JOURNAL OF THE BALKAN GEOPHYSICAL SOCIETY, Vol. 6, No. 1, February 2003, p. 46 – 52, 4 figs.

Short-period Q_{pP-P} in Vrancea area - Romania

Marian Ivan*

* Dept. of Geophysics, University of Bucharest, 6 Traian Vuia str., 70138 Bucharest 37, Romania.
e-mail: ivam@gg.unibuc.ro

(Received 18 November 2002; Accepted 17 February 2003)

Abstract: *An investigation of P-wave attenuation in the northwestern part of the Vrancea seismogenic body (Romania) is performed by using the two major seismic events 1986 August 30th and 1990 May 30th, recorded by the University of Utah Regional Network and Pacific Northwest Seismograph Network. The results suggest both horizontal and vertical Q_P variability, with values around 50 for the upper mantle above 89 km and no dependence on frequency. For the upper mantle above 137 km Q_P is around 90, with a possible frequency dependence. A decreasing of Q (increasing of attenuation) from SW to NE is suggested along the north-western side of the Vrancea seismogenic volume, above 89 km depth. It is a possible consequence of a horizontal variation of the thermal field and / or of the tectonic stress in the area.*

Key words: *P-wave attenuation, Q_P factor, frequency dependence of attenuation, Vrancea (Romania) seismic region.*

INTRODUCTION

The intra-continental active seismic area of Vrancea (Romania) is located at the contact of three major tectonic units, i.e. Moesian Platform, East-European Platform and Tisza-Dacia Block (Transylvanian Basin) (Fig. 1). The last major earthquakes in this area are the 1977 March 4th (M_w=7.5, h=86 km) event, the 1986 August 30th (M_w=7.2, h=137 km) and the 1990 May 30th (M_w=6.9, h=89 km) ones. All the above mentioned earthquakes are characterized by reverse faulting with T axis nearly vertical and P axis nearly horizontal, perpendicular to the Carpathian arc (Enescu, 1980; Oncescu and Trifu, 1987; Enescu and Zugrãvescu, 1990). Note that the slip angles are slightly above 90 degrees for most well constrained mechanism solutions. The intermediate depth seismogenic volume is highly confined; the epicentral area has a rectangular shape with the longer side oriented approximately 45°N, azimuth which is close to the fast wave direction derived from SK(K)S splitting measurements (Ivan, 2000). The physical mechanism of the subcrustal seismicity is question of debate, from basalt-eclogite phase change processes (Oncescu, 1980; Enescu, 1985) to mantle delamination (Chalot-Prat and Girbacea, 2000). An active continental subduction beneath the Carpathian arc has been also considered (e.g. Enescu and Enescu, 1993). Tomographic studies (Oncescu et al., 1984) outlined a low P-wave velocity zone between 40 and 80 km depth, which coincides with the 'aseismic gap', previously interpreted by Fuchs at al. (1979) as a low viscosity – high attenuation volume. According to the authors mentioned above, there is a high velocity anomaly between 80 and 250 km depth, the maximum depth of observed earthquakes being 221 km (1982, May, 16th, 45.42°N, 26.52°E). Similar results have been derived by regional tomography (Fan et al., 1998), indicating a near-vertical high velocity body located at depths between 100 and 170 km.

Various techniques have been used to estimate the Q factor for the upper mantle. Free oscillations indicate values in the range 120 (Sailor and Dziewonski, 1978) to 150 (Anderson and Hart, 1978). On a global scale, Q_P around 200, for the upper

400 km, has been obtained by Bhattacharyya et al. (1996). An average Q_P of 150 has been reported by Sipkin and Jordan (1980), for the upper mantle beneath the ocean basins. The values reported by Flanagan and Wiens (1998) for the Lau back arc basin, in the depth interval 0-200 km, are slightly lower, in the range 102-121. They are associated to the active oceanic subduction in Tonga-Fiji area, investigated in detail by Roth et al. (1999). Highest attenuation there (Q_P=90) has been found within the upper 100 km beneath the active portions of the Lau basin, well correlated with zones of low P wave velocity. Tomographic results indicate both horizontal and vertical variations of Q, the slab being a region with low attenuation material (Q_P > 900).

Previous estimations of quality factors in Vrancea area from local events have been performed by Oncescu (1986), using the two frequencies method of Granet and Hoang Trong (1980). The derived values, at 8 and 15 Hz respectively, suggested that the total decoupling is present only in the southwestern part of the slab, now sinking gravitationally. Other estimations of the body wave attenuation in Vrancea and adjacent areas have been performed in relation with coda-wave attenuation. Being mainly limited to the investigation of the upper crust (Oancea et al., 1991; Oancea, 1994), they show a good correlation with the shallow geological structures.

In this paper, some results related to the attenuation of short period P and pP phases (e.g. Bock and Clements, 1982) are presented, assuming that the amplitude spectrum of a certain wave decreases exponentially, according to

$$A(f) = A_0 \exp(-\pi \frac{t}{Q} f) \qquad (1)$$

where f is the frequency and t is the propagation time of the wave.

DATA AND METHODOLOGY

Digital waveforms available at the IRIS Data Base have been web-requested for the major Romanian earthquakes of 1986 August 30th and 1990 May 30th (Table 1). Their fault plane solutions and magnitudes are quite similar, indicating possible similar rupture processes. In each case, the data have been carefully inspected and the observed arrival times of the P and pP phases have been checked with respect to the values derived using IASP91 model (Buland and Chapman, 1983; Kennett and Engdahl, 1991). Generally, the (O-C) time residuals do not exceed 3 seconds. In order to allow better statistically based results, only the recordings at the University of Utah Regional Network (IRIS - UU) and Pacific Northwest Seismograph Network (University of Washington, IRIS – UW) have been selected, an example being shown in Figure 2. Most of the recording instruments have vertical short period sensors (Mark L-4 SP and Geotech S-13), with a natural frequency around 1 Hz.

The quality factor Q_P has been evaluated by using the spectral ratio method, closely following the methodology described by Roth et al. (1999). The arrivals of P and pP phases have been read on seismograms, to obtain the difference of the arrival times Δt_{pP-P}. The observed polarities have been also compared with the Harvard CMT fault plane solutions. When horizontal recordings have been available, the radial component has been obtained and its correlation with the vertical recording has been also used to pick up the times accurately. For each P or pP phase, a constant window of approximately 10 seconds (1024 points for instruments with sampling rates of approximately 100 Hz, and 512 points for sampling rates of approximately 50 Hz) has been used to evaluate the natural logarithm of the amplitude spectra (Fig. 3). This window length shows a very good focus of the wave energy and seems most appropriate to avoid the contamination with sP phase. However, sP phase could not be clearly identified. The dip angle of the nodal plane considered as the fault plane is 63° or 70°. For most stations, the take-off angle of P wave is around 22°, the similar value being around 158° for pP or sP waves. For the last waves, it follows the angle of the ray to the fault plane is close to 45 degrees, explaining the remarkable high amplitude of the pP wave and the absence of sP phase due to the radiation pattern. A noise window of the same length, preceding the arrival of the P wave has been selected to be used as reference. Spectral computations for P, pP and noise windows have been performed by using the (input bit reversal) FFT subroutine described by Stearn (1975). For almost all the processed windows, a clear change of the spectral slope around 2 Hz is observed (Fig. 4). Consequently, for each station the computations have been performed in two frequency windows (of equal lengths): 0.1 ÷ 1.8 Hz and 0.2 ÷ 1.9 Hz, where the amplitude spectra of both P and pP are clearly above noise spectrum, in almost all cases.

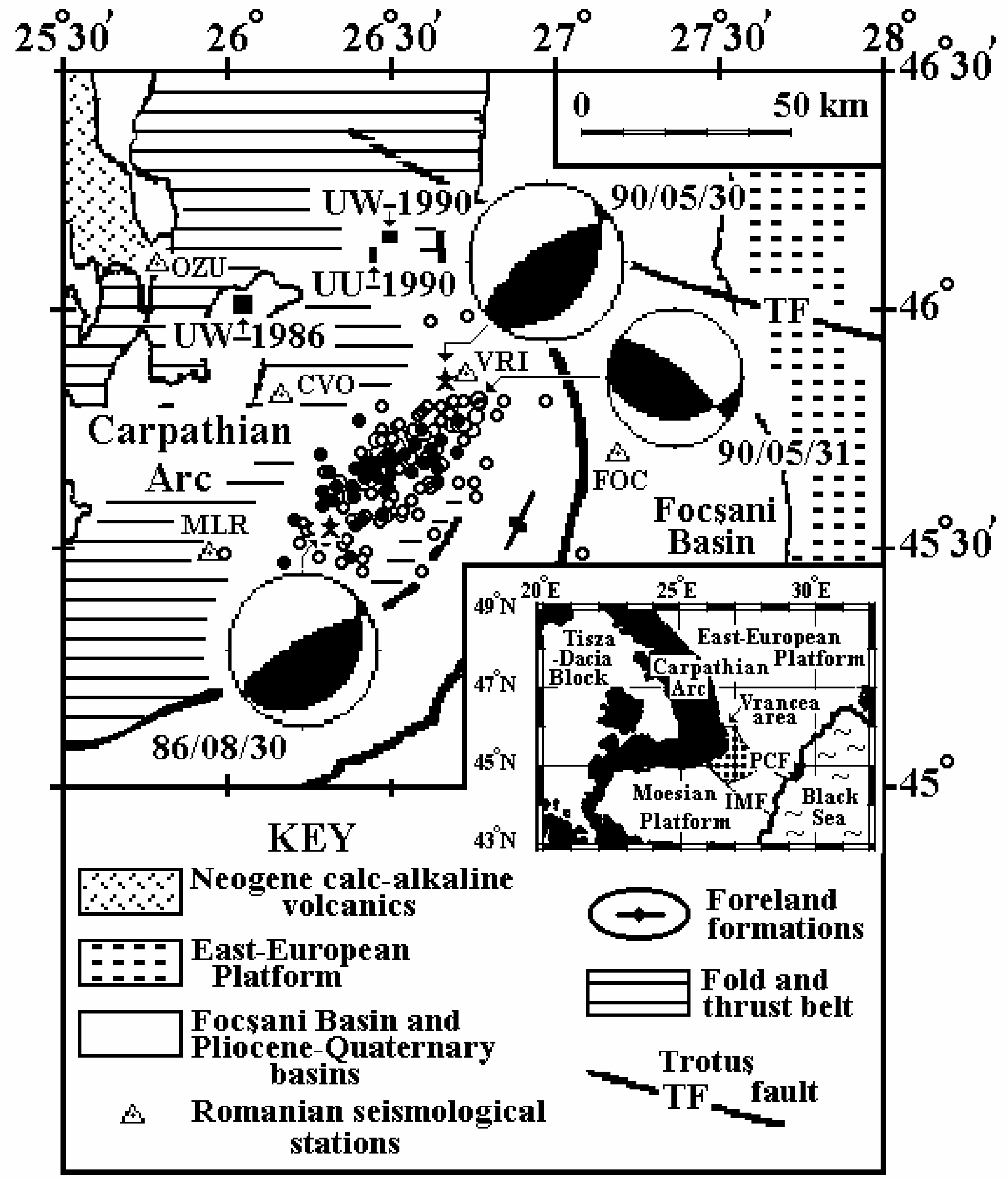

FIG. 1. Vrancea epicentral area (NEIC catalogue) with Romanian seismological stations. Open circles indicate earthquakes in the depth range 50–150 km and full circles earthquakes in the interval 150-250 km. Fault plane solutions are Harvard CMT ones. The tectonic setting is modified after Chalot-Prat and Girbacea (2000). Peceneaga-Camena fault is abbreviated as PCF and the Intra-Moesian one, as IMF. The surface bounce points corresponding to University of Utah Regional Network (UU) and Pacific Northwest Seismograph Network (UW) are indicated by full rectangles for each event.

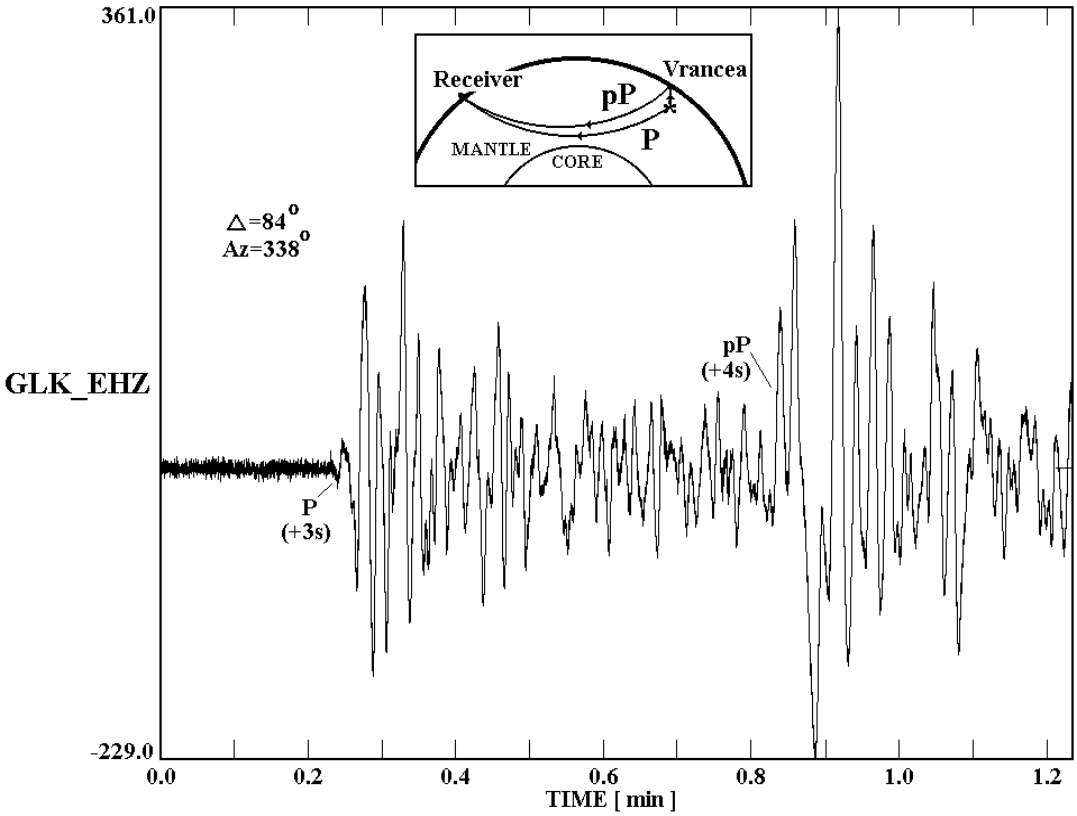

FIG. 2. Recording of 1986, August 30th event at GLK station. Teleseismic P and pP arrivals are indicated together with the O-C values (IASP91 model). A non-scaled sketch displaying ray paths of P and pP waves is also presented.

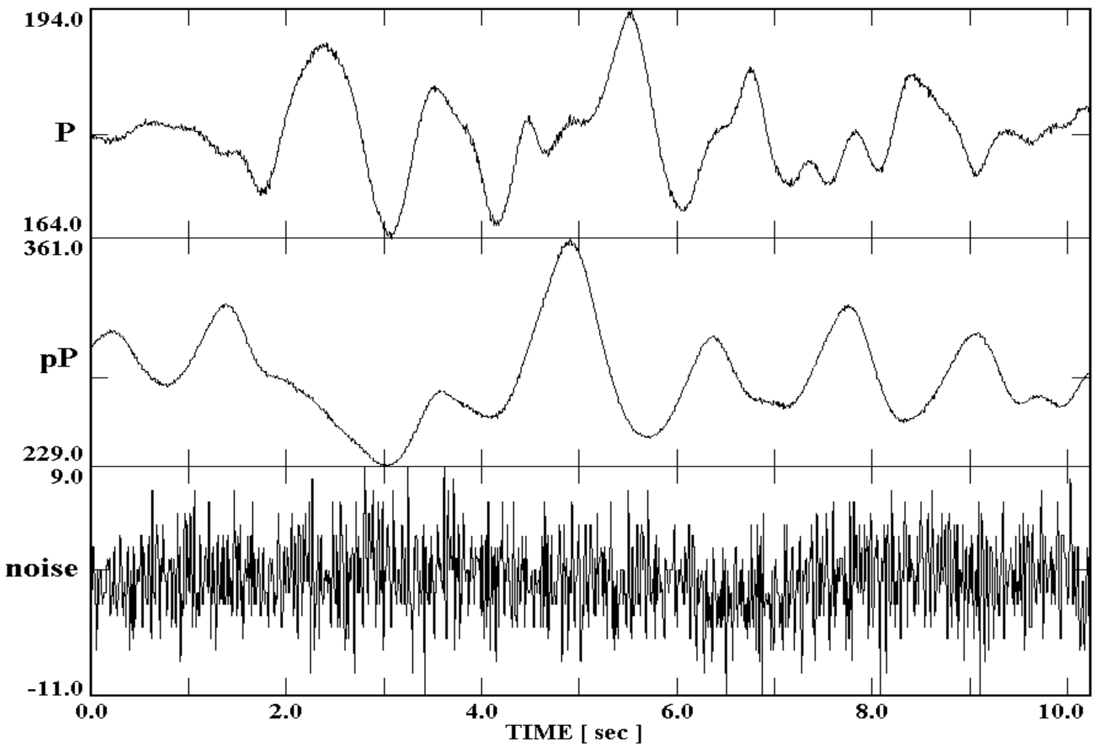

FIG. 3. Time windows processed by spectral ratio method. Note the depletion in higher frequencies of pP phase with respect to P phase, believed to be an effect of the attenuation.

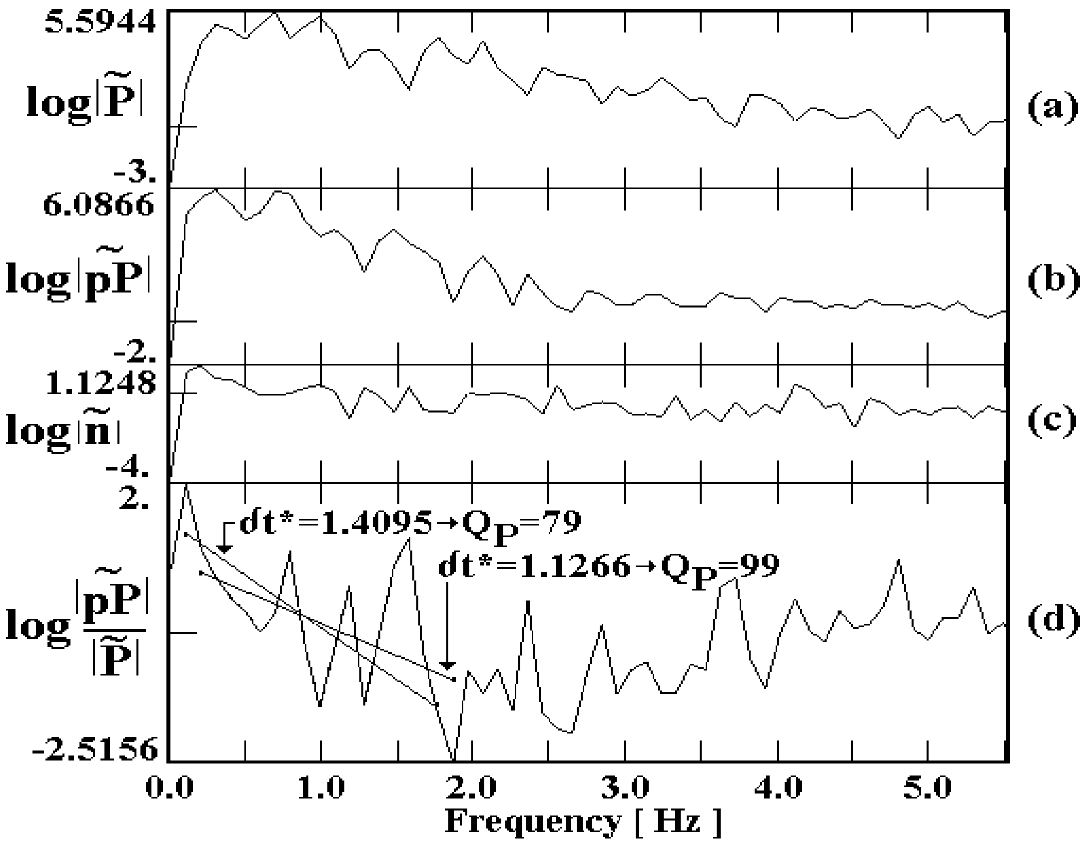

FIG. 4. Logarithm of the amplitude spectra for the time windows from Fig. 3 (below 5Hz). DC values are arbitrary. (a) P wave spectrum; (b) pP wave; (c) noise. (d) difference pP-P. Note the linear decreasing below 2 Hz. The slopes of the regression lines, for the two frequency windows analysed are presented together with the corresponding Q_P values.

Table 1: ISC location of the processed events. The preferred fault plane solutions (Harvard CMT) are indicated.

Event	Latitude	Longitude	Depth (km)	Seismic Moment	Strike (°)	Dip (°)	Slip (°)
1986-08-30	45.5373 N	26.3138E	137	7.91e+26	240	72	97
1990-05-30	45.8474 N	26.6625E	89	3.01e+26	236	63	101

The logarithm of the ratio of the spectral amplitudes of pP to P has been computed and a linear regression has been estimated for each of the two frequency windows. The Q factor has been evaluated by

$$Q = \pi \frac{\Delta t_{pP-P}}{\delta t^{*}} \qquad (2)$$

where δt^{*} is the absolute value of the slope of the regression line. Statistical analysis has been subsequently performed for each event and network.

RESULTS AND CONCLUSIONS

For the 1986 event (depth 137 km), only the recordings at UW network were available, resulting in a number of 144 evaluations for Q. The mean Q-value, obtained by averaging Q^{1}, is 89±4 (95% confidence level). However, for almost all the recording stations, a clear, slight increase of the Q value obtained in the second frequency window is observed, suggesting a possible dependence with frequency. The mean value of Q is 82±5 (95% confidence level) for the window 0.1 ÷ 1.8 Hz, and 98±6 for the window 0.2 ÷ 1.9 Hz. The small variation of the frequency window does not allow one to match accurately a power-law frequency dependence of Q (Anderson and Minster, 1979).

For the 1990 event (depth 89 km), recordings at both UU (41 stations) and UW (95 stations) networks were available. Mean values of 50±3, and 46±1 respectively, have been obtained for Q, when using the records of the UU network, and of the UW stations respectively. No clear frequency dependence of Q is observed.

In order to compare the Q values determined for different depth ranges (0-137 km, and 0-89 km respectively, in our case), the following relation for a horizontally stratified earth (e.g. Flanagan and Wiens, 1998) may be used:

$$\frac{t^{137-0}}{Q^{137-0}} = \frac{t^{137-89}}{Q^{137-89}} + \frac{t^{89-0}}{Q^{89-0}} \qquad (4)$$

where t^{137-0} is the propagation time of the P wave, vertically from 137 km depth to the surface, and Q^{137-0} is the mean value of Q on the same depth interval.

Propagation times have been evaluated by using IASP91 model, leading to $t^{137-0} = 18.43$ s and $t^{89-0} = 12.47$ s. The above values vary slightly to 18.63 s and 12.74 s respectively, for the 1-D velocity model used in routine location activity (Oncescu, 1984; Oncescu et al., 1984). According to eq. (4), for the average Q^{137-0} – value of 89, Q^{89-0} - values greater then 60 are expected in both cases. Lateral variations of Q_P in the area are suggested.

The mentioned tomographic results, the distribution of hypocenters and the fault plane solution of the 1990 May 31st event (Mw=6.3, h=90 km) suggest that the top of quasi-vertical seismogenic volume (around 90 km depth) is slightly dipping towards northwest on the northwestern side, and approximately towards southwest or south on the northeastern (or northern) side. The geometrical features of the seismogenic volume in the depth range from 90 to 170 km suggest the attenuation on the pP legs of the waves being responsible for that behaviour. Hence, most likely, an increasing of the attenuation (decreasing of Q) from SW to NE is indicated along the NW side of the Vrancea seismogenic volume, above 89 km depth. It is a possible consequence of a horizontal variation of the thermal field and / or of tectonic stress in the area.

ACKNOWLEDGEMENTS

The author is grateful to Dr. Luminita Ardeleanu (NIEP) and to Professor Vasile Marza (University of Brasilia) for useful suggestions and constructive reading of the manuscript.

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