Reference seismic crustal model of the Dinarides

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Abstract

Continental collision zones are structurally one of the most heterogeneous areas intermixing various units within relatively small space. The good example of this are the Dinarides where thick carbonates cover older crystalline basement units and remnants of subducted oceanic crust. This is further complicated by the highly variable crustal thickness ranging from 20 to almost 50 km. In terms of spatial extension, this area is relatively small, but covers tectonically very differentiated domains making any seismic or geological analysis complex, with significant challenges in areas that lack information on crustal structure.

Presently there is no comprehensive 3-D crustal model of the Dinarides (and the surrounding areas). Using the compilations of previous studies, we have created vertically, and laterally varying crustal model defined on a regular grid for the wider area of the Dinarides, also covering parts of Adriatic Sea and SW part of the Pannonian Basin. We have used kriging interpolation to obtain the model parameters. In addition to the seismic velocities (P- and S-) and density, three interfaces in our model were defined – Neogene deposits bottom, Carbonate rock complex bottom and Moho discontinuity. Neogene deposits and the Paleozoic to Eocene carbonate rocks are not present in all areas of the model whereas Moho discontinuity depth is defined for the entire model. To validate the newly derived model, we have calculated travel times for a regional earthquake recorded on a number of seismic stations in the Dinarides area. The calculated travel times have been compared with the travel times in the simple 1D model used for routine earthquake location in Croatia, and it proved to be a significant improvement.

The model derived in this work represents the first step towards improving our knowledge of the crustal structure in the complex area of the Dinarides. We hope that the newly assembled model will be useful for all the forthcoming studies (e.g. as a starting model for seismic tomography, as a model for earthquake simulations) which require knowledge of the crustal structure.

Introduction

First seismic investigation done in the wider Dinarides region can be traced to Mohorovičić's famous discovery of the boundary between Earth's crust and mantle (Mohorovičić, 1910).
The earliest deep seismic sounding experiments investigating the crustal structure beneath the Dinarides were conducted in the 1960s (Dragašević and Andrić, 1975; Aljinović, 1983; Aljinović et al., 1987). The most important results from those early investigations were about thickness of the upper crust (Aljinović, 1983). In more recent times, there was another set of active seismic experiments, the ALP 2002 (Brückl et al., 2007) focused on the investigation of the Eastern Alps but also covered the northernmost part of the Dinarides and the Pannonian basin. As part of the same international experiment, Šumanovac et al. (2009) modeled velocities in the crust and uppermost mantle from the measurements taken along the Alp07 profile located at the crossing between NW Dinarides and SW margin of the Pannonian basin. The results from the Alp07 profile were later extended by several studies including gravimetric modeling, P-wave receiver function analysis and local earthquake tomography (Šumanovac, 2010; Šumanovac et al., 2016; Orešković et al., 2011; Kapuralić et al., 2019), covering only the NW part of the Dinarides and SW part of the Pannonian basin. These studies reported a two-layer crust in the Dinarides, and one-layer crust in the Pannonian basin.

The first Moho map of the Dinarides was compiled by Skoko et al. (1987) utilizing gravimetric data from that area. Similarly, Šumanovac (2010) used results from active seismics (Alp07) to calibrate gravimetric data in the Dinarides and get a more accurate map of Moho depths in the region. Stipčević et al. (2011) were first to use direct seismic measurements in the central and southern Dinarides extending the analysis of Moho depths to that region. For this they employed the P-wave receiver functions (PRF) method and by modeling it found an intra-crustal reflector in the Internal Dinarides. Stipčević et al. (2020) extended the PRF analysis by including significantly more stations and created the map of Moho depths beneath the Dinarides. In that extended receiver function study, Stipčević et al. (2020) report significantly thicker crust under the southern Dinarides compared to the previous studies.

In line with the crustal exploration there have also been some recent investigations of the uppermost mantle. Using the S-wave receiver functions (SRF) Belinić et al. (2018) mapped the lithosphere–asthenosphere boundary (LAB) under the Dinarides. The most interesting feature of that LAB map is the lack of deep lithospheric root beneath the central Dinarides, which was interpreted as thinning of the lithosphere due to possible lithospheric mantle delamination. Two years later, there was a complementary study by Belinić et al. (2020), using the Rayleigh wave tomography to obtain the upper mantle S-wave velocity model for the greater Dinarides area. Similar to their first work, authors reported missing deep

lithospheric root under the central Dinarides, with a deep high velocity anomaly visible in
75 the south Dinarides.

From this short outline of the main geophysical investigations done in the wider Dinarides area it is obvious that the crustal structure of the region is fairly complex with quite a long history of geophysical exploration. Despite this, there has never been an attempt (as far as we know) to combine these results in order to create an extensive regional 3-D crustal model
80 for the Dinarides. The area was covered by the global and continental scale models, but the authors of these studies pointed out the lack of available data in the Dinarides area (e.g., Grad et al. 2009, Molinari and Morelli 2011, Artemieva et al., 2013). Regional scale crustal models are an important factor in all seismic studies relying on waveform modeling and seismic wave traveltime inversion. This is especially exacerbated in the areas with complex crustal
85 structure such as the Dinarides. Therefore, it is necessary to assemble a comprehensive, ready-to-use model based on all the available results.

In this study we focus on the crust and include all the data regarding crustal structure available to us, into a 3-D model covering the area of the Dinarides. The aim is to create vertically, and laterally varying crustal model defined on a regular grid for the wider area of
90 the Dinarides using the kriging interpolation. We estimate that this model will be a good starting point for both body and surface wave tomography studies and provide an excellent base model for local and regional physics-based earthquake shaking scenarios. Besides the velocity as the main parameter describing the crustal structure, we also include existing data on the Moho discontinuity depth, and the approximate Carbonate rock complex (CRC)
95 thickness in the Dinarides. Even though our crustal model is focused on the Dinarides, part of it also covers the SW margin of the Pannonian basin and Adriatic Sea (Fig. 1).

Tectonic and geological setting

The Alpine–Carpathian–Dinaridic–Albanide–Hellenic orogenic system is a part of a Circum-Mediterranean orogenic system. Encircling the Neogene Pannonian Basin System (PBS), this
100 orogenic system is constructed of tectonostratigraphic units derived from both Adriatic microplate and European plate that were separated by Neotethyan ocean (Schmid et al., 2008, 2020, and references therein). Tectonic units were amalgamated by fold-and-thrust systems of different polarity facing either European foreland (e.g., Western and Eastern Alps, and Carpathians) or Adriatic foreland (e.g., Southern Alps, Dinarides, Albanides, Hellenides)
105 as a result of change in subduction polarity between the European plate and Adriatic Microplate (Schmid et al., 2008, 2020; Ustaszewski et al., 2008).

The tectonic evolution of these large orogenic systems, i.e., Dinarides *sensu lato*, started with the Triassic (c. 220 Ma) opening of the Neotethys oceanic embayment between the African and Eurasian Plates. The Neotethys Ocean continued spreading during Late Triassic and

110   Early to Mid-Jurassic, being interrupted only by intra-oceanic subduction of Western Vardar oceanic domain and ophiolite obduction onto the eastern margin of the Adriatic Microplate (see Schmid et al., 2020 for details). According to Schmid et al. (2008) Neotethys Ocean, i.e., the Eastern Vardar oceanic domain remained open through the Late Jurassic–Cretaceous period (see Ustaszewski et al., 2009 and references therein).

115   The final closure of the Neotethys oceanic realm coincided with Adria Microplate and European foreland collision during Late Cretaceous–Early Paleogene (Schmid et al., 2020 and references therein) that initiated formation of the Dinarides (Środoń et al., 2018). During the Middle Eocene–Oligocene the regional NE–SW oriented compression caused NE-directed continental subduction and formation of complex fold-and-thrust sheets in Dinaridic region

120   due to SW-directed thrusting of Adria derived units, whereas in the internal Dinaridic domains, E–W oriented compression caused formation of the W-directed thrusting of the Tisza Mega-Unit over the Internal parts of the Dinarides (e.g., Handy et al., 2019; Schmid et al., 2020; Balling et al., 2021 and references therein). Continuous convergence of the Adria indenter towards the European foreland in the Late Oligocene–Early Miocene further

125   contributed to nappe stacking and thrusting in the Alps and Dinarides, but it also accommodated c. 400 km E-directed orogen-parallel lateral extrusion of the ALCAPA block (including the Eastern Alps, West Carpathians and Transdanubian ranges north of the Lake Balaton) and active "back-arc-type" extension in the PBS (Royden and Horváth, 1988; Ratschbacher et al., 1991a, b; Frisch et al., 1998; Fodor et al., 1998; Tari et al., 1999; Csontos

130   and Vörös, 2004; Horváth et al. 2006; Cloetingh et al., 2006; Schmid et al., 2008).

With an area of 80–200 km wide and nearly 700 km long, the Dinarides are a fold-and-thrust orogenic belt constructed from thrust sheets divided into internal and external tectonic domains (see **Fig. 1**). The lithological succession in the Internal Dinarides is characterized by units that comprise passive margin ophiolitic succession and pelagic derived sedimentary

135   rocks, while External Dinarides are dominated mainly by shallow marine carbonate platform deposits that in places reach 8000 m of thickness (Vlahović et al., 2005). Due to paleogeographic differences and tectonic complexity, Dinaridic lithological succession is spatially highly varied in its thickness and exposure (Vlahović et al., 2005, Schmid et al., 2020; Balling et al., 2021).

140   The oldest rocks cropping out in the External Dinarides are Carboniferous to Middle Triassic siliciclastic–carbonate deposits accumulated along the Gondwanian passive continental margin, which due to regional extensional tectonics marked by Middle Triassic volcanism differentiated the area forming isolated shallow marine carbonate platforms (Vlahović et al., 2005). Through the Middle Triassic–Cretaceous timespan carbonate platforms in the region

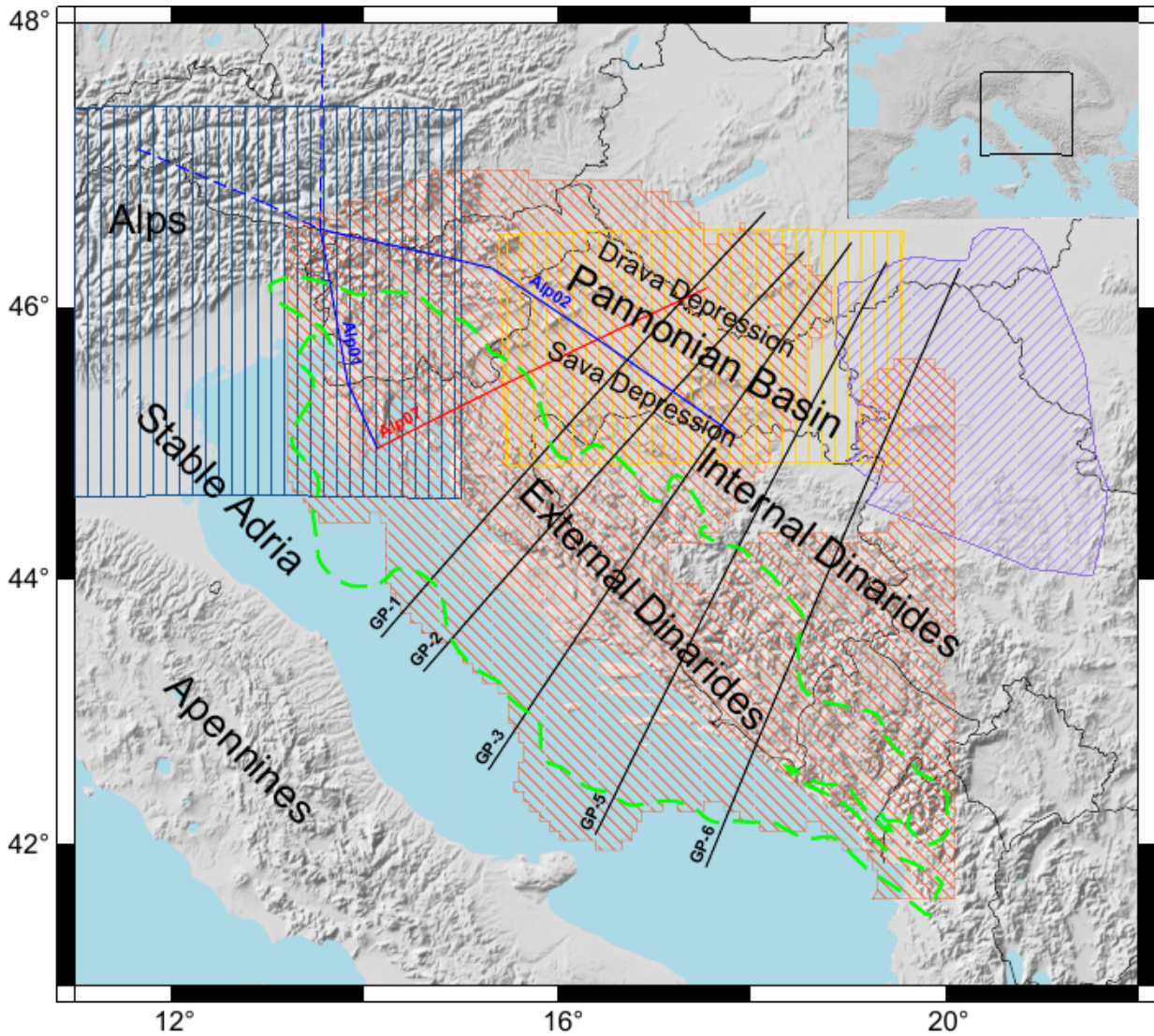
145   were surrounded by deeper marine areas and characterized by mostly continuous shallow-marine carbonate deposition, though the last extensional phase in Toarcian resulted in disintegration in several smaller platforms, including Adriatic Carbonate Platform,

Apenninic Carbonate Platform and Apulian Carbonate Platform (Vlahović et al., 2005 and references therein). During the 120 My of the AdCP's existence, i.e., from Early Jurassic to
150 Late Cretaceous (locally even to early Paleocene – Tešović et al., 2020), the thickness of deposited carbonate succession reached between 3500 and 5000 m (Vlahović et al., 2005 and references therein). The AdCP deposition ended due to regional emergence in the Late Cretaceous–Paleogene, which was linked and enhanced by tectonic deformations and continent collision of the Adria Microplate and European foreland yielding deposition of
155 synorogenic turbiditic deposits ("flysch") in newly formed foreland basins (Vlahović et al., 2005; Schmid et al., 2008 and references therein). The final uplift of the Dinarides took place from the Middle Eocene to Oligocene (Dinaric phase; see Schmid et al., 2008; Balling et al., 2021 with references).

Modeling approach

160 We have decided to create a one-layered crust with laterally and vertically variable parameters. The reason for a choice of one-layered crust was a fact that not all the input data had the interpretation of intra-crustal interfaces, and those that had, did the interpretation in different ways. We used seismic velocities and density as parameters, since that data was readily available. During data collection and integration within the new model we realized
165 that the Neogene deposits in the Pannonian Basin make up a very thick and distinct layer that cannot be ignored, and just incorporating it into a one-layered model would be an oversimplification. In the light of the Pannonian basin sedimentary complex being significantly different from the rest of the crust we included a Neogene deposits layer on top of a laterally and vertically varying crust. The same goes for the Paleozoic-Paleogene CRC in
170 the Adriatic–Dinarides region as there are numerous studies indicating distinct reflections from the bottom of the CRC in the seismic reflection studies (Dragašević and Andrić, 1975; Aljinović, 1983; Aljinović et al., 1987). Given that we had no available data on any of the layer's parameters (velocity or density) in the CRC, we only included the CRC bottom depth in our model. The velocity and density of the CRC were not interpolated separately but were
175 interpolated using the same input velocity (and density) data we had available for the rest of the crust. This choice seems reasonable, since each of the profiles used crosses the part of the investigated area covered with carbonates, and samples its features, even though none of the studies used explicitly interpreted carbonates as a separate layer.

To compile the data for the new crustal model, we used all available and published results.
180 Some of them were not available in a digital form (mostly the studies published before 2010) and had to be digitized manually. The datasets used are very diverse: including two 3-D crustal models (one consisting of two-layered crust with interface depths while other had single-layered crust), several 2D interface maps (Moho depths and Neogene deposits maps), three seismic refraction/wide angle reflection profiles (which were the basic source for

185 velocity data) and five gravimetric profiles. Coverage of the datasets used to compile the 3-D crustal model of the Dinarides is shown in **Fig. 1.** and details are listed in **Table 1**. The exact locations of data points used are shown in Appendix B.

190 **Figure 1.** Compilation of data used in this study. Blue lines mark the positions of Alp01 and Alp02 profiles (Brückl et al., 2007) – only the full line parts are used in the study; the red line is the position the Alp07 profile (Šumanovac et al., 2009); black lines are the positions of gravimetric profiles GP-1 to GP-6 (Šumanovac, 2010). Orange shaded area shows Moho depth data coverage from the receiver function study of Stipčević et al. (2020), blue vertically shaded area is the NAC model (Magrin and
195 Rossi, 2020). The yellow shaded area shows the extent of Saftić et al. (2003) data on the Pannonian basin Neogene deposits, and the purple shaded area shows the extent of data on the Pannonian basin Neogene deposits from Matenco and Radivojević (2012). Green dashed line marks the extension of

the AdCP carbonate rocks from Tišljar et al. (2002). The sources of the data shown in this map are listed in **Table 1**.

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Table 1. List of the data sources used for the Dinarides crustal model.

Profile, model, or project name	Data type	Reference	How were data processed	Parameter that data were used for
Alp01 and Alp02	Seismic refraction and wide-angle reflection	Brückl et al. (2007)	profiles manually digitized	Vp, Moho
Alp07	Seismic refraction and wide-angle reflection	Šumanovac et al. (2009)	profile manually digitized	Vp, Moho
GP-1, GP-2, GP-3, GP-5, GP-6	2D gravity modeling	Šumanovac (2010)	profiles manually digitized	Vp, density, Moho
	Receiver functions	Stipčević et al. (2020)	available in digital form	Moho
The Northern Adria Crust (NAC) model	Multiple data sets	Magrin and Rossi (2020)	available in digital form	Vp, density, Moho
	Isopach map	Saftić et al. (2003)	digitized manually	Neogene deposits bottom depth
	Multiple data sets	Matenco and Radivojević (2012)	digitized manually	Neogene deposits bottom depth
	Carbonate rock complex distribution	Tišljar et al. (2002)	digitized manually	CRC bottom depth

	Geological map	Basic Geological Map (1:100,000) of former Yugoslavia (Osnovna Geološka Karta SFRJ)	digitized manually	CRC bottom depth
	Geological map	1:200,000 Geological Map of Albania (Geological Map of Albania, 2002)	digitized manually	CRC bottom depth
	Constructed and balanced geological cross-sections	Balling et al., (2021)		CRC bottom depth
EPcrust		Molinari and Morelli (2011)	available in digital form	Vp, Moho
SRTM15+V2.0	Global elevation grid	Tozer et al. (2019)	available in digital form	Topography

We decided to use kriging for interpolation. Since kriging requires Cartesian coordinates, we transformed the data to ETRS89-extended/LAEA Europe[1] Cartesian coordinate system,
205 which is defined on the entire investigated area. The transformations were done using the pyproj package (Snow et al. 2021). Kriging interpolation was done using the gstat package (Pebesma, 2004). Interface data were interpolated on a regular 5 km × 5 km grid, and the velocity and density were interpolated on a slightly more complicated grid: the horizontal grid is the same as for the interfaces (5 km × 5 km), but the vertical spacing changes
210 depending on depth (in the first 10 km of depth, the spacing is 0.5 km; between depths of 10 km and 20 km, vertical spacing is 1 km, and at greater depths, vertical spacing is 2.5 km). This scheme was used to account for better sampling and more heterogeneous upper crust.

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[1] https://epsg.io/3035

Initially, we specified a relatively large area between 10° and 22° east longitude and 39° and 48° north latitude as the starting region of investigation. We performed interpolation in the entire initial region for each interface separately.

Kriging does not allow multiple values at a single point in space (i.e. there is no overlapping). Therefore, we needed to handle the overlapping of data from various sources, before starting the interpolation. We tried to reduce subjectivity as much as possible, and therefore included known and estimated variances into the data processing. In case of the data overlapping, we calculated the value which were interpolated (depth for interfaces or velocity/density for layer parameters) at the given point as a weighted mean of multiple values from different sources, with inverse of variances used as weights:

$$\hat{d}_m = \frac{\sum_i \frac{1}{\sigma_i^2} d_{m,i}}{\sum_i \frac{1}{\sigma_i^2}},$$

where $d_{m,i}$ is the value at point i, and σ_i^2 is variance at point i.

We also included error estimates in the final model. To calculate the total error of the model, we followed the procedure applied by Magrin and Rossi (2020) for the derivation of the NAC model. With the assumption of Gaussian distribution of errors, the total variance of the model is the sum of two terms: the variance of the input data, and the variance from the interpolation itself. The interpolation variance term was provided by the gstat package along with the interpolated data. In order to calculate the variance at each grid point, we interpolated the input data variances on the same grid as the data itself.

There are four interfaces defined in the presented model – CRC bottom, Neogene deposits bottom, Mohorovičić discontinuity (Moho), and topography/bathymetry. It should be noted that not all the interfaces are defined at all locations in the model (except for Moho and topography/bathymetry). The parameters of the model (seismic velocities and density) are defined on a regular grid and were calculated differently for the Neogene deposits layer and the rest of the crust.

In order to define the interfaces, we needed the data on their depths. The information about the Moho was compiled from many available studies. The main source of Moho data was the receiver function (RF) study of Stipčević et al. (2020). These data are digitally available as the crustal thickness map on a regular 8.3 km × 8.3 km grid (location shown in **Fig. 1** as an orange shaded area), and it includes error estimates on the same grid. We also included Moho data from seismic refraction and wide-angle reflection profiles (Brückl et al., 2007; Šumanovac et al., 2009; profiles' locations shown in **Fig. 1** as blue and red lines, respectively), and from the gravimetric profiles (Šumanovac, 2010; black lines in **Fig. 1**). All profiles (both

seismic and gravimetric) are available as figures in digital form but had to be digitized manually. Each profile was first georeferenced, and the interpreted Moho depths were digitized every 5 km along the profile's length. For the profile data there are no detailed error estimates but the authors report that the Moho interface was resolved to at least ±2–3 km for refraction and wide-angle reflection profiles. There are no such estimates for the gravimetric profiles. Since the reported errors are only general, we decided to include the error estimates as described in Grad et al. (2009). The authors had a similar problem while building the Moho model for the European continent but had much more input data to come up with reasonable error estimates. According to them, the error estimates for Moho obtained from the seismic profiles should be about 6% of the Moho depth. That estimate nicely fits with the error estimates reported by Brückl et al. (2007) and Šumanovac et al. (2009) for refraction and wide-angle reflection profiles used in this study. As for error estimates in gravimetric profiles, Grad et al. (2009) reported somewhat higher errors, of about 20% of estimated depths. Since there was no information on errors for gravimetric profiles, we used estimates from Grad et al. (2009). For the NW part of our model, the Moho data from the high-quality and digitally available Northern Adria Crust (NAC) model (Magrin and Rossi, 2020) was also included. NAC interfaces are defined on a regular ∼ 5 km × 5 km grid with error estimations on the same grid (location shown in **Fig. 1** as blue shaded area).

The data for the Neogene deposits base depth came from several geological studies. Largest volumes of the study area covered with unconsolidated Neogene deposits are situated in the SW part of the Pannonian Basin. As the basis for the definition of the Neogene sedimentary cover thickness in this region we used data from Saftić et al. (2003), which covers the southernmost part of the Pannonian Basin (yellow shaded area in **Fig. 1**). While preparing the data, we encountered the problem that missing deposits depth information for the eastern part of our study area is causing artifacts on the border of our model after the interpolation. To mitigate this, we included data from the study by Matenco and Radivojević (2012) (purple shaded area in **Fig. 1**). Data from both studies were obtained from georeferenced isolines of the Neogene deposits depths – the isolines were digitized every 5 km, which gives the impression of random spatial sampling. For the error estimation we again turned to the study of Grad et al. (2009), where they estimated that the error for Moho should be about 15% for manually digitized maps. Since there was no error estimate beforehand, we decided to use the same percentage for Neogene deposits depth estimate. For the NW part of the study area, we included deposits bottom information from the NAC model. The area of the Dinarides does not contain a Neogene soft deposits cover, at least not of significant thickness. The basement is overlain by a thick layer of Paleozoic-Paleogene CRC. Therefore, for the area of the Dinarides, we used the CRC distribution described in studies of Tišljar et al. (2002) and Vlahović et al. (2005) and marked it as a zero-Neogene deposits-thickness area, although there are locally some very restricted Neogene deposits

formed within few intramontane basins. Here, the CRC border was georeferenced and digitized manually (green dashed line in **Fig. 1**).

The interface with the least data at our disposal was the Carbonate rock complex (CRC) bottom depth. The CRC bottom depth was estimated combining geological and structural data published in available Basic Geological Map (1:100,000) of former Yugoslavia (Osnovna Geološka Karta SFRJ) with accompanying Explanatory Notes that cover entire Dinaridic area, Geological Map of Albania at the 1:200,000 scale (Geological Map of Albania, 2002) as well as geological-structural data published in studies of Tišljar et al. (2002), Vlahović et al., (2005) and Balling et al., (2021) (see Table A1). Based on the collected data, we determined the spatial extent of the Paleozoic–Paleogene CRC. Assessment of CRC thickness was initially performed at the scale of each of more than 80 geological maps covering the study area, using thicknesses presented in geological columns on each map. Derived values of CRC thickness were further considered in respect to the deformation styles and large-scale structural relations (e.g., Balling et al., 2021). Several regional carbonate nappe systems in the External Dinarides characterized by extensive folding and thrusting could reach a combined stacking thicknesses up to 12000 m, but thicknesses are not evenly spatially distributed. Significant variability of the CRC total thickness in the Dinarides is caused by combination of (1) initial differences in thickness due to significant paleogeographic differences along the Adria Microplate passive margin, since a total thickness of the Adriatic Carbonate Platform and thick underlying and thin overlying carbonates is in the range of 4500–8000 m (Tišljar et al., 2002; Velić et al., 2002; Vlahović et al., 2005), (2) structural position of individual nappe systems in respect to the active collision front, and (3) variable strain rates and stress orientation during the Cretaceous–Paleogene Adria–Europe collision. Nappe stacking systems in the central and southern part of the External Dinarides, where CRC is the thickest, locally incorporate up to four thrust sheets composed of different segments of the entire carbonate succession.

As for the physical characteristics, P-wave velocities were extracted from seismic refraction and wide-angle reflection studies (Brückl et. al, 2007; Šumanovac et al., 2009), densities from gravimetric profiles (Šumanovac, 2010), and P- and S-wave velocities and densities from the NAC model (Magrin and Rossi, 2020). We consider P-wave velocities as the best-defined layer property, since there is largest number of data sources for this parameter. As for the S-wave velocity, there was only data from the NAC model, which is just a border area of our model. Therefore, we did not interpolate S-wave velocity separately, but estimated it from the P-wave model.

The NAC model is defined on a regular grid and was easily included in our data set. The interpreted seismic reflection and refraction profiles (Brückl et al., 2007; Šumanovac et al., 2009) were digitized manually on a regular grid with the horizontal spacing of 5 km (along the length of each profile) and vertical spacing of 1 km. Since the gravimetric profiles are

interpreted in terms of homogeneous layers (density in a layer does not change vertically nor horizontally), we applied a slightly different logic. We used 5 km spacing along the profile and instead of regularly digitizing the densities in depth, we only picked one point in each layer, in the middle of the layer, with assigned layer density. It seemed the most logical course of action, assigning the density value to the middle of the layer, as a way to represent the entire layer. A simplified example of gravimetric profiles digitisation is shown in **Fig. 2**.

We should also mention the error estimation regarding the velocity and density data. The NAC model (Magrin and Rossi, 2020) had reported parameter errors for each grid point and for interfaces, so these were simply included in our data set. Seismic refraction and wide-angle reflection profiles (Brückl et al, 2007; Šumanovac et al., 2009) had a general estimate of velocity error of ± 0.2 km/s and ± 0.1 km/s, respectively. In those cases, we simply assigned that globally estimated error to each digitized data point. In the case of the density data calculated from the gravimetric measurements there was no error estimate. Therefore, we had to use other results to help us quantify error for that dataset. Since the only other source of density data was the NAC model, and since we had no reason to believe that the gravimetric profiles we used (Šumanovac, 2010) are of a lower quality than those included in the NAC model, we assigned the maximum error estimate (to make a conservative estimate) from the NAC model to each of the points from the digitized gravimetric profiles.

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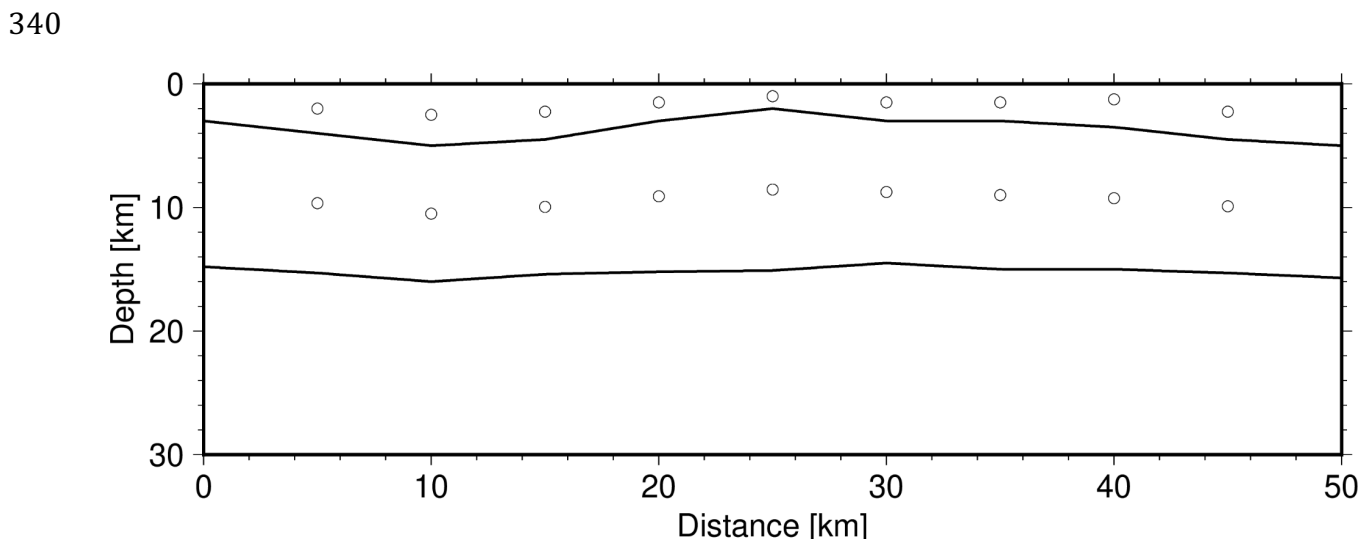


Figure 2. Simplified example of data sampling from one the gravimetric profiles with two interpreted layers. The dots represent digitized data points used to build our model. See text for details.

The most accurate source of velocity data were refraction and wide angle seismic reflection profiles (Brückl et al., 2007; Šumanovac et al., 2009), and the NAC model (Magrin and Rossi, 2020) but as can be seen in Fig. 1 these studies do not cover the central and southern area of the Dinarides. Therefore, we also included velocities calculated using the values of density from gravimetric profiles (Šumanovac, 2010). The differences in the digitization of the

gravimetric profiles have been described in the previous section. In order to calculate P-wave velocities from the available densities, we used Brocher's (2005) empirical equation:

$$V_P = 39.128\rho - 63.064\rho^2 + 37.083\rho^3 - 9.1819\rho^4 + 0.8228\rho^5,$$

where VP is P-wave velocity in km/s, and ρ is density in g/cm3. The equation is valid for densities between 2 and 3.5 g/cm3 (with the correlation coefficient of ~0.999), the condition which was satisfied for all the densities in gravimetric profiles used.

The velocity in the Neogene deposits is poorly known so it was estimated using Brocher (2008) empirical relations (eqs. 1, 3, 7, and 9 in the original article). These relations account for increasing burden pressure, but not for variations in other factors. Therefore, it is justified to use these relations as a first-order approximation, because there are no similar relations readily available for the Pannonian Basin. We used Brocher's Plio-Quaternary relations for the shallowest parts, and relations for Paleogene–Neogene sedimentary rocks for the rest of the Neogene deposits. The values obtained ranged from around 0.7 km/s for depths close to surface, to around 5.6 km/s at greatest depth in the Neogene deposits (which was 7.5 km below surface).

We used a regional EPcrust model (Molinari and Morelli, 2011) as the underlying model to fill the gaps in the data coverage. The EPcrust model is represented by three layers: sedimentary cover, upper crust, and lower crust, with a horizontal resolution of 0.5° × 0.5°. Each layer is characterized by laterally varying P- and S-wave velocity and density, and all the parameters are constant for each grid point. We include EPcrust Moho, Neogene deposits bottom depth and velocity information only in parts with no other sources of data. This was done to remove interpolation artifacts in transition areas between the local data described before and the underlying EPcrust model. Therefore, we implemented a condition to include EPcrust data in our dataset: each grid point defined in the EPcrust model had to be distanced more than 100 km in each direction distant from the other data. That way, the data from the regional EPcrust model will not have too much influence on more relevant, local data.

For topography the SRTM15+V2.0 model (Tozer et al., 2019) was used. It is an updated global elevation grid at a spatial sampling of 15 arc seconds, and it also includes bathymetry. Since the grid of the SRTM15+V2.0 model is much more refined than the grid we used for the definition of our model (15 arc seconds in the topography model compared to about 3 arc minutes in our model), a regridding has been performed by averaging all the values that fall inside our model cell.

After collecting and preparing the data, the next step was interpolation. Ordinary kriging was used to interpolate model interfaces and universal kriging when interpolating the layer properties (Vp velocity, density) as the layer properties are distinctly linearly dependent on depth. After interpolation, we filtered Moho discontinuity and layer properties with a 100-

km wide Gaussian filter to smooth the transitions between different data sources. The smoothing in case of the Neogene deposits and CRC interfaces was omitted because the data used in derivation of those interfaces came from similar sources, and smoothing would conceal known structures due to their relatively small spatial dimensions (e.g. Sava and Drava depressions would have been concealed with smoothing).

We observed how the smoothing influenced the crustal velocity, particularly in the area of the model for which most data was provided by gravimetric profiles. Given that the gravimetric profiles are interpreted in terms of isotropic sections and given that the smaller sections interpreted were roughly about 100 km in dimension along the profile, we chose this as the Gaussian width. It was also confirmed by trial and error that below this width we would observe some artifacts in the model. Model uncertainty was estimated as the sum of two factors: uncertainty in the input data and uncertainty from the interpolation.

The main characteristics of the model

Most of the available data was related to the Moho discontinuity, and therefore the results of that interpolation are considered most accurate. Because of that, we used Moho discontinuity depth errors in combination with lower uncertainty areas of the Vp model as guidelines to mark edges of our model. The final model covers only the area between roughly 13° E and 20° E, and 42° N and 47° N.

Interfaces embedded within the model are shown in the left-hand side of **Fig. 3** (panels a, c and e). The shaded areas mark the region where the model is not well defined. The right-hand side of Fig. 3 (panels b, d and f) shows the standard deviations (the positive square root of the variance) for CRC thickness, Neogene deposits thickness and Moho depth. For the deposits bottom error estimation, we had the following models available: Tišljar et al. (2002), Saftić et al. (2003), Matenco and Radivojević (2012), and Magrin and Rossi (2020) – the rest of the area was filled with regional EPcrust data and is therefore of lower accuracy. Since the model is mostly concentrated on the Dinarides where the topmost cover is predominantly made of Paleozoic–Eocene CRC, Neogene sedimentary cover is negligible for the large part of the model, but areas to the east and northeast of the Dinarides (i.e. SW Pannonian Basin) have thick Neogene deposits reaching thicknesses of more than 4 km. Most prominent features, clearly seen as two red bands in **Fig. 3a,** are the Sava and Drava depressions, with Drava depression being slightly deeper.

CRC bottom depth model (**Fig. 3c**) is almost a mirror image of the Neogene deposits bottom model. In the SW Pannonian Basin, where Neogene deposits are the thickest, the underlying carbonate layer is thin, and vice versa – in the Dinarides, where the CRC is the thickest, there are no prominent Neogene deposits. CRC thicknesses are well over 5 km in the northern part of the External Dinarides and they are getting even thicker going southwards along the

Dinarides chain strike (reaching cumulative thicknesses of almost 15 km in the southern part of the mountain chain). In the Adriatic Sea area, carbonates are thinning out going southwestwards, but that may also be partly caused by the relative lack of available data in that part of the model.

Greatest Moho depths in the investigated region are found in the SE part of the Dinarides mountain chain, where it reaches depths of over 45 km (see **Fig. 3e**). To the NW, along the External Dinarides mountain chain strike Moho becomes shallower, reaching depths of around 40 km. In the SW part of the Pannonian basin crustal thickness is between 20 and 30 km, becoming even shallower going further east. In the Adriatic Sea (within the part covered with our model), Moho is shallower than in the Dinarides, but deeper than in the SW Pannonian Basin, with crustal thicknesses between 30 and 35 km. At the transition from Adriatic Sea to the Dinarides mountain chain, the Moho depth change is gradual, whereas going towards the SW Pannonian basin from the Dinarides, the change is rather abrupt.

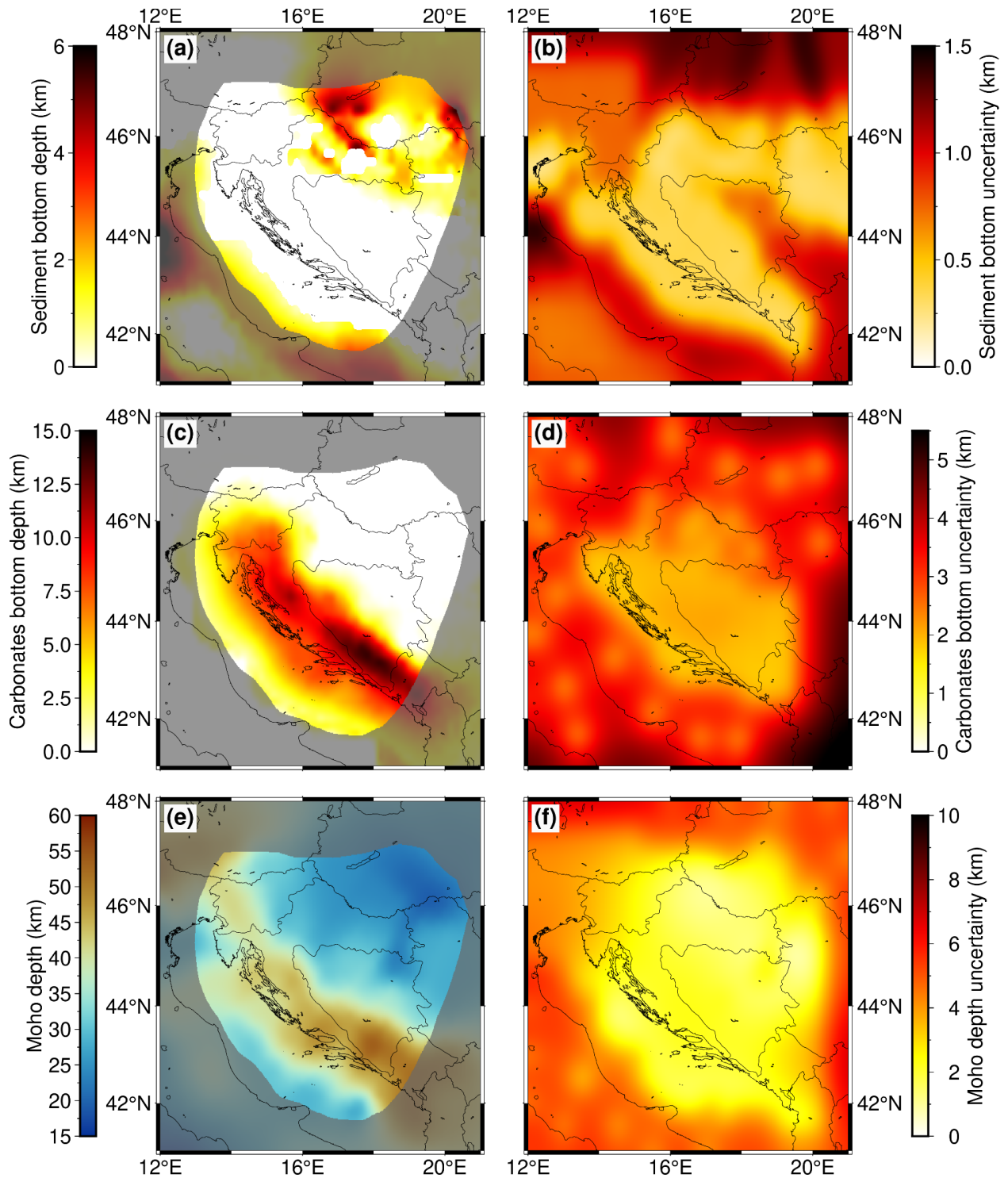
The right-hand side of **Fig. 3** (panels b, d and f) shows interface uncertainties (Neogene deposits, CRC bottom and Moho depth, respectively). Given that a significant contribution to the uncertainty value is the uncertainty from the interpolation itself (which is of greater value at grid points further away from the input data points), one can distinctly see the areas with less data coverage as areas with higher uncertainty. Moho depth uncertainty (**Fig. 3f**) is low in the entire area of interest, i.e. the wider Dinarides region. For Neogene deposits bottom the area of lower data coverage is in the eastern part of the Internal Dinarides where there is less information available on sedimentary thickness (see also **Fig. 1**). On the other hand, that part of the Internal Dinarides is mostly covered by the exposed bedrock largely composed of low-grade metamorphic Paleozoic–Mesozoic formations with thin cover of Mesozoic CRC (e.g. Schmid et al., 2008), so Neogene deposits thickness values here are mostly negligible. For CRC thickness, the area of least accuracy is in the NE part of the investigated area (junction zone between Dinarides-Pannonian basin-Southern Alps). Similar to the previous case, here the low accuracy is due to the lack of measurements on CRC thickness as the region is covered by a thick layer of Neogene deposits.

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Figure 3. Model interface depths and corresponding uncertainties: (a) Neogene deposits bottom depth, (b) Neogene deposits bottom uncertainty, (c) CRC bottom depth, (d) CRC bottom uncertainty, (e) Moho discontinuity depth, and (f) Moho depth uncertainty.

455 The seismic velocity distribution within the model is depicted in the left-hand side of **Fig. 4** at four depths (5, 10, 20 and 30 km) showing most prominent features of the model crustal structure. At a depth of 5 km (**Fig. 4a**), the P-wave velocity in a large part of the External Dinarides is about 6 km/s. Given that we had no independent estimate for the velocity for the CRC, we cannot discern it as a separate layer just looking at velocity values. Outside the

460 External Dinarides, one can see that the velocity in the SW Pannonian Basin at depth of 5 km is slightly lower (just below 6 km/s) than in the rest of the investigated area. At a depth of 10 km (**Fig. 4c**) the velocity values in the SW Pannonian Basin are considerably lower than in the rest of the model, with values around 6 km/s. In the area of the Internal Dinarides, velocity at 10 km depth is slightly higher than in the External Dinarides and beneath the

465 Adriatic Sea. It is hard to discern if this reflects the actual structure, or if it is the consequence of the higher uncertainty in that part of the model (the velocity here was estimated from the density values from the gravimetric profiles, given that there were no other data sources available). Similar situation can be seen in **Fig. 4e** for a depth of 20 km. For this depth the velocities in the SW Pannonian Basin are reaching values above 6 km/s whereas values for

470 the Dinarides are again higher (especially in the internal part) than in the rest of the investigated area, with values above 6.5 km/s. In the lower part of the crust (**Fig. 4g**) at a depth of 30 km in the central part of the Dinarides the P-velocity values are reaching 7 km/s. In the same image the mantle velocity values are shown in grayscale due to the considerable difference between crust and mantle values and the fact that the crustal thickness in SW

475 Pannonian basin is mostly less than 25 km. The mantle velocity variations are better seen in profile sections (e.g. see **Fig. 5f**). At the 30 km depth, the velocity is much higher in the south External Dinarides and below the Adriatic Sea (at least the part covered with our model) than in the northern External Dinarides. The mantle velocity shown in Figs. 4 and 5 is not estimated as part of this model, but was taken from Belinić et al. (2020), by estimating it from

480 the Vs model reported there using the standard P over S-velocity ratio for the upper mantle (Vp/Vs = 1.73).

Fig. 4 also shows error estimates for P-wave velocity at four depths: 5 km, 10 km, 20 km and 30 km (panels b, d, f and h, respectively). The lowest estimates are in the area where we used the NAC model input data (Magrin and Rossi, 2020), and in areas where data from active

485 seismic profiles were available (Brückl et al., 2007; Šumanovac et al., 2009). The disposition of the errors shown in Fig. **4** was expected given the fact that the digital NAC model and the active seismic profiles are of highest quality and that gravimetric data has higher uncertainty. Estimated uncertainty is highest in the area where gravimetric profiles (Šumanovac, 2010) are the main source of the data used to estimate P-velocity.

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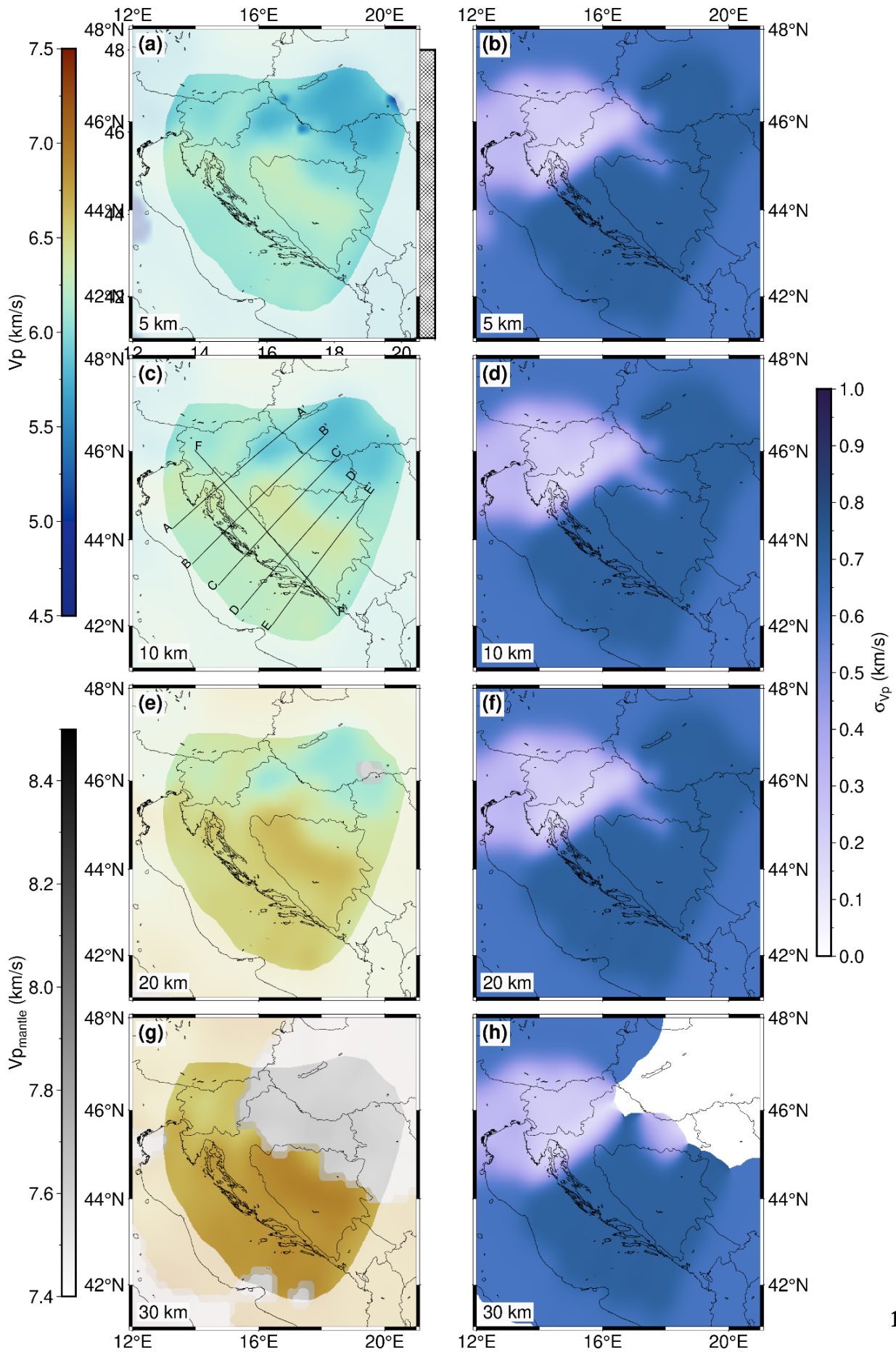
Figure 4. Velocity model depth slices and corresponding uncertainties for 5 km, 10 km, 20 km, and 30 km depth are shown in panels (a)(b), (c)(d), (e)(f) and (g)(h). In the panel (c) the positions of the profiles shown in **Fig. 5** are marked. The areas of lower resolution are shaded, and the gray color scale corresponds to the mantle velocity (see text for details).

495

Fig. 5 shows depth variations of velocity along the profiles (locations marked in **Fig. 4c**) crossing the Dinarides perpendicularly with one profile (FF)' running parallel to the main strike of the mountain chain. The profile AA' (**Fig. 5a**) crosses the Dinarides in their northern part. The maximum Moho depth below the Dinarides on this profile (around 40 km) is the

500 lowest compared to the other profiles. At a distance of ~270 km, the profile reaches the SW Pannonian Basin, which can be clearly seen as the thinning of the crust (between 20 and 30 km) and by the topmost layer of Neogene deposits of lower P-velocity. Although the CRC thickness is indicated with the dashed line, one cannot recognize it by looking just at the velocity values. Generally, the velocities in the part of the profile crossing the Dinarides are

505 larger than in the part of the profile crossing the SW Pannonian Basin. The velocity gradually increases with depth, reaching values of about 6.7–7.0 km/s in the deepest part of the crust, with the exception of the SW Pannonian Basin, where the velocity just above Moho is lower, about 6.5 km/s.

The maximum Moho depth seen on the profile BB' (**Fig. 5b**) is somewhat greater than in the

510 profile AA', a little over 40 km in the part of the External Dinarides. It is shallower in the Adriatic area (in the first 50 km of the profile) and in the Internal Dinarides (after about 220 km). At the very end of the profile, where it reaches the SW Pannonian Basin, one can see the same feature seen in the profile AA': thinner crust and slightly lower velocity just above Moho than in the rest of the profile. The CRC thickness, indicated by the dashed line, in the

515 part of the profile crossing the part where Moho is the deepest, in large parts of the profile coincides with the velocity isoline of about 6.3 km/s. That feature can also be observed on other profiles (CC' and DD') at similar locations.

The profile CC' (**Fig. 5c**) crosses the Dinarides in their central part. Here the maximum Moho depth is almost 50 km. The profile reaches the SW Pannonian Basin only in the last 100 km,

520 but it covers much of the Internal Dinarides. The CRC is of uniform thickness along the part of the profile covering the Dinarides (after the first 100 km, which cover the Adriatic area). The crust is thickest beneath the External Dinarides and is becoming thinner going both towards the Adriatic Sea and the Internal Dinarides. In this central part of the External Dinarides, there is also somewhat higher velocity recorded deeper in the crust, 7.0–7.2 km/s,

525 just above the Moho. In the Internal Dinarides (between 250 to 300 km from the start of the profile), the velocity just above the Moho is a little lower than in the external part, around 6.7–7.0 km/s. As noticed in the previous two profiles, in the SW Pannonian Basin, the velocity just above Moho is even lower than in the Internal Dinarides. Similarly, as for the profile BB',

the bottom of the CRC coincides with the velocity values of about 6.2–6.3 km/s, except in the
530   very beginning of the profile (first ~50 km).

Profiles DD' and EE' (**Fig. 5d and 5e**) cross the southern part of the Dinarides. Here, the Moho
reaches depths of over 50 km. Also, the crustal velocity at those depths is the largest of all
the profiles, reaching almost 7.5 km/s. Greater Moho depths and crustal velocity change can
be best seen in the profile FF' (**Fig. 5f**) running parallel to the Dinarides from northwest to
535   southeast. In this profile Moho depth increases from around 40 km in the northern part to
over 50 km in the southern part. Also, the crustal velocity just above Moho changes from just
below 7.0 km/s in the north to almost 7.5 km/s in the south. Similarly, velocity in the mid-
crustal zone (~20–25 km depth) is somewhat lower in the northern part of the profile, a little
over 6.5 km/s, but becomes higher in the SE part reaching values of about 7 km/s.

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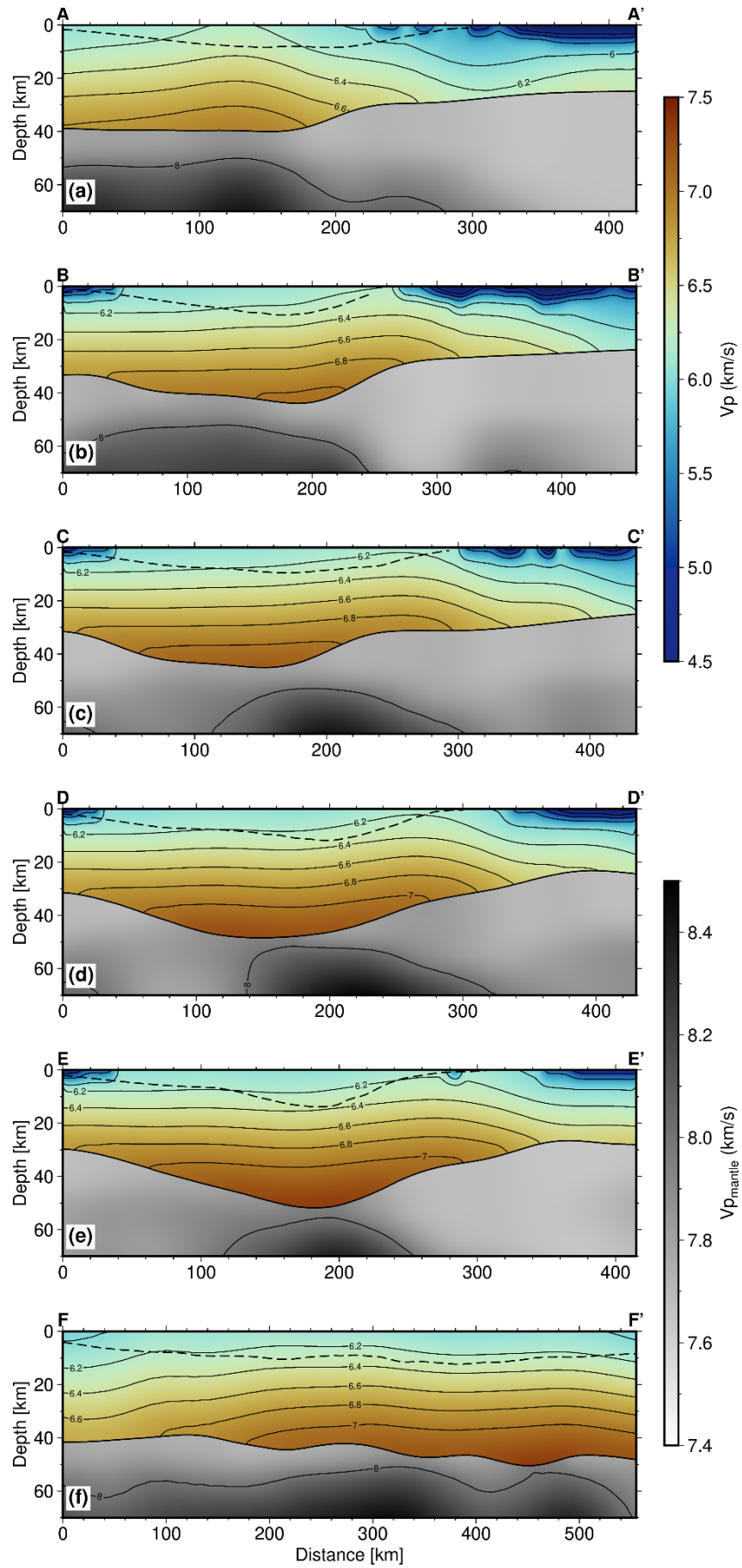
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From the smaller data set which includes the NAC model (Magrin and Rossi, 2020) and the gravimetric profiles (Šumanovac, 2010), we have interpolated the density values for the entire crust. The result is shown in **Fig. 6**. Keep in mind that for this interpolation there were only two sources of data, one of which had densities defined as isotropic layers (Šumanovac, 2010). Therefore, this parameter is much less accurate than P-wave velocity. As expected, this parameter reflects the results of Šumanovac (2010), since that was the main source of data. The density is slightly higher in the area of the Internal Dinarides than in the External Dinarides for all the depths considered here, possibly coinciding with higher density crystalline crust. In the SW Pannonian Basin, the density has much lower values.

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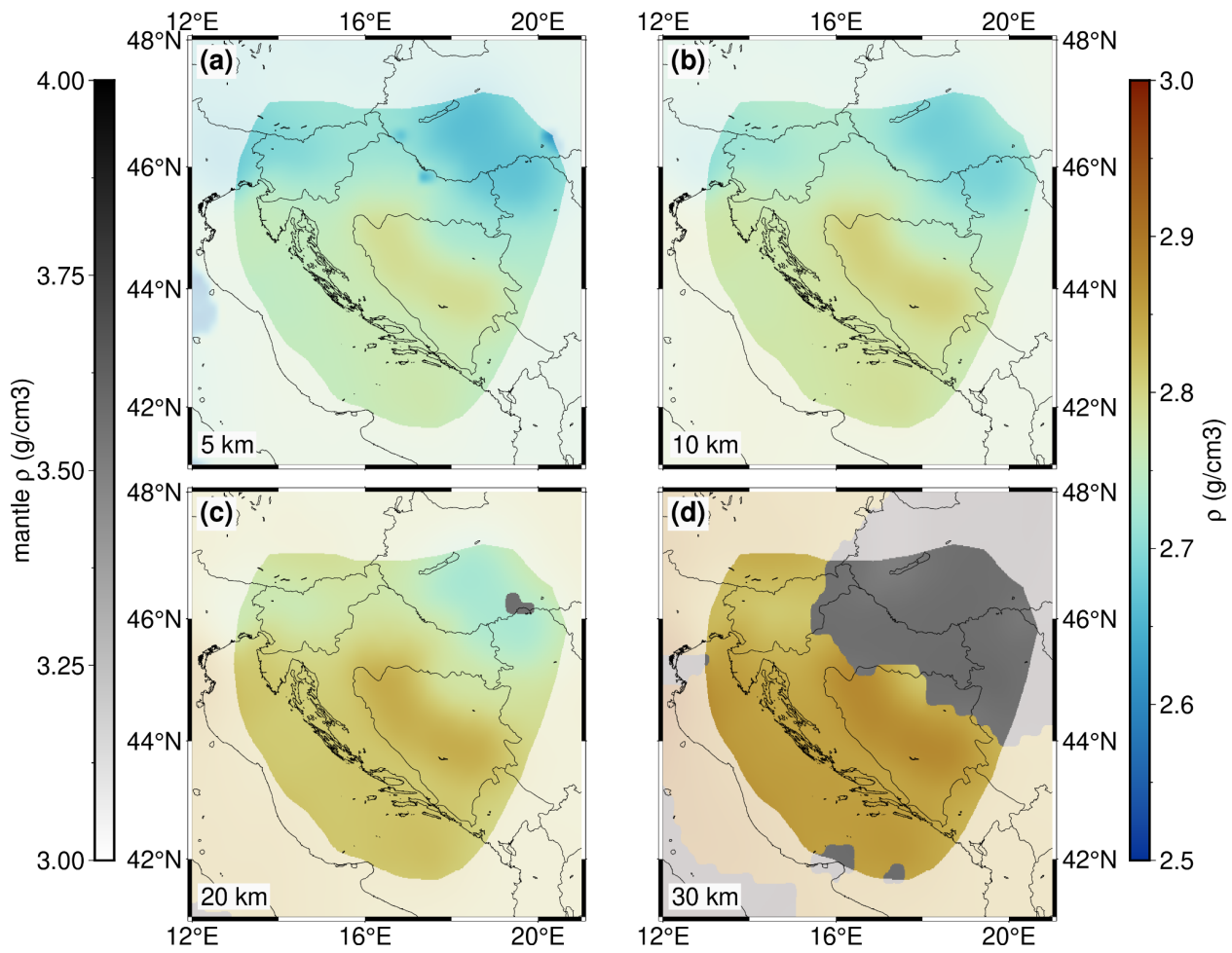


Figure 6. Density model depth slices for (a) 5 km, (b) 10 km, (c) 20 km, and (d) 30 km depth. The areas of lower resolution are shaded, and the gray color scale corresponds to the mantle density (see text for details)

Fig. 7 shows the S-wave velocity at four depths. In this case, there was no measured S-wave velocity data (the only available Vs results were from the NAC model which covers the western corner of our study area) so these values were estimated using the P-wave velocity and Brocher's (2005) empirical relation.

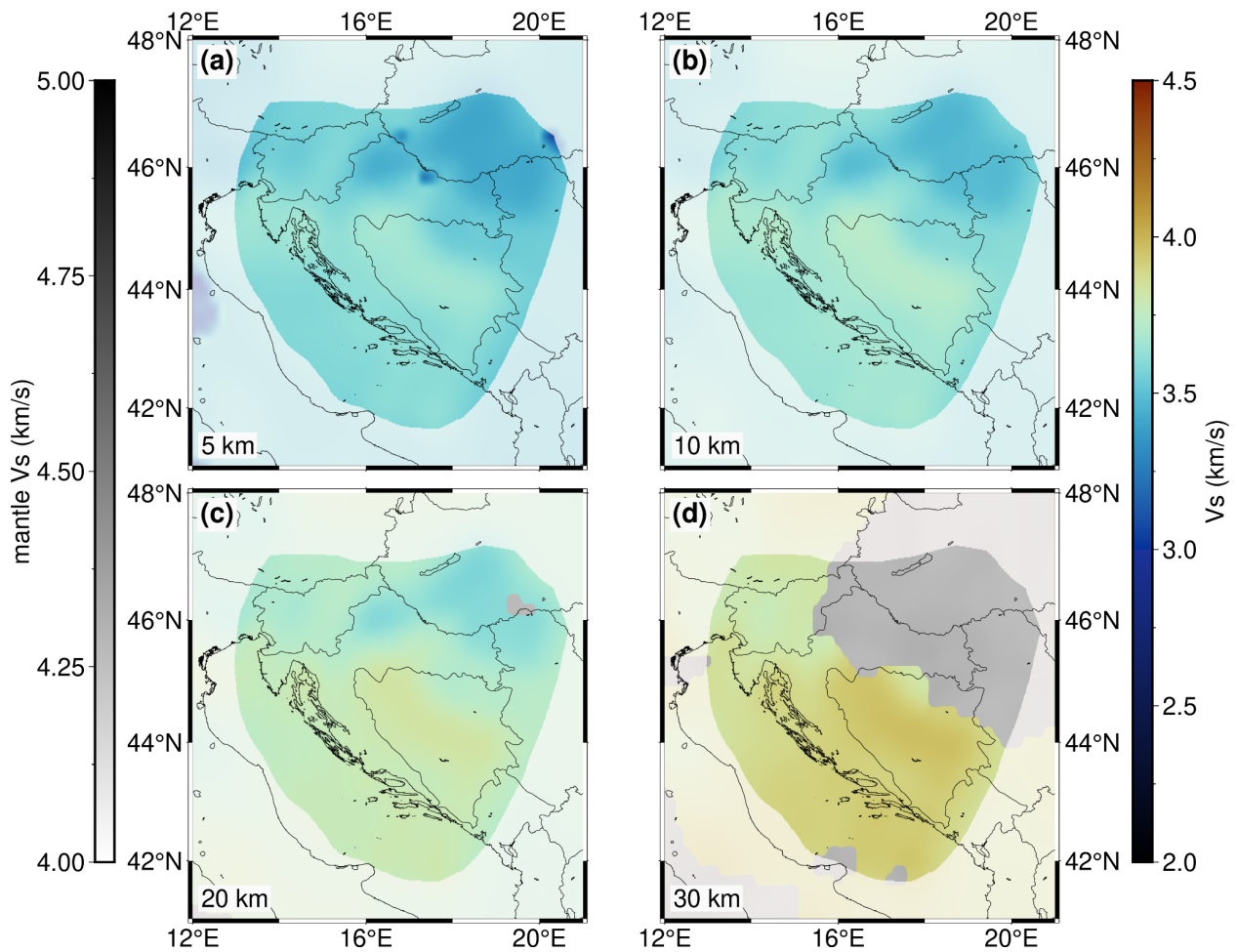



Figure 7. S-wave velocity depth slices for (a) 5 km, (b) 10 km, (c) 20 km, and (d) 30 km depth. The areas of lower resolution and the gray color scale corresponds to the mantle velocity (see text for details).

Discussion

575   The creation of the presented 3-D model was inspired by the need for the more complete seismic model of the Dinarides. Although there have been previous studies estimating various properties of the crust in the region, the complete seismic model of the Dinarides crust and upper mantle did not exist until this study. In this study we assembled data from previous investigations to create a first comprehensive model of the crust for the wider

580   Dinarides area. Moho depth is the best constrained parameter of our model, since there were several good sources of high-quality data regarding this parameter. It confirms what we already know about Moho in the Dinarides, but now it is presented as a comprehensive, ready-to-use model.

    For the Neogene deposit thickness, we used manually digitized maps, therefore having less

585   precise data, but which were originally created from a high number of active seismic profiles and thus strengthening our confidence that this parameter was adequately presented in our model. The thickest Neogene sedimentary cover can be found in the area of the Sava and Drava depressions with thinner cover in the rest of the SW Pannonian basin and almost non-existent in other regions, most notably in the Dinarides, as well as in some hilly areas of the

590   SW Pannonian Basin.

    As can be seen in **Fig. 1**, all the P-wave velocity high-quality data are concentrated in the NNW part of the study area. In the southern part, we relied on the inverted gravimetric profile data (Šumanovac, 2010), which is not the ideal data source due to the high uncertainties and lower resolution. Nevertheless, given the lack of other data sources for

595   South Dinarides, even the data from the gravimetric profiles proved to be of high value. It seems that the velocity in the Internal Dinarides, where we only had inverted gravimetric profile data available, is slightly higher than in the rest of the model. At this point, we cannot discern if it is an actual feature, or some artifact due to lower quality data. The fact is that this is a different tectonic unit, so it is not impossible that it has different features. If we have

600   omitted these data from interpolation, we would have even worse results, because in that case the values would be purely extrapolated. The approach we chose gave at least some constraint to the velocity values in this part of the model.

    It is interesting to note that the trend of velocity in the lower part of the crust corresponds to the velocity trends in the uppermost mantle. There is a positive velocity anomaly in the

605   part of the uppermost mantle right below (or slightly offset from) the part of the crust with the deepest Moho. These positive anomalies in the work of Belinić et al. (2020) have been interpreted as a signal of the subducting Adria Microplate. Our model is mere interpolation of what was already known, but perhaps what we see here is part of the Adria crust being dragged along with the uppermost part of the mantle in the subduction below the Dinarides.

610   As can be seen in the Fig. 5a, which is crossing into the SW Pannonian basin, the crustal

velocity in that part is much lower than in the Dinarides, a feature also observed by Šumanovac et al. (2009). Perhaps the lower velocity is a feature of the Pannonian crust, whereas the relatively higher crustal velocity is a feature of the Adria crust. To make a definitive conclusion, more investigation should be performed.

615   The velocity estimation for the Neogene deposits and the Mesozoic carbonates proved particularly challenging since there is little available data about this parameter. In the case of Neogene deposits, we used Brocher's (2008) relation for the deposits of similar age. For the CRC we could not derive any velocity–depth relation due to the lack of available data, so in this case, we simply used the same velocity interpolation as for the rest of the crust. It
620   seems, though, that at least in some parts of our new model, the CRC bottom depth coincides with the velocity of around 6.2 - 6.3 km/s.

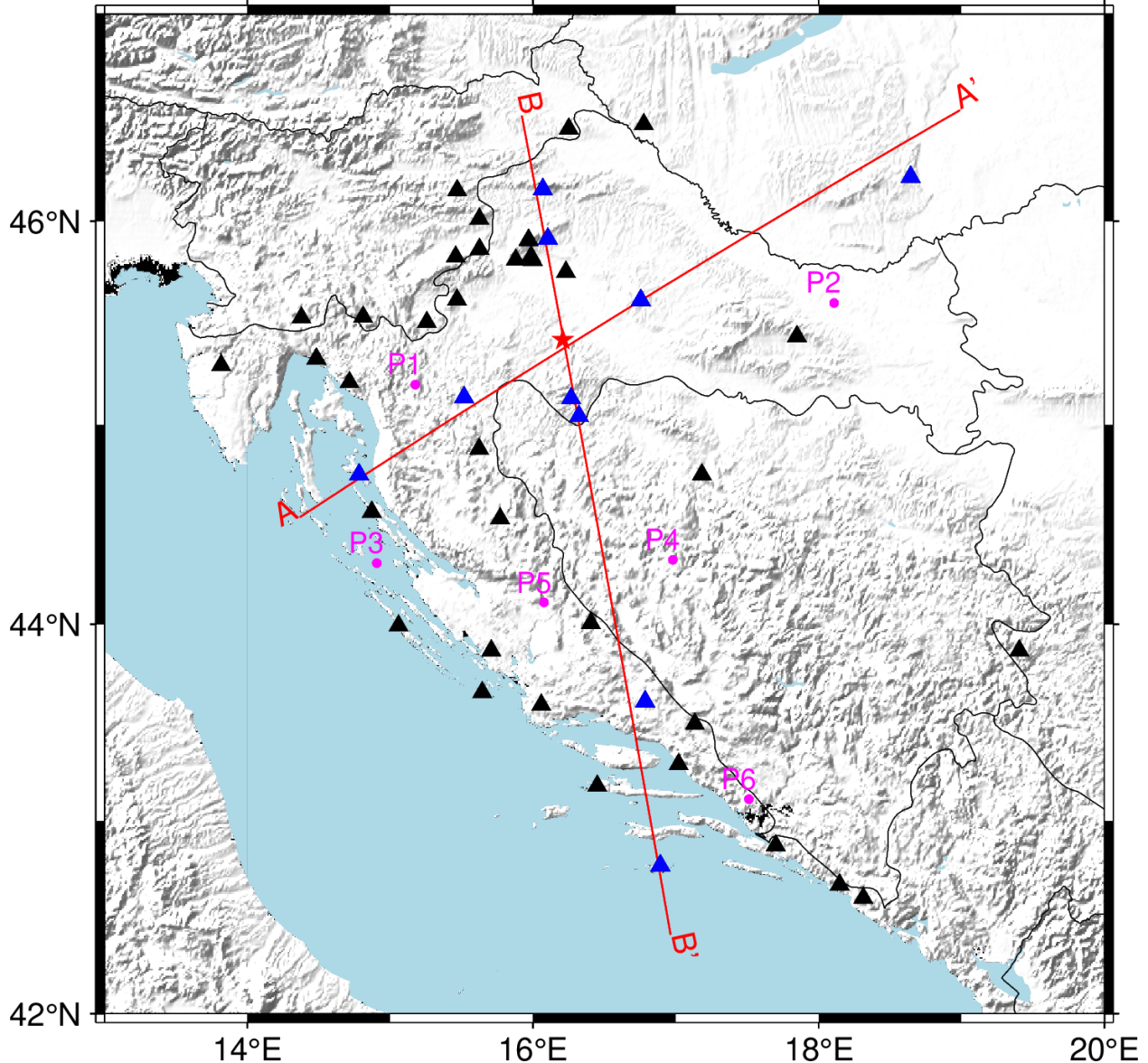
To test how well the newly derived 3-D model represents the true structure, we calculated the travel times for a regional earthquake recorded on representative seismic stations in the wider Dinarides area (**Fig. 8**). We also calculated travel times using the simple 1D model with
625   two isotropic crust layers currently employed for routine earthquake locating in Croatia (B.C.I.S. (1972), Herak et al., 1996). The 1D model's topmost layer is characterized by P-velocity of 5.8 km/s, and the deeper crustal layer has the P-wave velocity of 6.65 km/s. For the same model the uppermost mantle velocity is 8.0 km/s. We then compared the travel times from both models with the true measured travel times. We used the Pn and Pg phases
630   of the 2020 Petrinja Mw6.4 earthquake. The location of the earthquake (42.4188 °N, 16.2082 °E, 7.57 km) and the stations that recorded the wave onsets are shown in **Fig. 8**. For the same stations we calculated the travel times using the 1D and the new 3-D model. For travel time calculation we used the Fast Marching Method (de Kool et al., 2006) as implemented within the FMTOMO package (Rawlinson and Urvoy, 2006).

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Figure 8. A map showing the epicenter of the Petrinja 2020 earthquake sequence mainshock (red star) and stations we used for calculation of traveltimes (black and blue triangles). Travel times for all the stations are shown in **Figs. 9** and **10**. The red lines mark the positions of the cross sections shown in **Fig. 11** (section AA') and **Fig. 12** (section BB'). The stations coloured blue are the ones

640   shown in **Fig. 11** and **Fig. 12**. The points colored in magenta mark the position of 1D models shown in **Fig. 13**.

Calculated travel times are shown in **Figs. 9** and **10**, for Pg and Pn phase, respectively. Figures show the differences in travel times calculated between the models (both 1D and 3-D) and observed travel times. When looking at Pg phases (**Fig. 9**), we can see improvement

645   in calculated travel time accuracy when using the 3-D model (with respect to the 1-D model) for epicentral distances smaller than 50 km and over 100 km. For smaller epicentral distances, the more accurate travel times in the 3-D model are connected with better

specification of Neogene sedimentary cover with low P-wave velocity. On the other hand, for epicentral distances between 50 and 80 km 1D and 3-D models travel times are similarly offset compared to the observed travel times, with times calculated using the 1D model being slightly more accurate. We believe this is due to the less accurate velocity sampling in the upper crust in the transitional zone between Internal Dinarides and SW Pannonian Basin and lack of knowledge about spatial coverage of the CRC in this area. For greater epicentral distances we can see that travel times calculated using the 3-D model are much more accurate compared to those calculated using the 1D model. That means that the crustal velocity derived in our 3-D model is a considerable improvement of the simple 1D model.

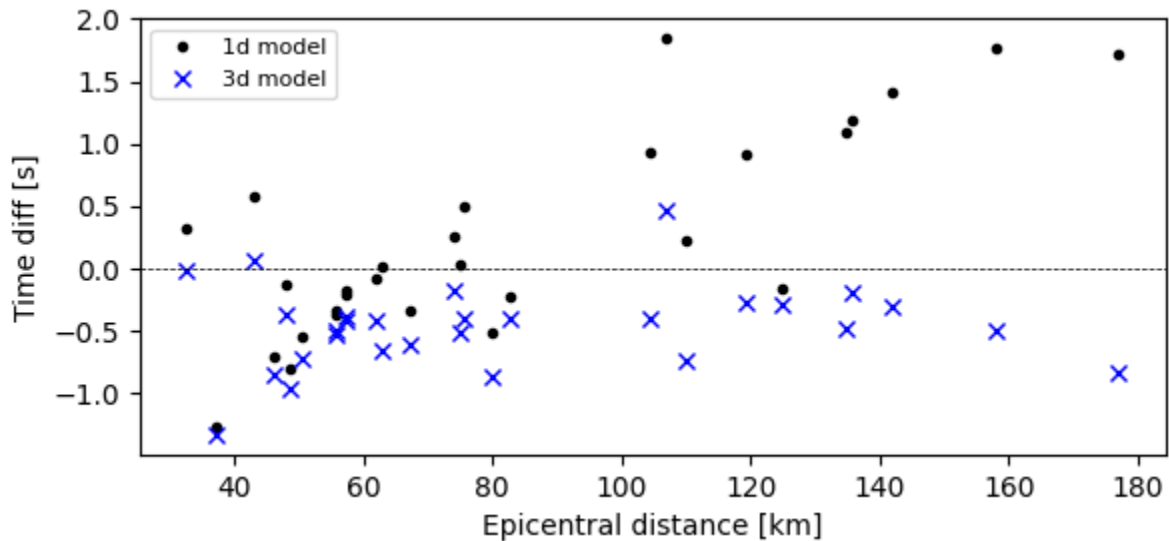



Figure 9. Pg phase travel times for the 2020 Petrinja earthquake: time difference between 1D model and observed travel times (black dots) and between 3-D model and observed travel times (blue crosses).

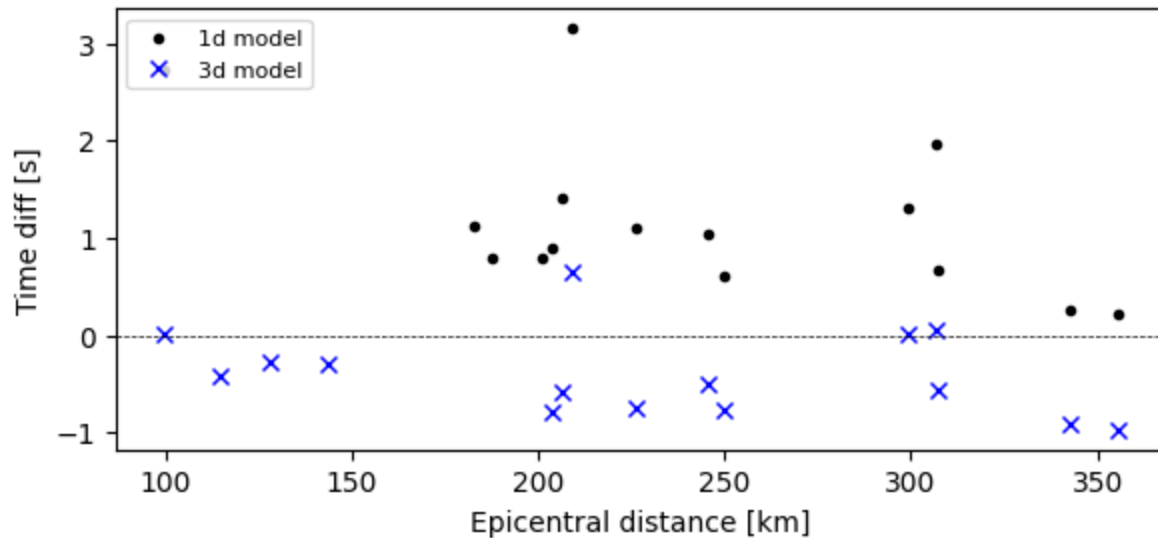
Concerning the Pn phases (**Fig. 10**), we can see that the travel times calculated using the 3-D model are generally closer to the actual observed travel times for all the epicentral distances shown than those calculated using the 1D model. In case of Pn phases, the uppermost mantle velocity plays a great part in the total travel times, so both the crustal model we derived here and the mantle model from Belinić et al. (2020) show improvement compared to the 1D model. There is still room for improvement in the uppermost mantle velocity since the model of Belinić et al. (2020) we used here is most accurate for greater depths (80–100 km).

670 **Figure 10.** Pn phase travel times for the 2020 Petrinja earthquake: time difference between 1D model and observed travel times (black dots) and between 3-D model and observed travel times (blue crosses).

Figure 11 shows a cross section AA' (from **Fig. 8**) of the newly derived 3-D model, with the calculated ray paths using the simple 1D and 3-D models. The section was chosen in a way
675 to show travel times for both Pg and Pn phases, and we also tried to position it in such a way that it crosses almost perpendicularly the main strike of the Dinarides. There is also another cross section shown in **Fig. 12** (BB'), oriented approximately north to south. Position of the stations shown in cross section BB' is also marked in **Fig. 8**. From both profiles it can be seen that the seismic rays cover completely different paths depending on whether they were
680 calculated using the 1D or the 3-D model. For example, the Pg phases, when calculated using the 3-D model, travel much deeper than their 1D counterparts. Also, since the Moho in the Pannonian Basin of our 3-D model is much shallower than the Moho in the simple 1D model, the calculated rays using either 1D or 3-D model travel on very different paths in the uppermost mantle. That is particularly visible in **Fig. 12**, in case of the ray path between the
685 source and the most distant station shown. The ray path calculated for the 3-D model reaches depths of almost 60 km, while the same ray path calculated for the 1D model reaches depths of only 40 km, which is a huge difference. Given all that, it can be concluded that precise knowledge about the crustal model is very important for all seismic applications.

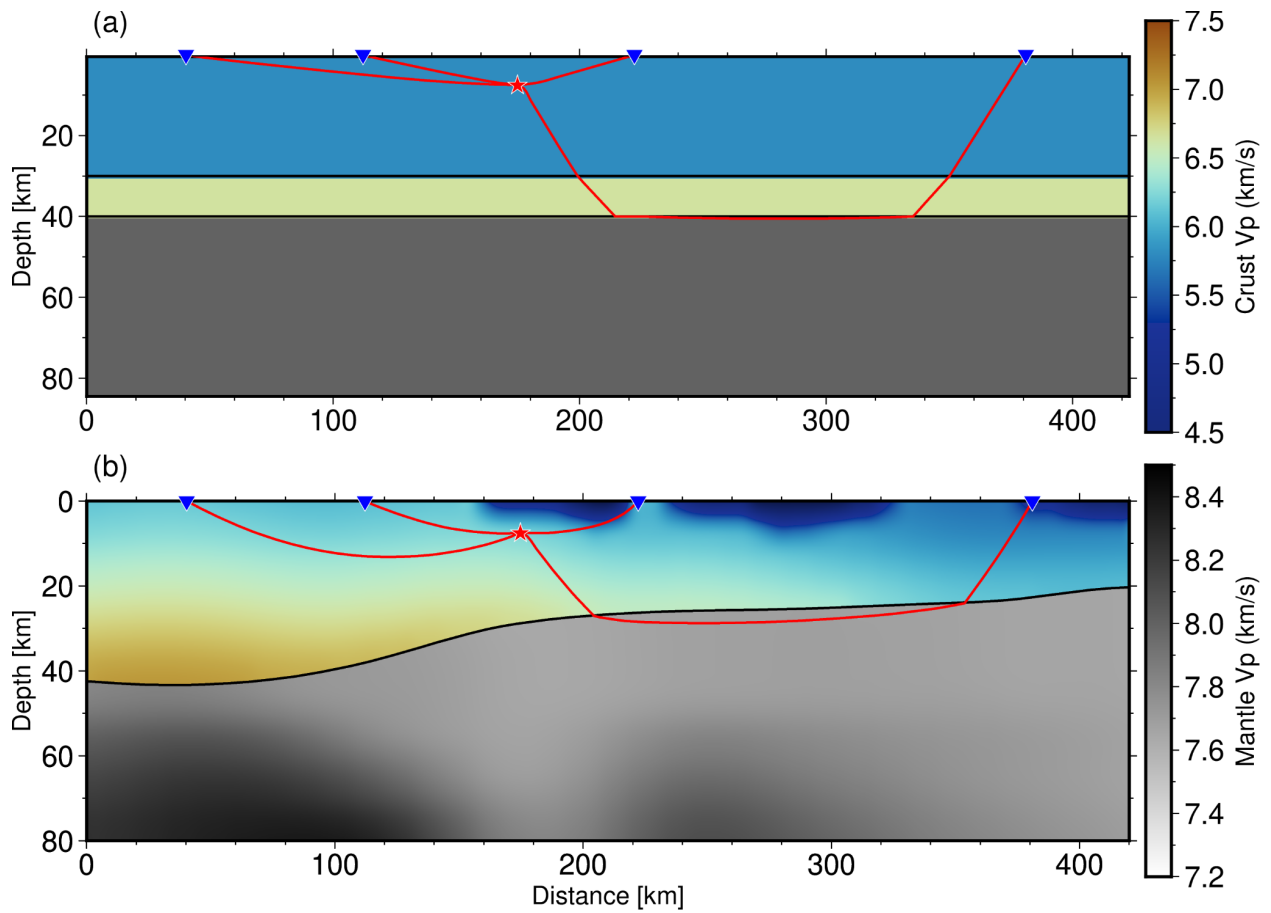



Figure 11. Earthquake ray paths in cross section AA' (for location see **Fig. 8**) for two models: (a) a simple 1D model with two isotropic crustal layers, and (b) our 3-D model with one anisotropic crustal layer. Colorbars are the same for both panels.

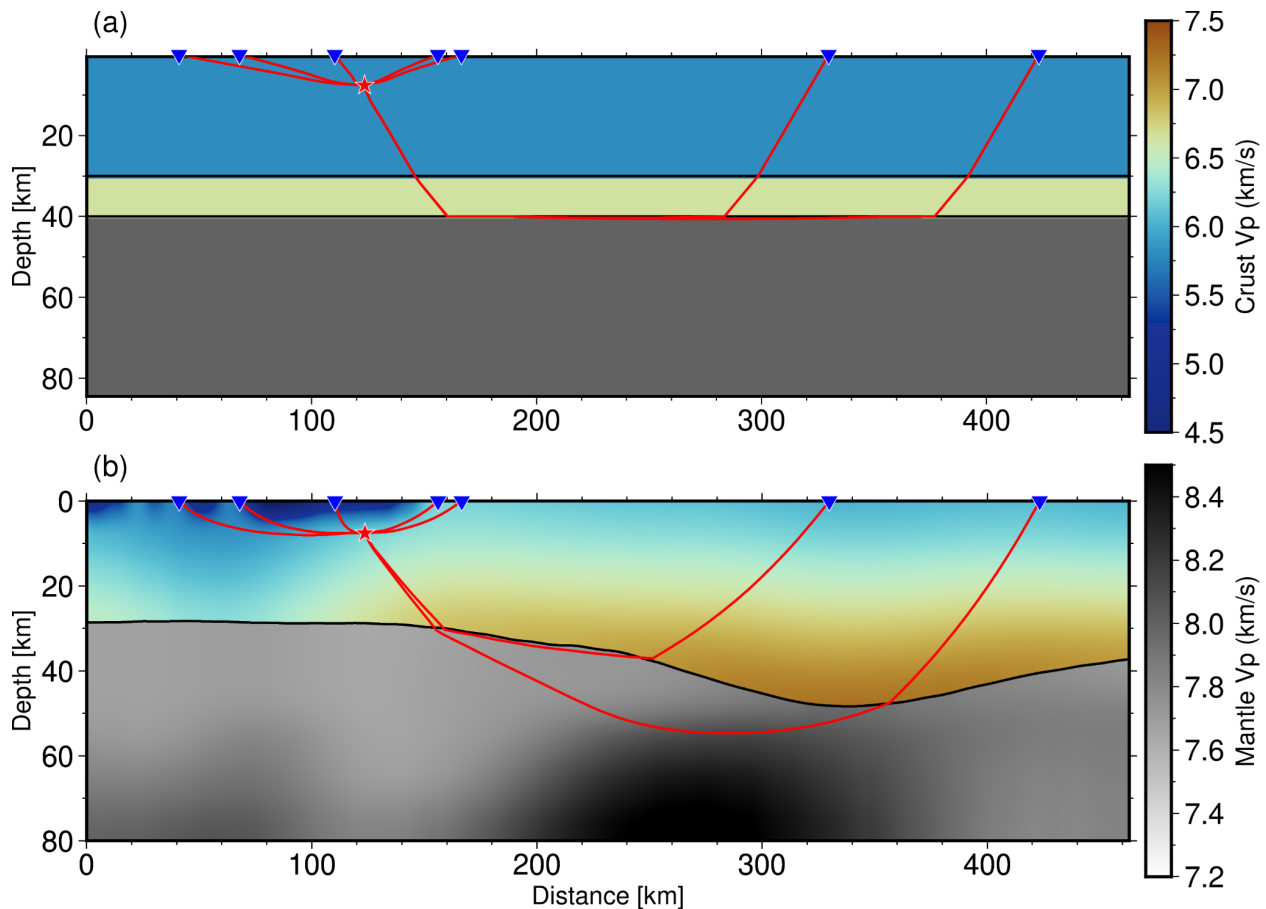



Figure 12. Earthquake ray paths in cross section BB' (for location see **Fig. 8**) for two models: (a) a simple 1D model with two isotropic crustal layers, and (b) our 3-D model with one anisotropic crustal layer. Colorbars are the same for both panels.

In addition to the profiles shown in **Figs. 11** and **12**, we have also extracted 1D depth velocity models for six points marked in **Fig. 8** in magenta. Those six points have been chosen to cover different domains of our model (Stable Adria, Internal and External Dinarides, and the SW Pannonian Basin). The velocity change with depth for the chosen six points is shown in **Fig. 13**. For example, at the P2 point, which is located in the SW Pannonian Basin, the velocity for the first couple of kilometers of depth is much lower than for the other points because there is a Neogene deposits layer on top. The outlook of each model shown in **Fig. 13** is generally similar at each point with obvious differences being Moho depth (see points P2 and P1) and rate of increase of velocity with depth (e.g., compare points P3 and P5).

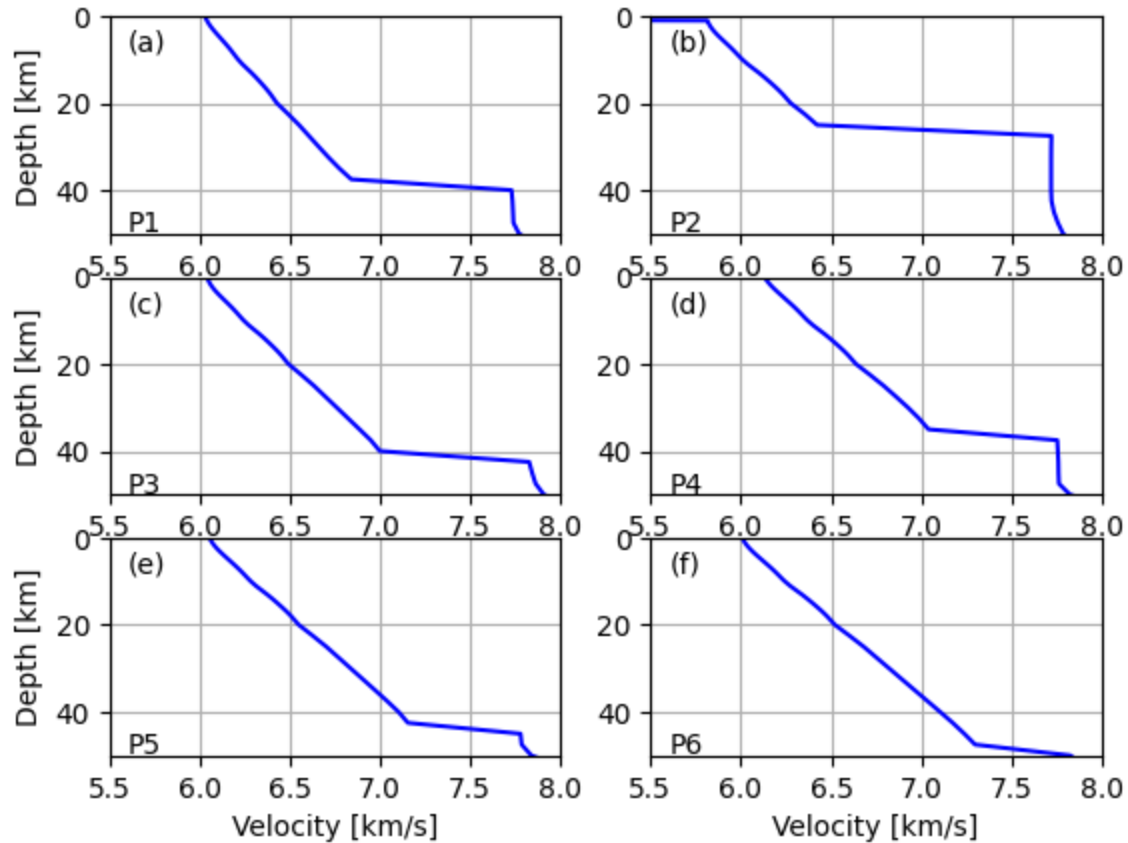



Figure 13. Velocity changes with depth for six points (P1 to P6; for locations see **Fig. 8**).

710

Conclusions

Having a complete 3-D model of the crustal structure is a major step forward in the study of the Dinarides and the surrounding areas. The newly derived model is defined on a regular dense grid for three key parameters (P- and S-velocity and density), and as such can be
715 readily used by seismologists who need information on crustal structure as input for their studies (earthquake locating, seismic tomography, shaking estimation, seismic hazard...). We tested the performance of the model in comparison with the commonly used 1D model and found significant improvements in time travel calculations. The model still has some inherent weaknesses that have been discussed in the previous sections which are mostly
720 connected with low number of measurements in some parts of the region. Nevertheless, the 3-D model represents a good first step towards improving the knowledge of the crustal structure in the complex area of the Dinarides.

725 The model clearly delimits several key areas (Dinarides, Pannonian Basin and Adriatic Sea) as well as distinguishing distinct layers in those regions (i.e., Neogene deposits and CRC). The most robust feature of the model is the depth to Mohorovičić discontinuity, but other parameters are also reasonably well defined. Inclusion of the CRC thickness is, to the best of our knowledge, the first attempt to estimate this parameter for the whole Dinarides region.

730 One of the new insights found during the creation of the model is the relatively high (P-wave) velocity for the lower crust in the southern part of the Dinarides. This feature needs to be confirmed by other studies, given the sparsity of information about velocity for that region. On the other hand, the high velocity feature fits nicely to the higher velocity of the uppermost mantle found in the same area thus corroborating the idea of continental subduction (and/or

735 lithospheric delamination) in south-central External Dinarides.

In conclusion, the model presented here represents the currently best and most complete crustal model for the wider Dinarides region. Model is assembled from all the available measurements on seismic velocity, density, layer composition and thickness to provide a full representation of the major variations of seismic wavespeeds in the regional crust.

740 Hopefully, the new 3-D model will help discover some new, previously unknown features of the crust and in turn, each new study that sheds some light on the crustal structure in this area may improve the 3-D model derived here.

Resources

The model derived in this work is available on the following link:
745 https://urn.nsk.hr/urn:nbn:hr:217:793485

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Appendix A

Table A1. Estimation of Carbonate Rock Complex (CRC) bottom depth in Dinarides. CRC bottom depths were estimated based on 93 sheets of Basic Geological Map (1:100,000) of former Yugoslavia (Osnovna Geološka Karta SFRJ) and 1:200,000 Geological Map of Albania (Geological Map of Albania, 2002). The CRC bottom depth values were estimated for each map centroid, in respect to External Dinarides structural relations, deformation styles that incorporate thrusting, folding (e.g., Balling et al., 2021), and thicknesses of deposits on individual maps. Regional nappe systems in External Dinarides incorporate up to six individual structural levels composed of thrust sheets of laterally and/or vertically variable thicknesses, enabling estimated combined absolute CRC depth up to 14,200 m.

No.	Basic Geological Map (BGM)	BGM centroid coordinates		Estimated average elevation	Estimated absolute bedrock depth (m)	Number of structural levels
		Latitude (φ)	Longitude (λ)	(m a.s.l.)		
1	Albania Nord	42.090	20.048	805	8200–14200	2–5
2	Banja Luka	44.834	17.247	300	1200	0–2
3	Bar	42.167	19.243	400	11100	2–4
4	Bihać	44.836	15.742	600	8400	2–3
5	Bijeljina	44.831	19.245	-	Saftić et al. (2003)	-
6	Biograd	43.830	15.244	-40	7900	1–2
7	Biševo	42.833	16.243	-70	7600	1–2
8	Bosanska Krupa	44.832	16.247	400	3700	0–2
9	Bosanski Novi	45.165	16.242	200	200	0

10	Brčko	44.831	18.739	-	Saftić et al. (2003)	-
11	Budva	42.167	18.747	-100	9100	2–4
12	Bugojno	44.160	17.233	900	6500	1–3
13	Cres	44.836	14.249	0	7200	1–2
14	Crikvenica	45.171	14.741	400	8300	2–3
15	Črnomelj	45.504	15.251	350	8200	1–2
16	Delnice	45.504	14.739	600	7900	1–2
17	Derventa	44.832	17.740	250	200	0–1
18	Doboj	44.837	18.246	-	Saftić et al. (2003)	-
19	Drniš	43.832	16.251	500	9000	2–3
20	Drvar	44.499	16.240	700	7800	2–3
21	Dubrovnik	42.497	18.244	200	9000	2–3
22	Foča	43.499	18.754	1000	300	0–1
23	Gacko	43.166	18.746	1300	10400	1–5
24	Glamoč	44.168	16.746	1000	8500	2–3
25	Gorica	45.830	13.747	350	7400	2–3
26	Gospić	44.496	15.247	500	10400	3–4

27	Ilirska Bistrica	45.502	14.240	400	7900	2–3
28	Imotski	43.502	17.246	800	13000	2–4
29	Jabuka	43.164	15.266	-150	7700	1–2
30	Jajce	44.492	17.246	600	4100	1–3
31	Jelsa	43.169	16.742	0	8900	1–2
32	Kalinovik	43.496	18.247	900	9000	1–4
33	Karlovac	45.502	15.750	200	1000	0–2
34	Ključ	44.501	16.743	700	7500	2–3
35	Knin	44.166	16.246	550	9500	2–3
36	Korčula	42.836	17.246	-50	6600	1–2
37	Kostajnica	45.167	16.751	250	200	0
38	Kotor	42.500	18.743	700	10300	2–5
39	Kranj	46.158	14.141	700	6800.00	2–3
40	Labin	45.169	14.245	300	6500	1–2
41	Lastovo	42.836	16.745	-80	8100	1–2
42	Livno	43.835	17.246	1100	11200	2–4
43	Lošinj	44.503	14.241	0	6200	1–2

44	Ljubovija	44.147	19.245	500	100	0–1
45	Metković	43.167	17.750	250	13100	3–4
46	Molat	44.163	14.745	-50	7200	1–2
47	Mostar	43.500	17.740	900	13500	4–5
48	Nevesinje	43.163	18.242	1000	14100	4–5
49	Nikšić	42.833	18.750	1000	12600	5–6
50	Nova Gradiška	45.167	17.247	-	Saftić et al. (2003)	-
51	Nova Kapela	45.165	17.743	-	Saftić et al. (2003)	-
52	Novo Mesto	45.837	15.240	250	7300	1–2
53	Obrovac	44.170	15.746	300	10000	2–3
54	Ogulin	45.171	15.250	400	9400	2–3
55	Omiš	43.502	16.751	450	10200	2–4
56	Otočac	44.829	15.248	500	10200	3–4
57	Ploče	43.169	17.246	250	10500	2–4
58	Pljevija	43.490	19.261	1000	200	0–1
59	Postojna	45.835	14.249	700	7700.00	2–3
60	Prača	43.814	18.749	1000	0	0–1

61	Prijedor	44.834	16.741	400	2400	0–2
62	Primošten	43.504	15.750	-100	7600	1–2
63	Prozor	43.824	17.743	1100	7500	0–4
64	Pula	44.831	13.743	0	4600	1
65	Rab	44.829	14.742	100	9200	2–3
66	Ribnica	45.837	14.751	600	7200	1–2
67	Rovinj	45.165	13.749	20	4500	1
68	Sarajevo	43.829	18.253	700	2600	1–3
69	Šavnik	42.833	19.252	1000	10600	1–5
70	Šibenik	43.837	15.741	50	8000	2–3
71	Silba	44.496	14.744	0	8000	2–3
72	Sinj	43.835	16.749	800	12000	3–4
73	Slavonski Brod	45.170	18.239	-	Saftić et al. (2003)	-
74	Slunj	45.169	15.746	350	4500	0–2
75	Split	43.500	16.244	100	7500	1–2
76	Ston	42.834	17.748	200	9000	1–3
77	Svetac	43.171	15.746	-150	7700	1–2

78	Teslić	44.491	17.736	700	700	0–3
79	Titograd	42.500	19.253	700	12200	4–5
80	Tolmin	46.153	13.610	700	6600.00	2–3
81	Trebinje	42.830	18.249	700	12000	2–4
82	Trst	45.497	13.741	200	6000	2–3
83	Tuzla	44.489	18.742	400	300	0–1
84	Udbina	44.503	15.750	800	12000	3–4
85	Ulcinj	41.825	19.244	-200	9200	2–3
86	Vareš	44.144	18.245	800	1600	0–1
87	Vis	43.166	16.250	–70	7600	1–2
88	Višegrad	43.813	19.271	700	700	0–1
89	Vlasenica	44.147	18.745	600	700	0–1
90	Žabljak	43.166	19.250	1200	2400	1–5
91	Zadar	44.163	15.245	10	8000	2–3
92	Zavidovići	44.495	18.252	350	0	0
93	Zenica	44.149	17.733	800	1200	0–3
94	Zvornik	44.489	19.245	250	800	0–1

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Appendix B

Point locations of data used

The following figure shows the exact location of data points used for interpolation of model interfaces and parameters. Data points used for each interface and parameter are shown in

955   a separate figure panel.

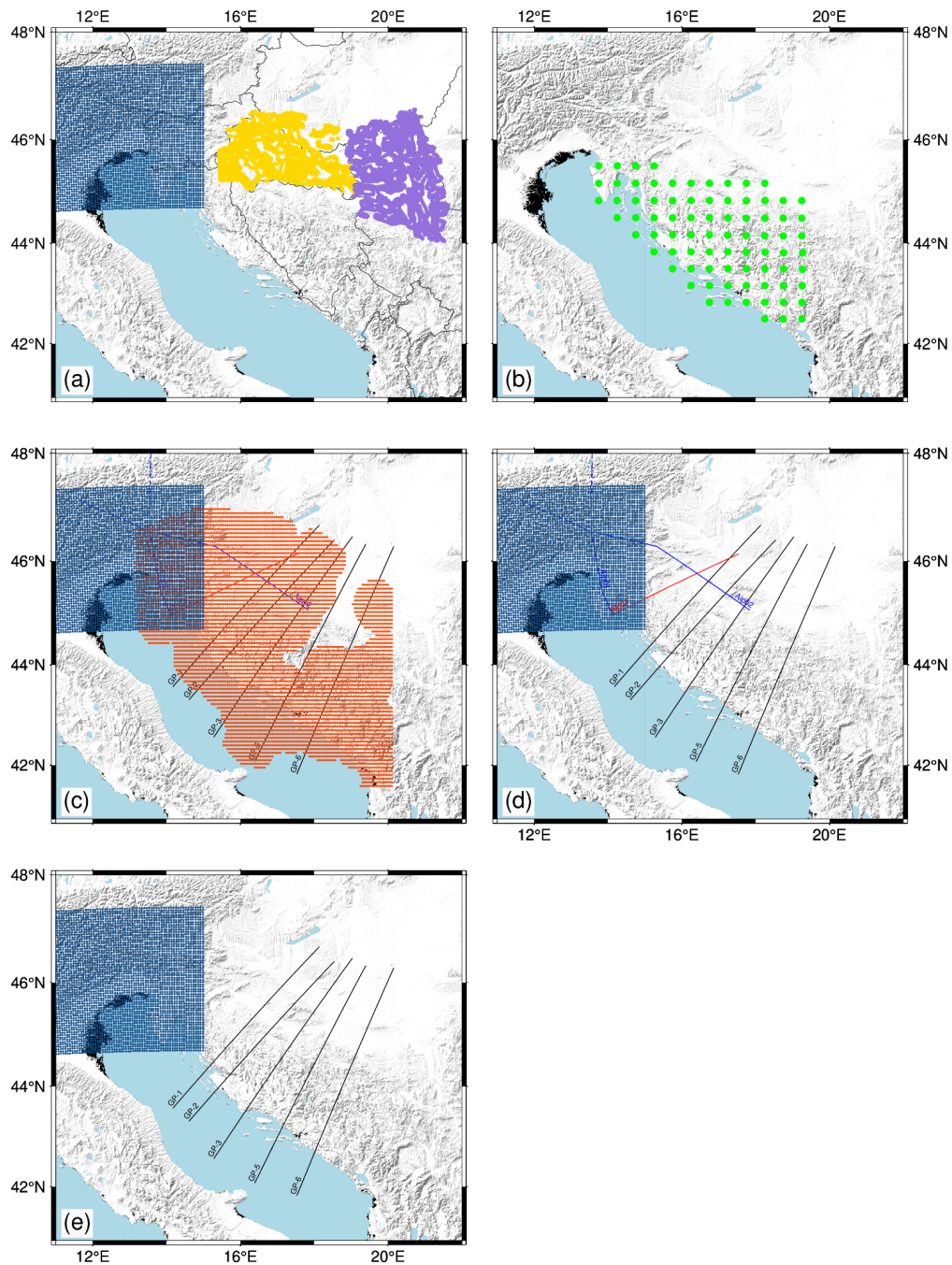



Fig. B1 Point locations of data used for interpolation of (a) sediment bottom depth, (b) CRC depth, (c) Moho depth, (d) P-wave velocity, (e) density. Yellow points mark data from Saftić et al. (2003), purple points mark data from Matenco & Radivojević (2012), green points mark data on CRC depths, blue points mark data from the NAC model (Magrin & Rossi, 2020), orange points mark data from Stipčević et al. (2011), blue lines mark the positions of profiles from Brückl et al. (2007), red lines mark the profile from Šumanovac et al. (2009), and black lines mark the gravimetric profiles from Šumanovac (2010).

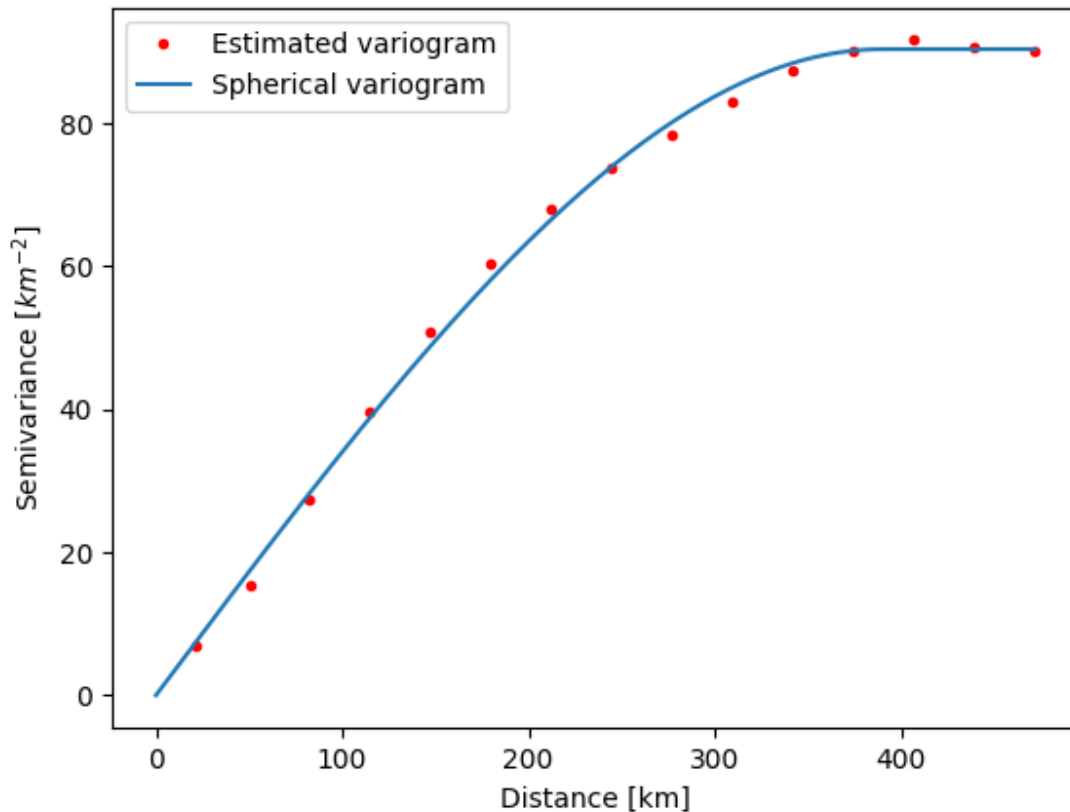
Appendix C

Kriging

Kriging is a method of interpolation formulated for the estimation of a continuous spatial attribute (e.g. depth to Moho interface) at an unknown site, using the limited set of data from sampled sites. It is a form of generalized linear regression used for the formulation of an optimal estimator in a minimum mean square error sense (Olea, 1999).

Generally, the value at a point of interest is calculated as follows:

$$\hat{Z}(x_0) = \sum_{i=1}^{n} \lambda_i Z(x_i),$$

where $\hat{Z}(x_0)$ is value estimation at point of interest x_o, λ_i are weights, and $Z(x_i)$ are known values at sites x_i. The kriging weights are derived from the covariance of the sampled values. The first step in kriging interpolation is estimation of the variogram (also called a semivariogram) – a statistic that assesses the average decrease in similarity between two random variables as the distance between them increases. It is the inverse of covariance – the covariance measures similarity, and the variogram measures dissimilarity. Unlike the other moments (e.g., the mean), the variogram is not a single number, but a continuous function of a variable h, called the lag. The variogram calculated from the sampled points is called the experimental variogram. The experimental variogram is not used in the calculation of kriging weights but is used to fit a theoretical variogram which in turn is used for calculation of the weights. When fitting, we can use limited types of semivariograms which have acceptable properties needed for solving the system of equations in order to obtain the weights. If we would use the experimental variogram directly, we might end up with an unsolvable system of equations. For example, **Fig. C1** shows an experimental and theoretical variogram used for interpolation of Moho interface depth. A variogram, such as the one in **Fig. C1**, that increases in dissimilarity with distance over short lags and then levels off is called a transitive variogram. The lag at which it reaches a constant value is called the range, and that constant value is called the sill. For the Moho depth interpolation, we had an abundance of data available, and we were able to estimate a theoretical variogram which fits the observed data almost perfectly.

995 **Fig. C1**. Estimated and theoretical (spherical) variogram used for interpolation of Moho interface
depth.

In case of other model parameters, we did not have such a perfect fit for larger lags. For
instance, **Fig. C2** shows the variogram used for Neogene deposits bottom interpolation. In
1000 this case, the theoretical variogram was not spherical, but exponential. In this case, for the
largest lags shown, the theoretical and estimated variograms do not fit. For the calculation
of the variogram pairs of measured values are used. Since for the greater distances (greater
lags) there are fewer numbers of such pairs, the estimates are less accurate for those lags.
Fortunately, for practical use in kriging, the variogram closer to the origin requires the most
1005 accurate estimation (Olea, 1999), and we had that condition met for all our model
parameters.

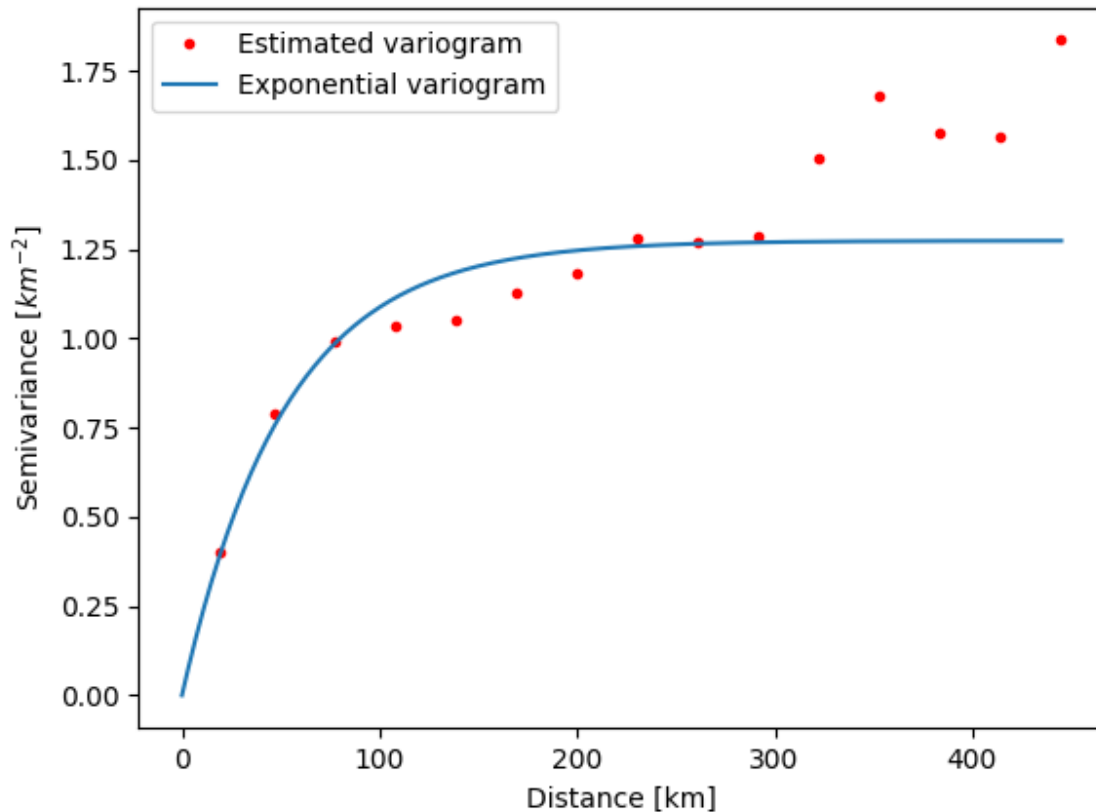



Fig. C2. Neogene deposits bottom estimated and theoretical (exponential) variograms.

1010

Variograms for CRC bottom depth and velocity estimation are shown in **Figs. C3** and C**4,** respectively. Crust velocity variogram shown in **Fig. C4**. is required to have a constant mean in the sample space in order to be able to estimate a variogram. In case of a gentle and
1015   systematic variation in the mean (called the drift), e.g. velocity increases with depth, it has to be removed prior to the estimation of the variogram. Such a drift was indeed observed and was removed prior to the estimation of the variogram shown in **Fig. C4**. Besides the drift, we estimated the experimental variogram in several directions, in order to check if it was dependent on the direction (i.e. if there was anisotropy), but we have not observed any
1020   anisotropy.

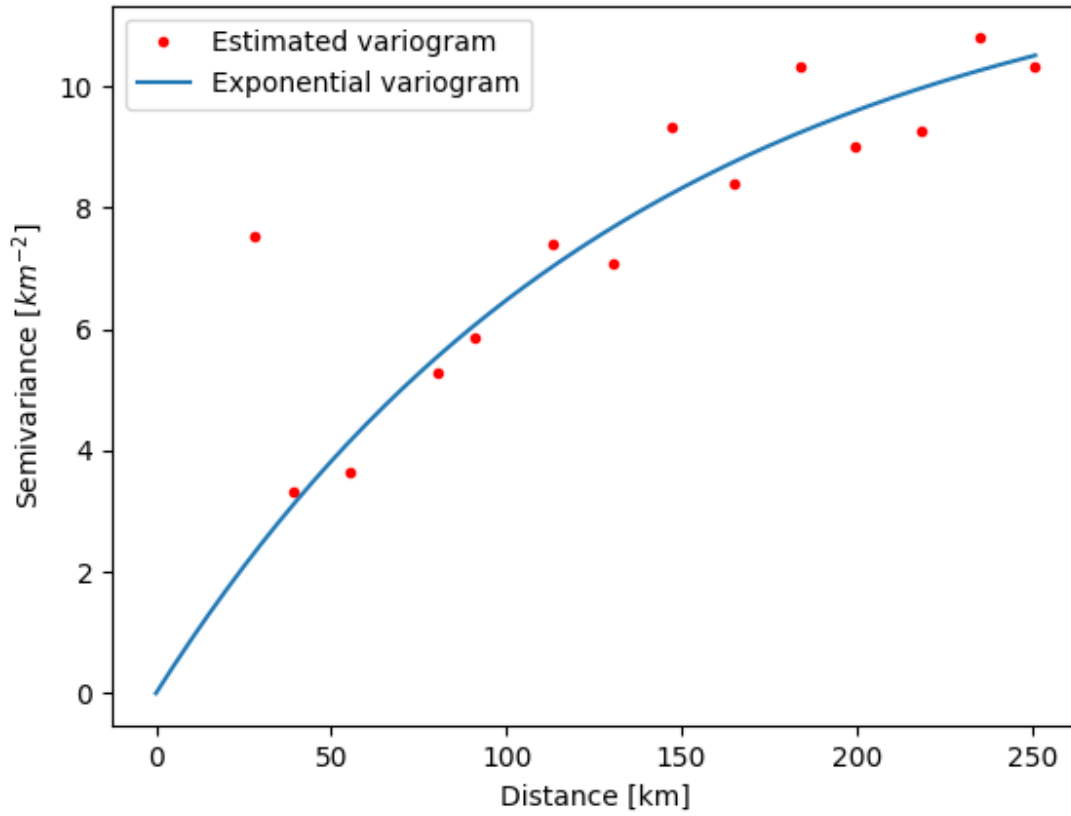



Fig C3. Carbonate Rock Complex (CRC) bottom estimated and theoretical (exponential) variogram.

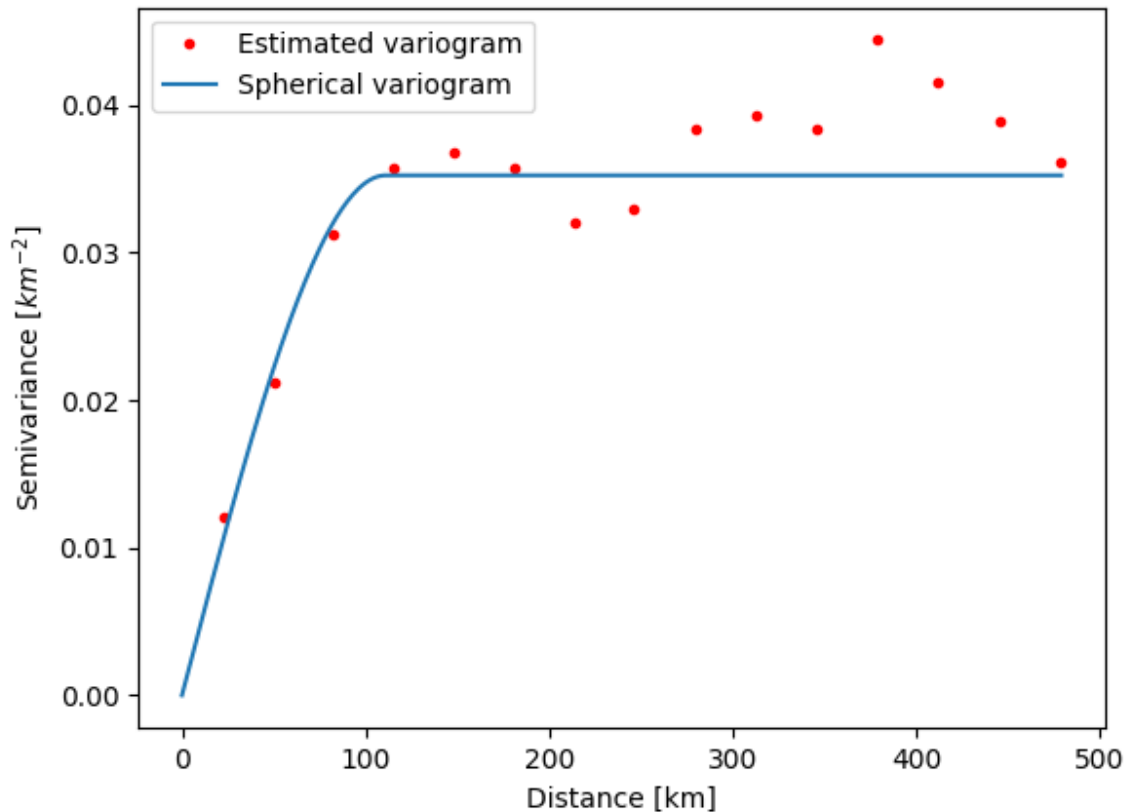



Fig. C4. Estimated and theoretical (spherical) variogram used for interpolation of the crustal velocity (Vp).

Once we had variograms estimated, we were able to obtain the weights and calculate the values of parameters (Moho, Neogene deposits bottom, CRC bottom, velocity) for each point in our grid. All the operations, both variogram estimation and the interpolation itself, were done using the gstat package (Pebesma, 2004). Alongside the interpolated values, the package also returns the variance estimates for each point in the grid.

The interface parameters (Moho, Neogene deposits bottom and CRC bottom depth) were interpolated using ordinary kriging. Ordinary kriging assumes that the mean of the value is unknown, but constant. For interpolation of the velocity, we had to use a more general type of kriging – the universal kriging. It relaxes the condition on the mean – it is no longer assumed constant. The other properties of kriging are shared between both types used. They are both minimum square error estimates. The estimation is not limited to the data interval (it is possible to extrapolate – although it is less accurate). They have, so called, declustering ability – the measurements that are spatially clustered have lower weights than isolated points. They are exact interpolators with zero kriging variance – meaning that if, for instance, we try to calculate the value exactly at a sampled point, kriging will return the exact value

and assign 0 variance to it. It can be nicely seen in **Fig. 5** showing the velocity variances. Since the variance at sampled sites is zero it is possible to discern the profiles that were sampled for data. Kriging is not able to handle duplicate points – it causes the insolvable systems of equations for the kriging weights – therefore we had to handle such points. It is also worth mentioning that the variance returned by the kriging software does not depend on variance or values of individual observations, but only on the sampling pattern. Therefore, we added the (estimated or available) data variance to the variance obtained from the kriging and called it the total variance.

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Appendix D

Comparison of the new model to existing regional models
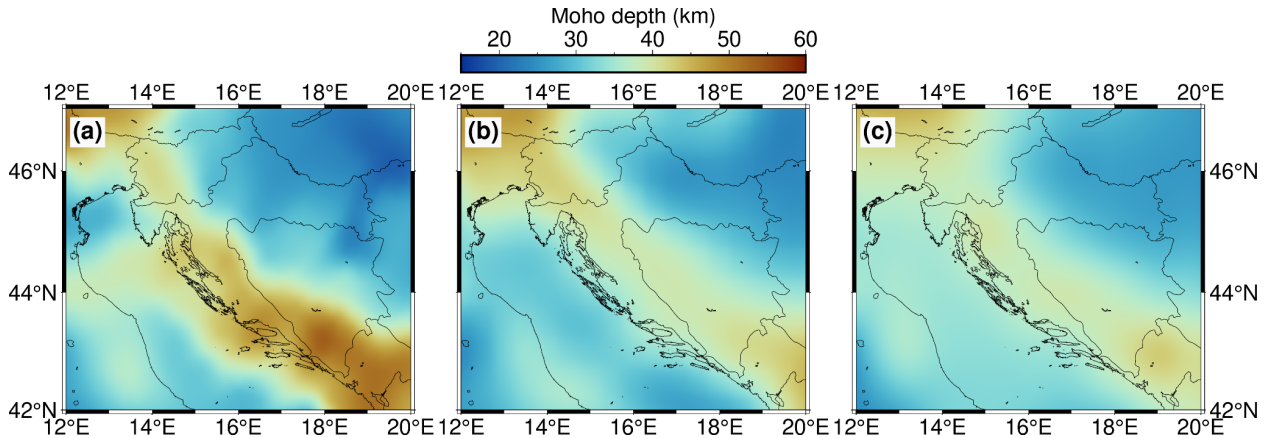
Fig. D1 Comparison of Moho depths among (a) our newly derived model, (b) Grad et al. (2009) european Moho model, and (c) EPcrust model (Molinari and Morelli, 2011).

Compared to existing regional models, Moho depth is considerably greater in the new model, especially in the southern part of the Dinarides.

The EPcrust model has sediment bottom depth defined. In our model, we discern CRC bottom and Neogene sediment bottom, which is not defined separately in the (regional) EPcrust model. Therefore, we put both in Fig. D2.
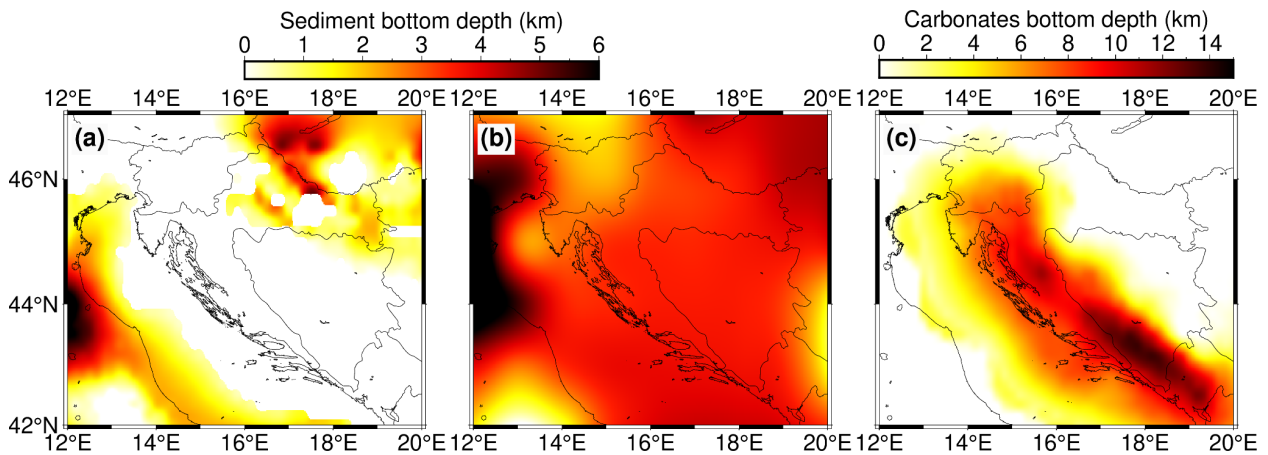
Fig. D2 Comparison of sediment bottom depth as defined as (a) Neogene sediment bottom depth in our model, (b) sediment bottom in EPcrust model, and (c) CRC bottom depth in our model.

The horizontal variations in sediment and CRC bottom depths are much more pronounced than in the regional EPcrust model.

In the EPcrust model, the velocity is defined as two layers (upper and lower crust) with constant velocity value (for a given grid point, it does not vary with depth; it varies horizontally, though). Therefore, we have picked two depths in our model, since we defined velocity as varying in all three dimensions. Even though we do not define crust as two layers (lower or upper crust), we have picked velocity at depth of 15 km to compare with EPcrust upper crust, and velocity at depth of 25 km to compare with EPcrust lower crust. The comparison is shown in Fig. D3.
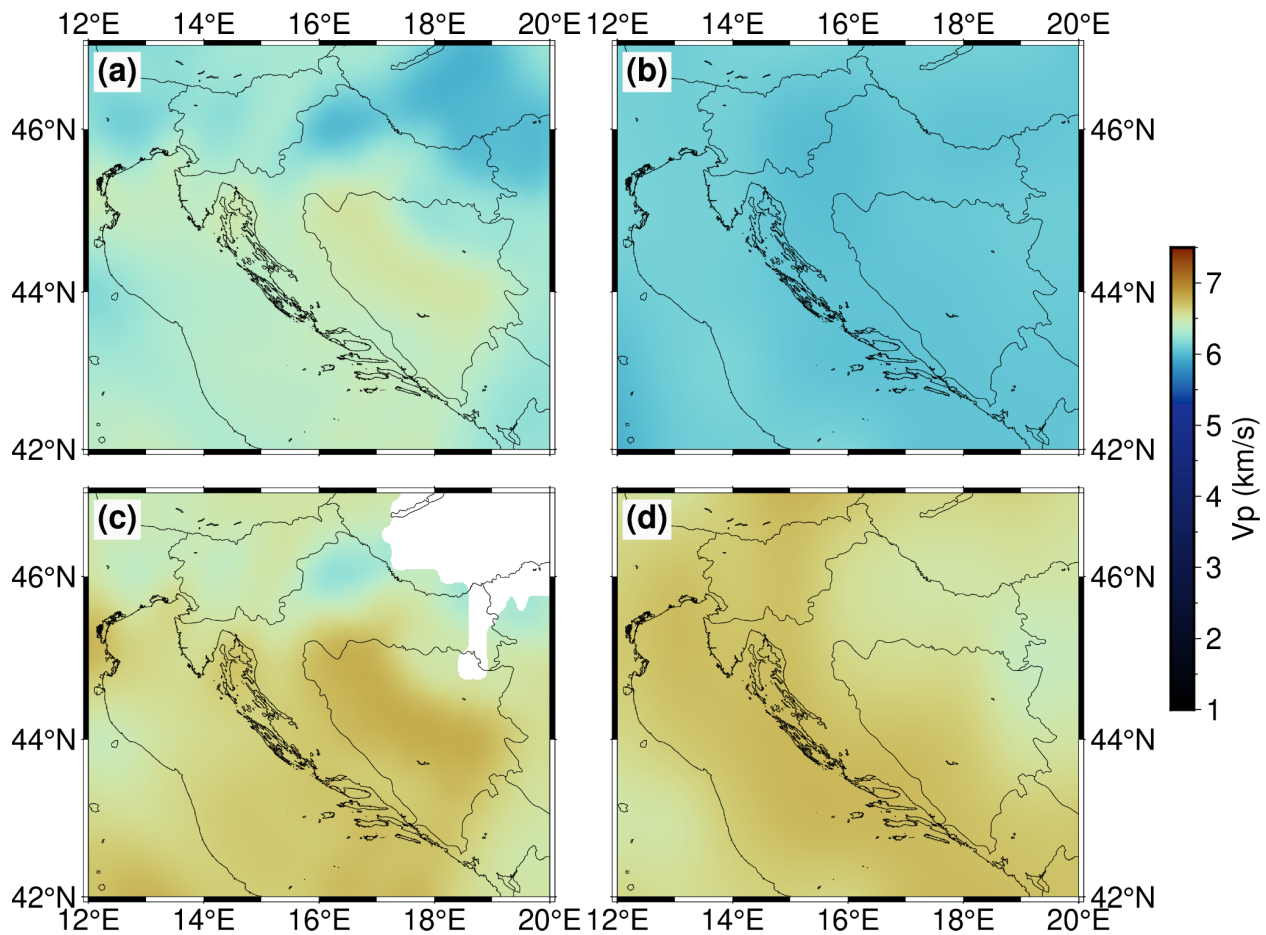
Fig. D3 Comparison of P-wave velocity: (a) our model at 15 km depth, (b) EPcrust upper crust, (c) our model at 25 km depth, (d) EPcrust lower crust.

There is noticeable horizontal variation of velocity values in our model compared to the velocities defined in the EPcrust model, due to inclusion of data from refraction and gravimetric profiles.