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Seismic anisotropy of the upper crust in southeastern Poland—effect of the compressional deformation at the EEC margin: Results of CELEBRATION 2000 seismic data inversion

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[1] Analysis of the data from CELEBRATION 2000 experiment in southeastern Poland revealed azimuthal variation of the Vp velocity, unlikely to be caused by crustal inhomogeneity. This is explained by upper crustal seismic anisotropy and was analysed by the anisotropic delay-time inversion. The result, indicating 8–10% anisotropy, fits to the geology of the area, where tightly folded metapelitic rocks are abundant. The fast velocity axis direction of 115° (WNW-ESE) coincides well with the azimuths of outcropping folds axes and other deformational structures. In order to assess the credibility of the result, synthetic tests have been performed. The traveltimes calculated for isotropic models were inverted, to check if an artificial anisotropy will be generated. The tests indicate that realistic isotropic velocity inhomogeneities can only account for a fraction of observed traveltime variations, and that the anisotropy cannot be an artifact resulting from the inhomogeneous crust. **Citation:** Środa, P. (2006), Seismic anisotropy of the upper crust in southeastern Poland—effect of the compressional deformation at the EEC margin: Results of CELEBRATION 2000 seismic data inversion, *Geophys. Res. Lett.*, *33*, L22302, doi:10.1029/2006GL027701.

1. Introduction

[2] The origin and tectonic evolution of Małopolska Unit (MU) and Łysogóry Unit (ŁU), located between the SW margin of the Precambrian East European Craton (EEC) in the north and Bruno-Silesian Unit (BSU) and Carpathian orogen in the south (Figure 1), is still a subject of controversy. They are considered to be separate units based on their different stratigraphy and evolution. However, contrasting interpretations exist concerning their Gondwana vs. Baltica provenance, the timing of their possible accretion at the EEC margin and the relative influence of Caledonian and Variscan deformation.

[3] *Pożaryski* [1990] interpreted these units as exotic terranes, forming a Caledonian strike-slip orogen. According to K-Ar age determinations of *Belka et al.* [2002], the MU is a Gondwana-derived terrane, located close to the EEC in the Late Cambrian, while the origin of the ŁU is enigmatic. *Dadlez et al.* [1994] consider the ŁU to be a part of the EEC passive margin deformed into a Caledonian thrust-and-fold belt. In their interpretation, MU is a proximal terrane, detached and re-accreted to Baltica. The studies of the Cambrian fauna of the ŁU indicate either Gondwana [*Belka et al.*, 2002] or Baltica [*Żylińska*, 2002] affiliation. *Jaworowski and Sikorska* [2006] consider both units as parts of the EEC Cambrian passive margin. A similar view is presented by *Mizerski* [2004].

[4] The crustal thickness, determined from wide-angle seismics, varies from 32–35 km in SW to 44–50 km in the east [*Środa et al.*, 2006; *Malinowski et al.*, 2005; *Janik et al.*, 2005]. The MU and ŁU contact along the Holy Cross Fault, which cuts the Holy Cross mountains (HCM) – an outcrop of deformed Palaeozoic rocks – in WNW-ESE direction, similar to the trend of most of the Palaeozoic structures. Outside the HCM, Palaeozoic strata are overlain by a Mesozoic/Cainozoic cover of variable thickness. In the MU, Neoproterozoic sequences have been probed by boreholes. The seismic modelling [*Środa et al.*, 2006; *Malinowski et al.*, 2005] indicates that beneath the MU and ŁU, the Neoproterozoic and older rocks, with Vp typical for low-grade metasediments, reach the depth of 18 km. In the NW Małopolska Unit, Vendian/Lower Cambrian sediments and greenschist facies metasediments (largely phyllitized claystones and siltstones), severely folded and faulted during Early Caledonian event, are discordantly covered by less deformed, Lower Ordovician and younger cover [*Buła et al.*, 1997; *Belka et al.*, 2002]. The cover thickness is <0.5 km to few km. In the SE part of the MU, strongly folded and thrusted Neoproterozoic/Lower Cambrian low-grade metasediments (largely shales, siltstones and claystones), in places unconformably overlain by less deformed Ordovician rocks [*Moryc and Jachowicz*, 2000], are discordantly covered by 0.5–7 km thick Miocene deposits of the Carpathian Foredeep and Carpathian nappes. The Łysogóry Unit, adjacent to the EEC, consists of Lower Palaeozoic rocks, overlain in the NE by the Mesozoic-Cainozoic cover. There is no direct evidence about the Neoproterozoic in the ŁU. The oldest known rocks are the Cambrian shales. According to *Belka et al.* [2002] and *Mizerski* [2004], an important difference between both units concerns the Early Caledonian deformation, observed in the MU (affected also by the weaker Late Caledonian event and by Variscan deformation) and lacking in the ŁU, affected only by the Variscan event. However, *Dadlez et al.* [1994] also argue for a Late Caledonian deformation of the ŁU.

[5] During the CELEBRATION 2000 experiment [*Guterch et al.*, 2003], a dense network of 3-D wide-angle seismic recordings was acquired in SE Poland. The modelling of the data revealed azimuthal variations of the seismic velocity, suggesting crustal anisotropy. In this study, the traveltime inversion was applied in order to quantify the anisotropy, obtain new information about the structure of

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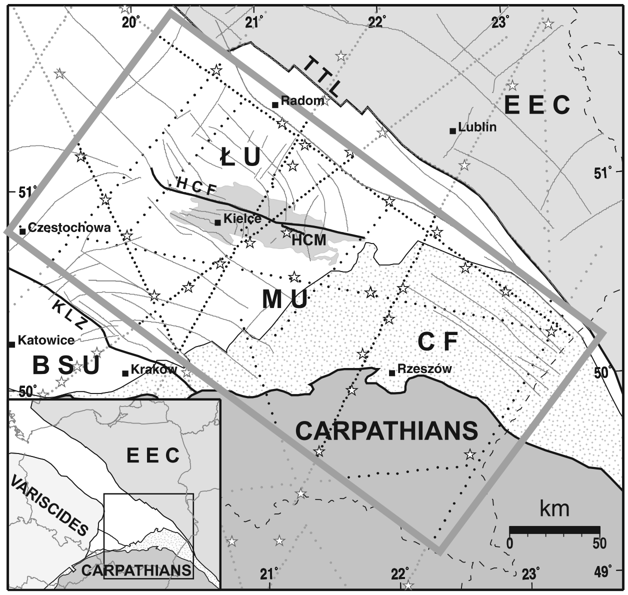

Figure 1. Tectonic sketch of SE Poland with location of CELEBRATION 2000 profiles. Gray rectangle - inversion area, black and grey dots - receiver locations, stars - shot points, grey and black lines - faults and tectonic boundaries. BSU - Bruno-Silesian Unit, CF - Carpathian Foredeep, EEC - East European Craton, HCF - Holy Cross fault, HCM - Holy Cross mountains, KLZ - Kraków-Lubliniec Zone, ŁU - Łysogóry Unit, MU - Małopolska Unit, TTL - Teisseyre-Tornquist Line.

the MU and ŁU and to improve our understanding of their tectonic evolution.

2. Data

[6] 2-D seismic modelling of the CELEBRATION 2000 data in SE Poland (MU and ŁU area) revealed large differences in the apparent velocity of the crustal refracted arrivals (Pg), depending on the profile orientation. Independently of the location, recordings along NW-SE azimuth show a higher apparent velocity (6.1–6.3 km/s) than perpendicular ones (5.4–5.7 km/s). Similarly, the models show discrepancies in the upper crustal Vp at their crossing points, suggesting an azimuthal velocity variation. As the differences could not be reconciled by any common isotropic model with realistic Vp distribution, the azimuthal variability of traveltimes was analysed by anisotropic inversion of data from a 300 × 152 km large rectangular area (Figure 1). The area has been designed to maximize the number of data from the MU and ŁU, without including rays from the areas with substantially different structure – the EEC and the BSU. The traveltime data used for inversion (4088 values of in-line and off-line Pg recordings in the 20–150 km offset range) are shown in Figure 2. The diagrams show a dependence on azimuth, with 180° periodicity and a trend roughly of $\cos 2\varphi$ form. The trend is clearly visible, thus its amplitude is substantially larger than the traveltime variations due to velocity inhomogeneities in the uppermost crust all over the area. The traveltime minimum is at about 100–120°. The present study attempts to explain these variations by a seismic anisotropy of the upper crust in the MU and ŁU area.

3. Modelling Method: Anisotropic Delay-Time Inversion

[7] In order to determine the anisotropy parameters, a traveltime inversion was performed using the delay-time method. In its original formulation [*Willmore and Bancroft*, 1960], the model consists of an upper layer with varying velocity and thickness, and a lower layer of unknown constant isotropic velocity V. The original formula for traveltime of a refracted ray, after modification for anisotropic medium with small azimuthal slowness perturbation after *Backus* [1965], has the form [*Song et al.*, 2001]:

$$t_{ij} = a_i + b_j + D_{ij}(S_0 + A\cos(2\varphi) + B\sin(2\varphi) + C\cos(4\varphi) + D\sin(4\varphi)), \quad (1)$$

where D_{ij} is the distance from source i to station j, S_0 is unknown average P-wave slowness (1/V) below the refractor, and a_i, b_j are unknown time delays for the i-th source and j-th receiver, respectively. The delays reflect the varying refractor depth and velocity above it. If the

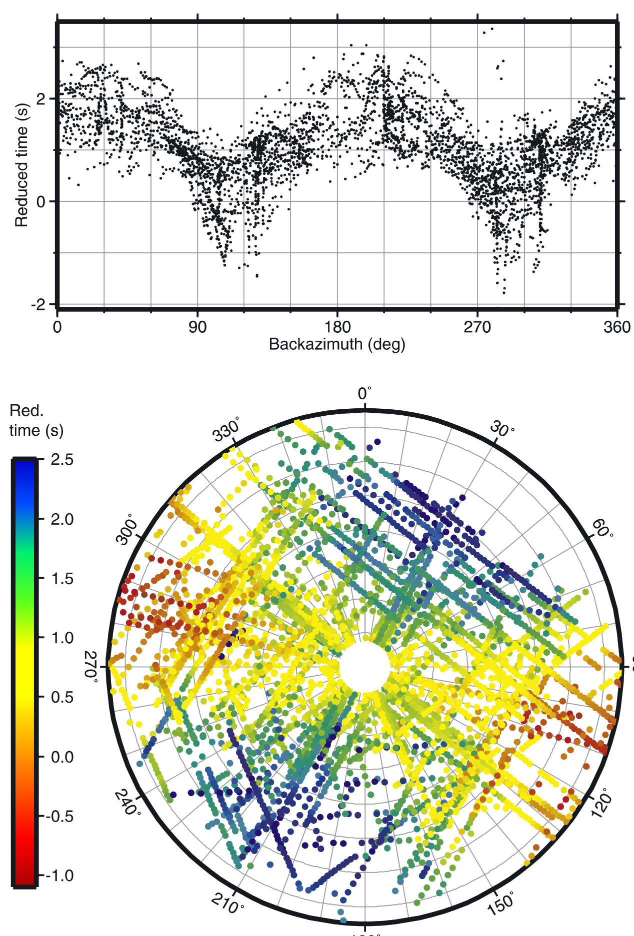

Figure 2. Azimuthal diagrams of Pg traveltimes for 20–150 km offset. Bottom: polar diagram of traveltimes. Polar coordinates r, φ are the offset and ray backazimuth, respectively. Reduction velocity is 5.8 km/s.

Table 1. Results of the Inversion and of the Significance Test[a]

Method	Vp, km/s	AN, %	$\varphi_{\max}$, deg	N_{data}	N_{par}	DF	RMS dT, s	$F_{calculated}$	F_{table}	Improvement
ISO (I)	5.9 ± 0.01	-	-	4088	487	3601	0.49	-		
ANI 2φ (A2)	5.7 ± 0.01	10.5 ± 0.2	115 ± 0.4	4088	489	3599	0.24	A2/I: 5458	4.61	YES
ANI 2φ,4φ (A4)	5.7 ± 0.01	10.2 ± 0.15	115 ± 0.4	4088	491	3597	0.23	A4/I: 3112	3.32	YES
								A4/A2:190	4.61	YES

[a]N_{data} - number of data points, N_{par} - number of parameters, DF - number of degrees of freedom ($N_{data} - N_{par}$).

coefficients C and D are small compared to A and B, anisotropy can be defined as $AN = (A^2 + B^2)^{1/2}$ and azimuth of maximum velocity as $\varphi_{MAX} = 0.5 \cdot \mathrm{atan}(B/A)$. The set of linear equations (1) for all source-receiver pairs can be written in a matrix form $\mathbf{d} = \mathbf{Am}$ ($\mathbf{m}$-model vector, $\mathbf{d}$-data vector) and solved using, for example, the damped least squares (DLS) inversion for a_i, b_j, S_0 and anisotropy by seeking the solution in the form:

$$\boldsymbol{m} = \left(\boldsymbol{A}^T\boldsymbol{A} + \lambda^2\boldsymbol{I}\right)^{-1}\boldsymbol{A}^T\boldsymbol{d}, \tag{2}$$

where λ is the damping coefficient. This technique has been used for the upper mantle anisotropy studies [e.g., *Hearn*, 1984; *Song et al.*, 2001]. *Růžek et al.* [2003] also applied it for a crustal anisotropy study.

[8] In the case of the upper mantle studies, the vertical velocity gradient is usually small and refracted rays are nearly horizontal, as it is assumed in the method. In this study, 2-D modelling revealed significant vertical Vp gradient in the upper crust. Therefore, a formula for traveltime in the gradient layer [*Enderle et al.*, 1996] has been used, instead of the linear one:

$$t_{ij} = a_i + b_j + (2/G)\mathrm{asinh}\left(\mathrm{GD_{ij}S}(\varphi)/2\right), \tag{3}$$

where G is the velocity gradient. Such a formula accounts properly for the traveltime beneath the refracting interface. The raytracing simulation shows that the inaccuracies due to the different ray geometry compared to a non-gradient medium are negligible. Also the fact that the rays sample the anisotropic volume at some angle, rather than horizontally as it is assumed in the method, does not cause significant errors.

[9] In the gradient case, the equations set is nonlinear in S, therefore it was solved using an iterative linearized inversion procedure: a) the initial model $\mathbf{m_0}$ was calculated by inverting the linear relation (1) using (2); b) the residuals $\boldsymbol{\delta t} = \boldsymbol{t_{obs}} - \boldsymbol{t_{ij}}$ were calculated for $\mathbf{t_{ij}}$ obtained using the nonlinear relation (3) for the current model; c) the model perturbations $\boldsymbol{\delta}$m were calculated as $\boldsymbol{\delta m} = (\boldsymbol{A^TA} + \lambda^2\boldsymbol{I})^{-1}\boldsymbol{A^T\delta t}$ by inverting the linear relation (1) and added to the previous model: $\boldsymbol{m_{i+1}} = \boldsymbol{m_i} + \boldsymbol{\delta m}$. The steps b) and c) were repeated until $\boldsymbol{\delta}$m was not significant. For this problem, 2 iterations were sufficient.

[10] In order to check if the anisotropy assumption was necessary and to evaluate the significance of adding new model parameters, three variants of inversion were solved: for isotropic Vp, for anisotropic Vp with 2φ dependence only and for anisotropic Vp with 2φ & 4φ terms. The latter produced a slightly better fit than inversion with 2φ terms, but the results are similar. Errors of the obtained parameters (Table 1), were calculated using the bootstrap method [*Efron*, 1979]. However, these estimates mainly reflect the effect of random data errors and do not take into account the inaccuracy resulting from assuming a simple model that cannot adequately image the inhomogeneous structure. This factor is hard to estimate reliably and therefore the actual errors may be substantially bigger.

[11] The *Backus* [1965] formula applies for near-horizontal rays in a weakly anisotropic medium and is not constrained to any particular symmetry. However, the only parameters obtained by the measurements constrained to a horizontal plane are mean velocity, anisotropy (AN) and φ_{MAX}. Therefore, for interpretation purposes, I simplify the medium by assuming an idealized transverse isotropic model with a horizontal symmetry axis, which can be uniquely defined by these parameters. Such an assumption is realistic, as layered or fractured rocks can exhibit transversal isotropy. The idealized case of the horizontal symmetry axis cannot be distinguished from a more likely case of a tilted axis—the observed anisotropy depends on rock properties and on a symmetry axis dip, and both factors are unknown. Another approximation is due to the representation of a large, geologically complex fragment of the crust by constant parameters for the whole area.

4. Results

[12] Results of the inversion are summarized in Table 1. The calculated magnitude of the anisotropy is ~10%, with azimuth of maximum Vp velocity 115° (Figure 3). The time delays are in range 0.1–1 s, which corresponds roughly to the 0.4–4 km thickness of the low-velocity cover. The distribution of the delays (Figure 3) reflects the variations of the cover thickness and velocity. The large delays at the upper (NE) edge of the area are caused by significant thickening of the Mesozoic/Cainozoic and partially also the Palaeozoic cover towards the EEC Margin, documented also by *Środa et al.* [2006], while the increase of the delays in the south results from the thickening of low-velocity sediments of the Outer Carpathians and the Carpathian Foredeep. Beneath the MU, the delays are minimal and are in general lower that beneath the ŁU, which suggests that the Neoproterozoic basement occurs here at shallower depths than in the latter unit.

[13] The final RMS residual is 0.24 s, which is considerably smaller than the residual for the isotropic model – 0.5 s. Even if this is a substantial improvement of the fit, it is still higher than the estimated data uncertainty (~0.1 s). This hints that more advanced algorithm, able to model the spatial variations of the velocity and of the anisotropy parameters, should be used. However, I believe that for a dataset where azimuthal traveltime variations

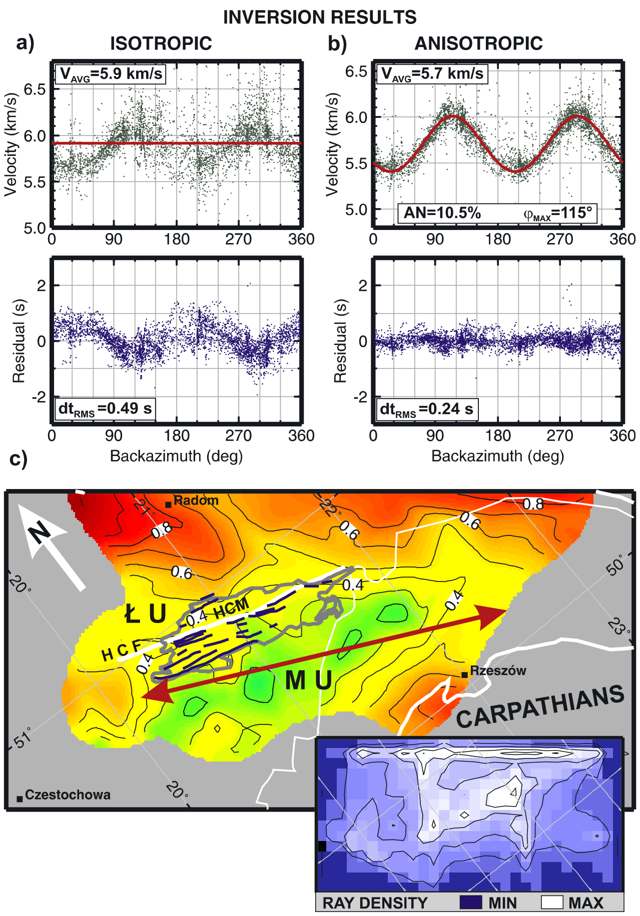

Figure 3. Results of the isotropic (a) and anisotropic delay-time inversion (b). Thick line - modelled velocity, points - observed velocity and traveltime residuals. c) - distribution of time delays for anisotropic inversion (isolines every 0.1 s) and map of ray density. Two-sided arrow - direction of fast velocity, thick lines in HCM area - axes of folds in the Palaeozoic strata [*Mizerski*, 2004]. Dark gray - area which is unresolved due to low data coverage.

clearly dominate (Figure 2), even the simple 1-D inversion used here gives a reliable and meaningful result, even if it leaves some crustal features unexplained.

[14] Analysis of the short-offset (0–20 km) subset of traveltimes showed that the near-surface Mesozoic/Cainozoic sequences are not anisotropic, in contrast to the Palaeozoic outcrops in the HCM, where anisotropy is observed in spite of scarce data. The average velocity at the top of the modeled anisotropic layer is 5.7 km/s (Vmin = 5.4 km/s, Vmax = 6.0 km/s) and at the bottom (for the Vp gradient of 0.03 s^{-1} the ray penetration depth was estimated to ~15 km) the average Vp is 6.2 km/s. These values are meaningful only for places with high ray density, shown in Figure 3 (most of the ŁU and the NW part of the MU). The test inversion of separate data subsets confined only to ŁU or MU shows that both units are anisotropic.

[15] In order to check if the anisotropic model results in a statistically significant data fit improvement with respect to isotropic one, an F-test was performed [*Hearn*, 1984]. If the obtained F-ratio is larger than the tabulated F-distribution value, the anisotropic model is assumed to significantly improve the fit. The results (Table 1) show a significant improvement due to the anisotropy assumption.

5. Synthetic Tests

[16] Determinations of the seismic anisotropy based on azimuthal variation of traveltimes have to be taken with caution, since a similar effect may be produced by an isotropic medium with elongated, ridge-like velocity inhomogeneities, particularly in the case of uneven ray coverage. In order to check the credibility of the models, tests involving inversion of synthetic data were performed. The finite difference code [*Vidale*, 1990] has been used to calculate the traveltimes for isotropic models containing ridge-like positive velocity anomalies of 1.2 km/s amplitude with a ridge axis azimuth of 115° and with a characteristic width of 80, 40 and 20 km (Figure 4). The geometry of sources and receivers was the same as for real data. The traveltimes, perturbed by a gaussian noise with 0.1 s standard deviation, were inverted using a delay-time method in the same way as the real data, in order to check if modelled anisotropy can be an artifact resulting from preferred orientation of velocity anomalies in an isotropic medium.

[17] The results show that the amount of artificial anisotropy varies depending on the characteristic width of anomalies, but even for unrealistically large amplitude anomalies, recovered anisotropy (about 3%) is not as large as that obtained in this study (10%). Also, comparing results of the anisotropic and isotropic test inversion, the RMS residual decrease (ca. 5% of the value for isotropic inversion, Figure 4) is much smaller than for real data (50%, Figure 3) The shape of the traveltime and residuals distributions for real (Figure 2) and synthetic data (Figure 4) show substantial differences. All this suggests that observed azimuthal velocity variations could not originate from an inhomogeneous, isotropic structure with preferred orientation. Therefore, the delay-time method can effectively separate anisotropy and inhomogeneity, especially when the characteristic length of the velocity anomalies is large (>40 km). However, velocity anomalies can bias obtained anisotropy parameters. Assuming a realistic amplitude of the anomalies, the bias should not exceed ~2% and therefore a good estimate for the final anisotropy value is 8–10%.

6. Discussion and Conclusions

[18] The seismic anisotropy occurs in both Małopolska and Łysogóry units, at a depth from 0–4 km to ~15 km, which suggests that it is a feature of Palaeozoic and Neoproterozoic (and older?) rocks, while the Mesozoic/Cainozoic cover is not anisotropic. The magnitude of the anisotropy (8–10%) is unusually large, considering that it was obtained in situ as an average for the c. 200 × 100 km large area where isotropic rocks as well as rocks of varying anisotropy and layer orientation coincide, which means that locally the anisotropy may be even higher. Nevertheless, the results of the laboratory measurements show that preferred mineral orientation in the metasedimentary rocks can result in even higher values than those observed in this study. Therefore, the seismic anisotropy observed in the Małopolska and Łysogóry units is interpreted as related to

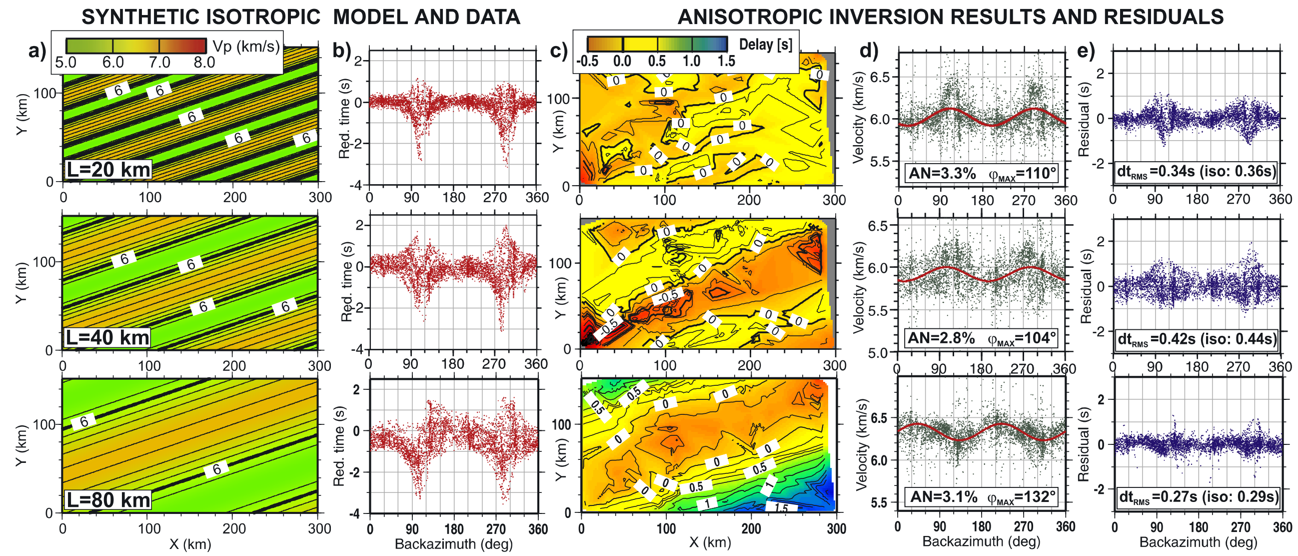

Figure 4. Results of the synthetic tests for models with characteristic anomaly widths 20 km (top), 40 km (middle) and 80 km (bottom). a) horizontal slice of the reference model at z = 10 km, b) synthetic traveltimes (V_{red} = 6.0 km/s), c) recovered delay times, d) reference (points) and recovered (line) velocities, e) traveltime residuals.

foliation fabrics and associated preferred mineral (mica) orientation in foliated rocks. *Christensen and Mooney* [1995] report values of 12% for phyllites, 16% for mica schists, 21% for slates and 5–10% for gneisses. *Godfrey et al.* [2000] obtained 9–20% anisotropy for similar types of rocks. *Johnston and Christensen* [1995] obtained 20–35% Vp anisotropy for shales, while measurements of shales in situ [*Leslie and Lawton*, 1999] resulted in 13–20% anisotropy. The average Vp of shales is very low (3–4.5 km/s), however, and therefore they cannot be the main anisotropic component of the crust, except in its uppermost part. Good candidates are phyllites with Vp of 6.0–6.3 km/s, mica schists (5.8–6.4 km/s), gneisses (5.5–6.2 km/s) or slates (6.0–6.2 km/s).

[19] In order to produce observed azimuthal velocity variations, the foliation planes of the anisotropic rocks have to be oriented subvertically (e.g., in tight, upright folds), or at a high angle from the horizontal, and at similar azimuth. This is very likely in the light of available data about the deformations and stratal dips in the MU and ŁU area. The Neoproterozoic dips measured in boreholes in the MU are usually in the 40–80° range, often reaching 90°. In the HCM, where outcrops of the Palaeozoic rocks were available for direct observations, the orientation of deformational structures and the attitude of strata are similar for both units: 100–110° for MU and 110–120° for the ŁU, with dips in the range of 30–90° [*Mizerski*, 1992]. Statistical analysis of the Lower Palaeozoic strata positions in the NW part of the HCM (Łysogóry Unit) by *Dębowska and Zawadzki* [2005] resulted in the strike range of 106–115°, with a stratal dip in the range of 37–47°. In The MU, according to *Stupnicka* [1986], the strike of Palaeozoic folds axes is 100–120°, and dips of 60–70° or more are common. Thus, the observed direction of fast velocity (115°) corresponds well with the strike of the main deformational structures in Palaeozoic and Neoproterozoic rocks building the upper crust. Other contributors to the anisotropy may also be abundant WSW-ENE oriented cracks, however, their effect seems to be secondary. The cracks close fast with increasing pressure at depths down to a few km, while anisotropy is also significant at greater depths.

[20] The anisotropic structure of the study area is likely to be a combined effect of deformations that occurred during both the Variscan and Caledonian events, characterized by a similar, SSW-NNE compression direction, orthogonal to the present fast velocity azimuth. It is hard to estimate which event contributed the most to the anisotropy. As mentioned above, several authors point out that the Neoproterozoic/Lower Cambrian rocks of the MU show traces of a much stronger deformation and steeper stratal dips than overlying Palaeozoic sequences. This would point to a major role of Early Caledonian deformation. However, as the stratigraphic studies show that the ŁU was not affected by this event, Late Caledonian and Variscan deformations have to be considered as an equally important contribution, at least in the ŁU area.

[21] Summarizing, the azimuthal dependence of upper crustal velocity observed in SE Poland can be explained by a transverse isotropic model with a horizontal symmetry axis and fast plane oriented in WNW-ESE direction. Observed crustal anisotropy will affect the passive and active source seismic data interpretation, in particular it will bias results of isotropic 2-D or 3-D wide-angle modelling. Petrologically, the anisotropic upper crust is likely to be composed of Neoproterozoic and Palaeozoic sediments and low grade metasediments, as shales (in the uppermost part), phyllites or slates, and by medium grade rocks (mica schists, gneisses) at larger depths, strongly folded due to the compression and crustal shortening during the Caledonian and/or Variscan deformational events.

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