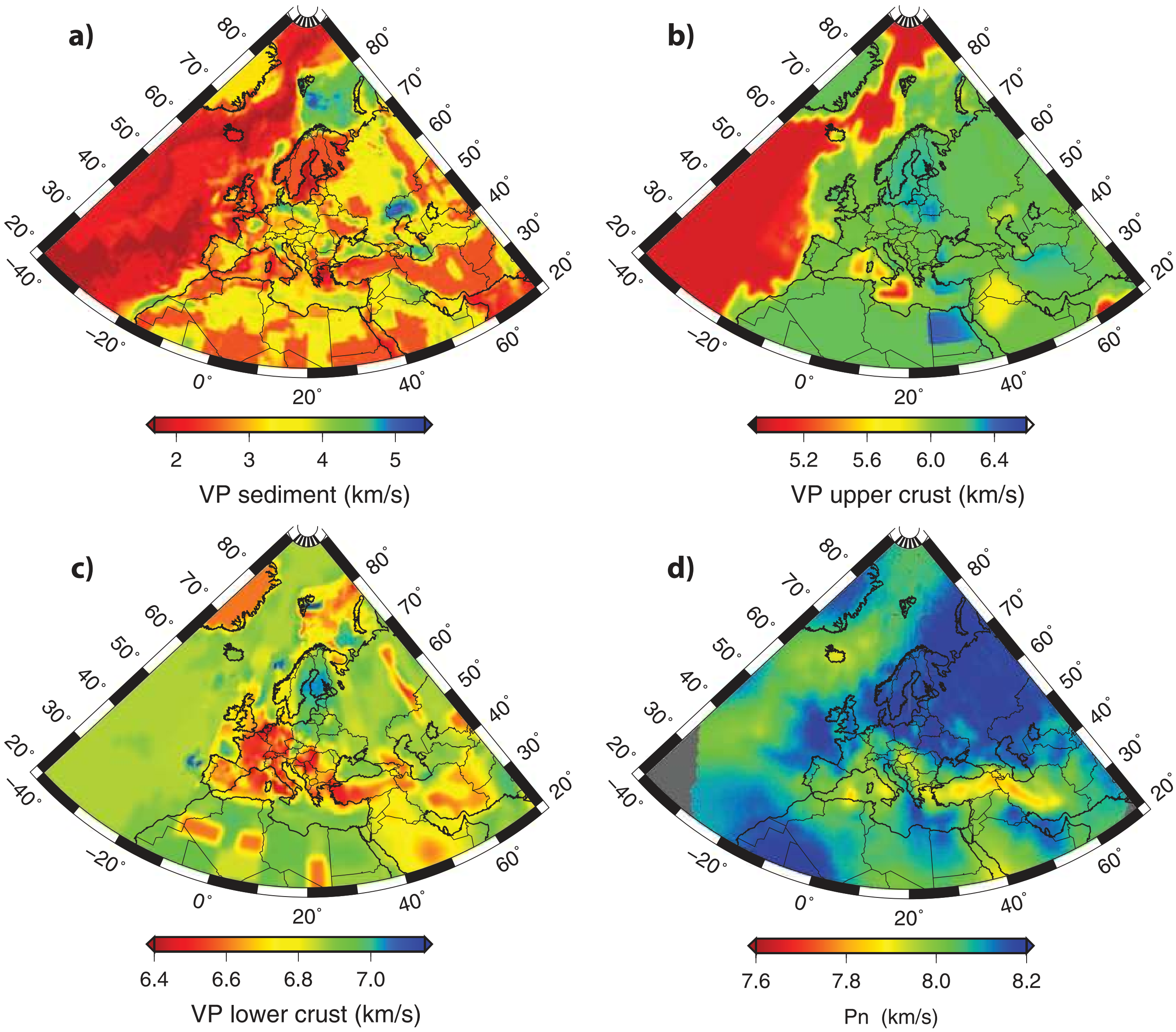

Figure 8. *P*-wave velocity in sediments (a), upper crust (b), lower crust (c) and uppermost mantle (*Pn* velocity, d).

the Pannonian basin (see Fig. 9) and the North-German Basin (4–6 km thickness).

In the Atlantic region Moho depth ranges from 8 km depth, near the Mid-Atlantic ridge, to 20 km in the ocean–continent transition zones. In the Barents Sea and along the continental margin near North Africa and the Norwegian coast, we find deep sedimentary basins with high velocity, in particular in the Barents basin where the local model BARENTS50 predicts $4.3 < V_P < 5.5$. If we compare our model with CRUST2.0, we find variations in Moho depth up to ±15 km. The main differences are in the North Atlantic region, along the continental margin and under the Alps, largely due to improved resolution of our model's Moho boundary. These features are all present in the original models, and their mapping is due to the use of more reliable and detailed data sets.

It is indeed difficult to estimate the uncertainties associated to the parameters of such a model, obtained form an assemblage of different kinds of informations, but it is nonetheless important to make an effort to evaluate them. Few of our constituent models provide an error estimate (Table 1), and a statistical evaluation of the variance of each parameters of the resulting model is impossible. However, considering that, our scheme allows for calculation of minimum and maximum values at each geographical point for each model parameter, we can define a range of variability of the different estimates at a specific location. For each gridpoint we generally have more than one value and, with our weighting scheme, we calculate our best value that will lie between the minimum and the maximum values of the original data set. In Fig. 11(b) we plot the available data for the Moho depth and the final EPcrust Moho (dashed red line) along cross-section C-D (Fig. 11a). In some locations the agreement between the different models is good, while in other regions we can have differences up to 10 km. This cross-section also reveals which are the regions with tighter constraints, and how large the variability could be in the estimations done in different models based on diverse data and approaches. Figs 5(b), (c), 6(b), (c) 7(b) and (c), show maps with minimum and maximum values for each gridpoint. This is a way to represent the variability of the original information of each parameter, and to estimate the maximum variability of the resulting value. Unfortunately, where

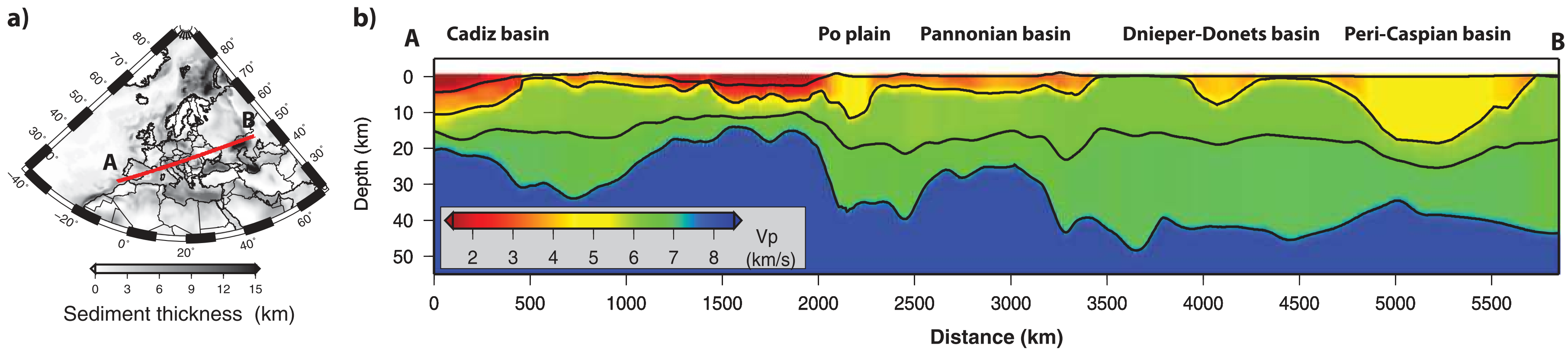

Figure 9. Cross-section of EPcrust (b) along the A-B profile (a). It is possible to recognize deep sedimentary basins, such as the Cadiz basin in the gulf of Cadiz (Spain), the Po plain in Northern Italy, the Pannonian and the Dnieper-Donets basins in Central-Eastern Europe, and the Peri-Caspian basin at the North of the Caspian Sea.

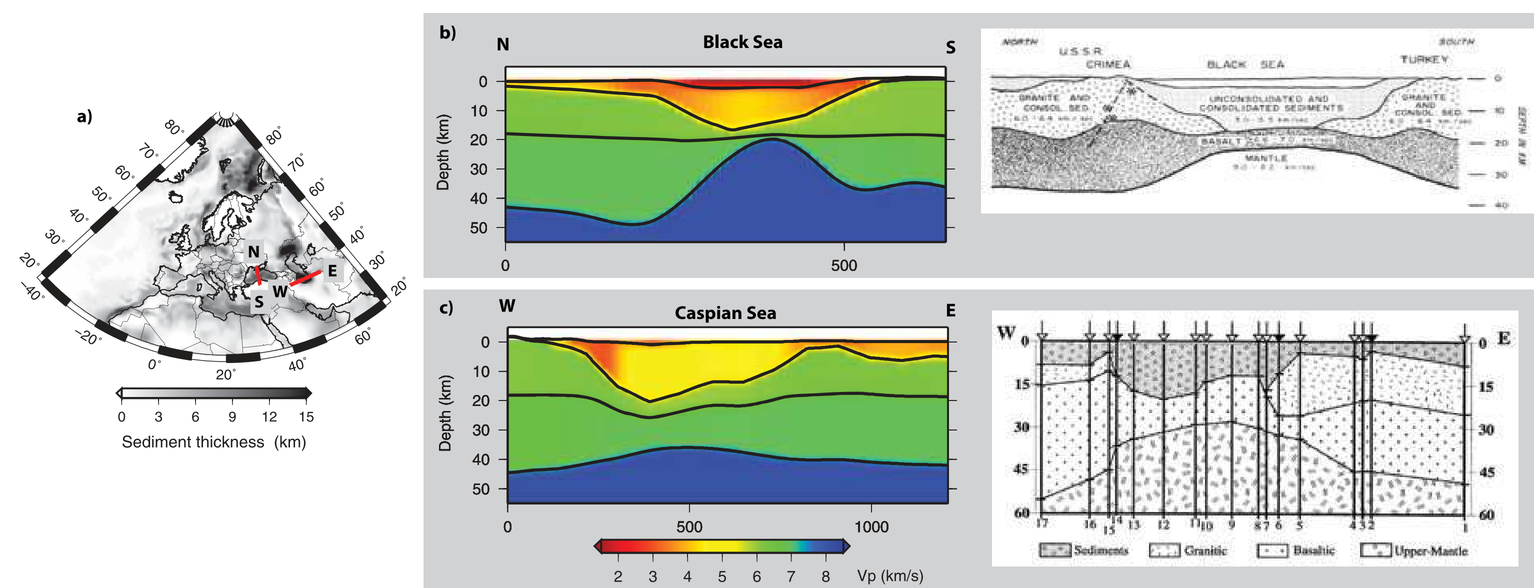

Figure 10. Comparison of two cross-sections of EPcrust along the N–S profile across the Black Sea (b) and W–E across the Caspian Sea (c) with corresponding profiles taken from Neprochnov & Ross (1978) and Mangino & Priestley (1998).

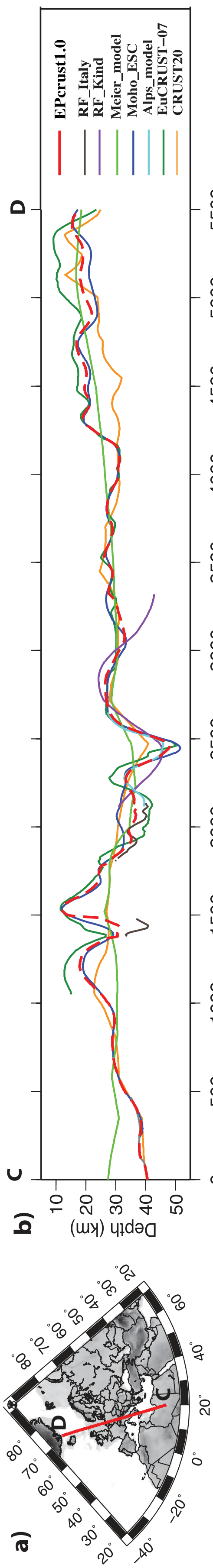

Figure 11. Moho variability along the cross-section C-D. The red dashed line shows Moho depth in EPcrust. The CRUST2.0 and Meier Moho depths are plotted interpolating values between adjacent gridpoints to avoid steps (a faithful visualization should be a staircase).

we have only one source of information, a zero range does not mean certainty.

5 CONCLUSIONS

We present a new crustal model of the European Plate, EPcrust, based on critical assemblage of information and previous models of European crustal structure. EPcrust has some important advantages with respect to earlier models: (i) it covers the whole European Plate, (ii) it is a complete and consistent model, with all the parameters provided, including for the sedimentary layer, (iii) it incorporates most of the recent result concerning the European crust, (iv) it is easy to update by adding new contribution, (v) it is available in a digital format and (vi) it is reproducible.

EPcrust is most suited for use at the broad European scale for a variety of research topics, including: wave propagation modelling at continental scale, crustal correction in tomography, gravity studies, dynamic topography inference and so on. It includes recent studies and it improves the knowledge of crustal properties at regional and continental scale. Recently, our model has been used as crustal correction in a surface wave tomography study to image the European upper mantel Schivardi & Morelli (2010). EPcrust, with respect to CRUST2.0, seems to improve the recovery of upper mantle structure at least in the area where it has a high resolution, such as in the Alps region. However, EPcrust also has some limits. In local studies (shake maps, seismic site response simulations), where a detailed knowledge of the crustal structure is required, this model will not be appropriate because of its low spatial resolution for such applications. Note however that, in the best constrained regions (Alps, Central Europe), EPcrust could be a good candidate for the representation of the crust, also at a smaller scale.

The sedimentary cover is a very important parameter to characterize the seismic response of a crustal model. Unfortunately, information on sedimentary thickness and v_P is not as detailed as one could desire, even at the scale we are working at. We made an effort to always use the most detailed information available about the thickness of the sedimentary cover, and had to make a plausible extrapolation to adjust v_P to the new depths where it was only available from global studies with lower detail.

The interface between upper and lower crust is often considered quite a controversial boundary: it is difficult to find reliable information about it, even from active source studies, as the velocity gap is smaller than the one at the Moho. We nonetheless include such a differentiation within the crystalline crust, following, for instance, EuCRUST-07 (Tesauro *et al.* 2008). This choice enables comparisons, and makes up for a simplified but still realistic representation. In EPcrust, upper and lower crust always have non-zero layer thickness, and their interface is always smoother than other boundaries.

A further improvement of EPcrust could be addressed including new, more refined and higher resolution models resulting from, for example, seismic noise studies, new refraction–reflection seismic experiments or receiver function results. We plan to do this in the future, as new high-quality descriptions of the European crust become available. Although it is not realistic to aim at an improvement in the whole, broad, region, at least the best instrumented areas will presumably see improvements in the near future.

EPcrust is available on http://www.bo.ingv.it/eurorem/EPcrust. The distribution format is based on the TomoJSON data exchange format described by Postpischl *et al.* (2010). In order to make the final model reproducible we also provide, on the website, all the individual contributions used to assembly EPcrust, to create a database for crustal structure in Europe. We plan to update this resource in the future as new local models and data sets will become available.

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