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Laser Doppler Interferometry (LDI) to obtain full stiffness tensor: A case study on a deformation zone in Sweden

Pouya Ahmadi
Previously: Department of Earth Sciences, Uppsala University, Uppsala, Sweden
Present Address: Department of Applied Geophysics, Curtin University, Perth, Western Australia
E-mail: pouya.ahmadi@postgrad.curtin.edu.au

Alireza Malehmir
Department of Earth Sciences, Uppsala University, Uppsala, Sweden
E-mail: alireza.malehmir@geo.uu.se

SUMMARY

Estimation of elastic anisotropy, which is usually caused by rock fabrics and mineral orientations, has an important role in exploration seismology and a better understanding of crustal seismic reflections. If not properly taken care of during data processing steps, it leads to wrong interpretation and/or distorted seismic image. In this work, a state-of-the-art under the development Laser Doppler Interferometer (LDI) device is used to measure phase velocities on the surface of rock samples from a major poly-phase crustal scale deformation zone (Österbybruk Deformation Zone) in the Bergslagen region of eastern Sweden. Then, a general inversion code is deployed to invert the measured phase velocities to obtain full elastic stiffness tensors of two samples from the deformation zone.

At the end, results are used to correct for the anisotropy effects using three dimensionless Tsvankin's parameters and a non-hyperbolic moveout equation. The resulting stacked section shows partial reflection improvement of the deformation zone compared with the traditional isotropic processing approach. This illustrates that rock anisotropy contributes to the generation of the reflections from the deformation zones in the study area although they do not show significant density contrast with their surrounding rocks.

Key words: Anisotropy, Laser Doppler Interferometry (LDI), Deformation Zone, Sweden.

INTRODUCTION

A new method for conducting laboratory measurements of P- and S-waves velocities and polarisations in rock samples using a Laser Doppler Interferometer (LDI) device was introduced by Lebedev et al. (2011). Using the equipment, one is able to measure particle velocities excited by a P- or S-wave transducer in a small portion of a sample along the direction of the laser beam and transform them into a 3-component (3C) particle/velocity vector (if three orthogonal measurements are done; Figure 1a). These measurements are useful because they can provide information about the stiffness tensor especially for anisotropic rock samples that combined with the density of the material could provide a better understanding of elastic wave propagation within the samples. Malehmir et al. (2013) recently investigated petrophysical properties of a major deformation zone (Osterbybruk Deformation Zone) in eastern Sweden. Using the LDI device (Figure 2c), they suggested up to 10% velocity-anisotropy associated with a major reflective package generated from the deformation zone in the seismic data (Malehmir et al. 2011). The origin of the package was linked to a combination of anisotropy and amphibolite lenses within the deformation zone (Malehmir et al., 2011 and 2013). However, their seismic data processing did not account for the anisotropy and a full exploitation of the LDI measurements was not provided.

The main objectives of the present study, therefore, are: (1) to invert phase velocity data from the LDI measurements for elastic stiffness tensor without assuming any kind of symmetry as *a priori* information, (2) to include anisotropy parameters in the processing for a comparison with the original processing work (Malehmir et al., 2011). We, for example, observe that including some percentage of anisotropy will partially improve the seismic image of the deformation zone, which further supports the idea that the deformation is anisotropic to also the seismic wave.

METHOD AND RESULTS

Different experiments and theoretical investigations have been carried out on different materials attempting to estimate their full elastic stiffness tensors. One of the successful ones was by Vestrum (1994).

He used a generalised linear inversion method to calculate all the elements of the tensor. He assumed no symmetry and obtained all components of the tensor. Here we used a modified version of Vestrum's original code. The code is relatively straightforward and uses no assumption about the symmetry of the sample in order to recover all the 21 components of the elastic stiffness tensor. Using LDI device, we measured phase velocities (P, S1 and S2) on the sample surface (Figure 1a). In the example shown in Figure 1, we clearly observe shear-wave splitting (S1 and S2), which is already an indication of anisotropy in the sample. The code starts with an elastic stiffness tensor as an initial guess after linearizing the problem around the initial guess; using a linear least-square inversion method, it will find out a new guess in

which the average squared phase velocity error (calculated with the measured ones) is minimized. This is one in an iterative manner until it reaches to a stage that the average squared error stops decreasing.

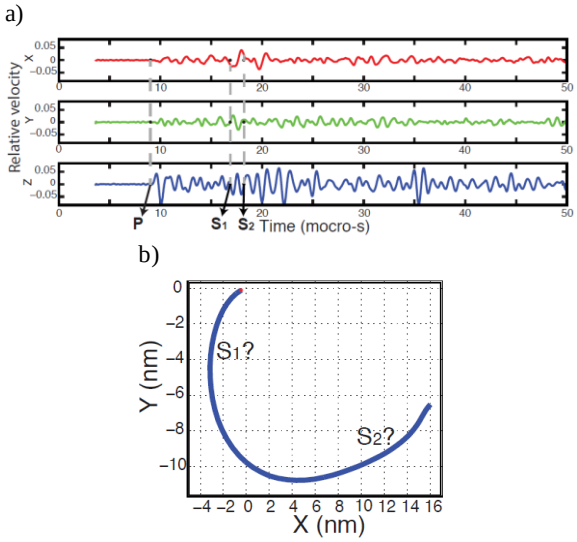

Figure 1 a) Velocities of the particle motion on the surface of the sample shown in Figure 2b at direction x, y and z against time. b) Hodogram showing S-wave splitting of S1 and S2 registered by the LDI device. We used a time window between 17.42 to 18.50 micro-s to produce this hodogram.

In the inversion part, Christoffel's equation is used for the forward modelling:

$$\Gamma_{il} A_l = \rho v^2 A_i, \tag{1}$$

where A_i are the components of the amplitude of the particle displacement vector, Γ_{il} are the components of a second order matrix that depends on the elastic constants via equation (2):

$$\Gamma_{il} = C_{ijkl} n_j n_k. \tag{2}$$

A plane wave propagates in the direction of a unit vector, n, and C is the fourth rank elastic stiffness tensor with 81 components, which has 21 independent components.

Tsvankin (1997) introduced the dimensionless anisotropic parameters (ε, σ and γ) preserving all benefits of Thomsen notation (Thomsen, 1986) in wave propagation for orthorhombic media, which is used here to introduce anisotropy into the processing steps and at the same time link the stiffness tensor to the real seismic data with the non-hyperbolic moveout equation as a correction for the NMO function in the case of dipping reflectors. Equations 3 and 4 summarize the link between these parameters:

$$\varepsilon \equiv \frac{C_{11} - C_{33}}{2C_{33}}, \quad \sigma \equiv \frac{(C_{13} - C_{55})^2 - (C_{33} - C_{55})^2}{2C_{33}(C_{33} - C_{55})}, \quad \gamma \equiv \frac{C_{66} - C_{44}}{2C_{44}} \tag{3}$$

$$\eta \equiv \frac{\varepsilon - \sigma}{1 + 2\sigma}, \quad t^2(x) = t_0^2 + \frac{x^2}{[V_{NMO}(0)]^2} - \frac{2\eta x^4}{[V_{NMO}(0)]^2 \{t_0^2 [V_{NMO}(0)]^2 + [1 + 2\eta] x^2\}} \tag{4}$$

where η is the anisotropic coefficient and $V_{NMO}(0)$ is the NMO velocity which can also be dip-dependent.

Results

Here, we present an example result for a data set on a sample from the deformation zone (Figure 2b). Figure 2a shows geological map of the study area and the locations of the seismic profile and the sample (number 2) used in this study. The sample is relatively felsic in composition and shows clear indications of lamination in the horizontal direction.

The elastic stiffness tensor of the sample was calculated using the inversion code in two-index Voigt notation and provided here (in GPa):

$$c = \begin{pmatrix} 89.9326 & 13.2261 & 23.4326 & -0.0076 & 0.0076 & 0.0000 \\ 13.2261 & 89.9247 & 23.4294 & -0.0033 & 0.0018 & 0.0000 \\ 23.4326 & 23.4294 & 79.7626 & 0.0041 & -0.0094 & 0.0000 \\ -0.0007 & -0.0033 & 0.0041 & 55.9497 & 0.0000 & 0.0028 \\ 0.0076 & 0.0018 & -0.0094 & 0.0000 & 54.4949 & -0.0010 \\ 0.0000 & 0.0000 & 0.0000 & 0.0028 & -0.0010 & 40.1881 \end{pmatrix}. \tag{5}$$

To simplify the result and prior to highlighting the effect of anisotropy from the lab data to the real seismic data, we take your attention in following:

- The importance of different frequencies used in the lab and in the seismic data should not be underestimated and its contribution (and to what extent) should further be investigated using real field 3C data;
- One or a few samples (we used five samples in our study) may not be representative of the whole deformation zone thus our results provide first order information about the anisotropy until more data and experiments are conducted;
- In this study, we assumed the nearest form of anisotropy, orthorhombic here, to the inversion result for the data processing to be able to use the dimensionless anisotropic parameters and the non-hyperbolic moveout equation.

By simplifying the elastic stiffness tensor shown above to an orthorhombic form, we obtained the following tensor and estimated the required corrections based on the equation (4):

$$C = \begin{pmatrix} 89.933 & 13.226 & 23.433 & & & \\ 13.226 & 89.925 & 23.429 & & 0 & \\ 23.433 & 23.429 & 79.763 & & & \\ & & & 55.950 & & \\ & 0 & & & 54.495 & \\ & & & & & 40.188 \end{pmatrix}. \tag{6}$$

Figures 3a and 3c show portions of the unmigrated stacked section crossing the deformation zone and the reflective package from and above it. Figures 3b and 3d show partially improved images of the same section but obtained by introducing the anisotropy parameters estimated from the LDI experiment and the inversion result.

DISCUSSION AND CONCLUSIONS

As shown in Figure 3, introducing anisotropy to the processing steps at least partially improves the unmigrated stacked section crossing the deformation zone. To further check the anisotropy signature of the deformation zone, velocities of the first arrivals using their offsets were considered and plotted against limited azimuth coverage (Figure 4a and 4b). The velocity versus azimuth plot is shown

in Figure 4b. A 2nd order sinusoidal curve can sufficiently fit the data.

In general, two prominent cycles can be recognized in the velocity-azimuth plot suggesting a high velocity pattern striking in NE-SW direction, which is consistent with the strike of the ÖDZ (Figure 2a) when it is crossed by the seismic data. Reprocessing results and these analysis both illustrate that the deformation zone is strongly anisotropic and introducing anisotropy into the processing steps not only provide improved image but also to some extent necessary and important to better image the internal reflectivity of the deformation zone.

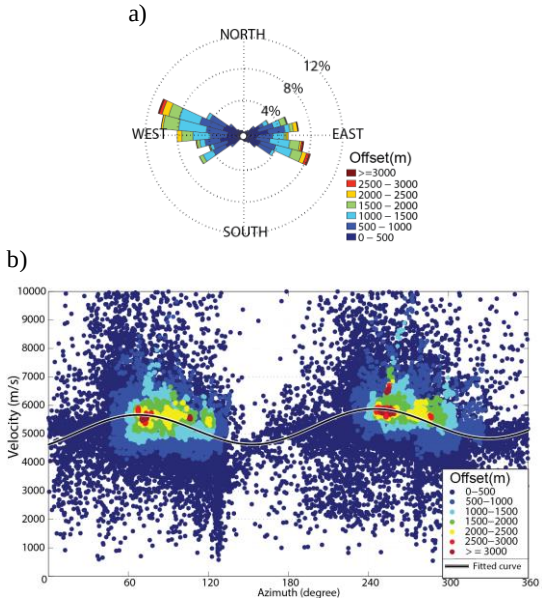

Figure 4: a) Azimuth-offset coverage of the Dannemora reflection seismic data along a portion of profile 1 that crosses the deformation zone. b) P-wave first arrival velocities calculated over the deformation zone with respect to their azimuths. Note that each dot represents one P-wave arrival velocity and the black line is a fitted curve using MATLAB curve fitting application. The graph suggests an NE-SW anisotropy direction that correlates well with the strike of the ÖDZ observed in the geological map (Figure 2a). Note that some of the very high velocities are likely due to the bedrock dip (apparent velocity) and not from the subsurface materials

To our knowledge, this is the first time that a general inversion code is used for LDI measurements to obtain stiffness tensor. The inversion was stable and fast likely due to the rich and often consistent input data. Inversion results suggest an orthorhombic media, consistent with the previous studies suggesting similar anisotropy system for major deformation zones. This was further supported by the new processing results showing partial reflection improvement when the orthorhombic anisotropy parameters were introduced. Future studies should aim at better understanding the relationship between the frequency and lab measurements for rock samples from crystalline rock environment often characterized by low-porosity and high-degree of solidification.

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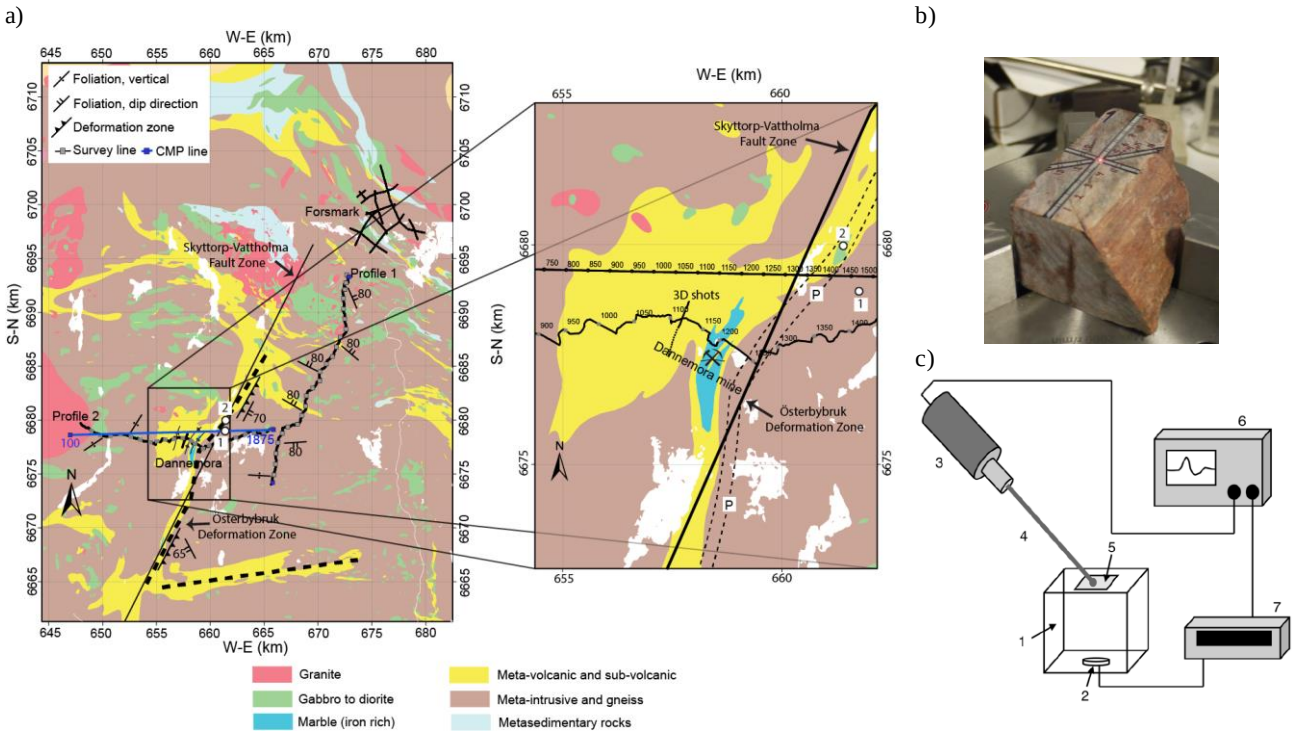

Figure 2 a) Geological map of the study area (zoomed in the middle figure). Crooked-black line is the location of the seismic profile; dashed lines show the inferred location of the deformation zone. A sample from location 2 is shown and investigated in this paper (modified from Malehmir et al., 2011). b) An example sample measured during the experiment (from Malehmir et al., 2013). c) Experimental setup, 1: sample, 2: source of elastic waves, 3: Laser Doppler Interferometer, 4: laser beams (emitted and backscattered), 5: reflective film, 6: acquisition system, and 7: pulse generator (modified from Lebedev et al., 2011).

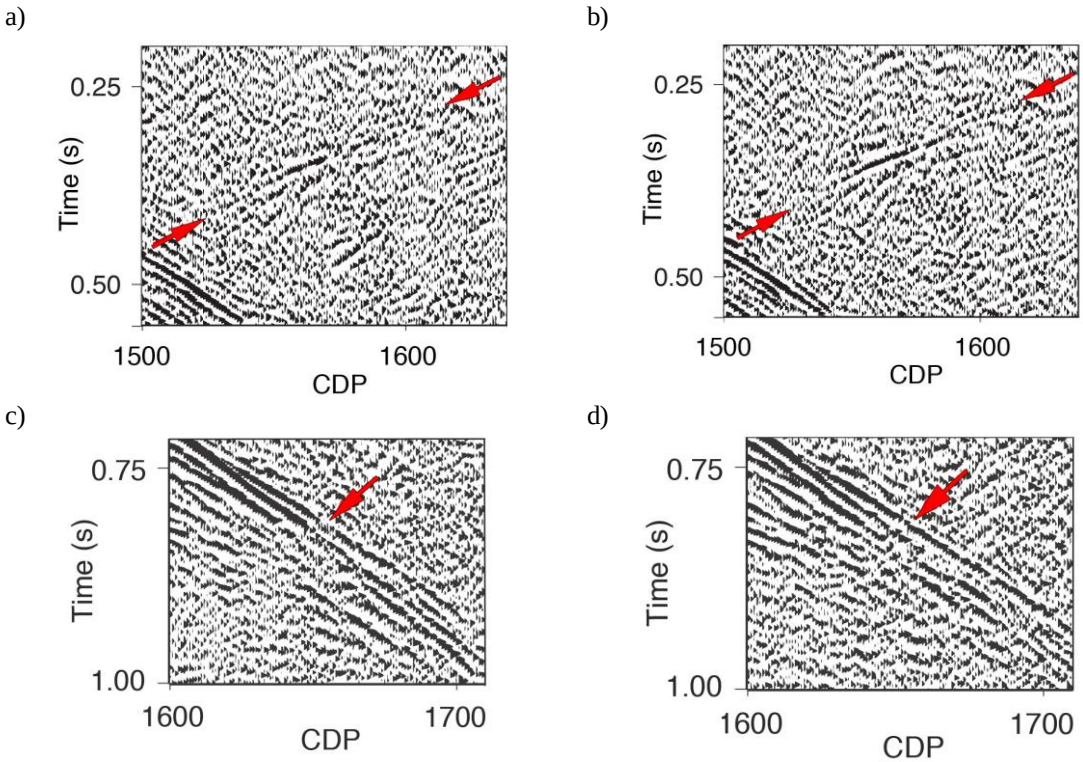

Figure 3 Portions of the unmigrated stacked section crossing the deformation zone a) and c) before, b) and d) after introducing the anisotropy parameters into the processing steps (this study). Note partial improvements marked on the section by the red arrows. CDP spacing is 10 m.