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Nature of the Vrancea seismic zone (Eastern Carpathians) – New constraints from dispersion of first-arriving P-waves

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ABSTRACT

The Vrancea region of the southeastern Carpathians is one of the most active seismic zones in Europe and it is well-known for its strong intermediate depth earthquakes. Seismic tomography had revealed a high-velocity body beneath Vrancea and the Moesian platform that extends to a depth of at least 350 km and can be interpreted as descending lithosphere. The strong earthquakes occur within the northeastern part of this high-velocity body, in a very limited seismogenic volume at intermediate depth (70-180 km). Several geodynamic models have been proposed for this area. They can be split into two main categories, in terms of the nature of the high-velocity anomaly, which may (a) be associated with descending relic oceanic lithosphere beneath the bending zone of the SE-Carpathians, either attached or already detached from the continental crust; or (b) it may represent continental lithosphere that has been delaminated, after continental collision and orogenic thickening. Based on currently available information, it appears difficult to distinguish between these two types of models. In this paper we attempt to shed more light on the nature of the seismic anomaly, as well as that of the origin of the intermediate depth seismicity in the Vrancea zone, by investigating the waveform character of P-waves excited by local earthquakes beneath this area, and in particular the dependence of group arrival times on frequency. We present observations of such a dispersion from stations situated at the bending zone of the SE-Carpathians. On the other hand, signals from the same earthquakes, but observed at reference stations outside of the anomalous zone do not show that frequency dependence. A natural explanation for these observations is that it is caused by the presence of a low-velocity channel at the top of the seismic anomaly, which is too thin to be resolved by classical seismic tomographic techniques. Similar observations of dispersed first-arriving P-waves have been made above subduction zones around the world, in which low-velocity layers with a thickness of several kilometers are known to exist. This suggests that a tabular slab of subducted oceanic crust is present within the seismic anomaly under the Vrancea region, and that the anomaly consists of subducted oceanic lithosphere rather than continental lithosphere, at least at depths shallower than the seismically active zone.

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1. Introduction

1.1. SE Carpathians – geology

The Carpathians (e.g. Fig. 1) were formed as a consequence of Mesozoic and Cenozoic closure of the Alpine Tethys during continental collision of the Eurasian and African plates, which has led to a lateral extrusion of the Tisza–Dacia block. This extrusion has led to an fan-shaped migration of the Carpathian collision front, accommodated by subduction of a last remnant of the Tethys Ocean (Csontos, 1995; Sperner et al., 2002; Stampfli and Borel, 2002). The Carpathian tectonic history might therefore be seen in association with a retreating subduction zone (e.g., Royden, 1988).

The subduction apparently ceased first in the northern part of the Eastern Carpathians, when the (thick and buoyant) lithosphere of the Eastern European craton came into contact with the subduction zone. Subduction in the southern and southeastern part continued to be active, until much more recently (Linzer, 1996). There is a systematic decrease in age of subduction-related volcanism from NW to SE and S, until 0.3 Ma ago, with changing chemical composition, perhaps related with asthenospheric upwelling (Seghedi et al., 2011). It is not fully clear however, when the subduction stopped (e.g., Ismail-Zadeh et al., 2012).

Compared with the Alps, where "hard collision" features persist, such as thickened crust of up to 50 km thick, high topography, and considerable uplift leading to erosion and denudation, the Eastern Carpathians rather show "soft collision" features, with a moderately thick crust, moderate topography, and not much denudation. The Vrancea region, in the bending zone of the Eastern Carpathians, is bounded by the Transylvanian and Focsani basins in the NW

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Fig. 1. Geologic map of Romania, emphasizing the position of mid-Cretaceous suture zone in the Transylvanian basin, the late Tertiary volcanic arc (red) and thrust nappes of Eastern Carpathians developed in Cretaceous to Miocene strata (green, orange, and yellow units) deposited on the East European/Moesian continental plates (after Săndulescu et al., 1978). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

and SE respectively, the Trotus and Peceneaga–Camena fault system to the NE and the Intramoesian fault to the S. Vrancea is the youngest part of the Carpathian subduction/collision process.

1.2. Seismicity

Vrancea is the site of strong intermediate-depth seismicity, down to 220 km depth. Different from the scattered crustal seismicity, most of the intermediate-depth earthquakes occur in a very limited seismogenic volume with about 30×70 km lateral extent and in a depth range from 70-180 km (Fig. 2). This zone is offset from the surface trace of the suture, to the southeast (Girbacea and Frisch, 1998). Seismic tomography has suggested that the earthquakes occur within a high-velocity anomaly, and that the seismogenic volume is thus part of the subducted lithosphere beneath the Carpathian bending zone (Wortel and Spakman, 2000; Martin et al., 2006). The P1 profile in Fig. 2 shows the almost vertical, fingershaped pattern of the intermediate-depth seismicity dipping slightly towards NW. The seismic moment rate of the intermediate-depth seismicity within the restricted seismic volume is about 0.8×10^{19} Nm/yr, which is comparable to that of southern California (Wenzel et al., 1998). The bending zone of the southeastern Carpathians thus constitutes one of the most active seismic zones in Europe, where intermediate-depth earthquakes of magnitude in excess of $M_w = 7.5$ occur with relative frequency

in a geographically restricted area (e.g., Wenzel et al., 2002). For centuries, these seismic events have resulted in a high toll of human casualties and property damage, making Bucharest one of the most threatened among the large population centers in Europe. Seismicity is recorded by the Romanian national seismic network (e.g., Neagoe and Ionescu, 2009), as well as temporary seismic networks in the area (e.g., Wenzel et al., 1998).

1.3. Geodynamic models

The nature of the seismogenic body beneath Vrancea is still subject of debate. Seismic tomography shows a nearly vertical high-velocity body beneath Vrancea and the Moesian platform that reaches a depth of about 350 km (e.g., Wortel and Spakman, 2000; Martin et al., 2006; Koulakov et al., 2010). This anomaly is colocated with the distribution of intermediate-depth seismicity and is interpreted as cold and dense material that is descending into the deeper mantle. However, when trying to explain this process the major question is whether the descending material represents subducted oceanic lithosphere (attached or already detached) or delaminated continental lithosphere (e.g., Ismail-Zadeh et al., 2012). Intermediate-depth earthquakes in the southeastern Carpathians have been studied (e.g., Fuchs et al., 1979), and attributed to oceanic lithosphere sinking steeply into the mantle. Wortel and Spakman (2000) and Sperner et al. (2001)



Fig. 2. Earthquake distribution in the southeastern Carpathians. (a) Shows the seismicity map, where events >70 km depth are represented by red circles, and shallower events by empty circles. Vertical cross-sections show projections of the hypocenter distribution, for the NW-SE profile P1 (b), and the SW-NE profile P2 (c). Only events with M > 2 are shown, from the NEIC event database for 1980–2013. Yellow symbols give positions of seismological stations used in this study. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

have proposed models of oceanic slab detachment with the break-off point migrating to the SE towards Vrancea, where it has now reached the final break-off stage, explaining the intermediate-depth seismicity and the history of volcanism in the East-Carpathian hinterland. Other oceanic-type models are the lateral migration of an oceanic slab (Girbacea and Frisch, 1998; Gvirtzman, 2002), subduction and lateral tearing of a slab (Martin et al., 2006; Wenzel et al., 1998; Wortel and Spakman, 2000). Knapp et al. (2005) believe that the Vrancea zone occupies a region overlain by continental crust and upper mantle, and accordingly suggest active delamination of continental lithosphere to explain the high-velocity body under Vrancea.

Gravitational instability has been proposed by Lorinczi and Houseman (2009), who show that the strain-rate profile in the seismogenic volume is consistent with predictions from numerical experiments for a downwelling of continental lithosphere rather than subducted oceanic lithosphere. Their model simulates mantle downwelling due to Rayleigh–Taylor instability that may explain the strain and stress localization which ultimately is the cause of the intermediate-depth seismicity beneath Vrancea. Using finitefrequency tomography of teleseismic P relative arrival time residuals, Ren et al. (2012) have obtained a high-resolution P-wave velocity model of the upper mantle beneath the Carpathian–Pannonian Region. They found that the Vrancea structure is broadly consistent with models based on either delamination of mantle lithosphere or lithospheric gravitational instability.

In this paper we want to shed more light on the question whether the observed high-velocity anomaly underneath Vrancea represents subducted lithosphere (continental or oceanic) or rather lithospheric delamination. This question has been identified as one of the key questions for the area (e.g., Ismail-Zadeh et al., 2012). The two types of heterogeneities would produce very similar tomographic images, and they can therefore not be distinguished by classical seismic tomography. We thus use an entirely different approach, based on the dispersion of body waves from local intermediate-depth earthquakes that travel nearly vertically, along the heterogeneity.

2. Method and observations

2.1. Structure of subducted oceanic lithosphere

Oceanic lithosphere comprises the mantle lid and the oceanic crust. The mantle lid consists of peridotites. Its thickness is controlled by conductive cooling. It therefore thickens with age. The crustal layer, on the other hand, is composed of a characteristic sequence of layers, where the topmost part (Layer 1) is composed of sediments that also thicken with age, due to deposition of sedimentary material in the ocean. The sediments are underlain by pillow lava and dikes (Layer 2), and the deepest part of the crust (Layer 3) consists of gabbros and cumulate ultramafic rocks. The oceanic crust is generally less than 10 km thick, with an average thickness of 6 km. In subduction zones, oceanic lithosphere descends into the deeper mantle. While a portion of the sediments may be scraped off during the convergence, forming accretionary wedges, much of the crust and mantle layering is most likely conserved during subduction. It therefore remains seismically distinct from mantle rocks until transformed at greater depth (Martin and Rietbrock, 2006), where basalt and gabbro from the former oceanic crust are eventually transformed to eclogite. This transformation may be complete only at depths exceeding 100 km (e.g., Abers, 2005). Once that process is completed, the crustal portion may then become difficult to distinguish from the surrounding mantle material.

2.2. Our technique

On top of subducting oceanic slabs we can expect a low velocity layer (LVL) that is generally less than 10 km thick, and which might carry guided waves, analog to surface waves that propagate within the low velocity layer at the Earth's surface. Tests with



Fig. 3. Wave propagation in a synthetic subduction zone that contains a low-velocity zone (after Martin et al., 2003). The two wavefield snapshots show the late arrival propagating upwards along the slab; (b) shows later time than (a). The seismic source (indicated by a star) is located within the slab, but below the low-velocity zone (see text).

full waveform synthetics (e.g., Martin et al., 2006) show that these waveguides indeed carry seismic energy at high frequency, which effectively produces a dispersion for the first-arriving body waves, as long as the raypath geometry is favorable, such that sources lie inside or close to the waveguides, and receivers are at the surface, in a position roughly in continuation of the slab structure. Fig. 3 shows a synthetic calculation of waves propagating along a slab that contains a low-velocity layer. Note the late-arriving highfrequency arrival that is propagating along the slab. The strength of dispersion depends mainly on the travel distance of the waves within the LVL, as well as its thickness (Gubbins and Snieder, 1991; Abers, 2005). That dispersion is usually in the sense that high frequencies are delayed by 0.5 to 1 s relative to lower frequencies in the frequency range between 0.5 and 4 Hz (Abers, 2005; Bokelmann and Maufroy, 2007; Martin et al., 2003). The illustration in Fig. 3 shows that high-frequent energy can be trapped in a low-velocity crustal channel also if the source is located just below the low-velocity zone. Bending of the subduction zone facilitates part of the trapped energy to escape from the slab, and to be observed at the Earth's surface.

Incidentally, subduction zones are regions of intense earthquake activity, from the surface down to several hundreds of kilometers. The earthquakes within the slabs are thus ideal sources to observe the dispersion phenomenon at receivers located near the projection of the slab toward the surface. This dispersion phenomenon allows testing for the presence of subducted oceanic lithosphere, if (part of) the crustal layer is still preserved at depth, and if it has not undergone major phase transformations. The transformation to eclogite increases P-velocities substantially, and it may render seismic velocities indistinguishable from typical mantle material. This phase transitions occurs progressively, depending on pressure and temperature. Most of the crustal material is eventually transformed to eclogite (Hacker, 1996), and the possibility for testing the nature of seismic anomalies in the mantle will therefore vanish at deeper depths in the mantle.

2.3. Data from the CALIXTO experiment

In 1999 the international CALIXTO field experiment (Carpathian Arc Lithosphere X-Tomography) was conducted from May to October 1999 in SE Romania (e.g., Wenzel et al., 1998). The CALIXTO temporary seismic network consisted of 143 broadband seismic stations. We examine waveforms of direct P-waves for local intermediate-depth earthquakes with M > 2 that have occurred within the operating time of the network, recorded at the stations situated in the area above the supposed slab (Fig. 2). Among all the recorded events, we focused on signals with a reasonable noise level and a clearly visible P-wave onset. That yields 61 events within the latitude-longitude box 45.2°-45.81°N and 26.13°–26.86°E that cover a depth range from 70 to 160 km. Direct P-wave arrival times at specific stations were compared to travel times calculated for a one-dimensional Earth model (Crotwell et al., 1999). In order to enhance the signal-to-noise ratio for the pick of the first arrival P-waves, we applied a band-pass filter from 2 to 10 Hz to the vertical-component seismograms. To measure dispersion we characterize the dependence of group velocity arrival times for band-pass-filtered wave packets on frequency. Verticalcomponent seismograms are band-pass-filtered in 1 Hz intervals between 0.5-8.5 Hz. The narrow-band records are squared and averaged with a boxcar smoothing operator with a half-width of 0.5 Hz in a procedure following Abers and Sarker (1996) and Bokelmann and Maufroy (2007).

3. Results

P-wave dispersion is quantified by examining the group arrival time variations with frequency. The time of the maximum amplitude in these smoothed envelopes is picked as an estimate of group arrival time for dispersive waves. Fig. 4 shows seismograms and their envelopes, for two events recorded at stations A10 and B05B in the center of the region, respectively, and at a reference station B02 further northwest (for station locations see Fig. 6). Both stations in the center show high frequencies delayed with respect to the low frequencies, by up to one second, between 0.5 and 7.5 Hz. The same two events recorded at the reference station B02, have all frequencies arriving at the same time, and thus "no dispersion". This clearly indicates that the dispersion phenomenon which we will discuss in more detail in the following cannot be due to source characteristics of the earthquakes, but must be associated of properties of the wave propagation path, which we will explain below.

We have inspected 64 out of 109 events that occurred between June and October 1999 in the Vrancea region. Nineteen of these events showed low-frequency energy arriving prior to higher frequencies (see Table 1). This phenomenon was observed only at stations in the center, where rays propagate steeply upwards, to the seven stations A08, A09, A10, B05A/B, E05, E06, E22. Seven of these eight stations are located only few kilometers northwest of the epicenters. For the same events, observed at stations further northwest or to the south-east, no frequency dependence was seen (Fig. 5). At the seven stations, a high proportion of observed events showed dispersion: A10 (37%), E06 (33%), E05 (33%), B05A (20%), A08 (18%) and B05B (17%).

To make the argument more complete, we compare in more detail station pairs, where we observe the same event. Fig. 6 shows the ray paths for the subset of observation that belong to pairs, where dispersion is observed at one station, but not at another.



Fig. 4. Waveform examples from two earthquakes. Each of the subfigures (a)–(d) shows a filtered vertical-component seismogram on the right, and the corresponding smoothed envelopes on the left. Both are broken into 8 frequency windows, from top to bottom at 0.5–1.5 Hz, 1.5–2.5 Hz, 2.5–3.5 Hz, 3.5–4.5 Hz, 4.5–5.5 Hz, 6.5–7.5 Hz, and 7.5–8.5 Hz. The times intervals are 1 s. Lines illustrate the observed dispersion, in red, where we conclude dispersion to be present, in blue, if not. (a) Shows event 31.8.1999 02:10:36 arriving at the central station A10 (for station location see Fig. 5); (c) shows event 27.7.1999 14:00:02 GTM arriving a central station B05B; (b) and (d) show the same two events recorded at the reference station B02. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Table 1

Earthquake observations showing dispersion. Date = Day of earthquake; event time in hrs:min:s (GMT); M_w = magnitude; Lon = longitude; Lat = latitude; Depth in km; Station name; Station Lon = station longitude; Station Lat = station latitude; Dt = dispersion time delay 8 Hz-1 Hz (in seconds).

Date	Event time	Mw	Lon	Lat	Depth (km)	Station name	Station Lon	Station Lat	Dt
	113.1111.3 (GWT)				(KIII)				(3)
1999-07-31	11:56:49	2.8	26.41	45.47	129.8	A08	26.572	45.952	0.6
1999-06-03	12:16:25	3.2	26.47	45.59	112.6	A08	26.572	45.952	1.0
1999-06-13	00:20:01	2.9	26.52	45.45	112.5	A08	26.572	45.952	0.6
1999-07-04	08:21:01	3.7	26.27	45.33	141.1	A08	26.572	45.952	0.8
1999-07-09	00:07:41	3.3	26.39	45.54	129.1	A10	26.271	45.760	0.6
1999-07-13	13:10:57	4.0	26.49	45.70	131.9	A08	26.572	45.952	0.6
1999-07-15	07:36:23	3.7	26.49	45.58	135.1	A08	26.572	45.952	0.7
1999-07-15	02:52:03	3.5	26.85	45.74	115.6	A10	26.271	45.760	1.0
1999-07-15	02:52:03	3.5	26.85	45.74	115.6	E06	26.875	45.989	0.5
1999-07-15	07:36:23	3.7	26.49	45.58	135.1	B05B	26.572	45.952	0.5
1999-07-22	10:07:50	3.7	26.33	45.45	136.7	E22	26.867	45.562	0.4
1999-07-27	14:00:20	3.2	26.57	45.61	135.9	B05B	26.125	45.609	0.8
1999-08-01	05:56:49	3.5	26.60	45.50	124.1	E05	26.656	46.000	0.9
1999-08-09	07:16:16	3.7	26.71	45.72	131.2	B05B	26.125	45.609	1.2
1999-08-19	01:55:49	3.0	26.65	45.69	120.8	A08	26.572	45.952	0.9
1999-08-27	21:53:19	3.4	26.53	45.67	153.4	A08	26.572	45.952	0.5
1999-08-31	02:11:12	3.4	26.49	45.61	140.3	A10	26.271	45.760	0.6
1999-09-10	03:12:30	2.8	26.59	45.67	115.2	B05B	26.125	45.609	1.0
1999-09-23	16:51:39	3.0	26.43	45.53	141.3	A08	26.572	45.952	0.5
1999-10-12	23:48:33	3.7	26.45	45.63	149.7	B05B	26.125	45.609	0.5
1999-10-16	09:33:18	3.4	26.45	45.52	130.4	A08	26.572	45.952	0.9

We observe the same features again. Dispersion is observed only for rays leaving the earthquakes on nearly vertical paths toward the surface, while there is a much wider range of departure angles, for which no dispersion is observed. Note that all of these events are showing dispersion on at least one station (on near-vertical paths).

All of this taken together suggests the presence of a waveguide above the earthquakes, along the ray paths shown in red in Fig. 6. We will discuss below, whether there are other wave propagation phenomena that might also lead to such a dispersion.

4. Discussion and interpretation

We have found dispersion for rays that travel nearly vertically, roughly through the supposed position of the slab under Vrancea (Fig. 7). All dispersion observations correspond to high frequencies, at 8 Hz, being delayed relative to 0.5 Hz by an average of 0.7 s in the sense of "normal dispersion". A similar effect had been observed at subduction zones around the world where a thin low-velocity layer on top of the slab acts as a waveguide for high frequencies but is too thin to be "recognized" by long wavelengths



Fig. 5. Distribution of events (filled circles) and stations used in this study (triangles). Events for which dispersion was observed are indicated by red color, others by blue color. Stations at which dispersion was observed are indicated by yellow filled triangles, others by empty triangles (see text). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)



Fig. 6. Map view (a) and vertical cross-sections of ray paths for the two profiles from P1 (b) and P2 (c) shown in Fig. 2. Event-station rays, where dispersion was observed are shown by red lines, while others (no dispersion) are shown by blue lines. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

(Abers, 2005; Bokelmann et al., 2011; Martin et al., 2003). The size of the observed dispersion in our study is between 0.5 and 1.3 s, with a mode (most frequent value) of 0.5 s. Since the crust

has a thickness in the area of 30–35 km (Diehl and Ritter, 2005; Ivan, 2011), and the events occur at 120 km depth or deeper, this leaves about 90 km of slab length, which might explain the



Fig. 7. Ray paths as in Fig. 6(a) (red for dispersion, blue for "no dispersion"), shown together with a tomographic model (after Koulakov et al., 2010). Note that the location of the rays that produce dispersion coincides with the position of high-velocity material (see text). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

observed effect following the arguments in Abers (2005; see also Bokelmann and Maufroy, 2007).

4.1. Are there alternative explanations for the dispersion?

In order to act as a waveguide, a low-velocity-layer must exhibit several properties: (1) the sources must lie within or very close (1-2 times the LVL thickness from its center) to the LVL, (2) propagation of trapped waves may possibly be disrupted by large jogs or heterogeneities within the LVL if their dimension exceeds the waveguide width, (3) details in source location or attenuation can affect the spectral content in ways that add uncertainty to layer thickness and (4) the boundary between high and low velocities must be relatively sharp in order to trap high-frequency energy. Gradual boundaries such as temperature variations are therefore unlikely to produce such a LVL that acts as a waveguide (Abers, 2000). Source effects are not likely to explain the observations, since for the same event dispersion is seen on central stations but not on other stations. Scattering and multipathing are not likely to explain the primary observation (e.g., Bokelmann and Maufroy, 2007). Attenuation would rather cause a weak dispersion effect of opposite sign. Scattering might explain why we sometimes observe the dispersion, and sometimes not, even though the paths are nearby - by waveform complexity (Abers and Sarker, 1996). Attenuation itself cannot explain the observed dispersion. The weak dispersion associated with attenuation rather acts in the opposite sense. Yet scattering is unlikely to conspire to cause the clear spatial pattern of dispersion itself.

We have shown that the observed dispersion is consistent with the presence of a subduction zone composed of oceanic lithosphere under the Eastern Carpathians. However, can continental material descending into the mantle also produce a dispersion such as the observed one? This is quite unlikely, both for geometrical reasons as well as the strong buoyancy of continental crust. Furthermore, the continental crustal material would need to maintain a tabular shape, similar to what an oceanic subduction zone produces naturally. It is therefore plausible to exclude descending continental material as a cause, even though this is not entirely impossible.

4.2. Which observations can constrain the nature of the anomaly in the upper mantle?

Geophysics does not offer many direct constraints on whether the lithosphere has oceanic or continental origin. Seismic tomography can constrain the shape of anomalies in the upper mantle, if the well-known limiting factors of ray smearing, damping, discretization effects, etc. can be overcome. However, tomography does not have the spatial resolution to resolve small-scale features with a scale length of only a few kilometers within anomalies that are steeply dipping and at a depth of around hundred kilometers. Therefore tomography can in principle not answer the question of the nature of the anomaly. Body-wave tomographic studies of the Vrancea region have (thus) arrived at different interpretations as to the nature of the seismic anomaly. Wortel and Spakman (2000) and Martin et al. (2006) prefer an oceanic lithosphere-type explanation, while Koulakov et al. (2010) prefer a continental delamination as an explanation, or a blob model (Lorinczi and Houseman, 2009). Surface-wave tomography (e.g., Boschi et al., 2009) has yielded relatively low velocities down to 100 km depth under Vrancea, and a faster mantle at larger depths, but spatial resolution is probably not good enough to really resolve features at scale lengths of subduction zones.

Seismic attenuation is helpful for imaging attenuating zones, e.g. due to infiltrating asthenospheric material. Such zones are common features in subduction zones, and are even a necessary feature for slab detachment models, where an influx of asthenospheric material is an inescapable consequence of detachment. Such an asthenospheric intrusion is a requirement of delamination models also however. Russo et al. (2005) have studied attenuation under the Vrancea region, and they suggest that delamination is unlikely to occur under the adjacent stable European Platform, the Scythian Platform, and the Easternmost Moesian Platform. They do not rule out delamination under the Transylvanian Basin or above the Vrancea seismic anomaly though.

Earlier tomographic studies (e.g., Martin et al., 2006) had suggested that the earthquakes occur within the high-velocity anomaly. Using the same data, but a more sophisticated tomographic approach, Koulakov et al. (2010) have recently suggested a more complicated relation between seismicity and the shape of the high-velocity anomaly: In their model, the shallower portion of the deep seismicity, at depths of 60-90 km, appears to occur in relatively low-velocity material. At larger depth, the earthquake zone dips nearly vertically, and seems to traverse the high-velocity anomaly, rather than being associated with its edge. Perhaps these small-scale features are at the resolving limit of tomographic models. If the velocity model does indeed resolve the small-scale features, one should perhaps expect a correlation between seismicity and structure for both types of lithosphere. Indeed, the tomographic models of Martin et al. (2006) and Koulakov et al. (2010) which are based on the same dataset, show a number of different features, except that they both are suggestive of a steeply-dipping and even slightly SE-dipping high-velocity anomaly.

Fig. 7 shows a NW-SE profile across the latter model, together with the rays in our study. It is striking that the rays showing dispersion indeed pass through the high-velocity anomalies at intermediate and shallow depths. The low-velocity block in between, with a size of about 30 km is probably not well-resolved. At the same location (and using the same data) Martin et al. (2006) had found a high-velocity anomaly that extends from 300 km depth upwards to at least 50 km depth.

The seismicity does not occur under the surface position of the suture, but is systematically displaced to the SE, by a distance of somewhere between 30 and 90 km, depending on the position of the suture in the different geological studies (e.g., Linzer et al., 1998; Müller et al., 2010; Radulescu and Sandulescu, 1973; Wenzel et al., 2002). Seismicity forms a near-vertical distribution, which corresponds to a typical feature of subduction zone seismicity. There is a "seismicity gap" between 40 and 70 km.

A characteristic of many subduction zones is however that seismicity occurs in two nearly parallel "layers", forming a "double



Fig. 8. Interpretative figure for the new observational constraints presented in this paper. The subducting slab is shown with the crustal portion in yellow. The initially moderate-angle subduction (a) of the oceanic slab steepened; (b) also illustrates the ray geometry for the dispersion observations in red, vs. non-observed dispersion in black (see text). The position of the suture is indicated at the top. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

seismic zone" (e.g., Kawakatsu, 1986). A similar phenomenon has been suggested to occur for Vrancea (Bonjer et al., 2005; Radulian et al., 2007), which, if confirmed, might itself be taken as evidence for the anomaly to be of oceanic origin. It appears that both sets of seismicity can in principle produce dispersion observations, since the distance between the two seismic zones under Vrancea is only about 10 km.

Seismic anisotropy may have a bearing on what the nature of the zone is (e.g., Bokelmann et al., 2011), since it can resolve mantle flow associated with continental lithosphere that is sinking into the deeper mantle. Ivan et al. (2008) have studied SKS splitting for the southeastern Carpathian Arc, and have found that fast orientations in the Eastern Carpathians are more or less arcparallel, what is also found in other parts of the Alpine mountain chains (e.g. Barruol et al., 2011; Bokelmann et al., 2013; Buontempo et al., 2008). Toward the northwest, fast orientations appear to turn into a strike-perpendicular orientation, which is a feature that is also found in a number of subduction zones (e.g., Long and Silver, 2008), although it might potentially be a feature associated with radial flow incited by a "blob"-model.

Beside our argument on in-situ structure, geochemistry and xenoliths studies probably have the best chance to address this question, and to test the oceanic-lithosphere model. The abundant andesites in the area are consistent with subduction (Downes and Vaselli, 1995) as well as xenolith studies (Rosenbaum et al., 1997; Kovács et al., 2012).

The proposition of a subducting oceanic slab is generally consistent with the focal mechanisms of earthquakes found in the region (e.g., Heidbach et al., 2007). At the moment of slab breakout, we would perhaps expect nearly horizontal fault planes, but these are in any case difficult to distinguish from near-vertical fault planes, due to the focal mechanism ambiguity. On the other hand, the observed diversity of focal mechanisms has been used to argue against subduction models (Enescu and Enescu, 1998).

4.3. What does the new constraint imply?

The new constraint suggests that there is oceanic lithosphere present in the upper mantle under Vrancea at depths shallower than 120 km, and probably at the full depth range up to the crustal levels. At these depths, tomographic models for the area generally show high velocity, although seismic body-wave tomography has only weak vertical resolution at these depths.

The seismicity gap between 40 and 70 km depth beneath Vrancea (Fuchs et al., 1979) does not necessarily indicate a different (continental) lithosphere at that depth. Perhaps there is simply less (seismic) deformation at that depth, if the slab has

detached at larger depth. The deeper seismicity is due to the increased seismic strain during the relatively rapid passage of that part of the slab into the deeper mantle. It had been argued before that the 40-70 km depth interval corresponds to the depth of the detachment horizon (e.g., Girbacea and Frisch, 1998; Russo et al., 2005; Sperner, 2005). The observations presented here require a long path through not-too-strongly perturbed oceanic crust toward the surface, to accumulate enough dispersion. If a detachment is occurring somewhere in the Vrancean subduction zone, which seems likely from the presence of strong earthquakes, this implies that the detachment is occurring within the oceanic slab, rather than at the former passive margin. The deeper depth levels of around 100 km (e.g., Gvirtzman, 2002) appears more likely for the detachment to occur, e.g. the depths where the largest earthquakes occur (e.g., Fig. 8). That depth range is also characterized by a "stress anomaly" (Ismail-Zadeh et al., 2000; Oncescu and Bonjer, 1997; Oncescu and Trifu, 1987), which is perhaps not a coincidence. The location where detachment may happen in this region is perhaps less controlled by the (light) weight of the (young) slab, but by former slab bending, e.g. outer-rise seismicity, or by "mantle wind", associated with the southwestward absolute plate motion of the Eurasian plate relative to the underlying mantle. That same mantle wind may have caused the slight southeasterly dipping, rather than toward the northwest, which is apparent in Martin et al. (2006) and Koulakov et al. (2010), see also the schematic representation of our constraint in Fig. 8(b).

As stated above, the end of subduction is not well-constrained (e.g., Ismail-Zadeh et al., 2012). Our observations are consistent with the conclusion from surface geology that there has been little convergence across the SE Carpathians in the last 9 million years (Leever et al., 2006; Matenco et al., 2010). This suggests that the convergence has stopped soon after the continents collided. This corresponds to the "soft continental collision" after the Miocene subduction which is often evoked (e.g., Linzer et al., 1998; Martin et al., 2006). The reason why the convergence ended is perhaps the relatively low weight of the young and short slab under Vrancea, which was overwhelmed by the new mechanical resistance encountered, after the continents collided. This would also explain why the detachment apparently occurred only much later than the continents collided, a fact which is also noted by other researchers (e.g., Koulakov et al., 2010). The foundering into the mantle seems to be occurring in a very narrow time window.

5. Conclusions

One of the most important open questions concerning the deep structure of the Vrancea region is that of what the high-velocity anomaly in the upper mantle under that zone represents (e.g., Ismail-Zadeh et al., 2012), i.e., whether it represents oceanic or continental lithosphere. In this paper we have approached that question directly, by studying seismic waveforms propagating through the anomalous feature, and have found dispersion characteristics that are representative for a classical subduction zone, which is steeply dipping, and which contains a low-velocity layer (the oceanic crust) together with a high-velocity mantle lid. This suggests that the tomographic anomaly represents subducted oceanic lithosphere, down to a depth of at least 120 km. This new constraint offers interesting insights into the behavior of "soft collisions" such as the one in the Eastern Carpathians. Apparently, part of the oceanic lithosphere has not been broken off from the surface region, but is still suspended in the topmost hundred kilometers under Vrancea. A detachment may have occurred, or be occurring, in the deeper portion of the zone, where the strongest earthquakes occur.

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