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

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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), 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, thrusted 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, Koulaiov 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 Palaeozoic-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 (Ciulevu 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 (Ball, 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, Panaïotu 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 (Mw < 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 Vp/Vs 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< Lat. (°N) < 47.50; 24.50 < Lon. (°E) < 28.00; 0.3 < M, < 5.7; 0 < h (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 ≥ 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) Vp~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 Vp (<4.5 km/s) follows the Carpathian front and the inner Carpathians side, although along-strike variations exist. Higher velocity patches (δVp = 8% with respect to the regional average) emerge in the EEC below ~10km depth, where presumably Precambrian crystalline basement is present. Throughout the upper crustal layer, the lowest velocities (Vp<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 (Vp>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 underthrusting westward dipping EEC crust (Matenco 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 (Vp~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 (Vp>6.5 km/s at 30 km depth, *Figure 5*).

![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 Vp and Vs) 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.](image-url)
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 ($V_s \approx 4$ 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 (Vs~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 (Vs<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 Vp 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 Vp/Vs ratios are indicated by letters: A, B, C, D, E, F, G, H and E while high Vp/Vs 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 Vp 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 1km 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 (Vp<5.4km/s; Vs~3.3km/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, Vp/Vs 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 Vs. Generally, low Vs, high Vp/Vs 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 Vp/Vs 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-Camera Faults (Figure 7). In the top 10 km, seismicity along the Trotus-Barlad-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, Matenco 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 Vp and the ~3.5 km/s contour for Vs. 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, Matenco 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 Vp (>6 km/s in the top 20 km) and thick (Moho~44km) crust from a lower Vp (<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 Vp > 6.3 km/s and Vs > 3.5 km/s in the eastern footwall, presumably the dipping Precambrian basement, to Vp < 6 km/s beneath the Carpathian Orogen to the west (Figures 5 and 8). This gently dipping seismic velocity transition is consistent in both Vp and Vs 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 Vp and Vs 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 Vp>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 Vp signatures (≤ 5.5 km/s) and are associated with high Vp/Vs 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 (Baffour 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 (Mw < 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 (Mw < 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, Karatson 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 3He/4He anomaly measured in natural gases, thermal and CO2-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 Vp and moderate Vs associated with low Vp/Vs (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-thrust 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 Vp > 6.5 km/s to the west in a region where Vp < 6 km/s. The transition is characterized also by a downward decrease in the Vp/Vs 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 Vp/Vs ratios up to 1.9, consistent with the presence of mafic material or partial melts.

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References


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

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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.