

The missing craton edge: Crustal structure of the East European Craton beneath the Carpathian Orogen revealed by double-difference tomography

Borleanu F.^(1,2), Petrescu L.⁽²⁾, Enescu B.^(1,3), Popa M.^(2,3), Radulian M.^(2,3)

1) Department of Geophysics, Kyoto University, Japan; 2) National Institute for Earth Physics, Romania; 3) Academy of Romanian Scientists, Romania

Highlights

The highest resolution 3-D seismic model of the crust was determined for the Eastern Carpathians, Romania.

The craton edge is traced at mid-crustal depths as a SW-dipping seismic velocity gradient beneath the Carpathian nappes.

Relocated seismicity in the SE Carpathians indicate that pre-existing active foreland faults extend towards the volcanic belt.

The Neogene volcanic belt shows slow Vp and Vs, and depth- varying Vp/Vs ratios indicating the existence of partial melts.

Abstract

The Trans-European Suture Zone (TESZ) is the most important and extensive continental suture in Europe, marking the edge of the East European Craton (EEC), from the North Sea to the Black Sea. It corresponds to significant changes in surface geology and deep crustal structure, evident in seismic, gravitational and magnetic studies. However, the TESZ disappears beneath the Eastern Carpathians accretionary nappes and Neotethys ophiolites, thrust over the subducted EEC passive margin in Romania that may have experienced progressive southward break-off generating post-collisional volcanism and anomalous seismicity in the Vrancea region of the South-East Carpathians. To illuminate the missing TESZ section and investigate the change in crustal properties from the Precambrian EEC across the collisional orogen and the impact of related volcanism, we determined a 3D seismic model of P and S wave velocities of the Eastern Carpathians in Romania. With the advent of new permanent broadband and short-period seismic stations of the Romanian Seismic Network, we were able to lower the earthquake magnitude detection threshold to 0.5 M_L , largely expanding the earthquake database, and create seismic images of the crust across the Carpathians. Using double-difference tomography and waveform cross-correlation differential times, we relocated local crustal earthquakes between 2010 and 2017 and jointly inverted for the 3D P and S -wave velocity structure down to the Moho discontinuity. Our study provides the highest resolution 3D crustal seismic model of this area to date and emphasizes the manifestation of surface tectonic boundaries at lower crustal depths. The TESZ is highlighted at mid-to lower-crustal depths beneath the Carpathian nappes, east of the post-collisional volcanic belt, as a gently-dipping transition from high Vp in the eastern footwall, the dipping Precambrian basement, to low Vp beneath the Carpathian Orogen to the west and a downward decrease in Vp/Vs. In parts of the volcanic region, Vp/Vs ratios increase suggesting the presence of fluids, mafic material and partial melts that may have altered the seismic structure of the TESZ. Relocated hypocenters are vertically distributed in the velocity transition zone from low to high especially beneath the youngest volcanoes in the south, consistent with a hypothetically active magmatic plumbing system. In the northern segment, where volcanoes are older, relatively high velocities are estimated and seismicity tends to cluster in the top 10 km, likely connected to a recently reactivated fault system. Seismicity in the foreland of the

East Carpathians aligns along the NW-SE trending major crustal faults and continues beneath the thrust belt, suggesting the East European Craton basement extends beneath the orogen as far as the volcanic belt.

Keywords: 3D velocity model, Double-difference tomography, Eastern European Craton, Trans-European Suture Zone, Crustal seismicity

1. Introduction

Cratons are remarkable tectonic archives that preserve evidence for Precambrian crustal formation and reworking at their margins. They are thick (~200 km) and buoyant continental cores that have survived tectonic reworking throughout geological history (*O' Reilly et al., 2001*). Studies of their crustal structure can provide fundamental insights into their internal structure and their interaction with younger continental crust. Imaging their margin also provides clues on the processes of lithospheric growth or destruction (*Snyder, 2002*) and their interaction with magmatism may constrain the way they are thermochemically eroded (e.g. *Eaton and Frederiksen 2007, Boyce et al., 2016*). The Eastern Carpathians (ECA), part of the larger Alpine Carpathian–Pannonian orogenic system, evolved within Eurasian Plate at the contact with the East European Craton (EEC) in the past 20 Ma (*Figure 1*). They are crossed by several major faults and along them descends the Trans-European Suture Zone (TESZ), the most important and extensive continental boundary in Europe (*Pharaoh et al., 2006*), marking the edge of the EEC (*Figure 1*). The TESZ corresponds to significant changes in surface geology and deep crustal structure (*Starostenko et al., 2013*). However, the TESZ disappears beneath the ECA accretionary nappes and Neotethys ophiolites, thrust over the subducted EEC passive margin in Romania. A probable north-to-south slab break-off may have caused age-progressive post-collisional volcanism along the orogen back-arc and is assumed to be an ongoing process beneath the South-Eastern Carpathians, where an isolated cluster of intermediate-depth seismicity persists (*Sperner et al., 2001, Radulian et al., 2002, Knapp et al., 2005, Koulakov et al., 2010, Ismail-Zadeh et al., 2012, Figure 1*). While most previous active and passive geophysical studies exclusively focused on imaging the South-East Carpathian corner, where most seismic activity is concentrated, (e.g. *Martin et al., 2003, Landes et al., 2004, Panea et al., 2005, Knapp et al., 2005, Hauser et al., 2007, Baron and Morelli 2017*), or imaging the TESZ where it is visible in the surface geology (e.g. *Pharaoh et al., 2006, Grad and Polkowski, 2015*), little work has been carried out to investigate the crustal and upper mantle structures of the East Carpathians and the transition from Phanerozoic to Precambrian Europe beneath the Carpathian Orogen (e.g. *Diehl et al., 2005, Petrescu et al., 2019*). The purpose of this research is twofold: (1) to estimate a local 3D seismic model of P and S wave velocities providing the highest resolution seismic images of the crust across the Eastern Carpathians to date, to illuminate the missing TESZ section highlighting the changes in crustal seismic properties from the East European Craton across the collisional orogen, and (2) to assess the impact of post-collisional volcanism on the craton margin. To reach our goal we carry out local double-difference (DD) tomography (*Zhang and Thurber, 2003*) using crustal events that occurred in the ECA (see *Figure 1*). This method is able to produce more accurate event locations and velocity structure near the source region, using both absolute and differential arrival times. In the following section, we briefly describe the main geological, geophysical and seismotectonic characteristics of the ECA as well as the main findings of the previous studies carried out in this area.

2. Regional seismotectonics and previous studies

The TESZ is the most prominent geological boundary in Europe delimiting the mobile Phanerozoic structures in the south and west from the stable Precambrian EEC, crossing the continent from the North Sea to the Black Sea (*Pharaoh et al., 1997, Figure 1*). While its surface geological and deep lithospheric signatures are clearer in the northern segment, the TESZ disappears under the Carpathian Orogen in Poland, Ukraine and Romania. The TESZ

represents a complex area of continental accretion at the edge of the EEC and shows variable seismic signatures and widths along its trajectory, earning broad definitions throughout the literature (McCann and Krawczyk, 2001, and references therein), such as a boundary (Pharaoh et al., 2006), a crustal domain comprising Neoproterozoic units overlying thinned EEC margin (e.g. Starostenko et al., 2013), or simply a broad zone of deformation that separates Precambrian terranes from younger accreted units (Grad et al., 2003a, and references therein). In Romania, the Precambrian passive margin of the EEC underthrusts the Eastern Carpathians, and the foreland crystalline basement is covered by 5-10 km thick Paleozoic-Mesozoic sediments and Neogene foredeep sediment strata as thick as 13 km near the Carpathian bend zone (Tărăpoancă et al., 2003). A Quaternary compressional pulse affected both the foredeep sediments (Mațenco et al., 2003) and the back-arc basin (Ciuvalu et al., 2000).

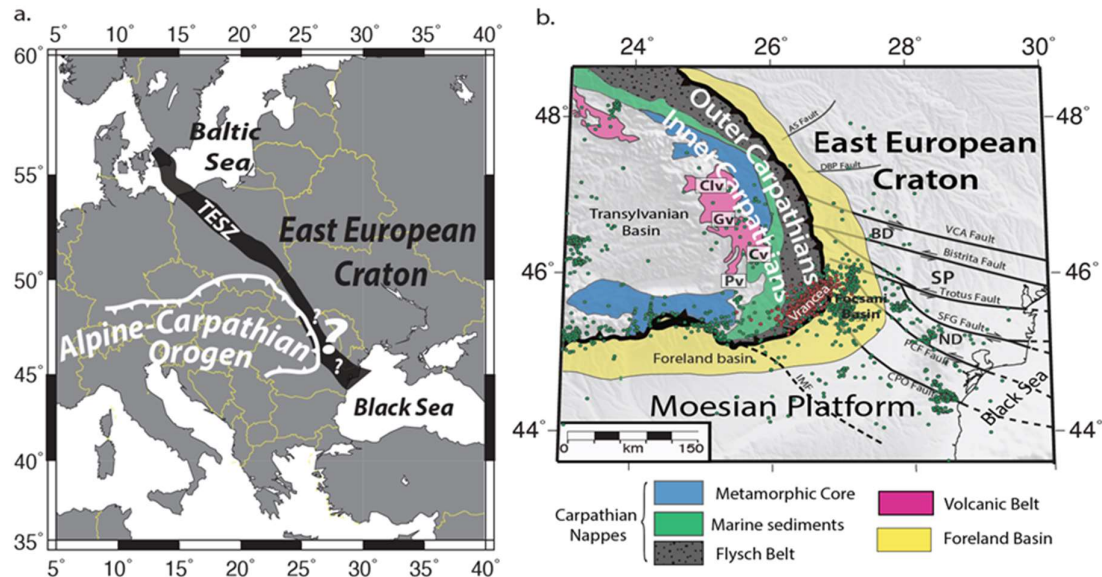


Figure 1. **a.** Simplified sketch after major tectonic units in Europe (TESZ-Trans European Suture Zone) and in the Eastern Carpathians region. **b.** Tectonic map of Romania after Sandulescu et al. (1981), Sandulescu et al. (1984) and simplified based on geological maps published by the Geological Institute of Romania (<http://geoportal.igr.ro>). Inner and outer branches of Carpathians are located towards the Transylvanian Basin, and towards the EEC and the Moesian Platform, respectively. The seismicity in the Eastern Carpathians region (between 2015 and 2019) according to the ROMPLUS catalogue (Oncescu et al., 1999, which is constantly updated) is plotted as circles coloured with respect to depth: red and green (for subcrustal and crustal events). Fault abbreviations are as follows: CPO: Capidava-Ovidiu Fault, IMF: Intramoesian Fault, SFG: Sfantul Gheorghe Fault, PCF: Peceneaga-Camena Fault, VCA: Vaslui-Cetatea Alba Fault, AS: Avramesti-Suceava Fault, DF: Draganesti-Belcesti-Probota Fault. Tectonic domains are ND: North Dobrogea, BD: Barlad Depression and SP: Scythian Platform. Volcanic centers are abbreviated as Clv: Calimani, Pv: Persani, Cv: Ciomadul, and Gv: Gurghiu.

The structure and location of the southwestern section of the TESZ is highly debated (Atanasiu et al., 2005), with widths reported between ~40 km (Bocin et al., 2013) and ~140 km (Starostenko et al., 2020) and trajectories considered to bend beneath North Dobrogea and the Black Sea based on electromagnetic methods (Figure 1, Stănică et al., 1999, Munteanu and Tatu 2003), or steer towards the south-west based on residual gravity anomaly interpretations (Ioane et al., 2019). The Carpathian arc represents the eastern continuation of

the Alpine orogeny, marking the collision between Phanerozoic European microplates and the passive margin of the EEC in the Miocene and the closure of the Neotethys (Balla, 1987, Ustaszewski et al., 2008). The ECA are a classic fold-and-thrust belt (Maţenco and Bertotti, 2000), comprising the allochthonous metamorphic core of Dacia (the colliding microplate), Neotethys ophiolites (Neugebauer et al., 2001) and accretionary wedge nappes thrust over the passive margin of the EEC (Figure 1). After subduction ceased ~9 Ma ago (Maţenco et al., 2010) the slab is thought to have gradually detached from north to south (Wortel and Spakman, 2000, Sperner et al., 2001) giving rise to post-collisional volcanism (Figure 1, Seghedi et al., 2011). Volcanoes in the ECA range in age from 11 Ma (the Calimani–Gurghiu–Harghita chain -CGH, Szakács and Seghedi, 1995, Mason et al. 1998, Seghedi et al., 2011) in the NW to 10 ka in the SE (Persani mountains, Seghedi and Szakács, 1994, Panaiotu et al., 2004, Harangi et al., 2006), although their genetic mechanism and connection with slab detachment remain unclear (e.g. Ismail-Zadeh, 2012). An anomalous cluster of intermediate-depth seismicity persists beneath the SE Carpathians in the Vrancea zone (Radulian et al., 2002) and is thought to mark the final stages of a slab break-off process (Wenzel et al., 1998). Although the seismicity in the region is dominated by the subcrustal events beneath the Carpathian bend zone, the crustal seismicity also shows interesting patterns that outline the fault configuration in the region. To the NE of the Vrancea region, in the Scythian Platform (Barlad Depression), the seismic activity consists of small to moderate-size events, up to 5.6 Mw (ROMPLUS, Oncescu et al., 1999) with a dominantly normal fault mechanism indicating an extensional stress regime (Radulian et al., 2000). Further to N and NE the seismic activity is increasingly sparse (Figure 1) consisting of only isolated events. In the inner Carpathian Orogen, seismicity is scarce, comprising mostly low magnitude events ($M_w < 4$) which partially correlate with the location of the youngest volcanic mountains (Figure 1). The crustal structure has been previously investigated by local passive and active seismic surveys as well as gravity and magnetotelluric experiments. Seismic refraction surveys of the Carpathians and the Vrancea area revealed a 40 km thick Precambrian foreland with a 20 km discontinuity and a crustal-scale ramp with an eastern uplifted section beneath the Focsani Basin (Radulescu et al., 1976). Enescu et al. (1992) identified Conrad and Moho discontinuities at 24 and 52 km for the Carpathian Orogen, and at 14 km and 34 km depth beneath the Transylvanian Basin, respectively. Landes et al. (2004) applied a finite-difference travel time algorithm to investigate crustal velocity structure and to detect the major crustal faults in the Vrancea region and its vicinity. They indicated distinct features in the ECA, characterized by higher velocities towards the north (5.6–5.8 km/s) than to the south (5.0–5.4 km/s) and lateral velocity variations associated with the Moesian and Scythian platforms. Diehl et al. (2005) and Petrescu et al. (2019) used receiver functions to highlight a significant Moho thinning across the ECA, placing the Moho at 27 km, in an area between Ciomadul and Gurghiu volcanoes, in contrast to refraction models. Hauser et al. (2007) analyzed the velocity structure along a 500 km long seismic profile, highlighting important similarities between velocity structures and seismic profiles across the TESZ in Poland (Dadlez et al., 2005) and Romania. The correlation resides mostly in a thick three-layered crust for the Precambrian Craton (42–45 km in Poland, 44 km in North Dobrogea), decreasing in thickness to the SW (29–32 km in the Palaeozoic terranes of Poland, 37–33 km in the Transylvanian Basin). The TESZ is covered by deep sedimentary basins with Permian origin or precursors (20 km thick Polish Basin, 22 km thick Focsani Basin area), suggesting an SW prolongation of the TESZ into Romania. By jointly inverting P and S-waves travel times from local earthquakes with local Bouguer anomaly data, Tondi et al. (2009) detected a low-velocity region in the upper crust located between the major Peceneaga-Camena (PCF) and Intramoesian (IMF) faults (Figure 1). Local and teleseismic earthquake tomography studies showed high-velocity structures beneath the SE Carpathians in the Vrancea area (e.g. Koulakov et al., 2010, Ren et al., 2012, Baron and Morelli, 2017), which correlate with intermediate-depth seismicity. Local earthquake tomography and ambient noise studies revealed low velocities to the NW of the Vrancea region beneath the volcanic chain (Popa et al., 2011, Ren et al., 2013), possibly associated with a mid-crustal magma chamber, and strong velocity contrasts suggesting a deep crustal fault (Popa et al., 2011). Ren et al., (2013), using ambient noise analysis and a non-linear 2-D tomographic inversion

technique estimate the group and shear wave velocities in the Carpathian Pannonian region. They showed low velocities down to ~6 km respectively 10–15 km depth in the Neogene volcanic region (northwest relative to Vrancea) and in the foredeep zone of the Focsani Basin. At the same time, they pointed out high-velocity anomalies in the upper crust of the Southern Carpathians, in contrast with the shallow low velocities of the Moesian Platform (corresponding to the Dacic foredeep basin). Using full-waveform local earthquake tomography, *Baron and Morelli, (2017)* emphasized a shallow low P and S velocity body under the Focsani Basin similar to the ambient noise study of *Ren et al., (2013)* and low V_p/V_s values in the upper-crust (10–30 km) of the inner Carpathians, to the NW beneath the Transylvanian Basin, similar to receiver function based estimates (*Petrescu et al., 2019*).

3. Methods and data

To investigate the velocity structure in ECA we applied a double-difference tomography technique that jointly determines event locations and the local 3D seismic velocity structure (*Zhang and Thurber, 2003*). The algorithm is based on hypoDD, a double-difference event relocation technique (*Waldhauser and Ellsworth, 2000*) and is able to produce more accurate locations and simultaneously recover the seismic velocity structure near the source region by using both absolute and differential travel times from local events. Data consist of P and S-waves arrivals gathered from 653 low- to moderate-size crustal events ($45.85 < \text{Lat. (}^\circ\text{N)} < 47.50$; $24.50 < \text{Lon. (}^\circ\text{E)} < 28.00$; $0.3 < M_L < 5.7$; $0 < h \text{ (km)} < 45$) recorded between 2010 and 2017 mostly by 100 Hz sampling rate short period and broadband seismic stations (*Figure 2*) of the Romanian Seismic Network (RSN) operated by the National Institute for Earth Physics (Măgurele, Romania) - NIEP (*Figure 2*). Each of the selected events has at least 8 recorded phases (P and S-wave) manually picked on vertical (Z) respectively horizontal components (N-S or E-W). The location is carried out within the Romanian Data Centre (RONDC) of NIEP, using IASP91, the Earth reference velocity model (*Kennett and Engdahl, 1991*) and the Locsat algorithm (*Bratt and Nagy, 1991*) embedded within the Antelope environment. Since the initial locations were obtained using a global velocity model, we firstly applied VELEST algorithm (*Kissling, 1995, 1988*) using initial locations to determine an optimum 1D local velocity model (*Figure 2*) and relocate those events based on the new velocity model.

The input data set for the DD tomography consists of 8215 high-quality absolute P and S-waves arrivals times, 36694 catalogue differential times as well 642 cross-correlation differential times computed for both P and S-waves. To estimate the catalogue absolute and differential travel times we applied the nearest-neighbour technique (*Waldhauser and Ellsworth, 2000*) designed to refine the link between the events and the number of recording stations.

The waveform cross-correlation (WCC) differential times were determined using the approach proposed by *Von Seggern, (2009)*, which showed that the differential times between the arrivals recorded at the same station may be determined with a precision of few milliseconds for the stations of 100 Hz sampling rate using an interpolation method. WCC times were determined only for the event pairs within 10 km distance separation and a time window of 1.0 s starting 0.2 s before phase arrival while the considered time lag was chosen between -0.3 s and 0.3 s. We selected only the phase pairs with a $WCC \geq 0.8$ to be further used within this study. In this way, the travel time errors of P and S-waves are neglected from the velocity model as well as from earthquake relocation. Therefore, using both data types the accuracy of earthquake location will be improved as well as the 3-D image of velocity structure. The weightings of absolute and differential data are harnessed by a hierarchical scheme, allowing for the absolute data to manage the broader scale velocity structure while the differential data performs the finer resolution near the sources. A 3-D velocity grid (*see Figure 2 a*) was used as an input velocity model. We established the best grid configuration after performing several resolution tests for different distance intervals and found the best resolution for a horizontal distance of 30 km (*Figure 2a*).

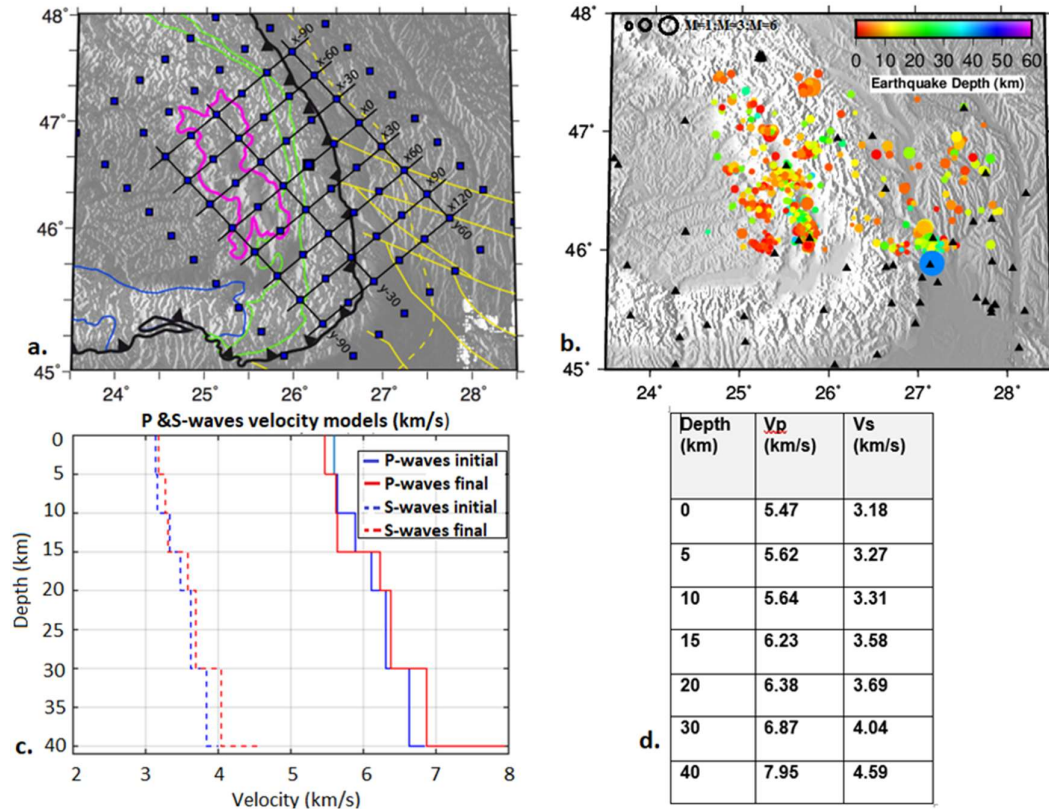


Figure 2. *a.* Distribution of model grid nodes (blue squares) and the main geological features across the Eastern Carpathians (the pink contour - the Neogene volcanic area, solid yellow lines - foreland faults system, dashed yellow line - the Carpathian Foreland Basin, blue contour - the Dacia allochthonous unit, green contour - Carpathian Thrust belt comprising folded marine sediments, black line - the Carpathian front). *b.* Distribution of seismic stations (black triangles) and selected epicenters colored as a function of depth and scaled with magnitude. *c.* P and S-wave velocity model slightly modified as was estimated by the VELEST algorithm and used as input for the inversion. *d.* Initial 1-D velocity model used in the DD tomography.

The vertical nodes are incremented from 5 km between 0 and 15 km and 10 km between 20 km and 40 km since most of the selected events occurred in the upper crust. Smoothing and damping values were selected from the knee of the trade-off curves (Eberhart-Phillips, 1986) between the model and data variance (see Figure S1 from Supporting information (SI) section). We set a smoothing value of 15 and a damping value of 100 and 65 for the last iterations since part of the data are rejected due to the residual weighting scheme (Waldhauser and Ellsworth, 2000, Zhang and Thurber, 2003).

4. Resolution and uncertainty

The system of equations is solved using an LSQR algorithm (Paige and Saunders, 1982), a popular iterative damped linear solver that does not provide the covariance matrix. This limits the ability to assess the uncertainty and resolution directly from the inversion. However, the use of both absolute and relative arrival time data should in theory provide more accurate results than standard tomography, where imprecise picks and correlated errors would result in more scattered event locations. We expect picking errors to be in the range of 0.04s, while differential travel time values to be less than 0.005s. With these simulated travel time errors, synthetic testing showed that differences between the true and recovered models have a

standard deviation of the order of 0.16 km/s, while event relocation accuracy has been shown to be less than 0.5 km (Zhang and Thurber, 2003).

The model resolution may be assessed using the derivative weight sum (DWS) values which represent the ray density coverage. We displayed in Figure S2 (SI) the DWS distribution computed for P-waves to better illustrate the best-recovered depth intervals. A higher DWS value will correspond to seismically active areas and good station coverage compared with regions with scarce seismicity or poor station coverage. Zhang and Thurber (2003) argued that the DWS may serve as a good measure of model resolution where the resolution matrix is not available. To assess the model resolution at different scales, we carried out classic checkerboard tests, which are important to estimate how well the amplitude and shape of velocity anomalies can be recovered using the selected data set and the parameterization (e.g. Rawlinson et al, 2014). The checkerboard velocity model was built by adding 5% alternately positive and negative velocity anomalies of 30 km width to the initial 3D velocity grid. To simulate the picking errors, we subsequently added random noise with zero mean and standard deviation of 0.15 s to the synthetic data. The standard deviation value was chosen as a mean value of picking uncertainty of P and S-waves determined through the correlation of the seismic waves for the pairs of events which occurred in a distance range of 10 km. The synthetic absolute and differential travel-times are determined using the pseudobending ray-tracing method of Um and Thurber, (1987), using the real receivers and event locations. The synthetic data are inverted using the same parameterization system as for the real data.

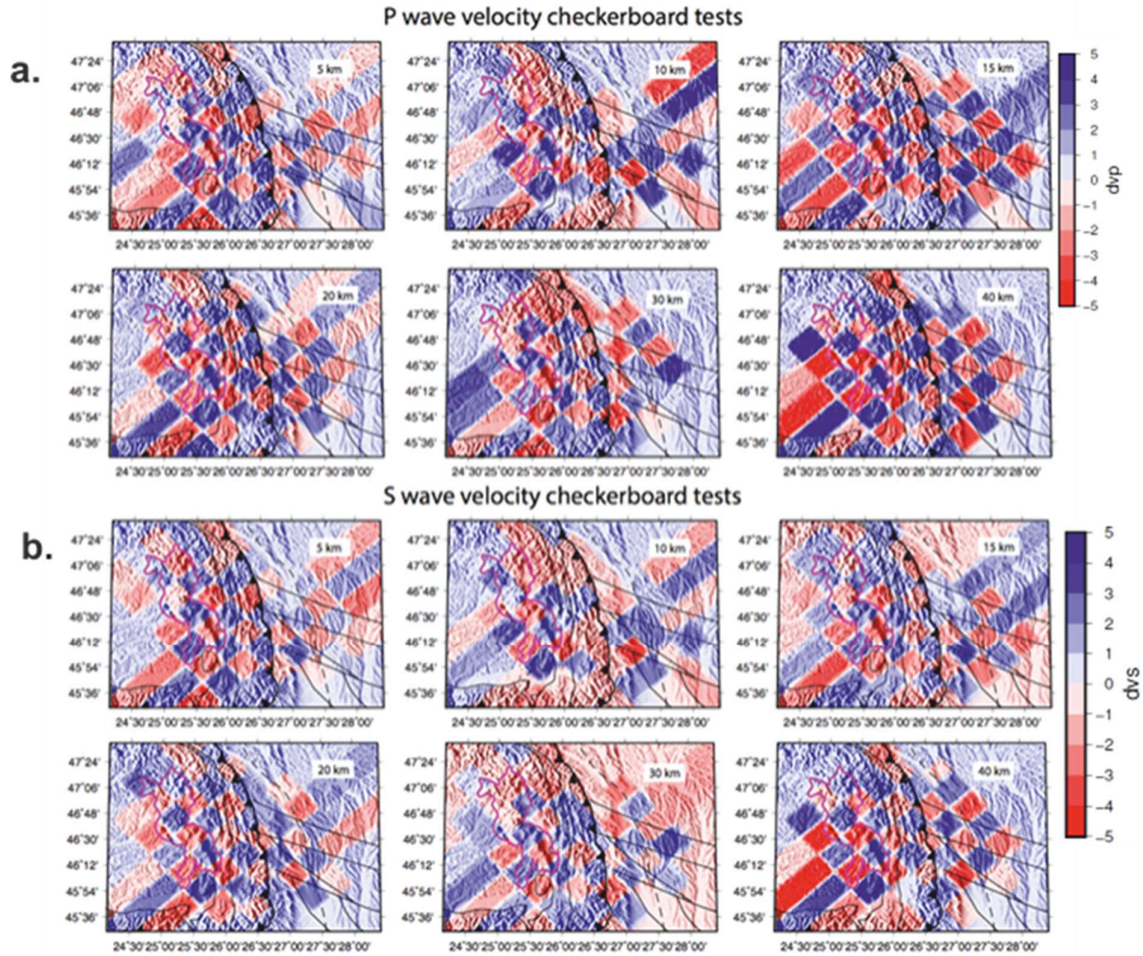


Figure 3. Results of the checkerboard test in horizontal sections for P (a.) and S (b.) waves at 5, 10, 15, 20, 30 and 40 km depth (from left to right).

The results of the checkerboard test for P and S-waves are displayed in *Figure 3 (and S3 figure from SI)*. The input pattern is well recovered for both P and for S-waves for almost the entire study area except at the edges where the resolution drops for several depth intervals. The best resolution is observed between 15 and 20 km depth where checkerboard patterns are recovered above 80% of the original input synthetic anomaly amplitude on a total of 88 selected nodes. To better assess the influence of parametrization, we performed an additional test, by moving the entire grid 15km towards NE (*Figure S4, in SI*). Shifting the grid inevitably results in the decay of resolution and time residuals and will produce some differences in the resulting model. However, the main velocity features are not affected significantly, although some differences appear at model edges. To highlight the reliability of the low-velocity anomalies we carried out spike tests (*Rawlinson & Spakman 2016*) which showed that a low-velocity anomaly (10%, top of the figure) is well-recovered by both P and S velocity structures with insignificant influence on the surrounding region (*Figure S5, in SI*).

5. Results

We show in *Figures 4 and 5* the horizontal sections of P and S-velocity structure obtained for the study area by applying double-difference tomography and in *Figures 6 and 7* the Vp/Vs ratios distribution and the relocated seismicity patterns. In the following, the results are presented briefly and in the next section, we discuss our findings in connection with the previous studies. Data residual distribution obtained for the last iteration indicates that the inversion was successful (*Figure 7d*).

5.1. Velocity models

The horizontal slices of P and S-wave velocity models in the ECA are shown in *Figures 4 and 5*. Both horizontal and vertical slices (*Figure 8*) of velocity structures, as well as of the Vp/Vs ratios (*Figure 6*) were displayed only for the grid nodes where the DWS values are greater than 50 as a good proxy for the resulting model. On the tomographic images, we overlay the main geological features, the fault system (gray lines) and the volcanic features (black triangles, Persani-bottom left, Ciomadul-bottom right and Gurghiu-middle and Calimani-top). The relocated epicentres (green dots) are displayed (function of magnitude) within ± 2 km depth relative to the horizontal layer. To better emphasize and interpret the results we represented in *Figure 9* a simplified tectonic sketch map of the Eastern Carpathians.

5.1.1 P-wave velocity

The horizontal slices of the P-wave velocity model (*Figures 4 and 5*) reveal heterogeneous velocity structures beneath the region. In the upper crust (≤ 10 km) $V_p \sim 5.4$ km/s dominates the entire region, with the lowest velocities estimated in the EEC, which is overlain with pre-Carpathian Paleozoic sediments (*anomalies G, H, E in Figures 4 and 8*). In the middle crust, low V_p (< 4.5 km/s) follows the Carpathian front and the inner Carpathians side, although along-strike variations exist. Higher velocity patches ($\delta V_p = 8\%$ with respect to the regional average) emerge in the EEC below ~ 10 km depth, where presumably Precambrian crystalline basement is present. Throughout the upper crustal layer, the lowest velocities ($V_p < 5$ km/s) are estimated beneath the volcanic region and portions of the ECA, at levels where Carpathian Nappes are still probably thrust over the craton margin and where seismicity is most intense (10 km depth, *Figure 4*). Higher velocities ($V_p > 6.5$ km/s) emerge beneath the Transylvanian Basin NW and SW of the volcanic chain at 15 km depth. The position of the high-velocity patch moved to the east at 10 km depth (*Figure 4*), marking a dip in the anomaly, consistent with the underthrust westward dipping EEC crust (*Maţenco and Bertotti 2000*), while the Carpathian nappes maintain the lowest velocities of the region (15 km depth, *Figure 4*). In the lower crust (~ 30 km) low-velocity features ($V_p \sim 6.0$ km/s) appear along the southern part of the Eastern Carpathians and towards Carpathians foredeep area as well as beneath the Gurghiu

volcanoes (*anomaly D*, *Figure 5*). In contrast, at the same depth, the much younger volcanic centres (Persani and Ciomadul mountains) located NW relative to the Vrancea region are dominated by relatively high-velocities ($V_p > 6.5$ km/s at 30 km depth, *Figure 5*).

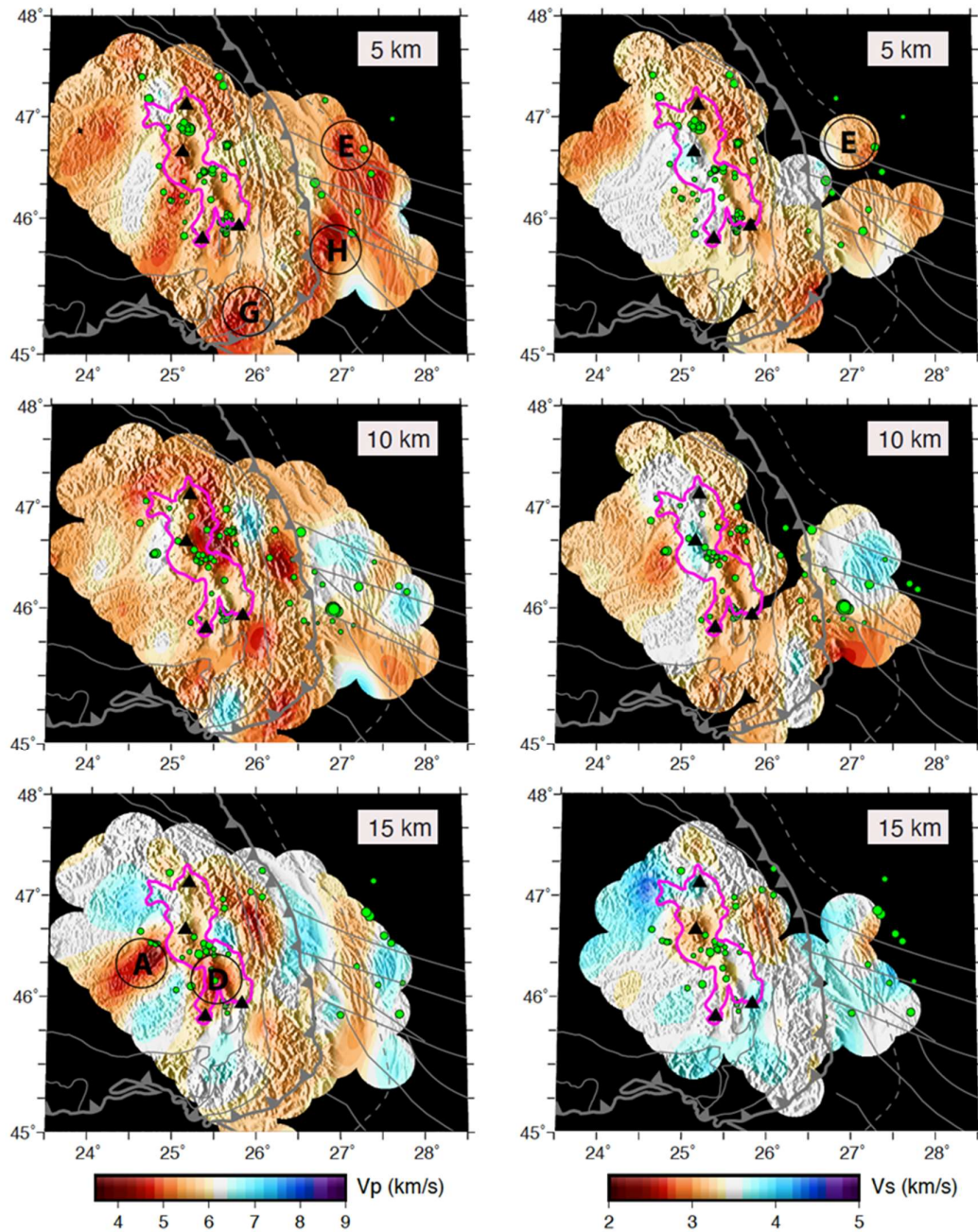


Figure 4. Horizontal sections of *P* (left) and *S* (right)-wave velocity models of upper crustal depths (5, 10 and 15 km) obtained using the double-difference algorithm. Black triangles show the volcanic centers and green dots relocated epicenters (size as a function of magnitude). Gray solid lines show geological features while dashed one represent the Carpathians foreland area. The pink contour shows the Neogene volcanic area. Low-velocity anomalies (for V_p and V_s) at the edge and in front of the ECA are shown by G, H and E letters while in the back of the ECA are marked by A and D letters.

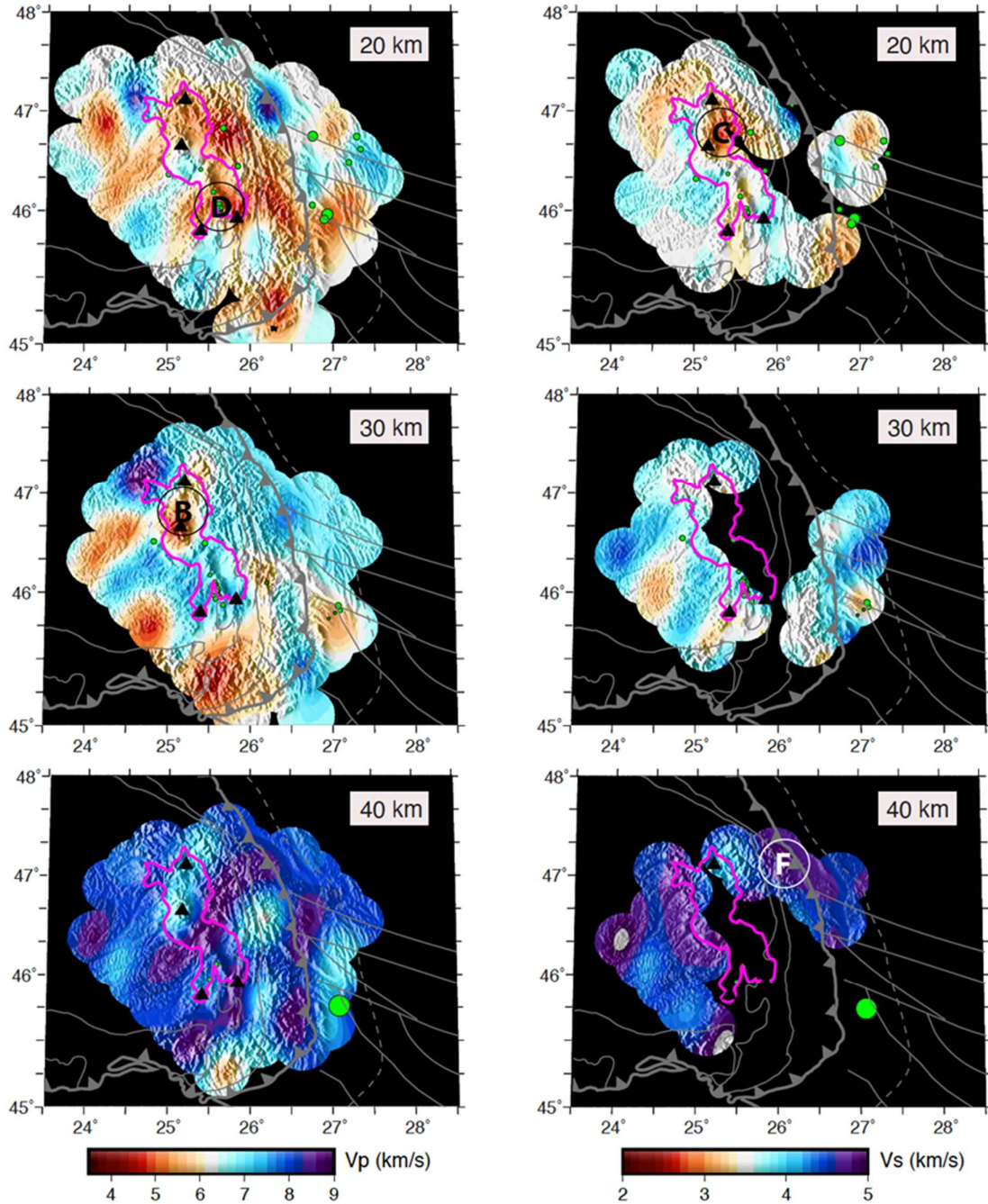


Figure 5. Horizontal sections of *P* (left) and *S* (right)-wave velocity models of lower crustal depths (20, 30 and 40 km) obtained using the double-difference algorithm. Legend and abbreviations are as in Figure 4. Low-velocity anomalies in the back of the ECA are marked by B, C and D letters while the high-velocity anomaly is shown by F letter.

5.1.2 S-wave velocity

The S- wave velocity structure reveals similar features as those revealed using P-waves with a few exceptions, despite the fact that the number of rays used for the analysis is ~2% lower. Prevailing high velocities ($V_s \sim 4$ km/s at 10 km and 15 km depths, Figure 4) dominate the EEC domain throughout the crust, while lower values characterize the Carpathian foredeep down

to ~10 km (*anomaly E in Figure 4*) and the Carpathian Nappes. Low velocities ($V_s \sim 3.4$ km/s) are estimated in the middle and lower crust (*Figure 5*) towards the western edge of the Peceneaga Camena Fault, the place where the strongest crustal earthquake recently occurred (November 2014) and triggered a seismic sequence by weakening neighboring fault systems. In contrast to the P-wave velocity distribution, S-wave velocities in the volcanic chain do not show a distinct anomaly with respect to the regional average in the uppermost crust, but the region beneath Gurghiu volcanoes becomes especially slow ($V_s < 3$ km/s, *anomaly C in Figure 5*) in the 15-30 km depth range (*Figures 4, 5*).

5.1.3 Vp/Vs ratios

Because the selected DD tomography algorithm does not determine directly for Vp/Vs ratios, we estimate these values by the division of the Vp by the Vs model. Although S-wave data are only 2% less than for P-waves, we note, however, that features may be dominated by anomalies in Vp since it is better resolved and the ray path coverage is denser. *Figure 6* shows the Vp/Vs distribution only for regions where the minimum DWS criteria is satisfied for the Vs and checkerboard anomalies are recovered above 80% for both Vp and Vs. The figure also displays a range of Vp/Vs ratio for common crustal lithologies (*after Christensen, 1996*) and the Vp/Vs ratios distribution obtained from previous studies, which show similarity to our results. Vp/Vs vary between 1.5 and 2.3, with the highest values estimated in the northern segment of the Neogene volcanic region (>2.2) at depths below 20 km, and the Focsani Basin at ~10 km depth (*Figure 6*). Low values (<1.7) can be observed in the volcanic region and the foreland basin (*Figure 6, anomaly H*) in the top 5-10 km, and in the southern volcanic segment (*Figure 6, anomaly D*), Carpathian nappes, and the foreland at ~15 km (*Figure 6*).

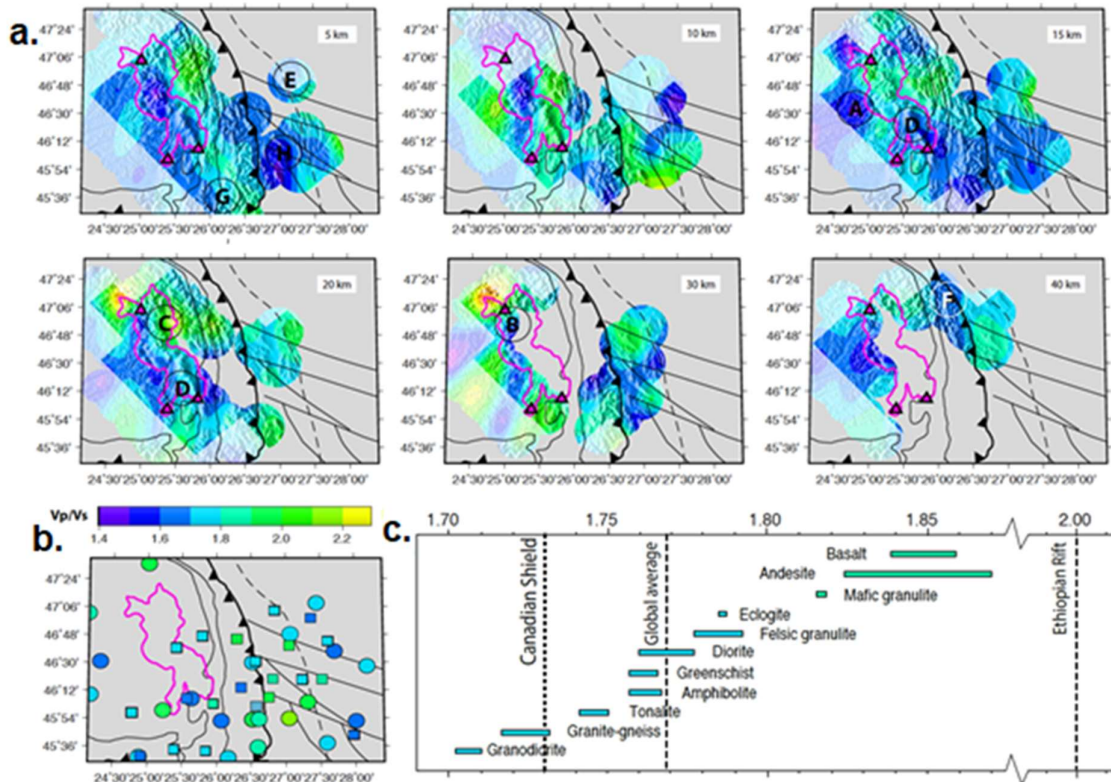


Figure 6. *a.* Depth sections through Vp/Vs resulted models. *b.* mean crustal Vp/Vs ratios determined using receiver functions from previous studies: circles (Petrescu et al., 2019), rectangles (Diehl et al., 2005). *c.* Variation in Vp/Vs ratios (bottom) for different petrological compositions based on laboratory experiments of Christensen et al. (1996). Figure modified

after Thompson et al. (2010), with values for specific locations taken from Petrescu et al., 2016 (Canadian Shield) and Stuart et al., 2006 (Ethiopian Rift). Low V_p/V_s ratios are indicated by letters: A, B, C, D, E, F, G, H and E while high V_p/V_s ratios are shown by C letter.

5.2. Event relocation

The relocated events (563) are shown in Figures 4, 5, 7 and 8. The weighted root mean square (RMS) residuals for catalogue and cross-correlation data decrease significantly from 0.63 and 0.56s to 0.004s respectively to 0.002s. A number of 90 events were rejected from the inversion due to residual limits and their location as air quakes. We show in Figure 7 the distribution of the relocated events using the velocity model resulted from the inversion and the fault plane solutions of the significant earthquakes as were considered by Petrescu et al. (2020b). The hypocenters distribution in the inner side of SE Carpathians, along the volcanic chain, as highlighted on the A-A' cross-section (Figure 7), shows a tendency of increasing focal depth towards the south (towards Vrancea intermediate-depth source). The hypocenters of the events located along and around A-A' segment are represented in the vertical cross-section. For the events around the Peceneaga-Camena and Trotus faults, we show an enlarged view of the area in order to better emphasize the association of these events with the crustal fault system. A few clusters of events are located in confined, narrow areas around the volcanic structures. Several industrial quarries are known to be operating in the study region and may contaminate our database. Unfortunately, we have no access to the operating schedule of the quarry blasts activity in order to discriminate them from the tectonic events. Taking into account the large percentage of events with hypocentral depths below 10 km, we may assume that these events are of tectonic nature. After relocation, many of the hypocenters are redistributed in the low-velocity areas especially beneath the southern volcanoes while in the northern segment they tend to cluster in the upper crust. Epicenter distributions show a tighter clustering after relocation and better correlation with the location of surface fault systems.

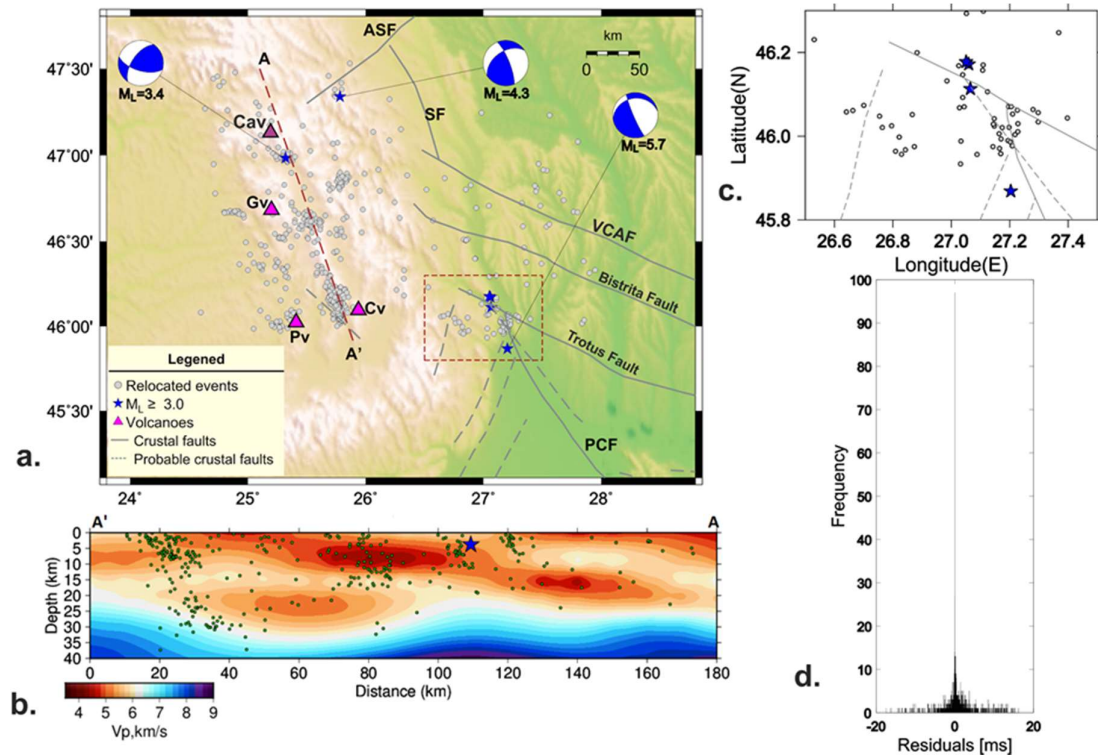


Figure 7. a. Relocated epicenters and fault plane solutions for the largest events (after Petrescu et al., 2020b). Solid gray lines are described in Figure 1 and the dashed gray lines

show the location of possible crustal faults (Diaconescu, 2017). The volcanic centers are displayed by pink triangles. **b.** Vertical cross-sections through the V_p along the A-A' profile and relocated hypocenters within a distance of ± 0.5 deg. relative to the same profile. **c.** Enlarged section of the area where the strongest earthquakes occurred. **d.** Data residual distribution obtained for the last iteration. Abbreviations are as for Figure 1.

6. Discussion

The ECA has been shaped by the significant tectonic structures aforementioned in the previous sections. We discuss in the following sections the resulting velocity structures with respect to the previous investigations that partially overlap the study area and with the main geophysical and geotectonic features.

6.1. Seismic structure of the upper crust: Carpathians orogen and sedimentary basins

Carpathian nappes are juxtaposed over the EEC passive margin, which was already covered with Paleozoic and Mesozoic sedimentary layers between 1 km and 6 km thick, increasing from north to south and east to west (Sandulescu 1984, Grasu et al., 2002, Răileanu et al., 2005). The Neogene foreland sedimentary cover related to the Carpathian orogeny added between 1 km and 13 km of sediments, the deepest point located in the Focsani Basin at the South-East Carpathian corner (Tărăpoancă et al., 2003). Relatively low-velocity features ($V_p < 5.4$ km/s; $V_s \sim 3.3$ km/s) observed in the upper crust of the foreland (anomalies G, H, E in Figures 4 and 8) may reflect the presence of sedimentary layers, although velocity variations do not seem to perfectly mirror these trends or the smoothly varying sedimentary layer thickness inferred from gravity anomaly analysis (Ioane and Ion, 2005), mostly due to insufficient vertical resolution limited to the model grid depth spacing. The foreland comprises various consolidated crustal blocks showing differences in the sedimentary cover (Seghedi et al., 1998, Hauser et al., 2001). These may be reflected in the strong spatial seismic velocity variation in the top 10 km (Figure 4), with the lowest velocities detected in the Focsani Basin (anomaly H, in Figure 4), confirming the deep nature of this anomalous foredeep basin (Tărăpoancă et al., 2004). Here, V_p/V_s ratios increase to ~ 2 at 10 km (Figure 6), compatible with results from previous studies (Diehl et al., 2005, Petrescu et al., 2019) and are associated with decreased V_s . Generally, low V_s , high V_p/V_s regimes detected in sedimentary basins likely indicate high fluid saturation and pressure (Dvorkin et al. 1999) and have often been interpreted as such (e.g. Reyners et al. 2006, Moretti et al., 2009), although elevated pore pressure commonly encountered in unconsolidated sediments may also play a role in rising V_p/V_s values (Zimmer et al. 2002).

6.2. Nature and clustering of seismic events in the foreland of the South-Eastern Carpathians

Relocated earthquakes tend to cluster in the upper crust in the foreland area, closer to the surface expression of the Vaslui-Cetatea Alba and Trotus Faults that fracture the Precambrian foreland crust, and at the junction between the Trotus and Peceneaga-Camena Faults (Figure 7). In the top 10 km, seismicity along the Trotus-Barlada-Vaslui Fault system appears to continue beneath the Carpathian nappes for at least 100 km west of the Carpathian Front, suggesting the continuation of the East European passive margin at least up to the volcanic chain (Figures 1 and 4). The relocated seismicity in the upper crust emphasizes better the northwestern trend of epicenters. The top 10 km may comprise both pre-Carpathian Paleozoic cover of the East European Craton and the juxtaposed Carpathian nappes. Gravity studies indicate that Paleozoic sedimentary layer thickness varies considerably along the strike from 1 km to 9 km towards the South and likely tapers out beneath the Carpathian nappes, placing earthquakes either along nappe faults and deeper within the Paleozoic cover or the crystalline

basement of the ECC. Near the northwestern edge of the Peceneaga Camena fault system, a series of small and moderate earthquakes occur at progressively deeper levels of the crust, down to ~40 km, confirming the crustal-scale of this seismogenic fault system (Leever *et al.*, 2006, Maţenco *et al.*, 2007, Craiu *et al.*, 2019). The significant seismic activity, its polikinetik character revealed in this region (Radulian *et al.* 2000) and also a rather low b-value obtained for the last significant seismic sequence (Craiu *et al.*, 2019) could indicate a high fluid pressure regime in the compressional forearc. Strong correlations between seismicity and high pore fluid pressure sediments have also been observed in Reno, Nevada (Jensen *et al.*, 2019). Towards Vrancea, in the Focsani Basin, relocated events show better clustering in the upper and mid-crustal levels. Here, their distribution is tighter and shifted deeper by ~4km. Events from the lower crust also display a better clustering and are shifted upwards (~3km) towards the middle crust. The progressive deepening of earthquake hypocentres towards the South-Eastern Carpathians suggests that these events are tectonic in nature and are likely caused by lithospheric bending forces associated with the actively detaching Vrancea slab (e.g. Petrescu *et al.*, 2020).

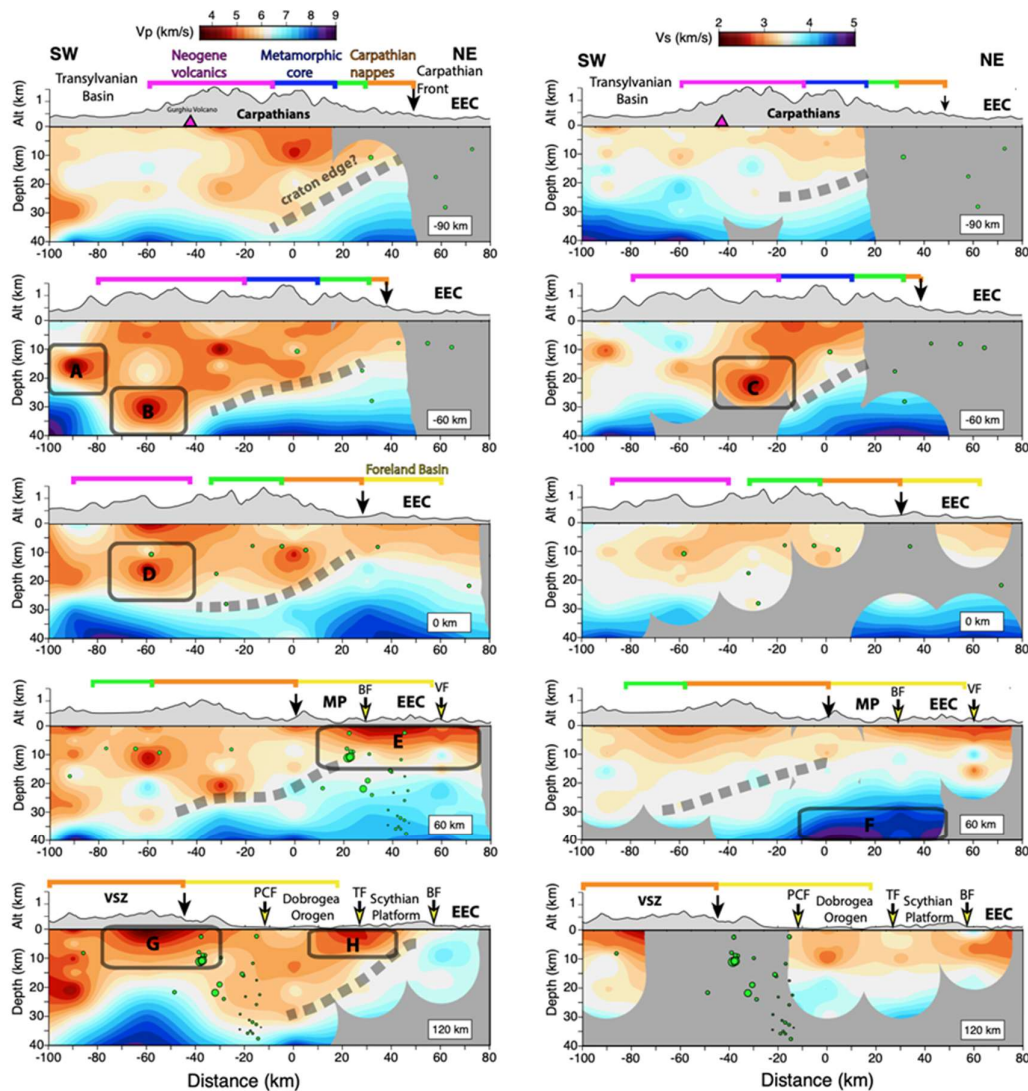


Figure 8. Vertical cross-sections through the Vp along horizontal X grid directions (see Figure 2) placed at (from top to bottom): -90 km, -60 km, 0 km, 60 km and 120 km. Green dots display

the relocated hypocenters within a distance range of 2km relative to the profile. The letters show the velocity anomalies as were mentioned in *Figures 4 and 5*. The gray dashed line shows the probable trace of the craton edge roughly following the ~ 6.3 km/s contour for V_p and the ~ 3.5 km/s contour for V_s . The abbreviations are as follows: MP: Moesian Platform; EEC: East European Craton; BF: Bistrita Fault; TF: Trotus Fault; VSZ: Vrancea Seismic Zone; PCF: Peceneaga Camena Fault

6.3. Possible TESZ structure across Romania

The top of the underlying Precambrian crystalline basement slopes gently from 6 km beneath the Carpathian Front to 10 km beneath the Carpathian nappes (*Ioane and Ion, 2005, Mațenco and Bertotti, 2000*). Thus, we would expect to see a heterogeneity associated with the craton margin below these depths. The TESZ is generally considered a lithospheric-scale boundary separating cratonic lithosphere from younger Phanerozoic accreted terranes down to ~ 140 km (*Zielhuis and Nolet, 1994, Ren et al., 2012*) and its crustal structure has been investigated with several active seismic surveys. In Poland, for example, the TESZ appears as a crustal-scale subvertical interface separating high V_p (>6 km/s in the top 20 km) and thick (Moho ~ 44 km) crust from a lower V_p (<5 km/s) thinner crust (*Grad and Polkowski, 2015; Narkiewicz et al., 2015*). Moreover, there is a wide range of possible TESZ trajectories suggested in Romania (*Stănică et al., 1999, Atanasiu et al. 2005, Bocin et al 2013, Ioane and Stanciu, 2018*). Specifically, the Miocene suture between younger accreted terranes and the Precambrian foreland may either follow the inner Carpathians, east of the Ciomadul volcano or further west (~ 100 km), in the South Carpathians. Others place it along the Peceneaga-Camena Fault (*Winchester, 2002, Pharaoh et al., 2006*) or along the Trotus-Barlad Faults in the foreland (*Bocin et al., 2013*).

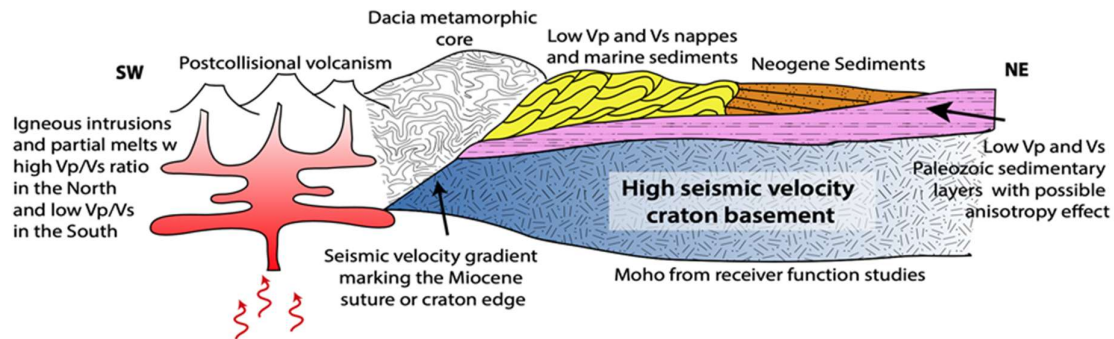


Figure 9. Simplified tectonic sketch of the East Carpathian crust, showing the main seismic properties and structural features of the crust from the old Precambrian terranes in the East across the orogen and the volcanic belt in the West.

Seismic velocity structure of the foreland (see anomalies E, H, F in *Figures 4, 5, 6 and 8*) shows heterogeneous structures at depth, seemingly independent of the surface fault systems, suggesting a more complex geological architecture at depth than the one outlined by the surface geology (*Figures 4, 5*). Relocated seismicity tends to migrate upwards to upper crustal levels. This suggests that most of the major seismogenic faults in the foreland do not extend down to the Moho, except the Peceneaga-Camena Fault (*Craiu et al., 2019*), and do not share the expected characteristics of the presumably lithospheric-scale TESZ. The craton edge is thus more likely to extend beneath the Carpathian nappes rather than turning SE-ward beneath the foreland, although our model has insufficient resolution at lowermost crustal to uppermost mantle depths, to reliably detect any corresponding Moho depth changes. Our results highlight a possible craton edge trajectory at the mid-to lower-crustal depths beneath the Carpathian nappes, east of the Ciomadul volcano (anomaly D, *Figures 4, 5, 6 and 8*),

where the Miocene suture is most likely to be located (*Knapp et al., 2005*). Here, a SW dipping velocity gradient becomes apparent in the mid-to-lower crust (15-35 km), marking a change from $V_p > 6.3$ km/s and $V_s > 3.5$ km/s in the eastern footwall, presumably the dipping Precambrian basement, to $V_p < 6$ km/s beneath the Carpathian Orogen to the west (*Figures 5 and 8*). This gently dipping seismic velocity transition is consistent in both V_p and V_s images (*Figure 8*) and is similar to the upper 25 km from the refraction seismic profile in Poland (*Grad and Polkowski, 2015*). Also, seismic refraction surveys in more northern sections reveal a high-velocity layer in the lowermost crust extending beneath the Elbe Line in Germany (*Bayer et al., 2002, and references therein*). Our interpretation is also consistent with inferences from gravity and magnetic studies (*Bocin et al., 2013*) that show a much denser and magnetically susceptible crust to the west of the Carpathians (*Figure 10*). We note, however, that the western trace of the TESZ beneath the Carpathians must have been reworked in collision and affected by subsequent metasomatism from post collisional volcanism and progressive slab break-off along the subducting craton edge (e.g. *Sperner et al., 2001*). Thus, its seismic characteristics would probably change as a result of increased pressure, temperature and chemical erosion.

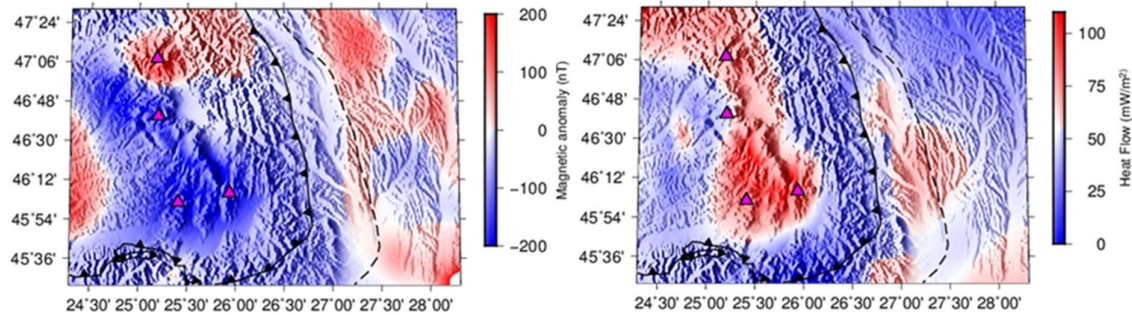


Figure 10. The magnetic anomaly (left side; after *Maus et al., 2009*) and heat flow (right side; after *Global Heat Flow Database; <https://www.ihfc-iugg.org/products/global-heat-flow-database>*) maps. Pink triangles display the volcanic centers (Persani-bottom left; Ciomadul-bottom right; Gurghiu-middle; Calimani-top).

6.4. Seismic heterogeneity in the Neogene volcanic chain

In the volcanic field that forms the inner Carpathians, low V_p and V_s patches located throughout the crust (*anomalies A, B, C, D in Figures 4 and 5*) collocate with negative magnetic anomalies (*Maus et al., 2009*) at surface, high heat flow measurements (*Figure 10*), and low Bouguer gravity anomalies (*Seghedi et al., 2019*). While the area is pervaded mostly by igneous volcanic rocks, several sediment-filled tectonic basins are also present in the upper crust (*Seghedi et al., 2019*), probably obscuring the magnetic signal from deeper igneous processes (e.g. *Blaikie et al., 2014*). *Demetrescu and Andreescu (1994)* suggested that the high heat flow anomaly may be reminiscent of the Miocene subduction and post-collisional volcanism. However, 2D geothermal models from the South China Sea (*Xiao-Yin et al., 2014*) show that the temperature disturbances caused by ~10 km thick vertical igneous intrusions almost completely disappear after 10 Ma. By analogy, any thermal effect of Neogene intrusions should have decayed by now, unless volcanic processes are ongoing below the surface, as was suggested by *Popa et al., (2011)* in the southernmost segment of the volcanic chain. Previous studies of classic locations of magmatism-pervaded crust showed some resemblance to our images, specifically regions of $V_p > 5.5$ km/s in the upper crust (e.g. *Toomey and Foulger, 1989, Foulger and Arnott, 1993, Foulger et al., 1995, Tryggvason et al., 2002*). Magmatic chambers with some degree of partial melt usually have low V_p signatures (≤ 5.5 km/s) and are associated with high V_p/V_s ratios (>1.8 , e.g. *Toomey and Foulger, 1989, Foulger and Arnott, 1993, Foulger et al., 1995, Benz et al., 1996, Dawson et al., 1999,*

Tryggvason *et al.*, 2002, Hansen *et al.*, 2004). Some exceptions include images of Etna, Italy, where high Vp bodies were detected beneath the volcano (Patane *et al.*, 2002, Patane *et al.*, 2006). In Romania, H-k stacking of receiver functions estimated low bulk crustal Vp/Vs ratios (<1.7) in the region (Diehl *et al.*, 2005, Petrescu *et al.*, 2019). While in the uppermost crust, we also estimate low Vp/Vs (<1.7), values increase up to 1.9 at ~10-15 km depth, consistent with the depth range of the previously inferred magma chamber (Popa *et al.*, 2011). Such high values are generally interpreted to indicate the presence of mafic compositions or even partial melts (e.g. Christensen *et al.*, 1996, Stuart *et al.*, 2006, Figure 6). Lower Vp/Vs on top of higher values could speculatively suggest differentiation of a low-temperature magma chamber into felsic and mafic layers (e.g. Zhang and Lin, 2014, Yukutake *et al.*, 2015). Moreover, an active seismic cluster is vertically distributed throughout the crust beneath the southern volcanoes at the transition between low and high Vp may be evidence of an active magmatic plumbing system (Figures 7 and 8). Other possible mechanisms affecting Vp/Vs distributions stem from possible anisotropy in the Earth's crust, most commonly thought to originate due to either preferential alignment of fractures (e.g. Boness and Zoback, 2004), fluid-filled microcracks (e.g. Crampin *et al.*, 1987, Pimienta *et al.*, 2018), sedimentary bedding layers (e.g. Kern and Wenk, 1990), enrichment in silica content or the presence of quartz in the crust (Chiarabba & Amato, 2003), mineral alignment in metamorphic terranes or shear zones (Balfour *et al.*, 2005, Boness and Zoback, 2006), or a combination thereof. Cumulative effects of crack anisotropy and high pore fluid pressure can lead to a significant increase in Vp/Vs ratios (up to 3.5) when ray paths are perpendicular to the crack fabric (Wang *et al.*, 2012). It is conceivable that a large number of sub-vertical ray paths travelling upwards and perpendicular on alternating sequences of rock layers of different seismic velocities, such as sedimentary bedding layers in the thick Focsani Basin, or the Paleozoic covers of the East European Craton, may contribute to increasing Vp/Vs alongside the presence of mafic material and partial melts.

6.5. Nature and clustering of seismic events in the inner Carpathians

The relocated seismicity within the inner Carpathians area highlights clusters of seismic events placed especially around volcanic structures (Figure 7). The seismicity consists of low to moderate-size events ($M_w < 4.0$) distributed within the crust (~1 to ~40 km within the selected dataset) and seems to be correlated with transitions from low to high-velocity features (Figure 8). The crustal events may be associated with the various faults and fractures that cross the Carpathians as well as possibly active magmatic processes revealed beneath the volcanic chain (Popa *et al.*, 2011). The lower crust seismic activity consists mainly of low magnitude ($M_w < 3.5$) events and maybe is related with Na-alkalic volcanism in the South Harghita mountains or maybe reflecting possible secondary extension of the post-collisional processes occurred at the bend of the South- Eastern Carpathians (Neagoe *et al.*, 2010). Specifically, the events which occurred close to Persani and Ciomadul volcanoes are distributed over a wide range of depths, crossing the Moho (~38 km) boundary (Răileanu *et al.*, 2012). This is in agreement with the findings of previous studies (Radulian 2000, Bala *et al.*, 2015) which showed a generally sparse seismic activity in the Transylvanian Basin while several clusters of seismic events have been identified beneath the Neogene volcanic region (Enescu *et al.*, 1996, Neagoe *et al.*, 2010, Popa *et al.*, 2011). These events might be influenced on one hand by the active slip along the crustal-scale faults that probably extend beneath the Carpathian nappes (Figure 7). On the other hand, volcanism may have a direct influence on seismic activity. The last eruption occurred ~30 Ka ago at Ciomadul volcano (Harangi *et al.*, 2015a, Karátson *et al.*, 2016, Molnár *et al.*, 2019), a seemingly inactive volcano (Harangi *et al.*, 2015a, Szakács *et al.*, 2015, Laumonier *et al.*, 2019). Although it is inactive there is some evidence like, high $^3\text{He}/^4\text{He}$ anomaly measured in natural gases, thermal and CO_2 -rich mineral waters (see Popa *et al.*, 2011 and the reference therein), the thermal anomaly (Demetrescu and Andreescu, 1994, see Figure 10), low electrical resistivity (Harangi *et al.*, 2015a), seismic wave attenuation (Russo *et al.*, 2005, Borleanu *et al.*, 2017), supporting an old system holding presumably active magma storage (Harangi *et al.*, 2015a). Moreover, our results are in agreement with previous studies (Oncescu *et al.*, 1999, updated, Popa *et al.*, 2011) supporting

the presence of an active magma chamber that probably generates the observed low-magnitude seismicity distributed from crust to the upper mantle around the Ciomadul volcano. The cluster of seismicity is not, however, confined to a specific depth range and cannot precisely constrain the location of the potential magmatic chamber. Towards the northern edge of the inner Carpathians, relocated earthquakes are shifted towards the upper crust, in a region of low V_p and moderate V_s associated with low V_p/V_s (Figures 6, 7b), indicative of low fluid content or dominant felsic composition, suggesting that active magmatic processes are likely not responsible for the observed seismicity. Here, a NW-SE oriented transpressional fault that acted as a plane of weakness for migrating magmas in the past (Seghedi *et al.*, 2019) may have been tectonically reactivated.

7. Conclusions

To investigate the structure of the Carpathian Orogen and the edge of the East European Craton in Romania, we analyzed local crustal earthquakes and used double-difference tomography to obtain 3D P and S-wave velocity models of the crust and improve the hypocentral distribution of local earthquakes. Our study provides the highest resolution 3D crustal seismic model of this area to date and emphasizes the manifestation of surface tectonic boundaries at lower crustal depths. Overall, a contrast between the seismically faster cratonic foreland crust and the over-thrusted Carpathian nappes (consisting of seismically heterogeneous accretionary sediments, metamorphic cores of Phanerozoic terranes that collided with the craton during the Miocene and a post-collisional volcanic belt) is obtained. Our results show an SW dipping velocity gradient in the mid- to lower crust (15-35 km) with a transition from a footwall where $V_p > 6.5$ km/s to the west in a region where $V_p < 6$ km/s. The transition is characterized also by a downward decrease in the V_p/V_s ratio consistent with increasing felsic composition with depth. According to these results, the craton edge is imagined as an SW dipping limit going beneath the Carpathian Orogen in the mid-to-lower crust. Relocated hypocentres tend to cluster in the top 10 km of the crust and along the major faults crossing the foreland from east to west. In the Inner Carpathians, relocated hypocentres tend to be vertically distributed at the transition from low to high-velocity features especially beneath the southern volcanoes and may be evidence of an active asymmetric magmatic plumbing system. Post-collisional volcanism displays a southward age progression and may have altered the seismic structure of the TESZ, thermochemically reworking the crust. Here, we estimated mainly low-velocity features which correspond to a region of high heat flow measurements and a downward increase in V_p/V_s ratios up to 1.9, consistent with the presence of mafic material or partial melts.

Acknowledgments

The present study was started within a Research and Development Project funded by Air Force Technical Application Center (AFTAC) (FA7022-11-C-0015) and continued during a stage of the first author at the Kyoto University, Japan, financed by a grant awarded by The Matsumae International Foundation (MIF) (18G18) in 2018. We thank Prof. Zhang for allowing us to use the TomoDD code and to David von Seggern for his support in applying the waveforms cross-correlation code. We are also grateful to Prof. Dr. Luca De Siena, Dr. Wolfram Geissler, an anonymous reviewer and the Editors who made critical and valuable comments on the original manuscript. Data used in the present study were provided by the National Institute for Earth Physics (Romania). Most figures were made using GMT (Wessel and Smith, 1998).

References

- Atanasiu, L., Manda, M., Zugravescu, D., Roharik, M., 2005. Trans-European suture zone over the Romanian territory in the light of new satellite data, *Rev. Roum. Géophysique*, 49, p. 49–61, București.
- Bala, A., Răileanu, V., Dinu, C., Diaconescu, M., 2015. Crustal seismicity and active fault systems in Romania, *Rom. Rep. Phys.*, 67(3), 1176–1191.
- Balla, Z., 1987. Tertiary paleomagnetic data for the Carpatho-Pannonian region in the light of Miocene rotation kinematics, *Tectonophysics*, 139, p. 67–98.

- Baron, J., and Morelli, A., 2017. Full-waveform seismic tomography of the Vrancea, Romania, subduction region, *Physics of the Earth and Planetary Interiors*, ISSN: 0031-9201, Vol: 273, Page: 36-49.
- Bayer, U., Grad, M., Pharaoh, T.C., Banka, D., Guterch, A., Laigle, M., Marotta, A.-M., Scheck, M., Thybo, H., 2002. The southern margin of the East European Craton, new results from seismic sounding and potential fields between the North Sea and Poland. *Tectonophysics* 360, 301 – 314
- Benz, H.M., Chouet, B.A., Dawson, P.B., Lahr, J.C., Page, R.A., Hole, J.A., 1996. Three-dimensional P and S wave velocity structure of Redoubt Volcano, Alaska. *Journal of Geophysical Research* 101 (B4), 8111–8128.
- Blaikie, T.N., Ailleres, L., Betts, P.G., Cas, R.A. F., 2014. Interpreting subsurface volcanic structures using geologically constrained 3-D gravity inversions: Examples of maar-diatremes, Newer Volcanics Province, southeastern Australia, *J. Geophys. Res. Solid Earth*, 119, 3857–3878, doi:10.1002/2013JB010751.
- Bocin, A., Stephenson, R., Mațenco, L., Mocanu, V., 2013. Gravity and magnetic modelling in the Vrancea Zone, south eastern Carpathians: Redefinition of the edge of the East European Craton beneath the south-eastern Carpathians, *Journal of Geodynamics* 71, 52– 64.
- Boness, N., and Zoback, M.D., 2004. Stress-induced seismic velocity anisotropy and physical properties in the SAFOD Pilot Hole in Parkfield, CA. *Geophys. Res. Lett.*, 31:L15S17, doi:10.1029/2003GL019020.
- Boness, N., and Zoback, M., 2006. A multiscale study of the mechanism controlling shear velocity anisotropy in the San Andreas Fault Observatory at depth, *Geophysics* 71, no. 5, F131–F146.
- Borleanu, F., De Siena, L., Thomas, C., Popa, M., Radulian, M., 2017. Seismic scattering and absorption mapping from intermediate-depth earthquakes reveals complex tectonic interactions acting in the Vrancea region and surroundings (Romania) *Tectonophysics*, 706-707 (2017), pp. 129-142.
- Boyce, A. Bastow, I. D., Darbyshire, F. A., Ellwood, A. G., Gilligan, A., Levin, V., and Menke, W., 2016. Subduction beneath Laurentia modified the eastern North American cratonic edge: Evidence from P wave and S wave tomography, *J. Geophys. Res. Solid Earth*, 121, doi:10.1002/2016JB012838.
- Bratt, S.R and Nagy, W., 1991. The LocSAT program. San Diego: Science Applications International Corporation.
- Chiarabba, C., and Amato, A. 2003. Vp and Vp/Vs images in the Mw 6.0 Colfiorito fault region (central Italy): A contribution to the understanding of seismotectonic and seismogenic processes. *Journal of Geophysical Research*, 108(B5), 1–17. <http://doi.wiley.com/10.1029/2001JB001665>.
- Christensen, N., 1996. Poisson's ratio and crustal seismology, *J. Geophys. Res.*, 101 (B2), 3139–3156, doi:10.129/95JB03446.
- Ciulavu, D., Dinu, C., Szakács, A., Dordea, D., 2000. Neogene kinematics of the Transylvanian Basin, Romania. *AAPG Bulletin* 84 (10), 1589-1615.
- Craiu, A., Ghita, C., Craiu, M., Diaconescu, M., Mihai, M., Ardeleanu, L., 2019. The source mechanism of the seismic events during the sequence of the moderate-size crustal earthquake of November 22, 2014 of Vrancea region (Romania). *Ann Geophys* 61(1):SE666.
- Crampin, S. 1987. The geological and industrial implication of extensive dilatancy anisotropy. *Nature* 328, 491–496, doi:10.1038/328491a0.
- Dadlez, R., Grad, M., Guterch, A., 2005. Crustal structure below the Polish Basin: is it composed of proximal terranes derived from Baltica?. *Tectonophysics* 411, no. 1-4, 111-128, <https://doi.org/10.1016/j.tecto.2005.09.004>.
- Dawson, P.B., Chouet, B.A., Okubo, P.G., Villasenor, A., Benz, H.M., 1999. Three-dimensional velocity structure of the Kilauea caldera, Hawaii. *Geophysical Research Letters* 26 (18), 2805–2808.
- Demetrescu, C., Andreescu, M., 1994. On the thermal regime of some tectonics units in a continental collision environment in Romania. *Tectonophysics*, 230: 265-271.
- Diaconescu, M., 2017. Sisteme de fracturi active crustale pe teritoriul României, Ph.D. thesis, University of Bucharest, România.
- Dias, N. A., Matias, L., Lourenço, N., Madeira, J., Carrilho, F., & Gaspar, J. L., 2007. Crustal seismic velocity structure near Faial and Pico Islands (AZORES), from local earthquake tomography, *Tectonophysics*, 445(3-4), 301-317.

- Diehl, T., Ritter, J.R.R., the Calixto group 2005. The crustal structure beneath SE Romania from teleseismic receiver functions, *Geophys. J. Int.*, doi: 10.1111/j.1365-246X.2005.02715.x.
- Dvorkin, J., Mavko, G. & Nur, A., 1999. Overpressure detection from compressional and shear wave data, *Geophys. Res. Lett.*, 26, 3417–3420
- Eaton, D., and A. Frederiksen 2007. Seismic evidence for convection-driven motion of the North American plate, *Nature*, 446(7134), 428–431, doi:10.1038/nature05675.
- Eberhart-Phillips D., 1986. Three-dimensional velocity structure in Northern California Coast Ranges from inversion of local earthquake arrival times, *Bull. Seismol. Soc. Am.*, 76(4), 1025–1052.
- Enescu, D., Danchiv., D., Bala., A., 1992. Lithosphere structure in Romania. II. Thickness of Earth's crust. Depth-dependent propagation velocity curves for the P and S waves. *Stud Cercet Geol Geofiz Geogr Ser Geofiz.* 1992; 30: 3–19.
- Enescu, D. and Enescu, B., D., 1996. Focal mechanism and Vrancea (Romania) earthquake prediction. A model for predicting these earthquakes, *Rev. Roum. de Géophysique* 40, 17–32.
- Foulger, G.R., Arnott, S.K., 1993. Local tomography: volcanoes and the accretionary plate boundary in Iceland. In: Iyer, H.M., Hirahara, K. (Eds.), *Seismic Tomography: Theory and practice*. Chapman & Hall, London, pp. 644–675.
- Foulger, G.R., Miller, A.D., Julian, B.R., Evans, J.R., 1995. 3- Dimensional Vp, and Vp/Vs Structure of the Hengill Triple Junction and Geothermal Area, Iceland, and the Repeatability of Tomographic Inversion. *Geophysical Research Letters* 22 (10), 1309–1312.
- Global Heat Flow Database, <https://www.ihfc-iugg.org/products/global-heat-flow-database> (last time accessed 13 May 2018).
- Grad M., Jensen S.L., Keller G.R., Guterch A., Thybo H., Janik T., Tiira T., Yliniemi J., Luosto U., Motuza G., Nasedkin V., Czuba W., Gaczyński E., Środa P., Miller K.C., Wilde-Piórko M., Komminaho K., Jacyna J., Korabliova L., 2003a. Crustal structure of the Trans-European suture zone region along POLONAISE'97 seismic profile P4. *J Geophys Res* 108(B11):2541. doi:10.1029/2003JB002426
- Grad, M., Polkowski, M., 2015. Seismic basement in Poland. *International Journal of Earth Sciences*, 105: 1199–1214.
- Grasu, C., Miclăuș, C., Brânzilă, M., Bobos, I., 2002. The Sarmatian from the foreland basin system of the Eastern Carpathians, *Editura Tehnică, București*, 407 pp. (in Romanian with English summary, 368-373).
- Hansen, S., Thurber, C., Mandernach, M., Haslinger, F., Doran, C., 2004. Seismic velocity and attenuation structure of the east rift zone and south flank of Kilauea volcano, Hawaii. *Bulletin of the Seismological Society of America* 94 (4), 1430–1440.
- Harangi, S., Downes, H., and Seghedi, I., 2006. Tertiary-Quaternary subduction processes and related magmatism in the Alpine-Mediterranean region, in Gee, D., and Stephenson, R., eds., *European Lithosphere Dynamics: Geological Society of London Memoir* 32, p. 167–190.
- Harangi, S., Lukács, R., Schmitt, A.K., Dunkl, I., Molnár, K., Kiss, B., Seghedi, I., Novothny, Á., Molnár, M., 2015a. Constraints on the timing of Quaternary volcanism and duration of magma residence at Ciomadul volcano, east-central Europe, from combined U–Th/He and U–Th zircon geochronology. *J. Volcanol. Geotherm. Res.* 301, 66–80.
- Hauser, F., Răileanu, V., Fielitz, W., Bala, A., Prodehl, C., Polonic, G., Schulze A., 2001. VRANCEA99: the crustal structure beneath the southeastern Carpathians and the Moesian Platform from a seismic refraction profile in Romania, *Tectonophysics*, 340 (3–4) (2001), pp. 233-256, [https://doi.org/10.1016/S0040-1951\(01\)00195-0](https://doi.org/10.1016/S0040-1951(01)00195-0).
- Hauser, F., Răileanu, V., Fielitz, W., Dinu, C., Landes, M., Bala, A., and Prodehl, C., 2007. Seismic crustal structure between the Transylvanian Basin and the Black Sea, Romania, *Tectonophysics*, 430: 1-25.
- Ioane, D., Ion, D., 2005. A 3D crustal gravity modelling of the Romanian territory, *J. Balkan Geophys. Soc.*, 8(4), 189–198.
- Ioane, D., Stanciu, I.M., 2018. Extensional tectonics in Vrancea zone (Romania) interpreted on recent seismicity, geophysical and GPS data, *SGEM 2018 Conference Proceedings*, Volume 18, Issue 1.1 Applied and Environmental Geophysics.

Ismail-Zadeh, A., Mațenco, L., Radulian, M., Cloetingh, S., Panza, G., 2012. Geodynamics and intermediate-depth seismicity in Vrancea (the south-eastern Carpathians): current state-of-the-art. *Tectonophysics*. 530-531: 50-79.

Jansen, G., Ruhl, C. J., & Miller, S. A., 2019. Fluid pressure-triggered foreshock sequence of the 2008 Mogul earthquake sequence: Insights from stress inversion and numerical modeling. *Journal of Geophysical Research: Solid Earth*, 124, 3744– 3765. <https://doi.org/10.1029/2018JB015897>

Karátson, D., Wulf, S., Veres, D., Magyari, E.K., Gertisser, R., Timar-Gabor, A., Novothny, Á., Telbisz, T., Szalai, Z., Anechitei-Deacu, V., 2016. The latest explosive eruptions of Ciomadul (Csomád) volcano, East Carpathians a tephrostratigraphic approach for the 51–29 ka BP time interval. *J. Volcanol. Geotherm. Res.* 319, 29–51.

Kennett, B. L. N., and Engdahl, E.R., 1991. Travel times for global earthquake location and phase identification, *Geophys. J. Int.*, 105, 429– 465.

Kern, H., and Wenk H., R., 1990. Fabric-related velocity anisotropy and shear wave splitting in rocks from the Santa Rosa Mylonite Zone, California, *J. Geophys. Res.*, 95, 11,213–11,223, doi:10.1029/JB095iB07p11213.

Kissling, E., 1995. Program VELEST User's Guide – Short Introduction (Second Draft Version), Institute of Geophysics, ETH Zurich, Switzerland, 26 pp.

Kissling, E., 1988. Geotomography with local earthquake data, *Rev. Geophys.* 26, 659–698.

Knapp, J.H., Knapp, C.C., Răileanu, V., Mațenco, L., Mocanu, V. and Dinu, C., 2005. Crustal constraints on the origin of mantle seismicity in the Vrancea Zone, Romania: The case for active continental lithospheric delamination, *Tectonophysics*, 410(1-4), pp.311-323.

Koulakov, I., Zaharia, B., Enescu, B., Radulian, M., Popa, M., Parolai, S., Zschau, J., 2010. Delamination or slab detachment beneath Vrancea? New arguments from local earthquake tomography, *G-cubed*, doi:10.1029/2009GC002811.

Landes, M., Fielitz, W., Hauser, F., Popa M., and the CALIXTO Group, 2004. 3-D upper-crustal tomographic structure across the Vrancea seismic zone, Romania. *Tectonophysics*, 382: 85-102.

Laumonier, M., Karakas, O., Bachmann, O., Gaillard, F., Lukács, R., Seghedi, I., Menand, T., Harangi, S., 2019. Evidence for a persistent magma reservoir with large melt content beneath an apparently extinct volcano. *Earth and Planetary Science Letters* 521 (2019) 79–90, <https://doi.org/10.1016/j.epsl.2019.06.004>.

Leever, K., Mațenco, L., Bertotti, G., Cloetingh, S., Drijkoningen, K.G., and Vasiliev, I., 2006. Pliocene to Recent kinematics in the Focșani basin, Romania: New constraints from shallow seismic and paleomagnetic data, *Basin Res.*, 18, 521– 545, doi:10.1111/j.1365-2117.2006.00306.x.

Martin, M., the CALIXTO working group, 2003. High resolution teleseismic P-wave tomography for SE-Romania. *Geophysical Research Abstract* 5, 10512.

Mason, P.R.D., Seghedi, I., Szakacs, A., Downes, H., 1998. Magmatic constraints on geodynamic models of subduction in the Eastern Carpathians, Romania. *Tectonophysics* 297, 157– 176.

Mațenco, L., Bertotti, G., 2000. Tertiary tectonic evolution of the external East Carpathians (Romania), *Tectonophysics*, 316, 255-286.

Mațenco, L., Bertotti, G., Cloetingh, S., Dinu, C., 2003. Subsidence analysis and tectonic evolution of the external Carpathian–Moesian Platform region during Neogene times. *Sediment. Geol.* 156, 71–94.

Mațenco, L., Bertotti, G., Leever, K., Cloetingh, S., Schmid S., Tărăpoancă, M., Dinu, C., 2007. Large-scale deformation in a locked collisional boundary: interplay between subsidence and uplift, intraplate stress, and inherited lithospheric structure in the late stage of the SE Carpathians evolution *Tectonics*, 26, p. TC4011, 10.1029/2006TC001951.

Mațenco, L., Krézsek, C., Merten, S., Schmid, S., Cloetingh, S., Andriessen, P., 2010. Characteristics of collisional orogens with low topographic build-up: an example from the Carpathians. *Terra Nova* 22 (3), 155–165.

Maus, S., Barckhausen, U., Berkenbosch, H., Bournas, N., Brozena, J., Childers, V., F., Dostaler, F., Fairhead, J. D., Finn, C., von Frese, R. R. B., Gaina, C., Golynsky, S., Kucks, R., Lühr, H., Milligan, P., Mogren, S., Müller, R. D., Olesen, O., Pilkington M., Saltus, R., Schreckenberger, B., Thébaud, E., Caratori Tontini, F., 2009. EMAG2: A 2–arc min resolution Earth Magnetic Anomaly Grid compiled from satellite, airborne, and marine magnetic measurements *Geochem. Geophys. Geosyst.* vol. 10 pg. Q08005, doi:10.1029/2009GC002471.

- McCann, T., Krawczyk, C. M., 2001. The Trans-European Fault: a critical reassessment. - *Geological Magazine*, 138, 1, 19-29, DOI: 10.1017/S0016756801004915
- Molnár, K., Lukács, R., Dunkl, I., Schmitt, A.K., Kiss, B., Seghedi, I., Szepesi, J., Ha-rangi, S., 2019. Episodes of dormancy and eruption of the Late Pleistocene Ciomadul volcanic complex (Eastern Carpathians, Romania) constrained by zir-con geochronology. *J. Volcanol. Geotherm. Res.* 373, 133–147. <https://doi.org/10.1016/j.jvolgeores.2019.01.025>.
- Moretti, M., De Gori, P. and Chiarabba, C., 2009. Earthquake relocation and three-dimensional V_p and V_p/V_s models along the low angle Alto Tiberina Fault (Central Italy): evidence for fluid overpressure. *Geophysical Journal International*, 176(3), pp.833-846.
- Munteanu, M., Tatu, M., 2003. The East-Carpathian Crystalline-Mesozoic Zone (Romania): Paleozoic Amalgamation of Gondwana-and East European Craton-derived Terranes, *Gondwana Research Volume 6*, Issue 2, April 2003, Pages 185-196, [https://doi.org/10.1016/S1342-937X\(05\)70969-2](https://doi.org/10.1016/S1342-937X(05)70969-2).
- Narkiewicz, M., Maksym, A., Malinowski, M., Grad, M., Guterch, A., Petecki, Z., Probulski, J., Janik, T., Majdański, M., Środa, P., Czuba, W., Gaczyński, E., Jankowski L., 2015. Transcurrent nature of the Teisseyre–Tornquist Zone in Central Europe: results of the POLCRUST-01 deep reflection seismic profile. *Int. J. Earth Sci.*, 104 (3), pp. 775-796, <https://doi.org/10.1007/s00531-014-1116-4>.
- Neagoe, C., Popa, M., Diaconescu, M., Radulian, M., 2010. Possible deep lithospheric roots beneath South-Eastern Carpathians back-arc region, *Acta Geodaetica et Geophysica Hungarica* 45, 340–355.
- Neugebauer, J., Greiner, B., Appel, E., 2001. Kinematics of the Alpine–West Carpathian orogen and paleogeographic implications. *J. Geol. Soc. (Lond.)* 158, 97–110.
- Oncescu, M., Marza, V.I., Rizescu, M., Popa, M., 1999. The Romanian earthquake catalogue between 1984–1997. In: Wenzel, F. (Ed.), *Vrancea Earthquakes: Tectonics, Hazard and Risk Mitigation*. Springer, Netherlands, pp. 43–47.
- O'Reilly, S.Y., Griffin, W.L., Poudjom Djomani, Y., Morgan, P., 2001. Are lithospheres forever? Tracking changes in subcontinental lithospheric mantle through time. *GSA Today* 11, 4–9.
- Paige, C.C. and Saunders, M.A., 1982. LSQR: an algorithm for sparse linear equations and sparse least squares: *ACM Trans. Math. Soft.*, 8, 195-209.
- Panaiotu, C., Pécskay, Z., Hambach, U., Seghedi, I., Panaiotu, C.E., Itaya, T., Orleanu, M., Szakács, A., 2004. Short lived Quaternary volcanism in the Persani Mountains (Romania) revealed by combined K-Ar and paleomagnetic data, *Geologica Carpathica*, 55, 333-339.
- Panea, I., Stephenson, R., Knapp, C., Mocanu, V., Drijkoningen, G., Maţenco, L., Knapp J., and Prodehl, C., 2005. Near-vertical seismic reflection image using a novel acquisition technique across the Vrancea Zone and Foscani Basin, south-eastern Carpathians (Romania), In: Cloetingh, S., Maţenco, L. Bada, G. Dinu, C., and Mocanu, V. (eds.), *The Carpathians-Pannonian Basin System: Natural Laboratory for Coupled Lithospheric -Surface Processes*. Tectonophysics, 410 (1-4): 293-309.
- Patane, D., Chiarabba, C., Cocina, O., De Gori, P., Moretti, M., Boschi, E., 2002. Tomographic images and 3D earthquake locations of the seismic swarm preceding the 2001 Mt. Etna eruption: Evidence for a dyke intrusion. *Geophysical Research Letters* 29 (10).
- Patane, D., Barberi, G., Cocina, O., De Gori, P., Chiarabba, C., 2006. Time-resolved seismic tomography detects magma intrusions at Mount Etna. *Science* 313 (5788), 821–823.
- Petrescu, L., Bastow, I. D., Darbyshire, F. A., Gilligan, A., Bodin, T., Menke, W., and Levin V., 2016. Three billion years of crustal evolution in eastern Canada: Constraints from receiver functions, *J. Geophys. Res. Solid Earth*, 121, 788–811, doi:10.1002/2015JB012348.
- Petrescu, L., Stuart, G.W., Tataru, D., Grecu, B., 2019. Crustal structure of the Carpathian Orogen in Romania from receiver functions and ambient noise tomography: how craton collision, subduction and detachment affect the crust, *Geophysical Journal International*, ggz140, <https://doi.org/10.1093/gji/ggz140>.
- Petrescu, L., Borleanu, F., Radulian, M., Ismail-Zadeh, A., Maţenco, L., 2020b Earthquake mechanisms and tectonic stress in the Vrancea Seismic Zone: implications for crust-mantle coupling in intraplate earthquake nests, accepted for publication in *Tectonophysics*.

- Pharaoh, T. C., England, R. W., Verniers, J., and Zelainiewicz, A., 1997. Introduction: geological and geophysical studies in the Trans-European Suture Zone, *Geol. Mag.*, 134, 585–590.
- Pharaoh, T.C., Winchester, J.A., Verniers, J., Lassen, A., Seghedi, A., 2006. The Western Accretionary Margin of the East European Craton: an overview. *Mem. Geol. Soc. Lond.* 32 (1), 291–311.
- Pimienta, L., Schubnel, A., Violay, M., Fortin, J., Guéguen, Y., & Lyon-Caen, H. 2018. Anomalous Vp/Vs ratios at seismic frequencies might evidence highly damaged rocks in subduction zones. *Geophysical Research Letters*, 45, 12,210–12,217. <https://doi.org/10.1029/2018GL080132>.
- Popa, M., Radulian, M., Szakács, A., Seghedi, I., Zaharia, B., 2011. New seismic and tomography data in the Southern part of the Harghita Mountains (Romania, Southeastern Carpathians): connection with recent volcanic activity *Pure and Applied Geophysics*, 10.1007/s00024-011-0428-6.
- Radulescu, D., P., Cornea, I., Sandulescu, M., Constantinescu, P., Radulescu, F., Pompilian, A., 1976. Structure de la croûte terrestre Roumanie, *Essai d'interprétation des études sismiques profondes*. *Anu. Inst. Geol. Geofiz.* L(50), 5–36.
- Radulian, M., Mandrescu, N., Panza, G., Popescu, E., Utale, A., 2000. Characterization of Seismogenic Zones of Romania *Pure appl. geophys.* 157: 57, <https://doi.org/10.1007/PL00001100>.
- Radulian, M., Popescu, E., Bala, A., and Utale, A., 2002. Catalog of the fault plane solutions for the earthquakes occurred on the Romanian territory, *Romanian Journal of Physics*, 47, 663 - 670.
- Răileanu, V., Bala, A., Hauser, F., Prodehl, C., Fielitz, W., 2005. Crustal properties from S-wave and gravity data along a seismic refraction profile in Romania. *Tectonophysics* 410 (1-4), 251–272.
- Răileanu, V., Tataru, D., Grecu, B., Bala, A., 2012. Crustal models in Romania – II. Moldavian Platform and adjacent areas, *Rom. Journ. Phys.*, Vol. 57, Nos. 9–10, P. 1438–1454, Bucharest.
- Rawlinson, N., Fichtner, A., Sambridge, M. and Young, M.K., 2014. Seismic tomography and the assessment of uncertainty, *Adv. Geophysics*, 55, 1–76.
- Rawlinson, N., and Spakman, W., 2016. On the use of sensitivity tests in seismic tomography, *Geophysical Journal International*, 205, 1221–1243, <https://doi.org/10.1093/gji/ggw084>.
- Ren, Y., Stuart, G.W., Houseman, G.A., Dando, B., Ionescu, C., Hegedüs, E., Radovanović, S., Shen Y., South Carpathian Project Working Group, 2012. Upper mantle structures beneath the Carpathian–Pannonian region: implications for geodynamics of the continental collision, *Earth planet. Sci. Lett.*, vol. 349–350 (pg. 139-152).
- Ren, Y., Grecu, B., Stuart, G.W., Houseman, G.A., Hegedüs, E., South Carpathian Project Working Group, 2013. Crustal structure of the Carpathian–Pannonian region from ambient noise tomography, *Geophysical Journal International*, Volume 195, Issue 2, 1 November 2013, Pages 1351–1369, <https://doi.org/10.1093/gji/ggt316>.
- Reyners, M., Eberhart-Phillips, D., Stuart, G. and Nishimura, Y., 2006. Imaging subduction from the trench to 300 km depth beneath the central North Island, New Zealand, with Vp and Vp/Vs. *Geophysical Journal International*, 165(2), pp.565-583.
- Russo, R. M., Mocanu, V., Radulian, M., Popa, M., and Bonjer, K.-P., 2005. Seismic attenuation in the Carpathian bend zone and surroundings. *Earth Planet. Sci. Lett.*, 237:695–709.
- Săndulescu M., H. G. Krautner, I. Balintoni, D. Russo-Sandulescu M. Micu, 1981. The structure of the East Carpathians. *Guide Exc. B1. Carp.-Balk. Geol. Ass.*, XII Congr. Romania, 1-92, Bucharest.
- Săndulescu M., (1984). *Geotectonics of Romania* (in Romanian), Technical Publishing House, Bucharest, Romania
- Seghedi, I., Szakács, A., 1994. Upper Pliocene to Quaternary basaltic volcanism in the Perșani Mountains, Romania. *Rom. J Petrol* 76:101–107.
- Seghedi, I., Balintoni, I., Szakacs, A., 1998. Interplay of tectonics and Neogene post-collisional magmatism in the intra-Carpathian region, *Lithos* 45, 483 – 497.
- Seghedi, I., Mațenco, L., Downes, H., Mason, P.R.D., Szakács, A., Pécskay, Z., 2011. Tectonic significance of changes in post-subduction Pliocene–Quaternary magmatism in the south east part of the Carpathian–Pannonian Region. *Tectonophysics* 502 (1–2), 146–157.

Seghedi, I., Besutiu, L., Mirea, V., Zlagnean, L., Popa, R.G., Szakács, A., Atanasiu, L., Pomeran, M., Visan, M., 2019. Tectono-magmatic characteristics of post-collisional magmatism: Case study East Carpathians, Călimani Gurghiu-Harghita volcanic range, Physics of the Earth and Planetary Interiors Volume 293, August 2019, 106270, <https://doi.org/10.1016/j.pepi.2019.106270>.

Snyder, D., 2002. Lithospheric growth at margins of cratons, *Tectonophysics*, 355(1), 7–22.

Sperner, B., Lorenz, F., Bonjer, K., Hettel S., Müller B. and Wenzel F., 2001. Slab break-off – abrupt cut or gradual detachment? New insights from the Vrancea Region (SE Carpathians, Romania). *Terra Nova*, 13, 172–179.

Starostenko, V., Janik, T., Kolomiyets, K., Czuba, W., Sroda, P., Grad, M., Kovacs, I., Stephenson, R., Lysynchuk, D., Thybo H., Artemieva, I.M., Omelchenko, V., Gintov, O., Kutas, R., Gryn, D., Guterch, A., Hegedus, E., Komminaho, K., Legostaeva, O., Tiira, T. and Tolkunov, A., 2013. Seismic velocity model of the crust and upper mantle along profile PANCAKE across the Carpathians between the Pannonian Basin and the East European Craton. *Tectonophysics*, 608, 1049–1072.

Starostenko, V., Janik, T., Mocanu, V., Stephenson, R., Yegorova, T., Amashukeli, T., Czuba, W., Sroda, P., Murovskaya, A., Kolomiyets, K. and Lysynchuk, D., 2020. RomUkrSeis: Seismic model of the crust and upper mantle across the Eastern Carpathians—From the Apuseni Mountains to the Ukrainian Shield. *Tectonophysics*, 794, p.228620.

Stănică, M., Stănică, D., Furnica, C. 1999. The placement of the Trans-European Suture Zone on the Romanian territory by electromagnetic arguments. *Earth Planet Space*, 51, 1073–1078.

Stuart, G.W., Bastow, I.D., Ebinger, C.J., 2006. Crustal structure of the Northern Main Ethiopian rift from receiver function studies. *Geol. Soc. Spec. Publ.* 259 (1), 253–267. doi:10.1144/GSL.SP.2006.259.01.20.

Szackacs, A., Seghedi, I., 1995. Time-space evolution of Neogene–Quaternary volcanism in the Calimani–Gurghiu-Harghita volcanic chain, *Rom. J. Stratigr.* 76 (Suppl. 4) 24 pp.

Szakács, A., Seghedi, I., Pécskay, Z., Mirea, V., 2015. Eruptive history of a low-frequency and low-output rate Pleistocene volcano, Ciomadul, South Harghita Mts. *Bull. Volcanol.* 77 (12).

Tărăpoancă, M., Bertotti, G., Mațenco, L., Dinu, C., Cloetingh, S., 2003. Architecture of the Focsani Depression: a 13 km deep basin in the Carpathians Bend Zone (Romania) *Tectonics*, 22, p. 1074.

Tărăpoancă, M., Garcia-Castellanos, D., Bertotti, G., Mațenco, L., Cloetingh, S., Dinu, C., 2004. Role of the 3-D distributions of load and lithospheric strength in orogenic arcs: polystage subsidence in the Carpathians foredeep, *Earth planet. Sci. Lett.*, 221(1–4), 163–180.

Thompson, D.A., Bastow, I. D., Helffrich, G., Kendall, J.-M., Wookey, J., Snyder, D.B., and Eaton D.W., 2010. Precambrian crustal evolution: Seismic constraints from the Canadian Shield, *Earth Planet. Sci. Lett.*, 297, 655–666, doi: 10.1016/j.epsl.2010.07.021.

Toomey, D.R., Foulger, G.R., 1989. Tomographic inversion of local earthquake data from the Hengill–Grensdalur central volcano complex, Iceland. *Journal of Geophysical Research* 94 (B12), 17497–17510.

Tondi, R., Achauer, U., Landes, M., Davì, R., Besutiu, L., 2009. Unveiling seismic and density structure beneath the Vrancea seismogenic zone, Romania. *J. Geophys. Res.* 114, B11307, <http://dx.doi.org/10.1029/2008JB00592>

Tryggvason, A., Rognvaldsson, S.T., Flovenz, O.G., 2002. Three- dimensional imaging of the P- and S-wave velocity structure and earthquake locations beneath Southwest Iceland. *Geophysical Journal International* 151 (3), 848–866.

Um, J., Thurber, C. H., 1987. Fast algorithm for two-point seismic ray tracing, *Bulletin of the Seismological Society of America*, vol. 77, no. 3, pp. 972–986, 1987

Ustaszewski, K., Schmid, S., M., Fügenschuh, B., Tischler, M., Kissling, E., Spakman, W., 2008. A map-view restoration of the Alpine–Carpathian–Dinaridic system for the Early Miocene. *Swiss J Geosci* 101(1):273–294.

Von Seggern, D. H., 2009. Direct empirical estimation of arrival-time picking error from waveform cross-correlations, *Bull. Seismol. Soc. Am.*, 99 (5), 2749–2758, doi:10.1785/0120080304.

Waldhauser, F., Ellsworth W. L., 2000. A double-difference earthquake location algorithm: Method and application to the northern Hayward Fault, California, *Bull. Seismol. Soc. Am.*, 90, 1353–1368.

- Wang, X.Q., Schubnel, A., Fortin, J., David, E.C., Guéguen, Y. and Ge, H.K., 2012. High Vp/Vs ratio: Saturate cracks or anisotropy effects? *Geophysical Research Letters*, 39(11).
- Winchester, J. A., Pharaoh, T.C., Verniers, J., 2002. Palaeozoic Amalgamation of Central Europe, Geological Society, London, Special Publication, vol. 201, pp. 1-353.
- Wenzel, F., Lorentz, F.P., Sperner, B., Oncescu, M.C., 1998. Seismotectonics of Romanian Vrancea area. In: Wenzel F., Lungu D., (Eds), *Vrancea Earthquakes*, Kluwer Academic Publishers, Dordrecht, Netherlands.
- Wessel, P., & Smith, W. H. F., 1998. New, improved version of Generic Mapping Tools released. *Eos, Transactions American Geophysical Union*, 79(47), 579. <https://doi.org/10.1029/98EO00426>.
- Wortel, M.J.R., and Spakman, W., 2000. Subduction and slab detachment in the Mediterranean-Carpathian region, *Science*, 290:1910-1917.
- Xiao-Yin, T., Gong-Cheng, Z., Jian-She, L., Shu-Chun, Y., Song, R., Sheng-Biao, H., 2014. Modelling of thermal effects of igneous intrusions on the temperature field and organic maturity in the Changchang Sag, Qiongdongnan Basin, SouthChina Sea. *Chin. J. Geophys.* 57 (2), 219-229. <http://dx.doi.org/10.1002/cjg2.20098>.
- Yukutake, Y., Honda, R., Harada, M., Arai, R. and Matsubara, M., 2015. A magma-hydrothermal system beneath Hakone volcano, central Japan, revealed by highly resolved velocity structures. *Journal of Geophysical Research: Solid Earth*, 120(5), pp.3293-3308.
- Zhang, H., Thurber C. H., (2003), Double-difference tomography: The method and its application to the Hayward fault, California, *Bull. Seismol. Soc. Am.*, 93, 1875–1889.
- Zhang, Q. and Lin, G., 2014. Three-dimensional Vp and Vp/Vs models in the Coso geothermal area, California: Seismic characterization of the magmatic system. *Journal of Geophysical Research: Solid Earth*, 119(6), pp.4907-4922.
- Zielhuis, A., and G. Nolet, 1994. Deep seismic expression of an ancient plate boundary in Europe, *Science*, 265 (5168), 79–81, doi:10.1126/science.265.5168.79.
- Zimmer, M., Prasad, M. and Mavko, G., 2002. Pressure and porosity influences on VP–VS ratio in unconsolidated sands. *The Leading Edge*, 21(2), pp.178-183.

Supporting Information for "The missing craton edge: crustal structure of the East European Craton beneath the Carpathian Orogen revealed by double-difference tomography"

Borleanu F.^(1,2), Petrescu L.⁽²⁾, Enescu B.^(1,3), Popa M.^(2,3), Radulian M.^(2,3)

1) Department of Geophysics, Kyoto University, Japan; 2) National Institute for Earth Physics, Romania; 3) Academy of Romanian Scientists, Romania

Contents of this file:

Figures S1 to S5

The supplemental material comprises alternate plots in order to enhance different manuscript sections.

Figure S1 is to support the Methods and data section of the manuscript, to better illustrate our choice for damping and smoothing.

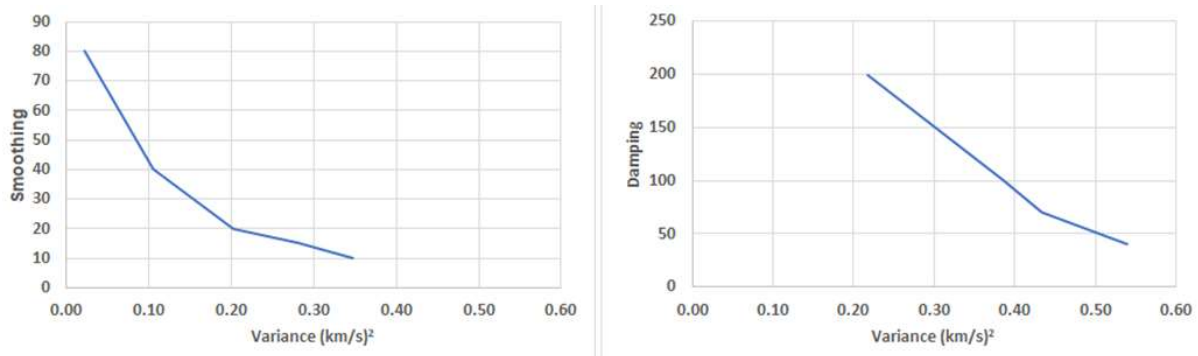


Figure S1. Trade-off curves between final model variance and smoothing (left) and damping (right).

Figure S2 is to support the Resolution and uncertainty section of the manuscript, to display the rays derivative weight sum (DWS) distribution for each depth.

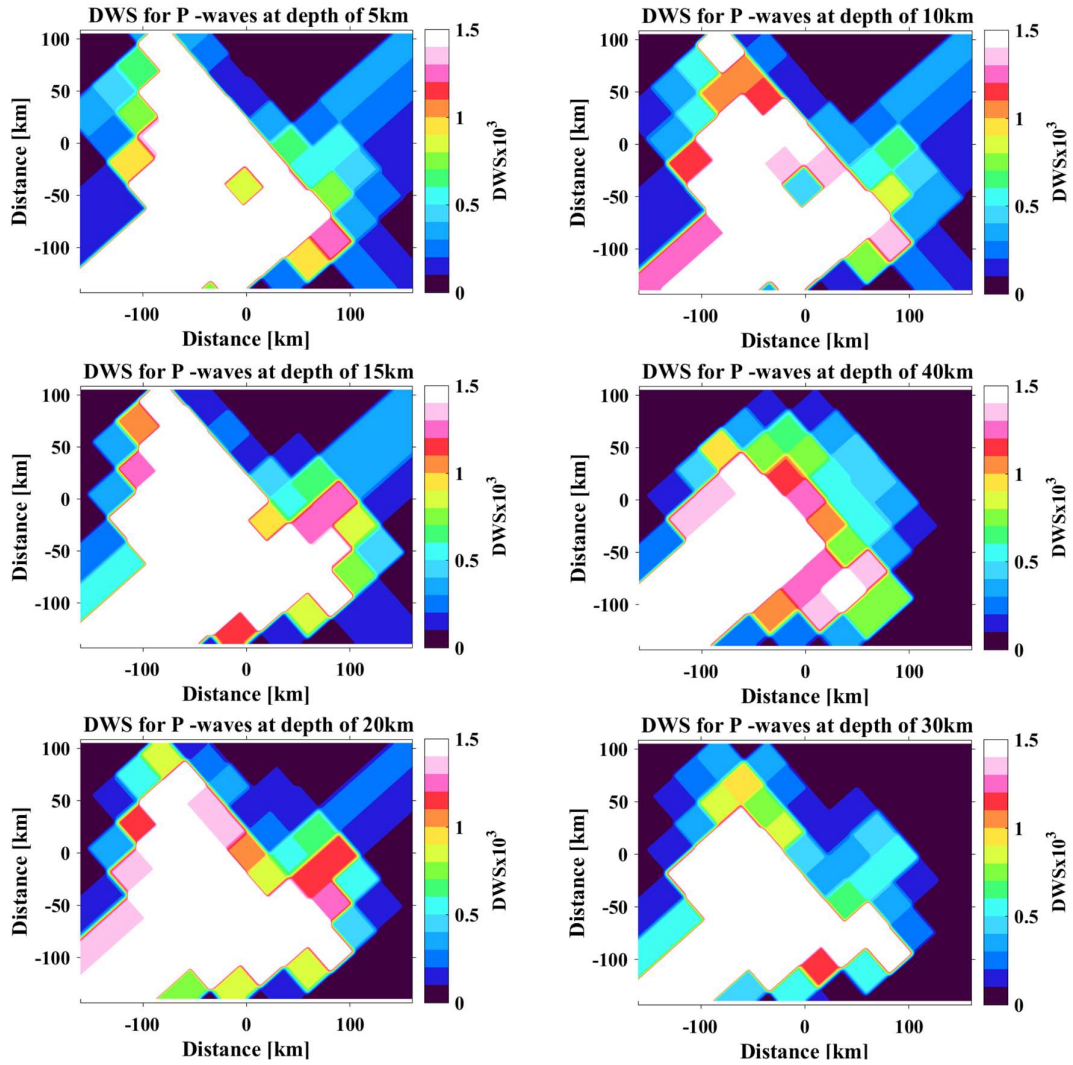


Figure S2. Horizontal sections showing the P-waves rays derivative weight sum (DWS) distribution computed for each depth.

Figure S3 is to support the Resolution and uncertainty section of the manuscript, showing results of the P-waves checkerboard test also for cross-sections.

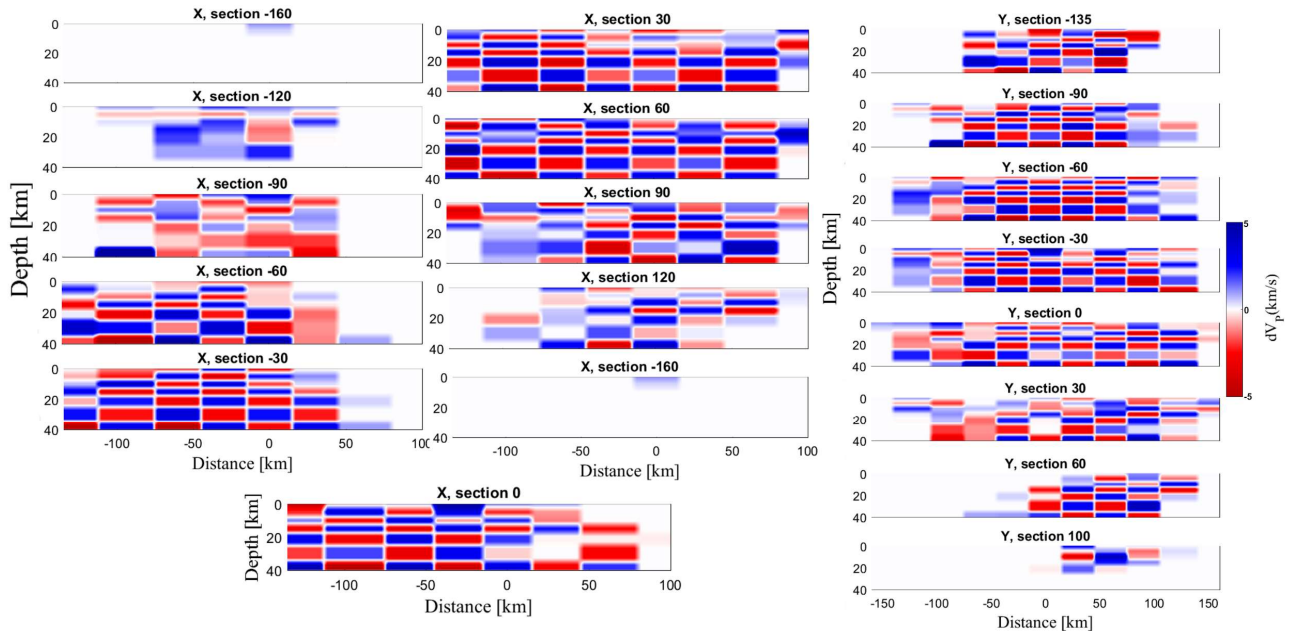


Figure S3 Results of the P -waves checkerboard test for the vertical cross-sections.

Figure S4 is to support the Resolution and uncertainty section of the manuscript, to assess the influence of parametrization.

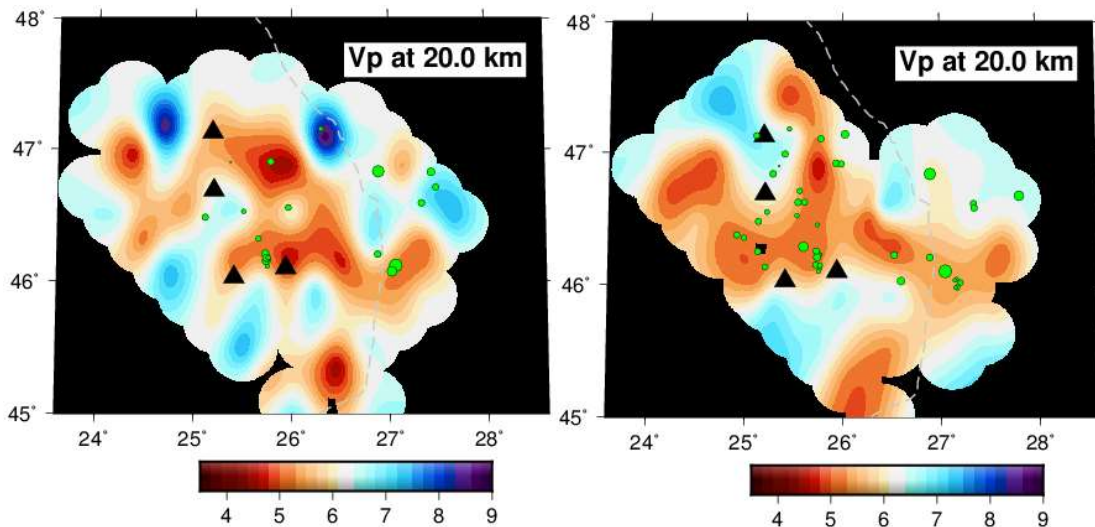


Figure S4. Horizontal sections of P-wave velocity model at 20 km depth for the initial grid (left) and for 15 km shifted grid towards NE (right)

Figure S5 is to support the Resolution and uncertainty section of the manuscript, to assess the reliability of the low-velocity anomaly.

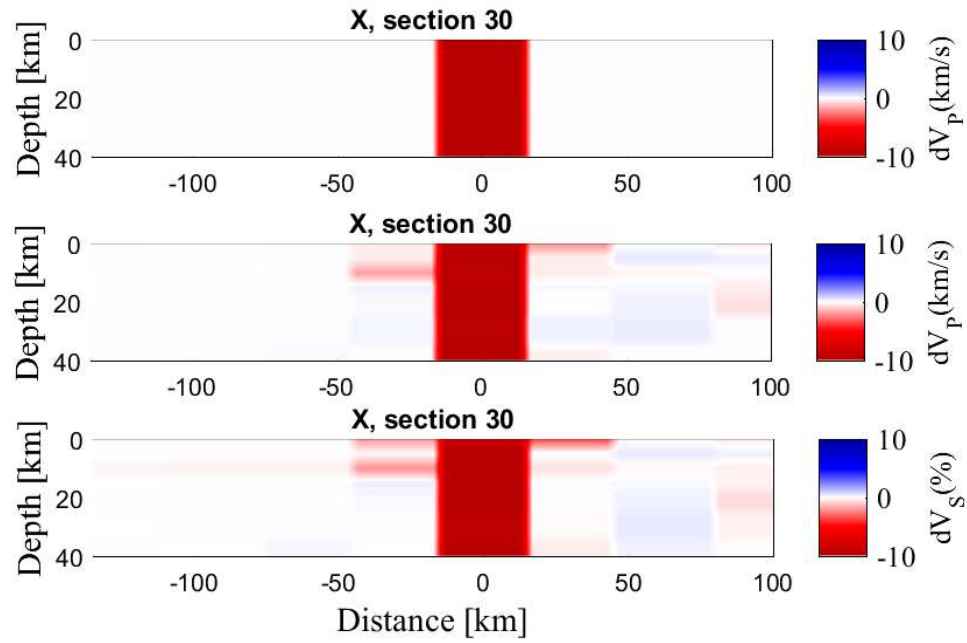


Figure S5. Cross-sections showing how a low-velocity anomaly (10%, top of the figure) is recovered by both P (middle in the figure below) and S (bottom in the figure below) velocity structures