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Interpretation of Pn-Velocity Residuals in the Baltic Shield

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Abstract

Controlled travel time data from the Fennolora Long-Range Profile, observed at the Swedish and Finnish permanent station networks, are employed to study Pn-travel time residuals in the Baltic Shield with reference to the crustal model of Båth (1979). Interpretation of Pn-velocity variations in terms of a uniform dipping Moho discontinuity or anisotropic wave propagation, or combinations of both, is briefly discussed.

Pn-travel time residuals, relating observations to Båth's (1979) model, tend to decrease with increasing epicentral distance. Least-squares fit provides a new Pn-velocity of 8.17 km/s and crustal thickness of 43 km, to be compared to 7.84 km/s and 38 km, respectively, in Båth's model. Pn-travel time residuals, relating observations to the new model, do not provide any further information about properties (dip, anisotropy) at the base of the crust (Moho).

1. *Introduction*

The Fennolora Long-Range Profile experiment was carried out during August 1979, supplying seismic refraction data for detailed investigations of the crust and upper mantle in Northern Europe. Technical data of the Fennolora shots are provided by *Guggisberg* (1986). The interpretation of Fennolora data by *Guggisberg* (1986) should be considered most comprehensive. The 27 explosions from 8 different explosions sites were also recorded at the permanent seismograph stations in Sweden and in Finland, the sites of which are geographically well distributed over the Baltic shield area.

The average crustal model of *Båth* (1979), which is usually applied by the Seismological Department in Uppsala for epicentral location of seismic events in Fennoscandia, includes a *Pn*-velocity of 7.84 *km/s*. This *Pn*-velocity appears rather low in comparison to *Pn*-velocities given by other authors. Velocities of 8.55 *km/s* by *Brown et al.* (1971) for southern Fennoscandia, 8.0-8.3 *km/s* by *Sellevoll* (1973) for Fennoscandia and adjacent parts of the Norwegian Sea, 8.32 *km/s* by *Lund* (1979) for the Blue Road profile, 8.27 *km/s* by *Båth* (1984) for a profile in Swedish Lapland, 8.0-8.2 *km/s* by *Luosto et al.* (1985) for

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the *BALTIC* profile in *SE* Finland, 8.04 *km/s* by *Luosto* (1986) for the Sylen-Porvoo profile in *SW* Finland, 8.0-8.05 *km/s* by *Grad and Luosto* (1987) for the *SVEKA* profile in central Finland or 8.1-8.3 *km/s* by *Lund* (1987) for the northern part of the Fennolora profile, may serve as some examples.

The Fennolora experiment offers a large number of combinations between explosions sites and seismic stations, thus covering an appreciable amount of different azimuths and epicentral distances in the Baltic shield area. Our aim was to find an average *Pn*-velocity in comparison to the low average *Pn*-velocity given by *Båth* (1979), and possible lateral variations of *Pn*-velocities. Lateral variations of *Pn*-velocity are of certain interest, since they reflect physical properties at the base of the crust, such as dipping Moho discontinuity and/or anisotropic wave propagation. For instance, *Brown et al.* (1971) interprete variations of apparent *Pn*-velocities, observed along seismic profiles with different azimuthal directions, by an appreciable regional Moho dip in South-Central Sweden. Furthermore, the anisotropic nature of the crust and upper mantle, including even temporal changes, is today widely accepted. Convincing evidence is, for instance, provided by *Crampin* (1987), giving a comprehensive summary of extensive dilatancy anisotropy (*EDA*), deduced from the method of shear wave splitting. There is today very few geophysical evidence for anisotropy in Fennoscandia. Some indication of an anisotropic lower crust in the Baltic Shield is mentioned by *Rasmussen* (1987), from the interpretation of apparent resistivities derived from the magnetotelluric method. *Brooks et al.* (1987) observed shear wave splitting at the northern section of the Fennolora experiment which they interpreted as a crack-induced anisotropy according to the *EDA* model (*Crampin et al.*, 1984). Seismological support for *Pn*-velocity anisotropy in northwestern Europe is presented by *Bamford* (1977) for Western Germany and by *Bamford et al.* (1979) for Northern Britain.

2. *Notation*

V_p	apparent *Pn*-velocity
C_p	isotropic *Pn*-velocity
$V_{1,2}$	crustal *P*-velocities, 6.22 *km/s* and 6.64 *km/s*, respectively
t_{Pn}	travel time, *Pn*
t_o	time intercept, *Pn*
Δt	travel time residual
Δ	epicentral distance
Δc	critical distance = minimum distance of *Pn*-phases
Z_1, Z_2	crustal layer thicknesses

$$A = \frac{1}{8}(3C_{1111} + 3C_{2222} + 2C_{1122} + 4C_{1212})$$

$$B = \frac{1}{2}(C_{1111} - C_{2222})$$

$$C = C_{2111} + C_{1222}$$

$D \quad = \frac{1}{8}(C_{1111} + C_{2222} - 2C_{1122} - 4C_{1212})$

$E \quad = \frac{1}{2}(C_{2111} - C_{1222})$

θ azimuth, clockwise direction from the north

θ_1 azimuth, clockwise direction from plane of symmetry (updip direction)

ε Moho dip

3. *Observational material*

Pn observations employed in this work are deduced from recordings of the permanent seismograph station networks of Sweden and Finland. The azimuthal coverage, combining the 8 explosion sites with the 10 seismic stations is not as complete as one would desire. Especially the range 270° - 360° is poorly covered. First *P*-arrivals were measured by experienced seismologists independently from each other. Some few cases which caused disagreement (weak onsets) were discussed, either homogenized or excluded. The epicentral distance at which *Pn*-phases preced P_g-phases is expected at about 200 *km*. Thus, only data for $\Delta \geq 220$ km are considered.

4. *Data interpretation*

Travel time residuals Δt of *Pn*-phases are defined by the difference of observed travel time minus travel times according to the crustal model of *Båth* (1979). In Fig. 1 Δt values are displayed versus azimuth (*a*) and versus epicentral distance (*b*). From Fig. 1b the tendency of decreasing Δt with increasing epicentral distance is obvious. This fact certainly reflects the low *Pn*-velocity of 7.84 *km/s* in Båth's model. With regard to other higher *Pn*-velocities for Fennoscandia mentioned earlier and considering the data in Fig. 1b we shall propose a revision of Båth's model. *Pn*-travel times t_{Pn} may be written

$$t_{Pn} = t_o + \frac{\Delta - \Delta_c}{C_p}\ [s]. \qquad (1)$$

Numerical values for the critical distance Δ_c, corresponding time intercept t_o and *Pn*-velocity C_p are in Båth's model equal to 110 *km*, 20.8 *s* and 7.84 *km/s*, respectively. Generally it is preferred to include the term Δ_c/C_p into the time intercept. In that case the time intercept is defined at $\Delta = 0$. In the present work, however, we prefer the formulation according to Eq. (1), for the sake of the subsequent computations.

Following Båth's model we write

$$t_{Pn,Båth} = 20.8\, s + \frac{(\Delta\, km - 110)}{7.84\, km/s}\, [s]. \tag{2}$$

A linear regression $a+b\,(\Delta\text{-}110)$ to the data in Fig. 1b can directly be added to Eq. (2). We obtain

$$a = 0.7\, s$$

$$b = -5.15 \cdot 10^{-3}\, s/km$$

Adding this result to Eq. (2) we can write (NM = new model)

$$t_{Pn,NM} = 21.5\, s + \frac{(\Delta\, km - 110)}{8.17\, km/s}\, [s]. \tag{3}$$

This formula gives the best fit to the observations. However, the critical distance Δ_c has now been changed from 110 *km* to 98 *km*, which is obtained from the relation

$$\Delta_c = 2\,(Z_1 \frac{V_1}{C_p \sqrt{1-(V_1/C_p)^2}} + Z_2 \frac{V_2}{C_p \sqrt{1-(V_2/C_p)^2}}). \tag{4}$$

Here we accept the crustal velocities V_1 = 6.22 *km/s* and V_2 = 6.64 *km/s* given by *Båth* (1979), thus changing only the *Pn*-velocity. Similarly we find the new time intercept, t_o = 19.2 *s*, according to

$$t_o = 2\,(\frac{Z_1}{V_1 \sqrt{1-(V_1/C_p)^2}} + \frac{Z_2}{V_2 \sqrt{1-(V_2/C_p)^2}}). \tag{5}$$

While still accepting Båth's crustal velocities V_1 and V_2 we now find a crustal thickness which fits to Eq. (3). Using the same model of a two layered crust with equal thicknesses as proposed by *Båth* (1979), i.e. $Z_1 = Z_2$, we obtain the new layer thicknesses from the ratio of time intercepts, or the ratio of critical distances, multiplied by previous layer thickness.

Thus

$$Z_1 = Z_2 = \frac{21.5}{19.2} \cdot 19\, km = 21.3\, km$$

or

$$Z_1 = Z_2 = \frac{110}{98} \cdot 19\, km = 21.3\, km.$$

Summarizing this part we conclude that a *Pn*-velocity of 8.17 *km/s* and a crustal thickness of about 43 *km* (21.3 *km* + 21.3 *km*) provide the best fit to the observations.

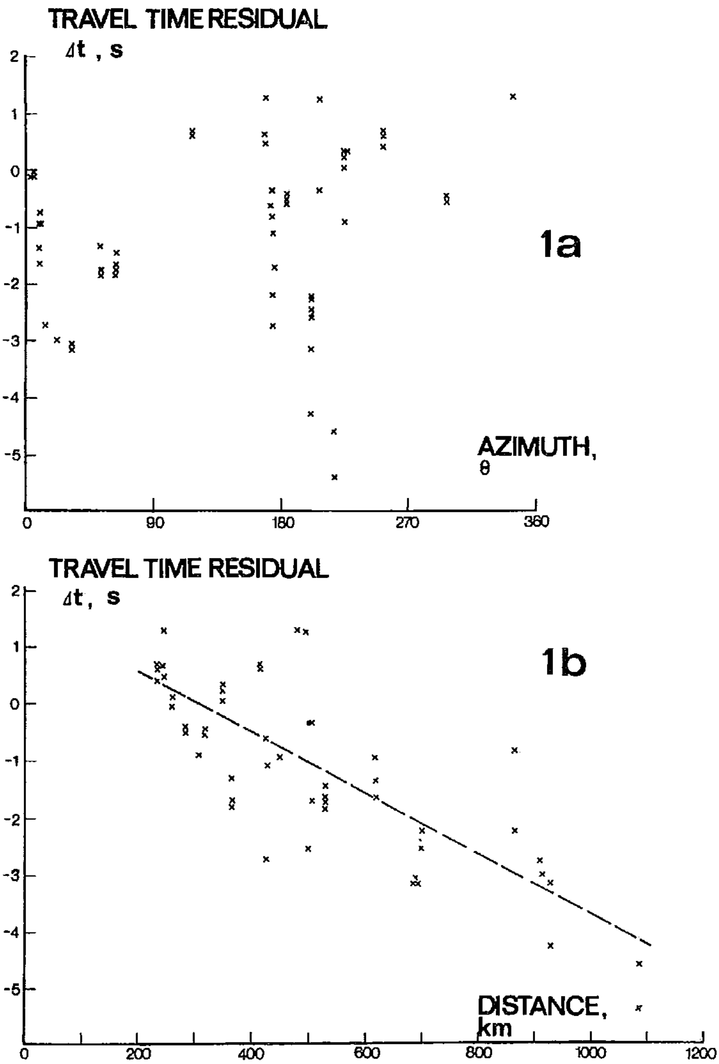

Fig. 1. *Pn*-travel time residuals Δt, observed minus calculated due to crustal model of *Båth* (1979).
a) Δt versus azimuth θ.
b) Δt versus epicentral distance Δ. The dashed line is the least squares regression, $\Delta t = 0.7 - 0.00515\ (\Delta - 110)$.

5. *Lateral variation for apparent Pn-velocity*

5.1 *Dipping layer*

The simplest case of azimuthal changes of apparent *Pn*-velocities can be described by the model of a dipping Moho discontinuity. Following *Brown et al.* (1971) we find that

$$V_p^2\left(\left(C_p^2/V_2^2-\cos^2\theta_1'\right)\sin^2\varepsilon-\left(C_p^2/V_2^2-1\right)^{1/2}\cos\theta_1'\sin^2\varepsilon+\cos^2\varepsilon\right)=C_p^2 \tag{6}$$

The azimuth is determined by θ_1', which is the angle between the ray and the updip direction. Since this angle is measured in the dipping plane (*cf. Brown et al.*, 1971) it is not exactly the true azimuth. However, the true azimuth θ_1 differs from θ_1' by less than 0.5° if e < 10°. Thus, θ_1' ¢ θ_1

The azimuthal variation of apparent *Pn*-velocity can be expressed by the ratio V_p/C_p. It obtains a maximum in the direction of updip of the Moho discontinuity and a minimum in downdip direction (See Fig. 2). The magnitude of V_p/C_p depends on the dip angle (ε). For an isotropic *Pn*-velocity of 8.17 *km/s* the ratio V_p/C_p changes by ± 1.5 % for $\varepsilon = 1°$ and by ± 7 % for $\varepsilon = 5°$ (cf. Fig. 3).

5.2 *Anisotropic wave propagation*

The general case of body wave propagation in anisotropic elastic media (*Crampin*, 1977) can be simplified for the case of a weak anisotropy of *P* refractor velocity. This has been shown by *Backus* (1965) and by *Crampin* (1977). In this particular case the anisotropic *Pn*-wave velocity, V_p, can be written

$$V_p^2 = C_p^2 + A + B\,cos(2\theta) + C\,sin(2\theta) + D\,cos(4\theta) + E\,sin(4\theta) \tag{7}$$

The variation of *Pn*-velocity, $V_p^2 - C_p^2$, is represented by the first five terms of the Fourier series of a function repeating itself every 180°. The coefficients *A*, *B*, *C*, *D* and *E* represent linear combinations of elastic constants which connect stress and strain tensors.

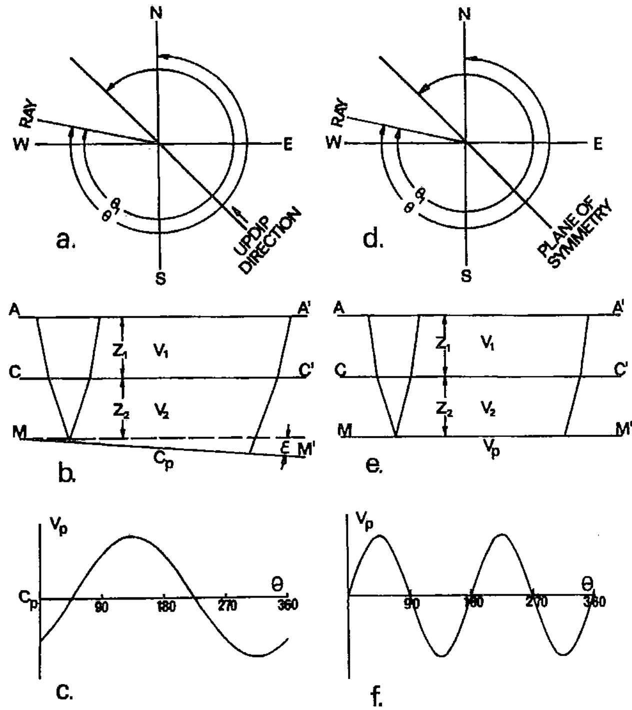

Fig. 2. a) Horizontal crossection with dipping plane (Moho), arbitrary updip direction E 45° S to W 45° N. θ denotes azimuth.

b) Vertical crossection (2-layered crust) along the direction of dip and propagation path of Pn-waves; A-A' = earth surface, C-C' = Conrad discontinuity, M-M' = Moho discontinuity; V_1, V_2 are crustal velocities, Z_1, Z_2 layer thicknesses, ε denotes dip angle and C_p is the Moho phase (Pn) velocity.

c) Schematic apparent Pn-wave velocity versus azimuth θ according to updip in a.

d) Horizontal crossection for anisotropic Pn-wave velocity, V_p, arbitray plane of symmetry E 45° S to W 45° N. θ_I is the azimuth measured clockwise from the plane of symmetry.

e) Vertical crossection, notation as in b., V_p is anisotropic Pn-wave velocity.

f) Schematic apparent Pn-wave velocity versus azimuth θ according to the assumed direction of the plane of symmetry.

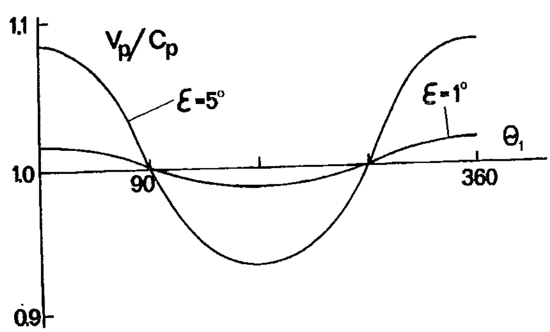

Fig. 3. Azimuthal variations of the ratio of apparent, V_p, and isotropic, C_p, Pn-wave velocities versus azimuth θ_1 (clockwise updip direction), for different Moho dip, ε = 1° and ε = 5°.

5.3 *Complex Moho structures*

In a real case it seems rather unlikely to assume that either a large scale dip of the Moho discontinuity (case 1) or anisotropic wave propagation (case 2) prevail. Recent work about Moho depth in Fennoscandia (*Jentzsch*, 1986) suggests quite inhomogeneous distribution of crustal thicknesses, implying a non-uniform, undulating structure of the Moho. If anisotropic wave propagation due to the geological microstructure is superimposed onto a widely varying, inhomogenous distribution of Moho depth, the exact interpretation of lateral variations of *Pn*-velocities becomes certainly difficult, maybe impossible. Assuming this complex type of Moho structures one may attempt to minimize the travel time residuals of *Pn*-waves by a least-squares procedure, thus changing the unknown parameters in Eqs. (6) and (7) until a minimum residual is achieved.

5.4 *Interpretation of travel time residuals*

After the first approach to minimize *Pn* travel time residuals by changing the crustal model in the previous part, we shall now examine the remaining travel time residuals in terms of lateral *Pn*-velocity variations mentioned above. The travel time residual is now defined as the time difference of observed *Pn*-phases and calculated *Pn*-phases according to the new model. These residuals are plotted in Fig. 4. There is no obvious θ (uniform Moho dip) or 2 θ (*Pn* anisotropy) dependence. Since the remaining residuals are also small, about $|1|s$ in average, no further interpretation seems justified.

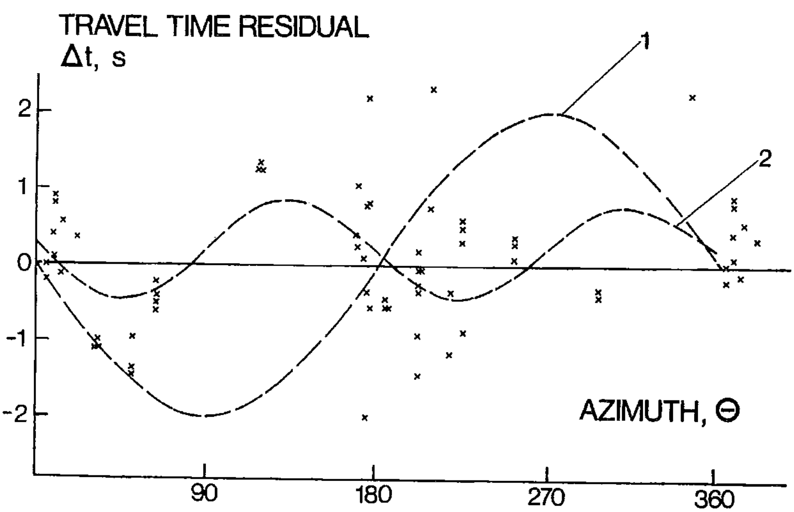

Fig. 4. *Pn*-travel time residuals Δt, observed minus calculated due to the new model. The dashed lines demonstrate (non-significant) θ (1) and 2θ (2) dependent least-squares solutions.

6. *Discussion*

Recent results about Moho depth variations in Fennoscandia suggest rather inhomogeneous distribution of the crustal thickness (*Guggisberg*, 1986; *Jentzsch*, 1986). This is obviously not supporting any global uniform Moho dip hypothesis which has been preferred earlier. If an undulating Moho discontinuity also includes anisotropic features the interpretation of lateral *Pn*-velocity variations becomes a difficult task to solve. The insufficient amount of data, including lack of observations in the azimuth range from 270° to 360°, are contributing factors and thus no conclusions can be reached concerning the nature of lateral variations of *Pn* travel time residuals.

On the other hand, the dependence of *Pn*-travel time residuals on epicentral distance is obvious, implying a larger *Pn*-velocity of 8.17 *km/s* compared to the *Pn*-velocity of 7.84 *km/s* given in Båth's model (1979). In the presence of larger *Pn*-velocities proposed by other authors (for instance, *Brown et al.*, 1971; *Sellevoll*, 1973; *Lund*, 1979; *Båth* 1984; *Luosto et al.*, 1985; *Luosto*, 1986; *Grad and Luosto*, 1987; *Lund*, 1987) the change from 7.84 *km/s* to 8.17 *km/s* seems also justified. Concerning the proposal of a crustal thickness of 43 *km* this value is somewhat lower than the average Moho depth given by *Jentzsch* (1986).

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