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REINTERPRETATION OF THE TRAVEL TIMES OF *P* WAVES GENERATED BY QUARRY BLASTS IN WESTERN BOHEMIA

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ABSTRACT. Using travel times of seismic body waves generated by quarry blasts, (Horálek et al., 1987) proposed a homogeneous half-space as a model for the region of the West-Bohemian earthquake swarms. In the present paper, the same experimental data are analyzed in a greater detail. It is shown that the assumption of a homogeneous half-space is not fully compatible with the observed data. Simple vertically inhomogeneous models are proposed as better models of the medium. Consequences for the accuracy of the earthquake location are discussed.

KEYWORDS: Western Bohemia, body wave, travel time, homogeneous model, vertically inhomogeneous model, earthquake location

1. INTRODUCTION

A homogeneous half-space model was suggested by (Horálek et al., 1987) to represent the real medium in locating the shocks from the West–Bohemian earthquake swarms. To determine the compressional velocity v for the mentioned model, they used the travel times of the corresponding waves from quarry blasts of known position and origin time to the local seismic network.

The position of the shot–points and seismic stations is shown in Fig. 1. Their geographic coordinates and altitudes are given in Tab. 1. Moreover, this table contains also the coordinates of these points in a local Cartesian system. The origin of the local system is selected at the point of 50° N and 12° E, the x–axis is oriented eastwards and the y–axis northwards. The arc of 1° along the parallel (x–axis) is assumed to have a length of 71.322km, and along the meridian (y–axis) to have a length of 111.248 km; these lengths correspond to a mean latitude of 50.25° N according to the tables in (Valouch, 1958).

The P–wave travel times obtained by (Horálek et al., 1987) are given in Tab. 2 and shown in Fig.2. Using a straight–line approximation of the travel times, these authors arrived at a velocity of $v = 5.757$ km/s. The homogeneous half-space with this velocity has been used in routine location of many shocks and, despite its simplicity, it seems to give acceptable results (Horálek et al., 1995).

The deviation of the observed travel–time points from a mean line is rather high (Fig. 2). The reason is that due to a small number of the measurements, the authors had to put together data from different parts of the investigated region with

TABLE 1. Seismic stations and shot–points: names, abbreviations, latitudes φ and longitudes λ (in degrees), heights above sea level h, local Cartesian coordinates x and y described in the text

Name	Abb.	φ	λ	h (m)	x (km)	y (km)
Oloví	OL	50.2453 N	12.5588 E	528	39.85	27.29
Vackov	VA	50.2353 N	12.3786 E	525	27.00	26.18
N.Kostel	NK	50.2328 N	12.4483 E	566	31.97	25.90
Nejdek	NE	50.3136 N	12.6888 E	610	49.13	34.89
Vernéřov	VE	50.2225 N	12.2408 E	588	17.17	24.75
Wernitzgrün	WR	50.2881 N	12.3611 E	620	25.75	32.05
Klingenthal	KL	50.3711 N	12.4700 E	570	33.52	41.28
Rotava	RO	50.2994 N	12.5864 E	640	41.82	33.31
Libá	LI	50.1264 N	12.2289 E	617	16.33	14.06
Vintířov	VI	50.2094 N	12.7164 E	366	51.10	23.30

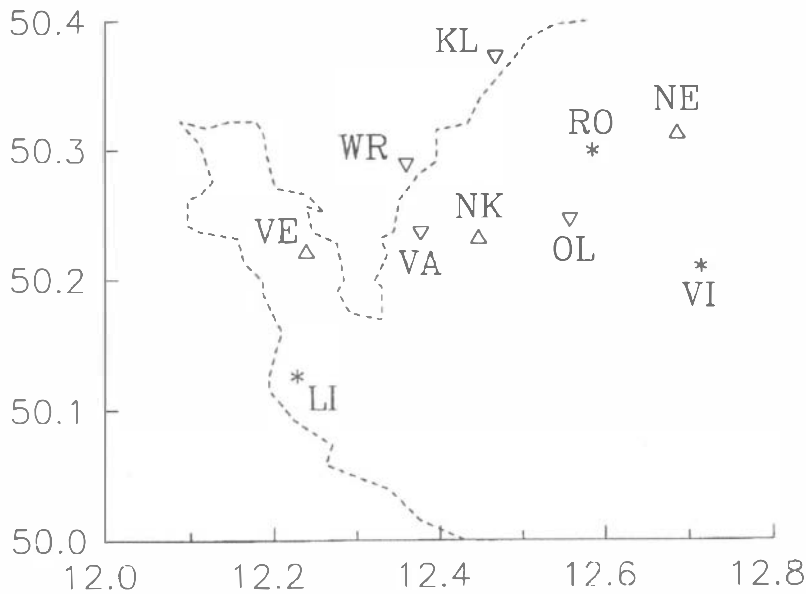

FIG. 1. The position of the shot–points (*), digital seismic stations (△) and analog seismic stations (▽). For the abbreviations see Tab. 1

slightly different structures. The velocity obtained must therefore be considered as an average representation of the whole region.

Although Horálek et al. (1987) investigated also the S waves, in this paper we shall restrict ourselves to the P waves only. We shall not consider the S waves, as their travel times are not known for short epicentral distances and, moreover, their onsets may be contaminated by transformed Sp waves, which will require a special analysis (Horálek et al., 1995). The P–wave travel–time data will be analyzed here in detail and used for the derivation of alternative structural models. The models considered in this paper are summarized in Tab. 3 and discussed in the following sections. Models A and B are homogeneous half–spaces, models C to E represent a homogeneous layer on a homogeneous half–space, and in model F the velocity increases continuously with depth. These models will also be used for estimating the accuracy of earthquake location.

2. AN ALTERNATIVE HOMOGENEOUS MODEL

Let us approximate the travel time values in Tab. 2 by a straight line of a general form, i.e. $t = k \cdot D + q$ where t is the travel time, D is the epicentral distance, k and q are unknown coefficients. Determining the coefficients by means of the least–squares method we arrive at the equation

$$t = 0.1737 \cdot D + 0.2550\,. \tag{1}$$

The reciprocal value of the first coefficient $v = 1/k = 5.757\,\mathrm{km/s}$ is the velocity adopted by (Horálek et al., 1987) as the velocity in a homogeneous half–space. This model is denoted as model A in Tab. 3.

TABLE 2. Epicentral distances and P–wave travel times

Dist. (km)	t_P (sec)	Profile
6.34	1.16	Rotava — Oloví
16.08	3.0	Rotava — Wernitzgrün
19.28	3.70	Vintířov — N.Kostel
19.28	3.72	Vintířov — N.Kostel
19.64	3.92	Libá — N.Kostel
24.25	4.58	Vintířov — Vackov
25.14	4.5	Vintířov — Klingenthal
25.14	4.7	Vintířov — Klingenthal
26.78	4.8	Vintířov — Wernitzgrün
26.78	4.9	Vintířov — Wernitzgrün
33.93	6.01	Vintířov — Vernéřov
33.93	6.03	Vintířov — Vernéřov
38.85	7.08	Libá — Nejdek

TABLE 3. Models of the medium: v_1 is the P–wave velocity (in km/s) and d_1 is the thickness (in km) of the first layer, v_2 and d_2 are the parameters of the second layer (if any)

Model	v_1	d_1	v_2	d_2
A	5.757	∞		
B	5.461	∞		
C	2.066	0.282	5.757	∞
D	4.507	0.924	5.757	∞
E	5.092	1.391	5.757	∞
F	model with a velocity gradient			

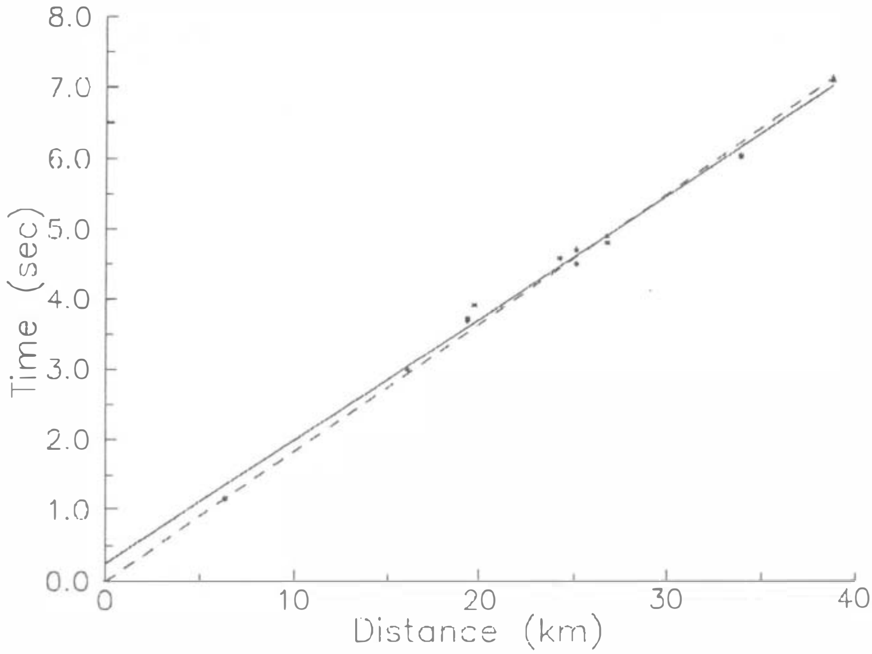

FIG. 2. Observed travel times (points) and their straight–line approximations: the solid line shows the line given by formula (1), the dashed line shows the line given by formula (2)

The mentioned straight line has, however, an intercept time of 0.255 s at the epicentral distance $D = 0$. This fact has some consequences, in particular, it influences the accuracy of earthquake location.

As a simple illustration, let us try to locate back the position of the source (quarry blast) from the travel times in Tab. 2, assuming all stations are arranged along a straight line (e.g., the x-axis) according to their epicentral distances. It means that we simultaneously neglect the altitude differences of the stations. Using the half-space velocity of 5.757 km/s, let us locate the source by means of the gradient method (Bakhvalov, 1973; Demidovich, Maron, 1966). For the length of the step along the gradient we use the value which is usually adopted in linear inverse problems (Ralston, 1973), although our problem is not linear. The same algorithm was used for interpreting dispersion curves of surface waves by (Novotný et al., 1995); a more detailed description of the method can be found there. In this case, for the fixed source depth $z_s = 0$ and origin time $t_s = 0$ we get a shift of the source from the correct value $x_s = 0$ to the position $x_s = -1.47$ km. For fixed $x_s = 0$ and $t_s = 0$ we even get much larger shift of the source from the correct depth $z_s = 0$ to the depth $z_s = 6.91$ km. For fixed $x_s = z_s = 0$ we get, instead of the correct origin time $t_s = 0$, the value $t_s = 0.255$ s, which is equal to the intercept time mentioned above. If all three parameters x_s, z_s and t_s are variable, the best results are again obtained for t_s close to 0.255 s and x_s and z_s close to zero. However, in this case the location process is less stable, as the source depths up to about 1 km yield also satisfactory results. By this shift of the origin time, the location process tries to eliminate the inconsistency between the assumption of the homogeneous half-space and the non-zero intercept time.

To cope with the problems described above, let us now approximate the travel time values in Tab. 2 by a straight line passing through the origin, i.e. by the line $t = k \cdot D$. Using again the least-squares method we arrive at

$$t = 0.1831 \cdot D\,, \tag{2}$$

which now yields a lower value of the half-space velocity, $v = 5.461$ km/s; see model B in Tab. 3 and the dashed line in Fig. 2. This model diminishes substantially the errors in the same backward location of the quarry blasts, discussed above. For fixed $z_s = t_s = 0$ we get only $x_s = -0.14$ km, for $x_s = t_s = 0$ we get $z_s = 2.61$ km and for $x_s = z_s = 0$ we get $t_s = 0.026$ s. Nevertheless, it can be seen that some problems still remain, especially in the source depth determination.

3. A Homogeneous Layer on a Homogeneous Half-Space

The straight line (1) approximates the observed travel times better than the straight line (2). The sum of squares of residuals is equal to $0.212\,\mathrm{s}^2$ for line (1), and $0.299\,\mathrm{s}^2$ for line (2). On the other hand, the model B, based on approximation (2), yielded better results in the location tests mentioned above.

We think that a simple solution of these problems is to keep the approximation of the observed travel times by the line (1), but to consider this line as the travel-time curve of a head wave, and not as the travel-time curve of the direct wave. Since the simplest model of the medium, in which a head wave can propagate, is a homogeneous layer on a homogeneous half-space, we shall consider such a model in this section.

Denote by v_1 and v_2 the P-wave velocities in the layer and in the half-space, respectively, and by d_1 the thickness of the layer. The travel times of the head wave then read the equation

$$t(D) = 2d_1\sqrt{v_1^{-2} - v_2^{-2}} + D/v_2\,, \tag{3}$$

where D is the epicentral distance. Extrapolating this travel–time curve to the zero epicentral distance we arrive at the intercept time

$$t_0 = 2d_1\sqrt{v_1^{-2} - v_2^{-2}}\,. \tag{4}$$

Assuming equation (1) to be the travel–time curve of the head wave, we get

$$v_2 = 5.757\,\mathrm{km/s}\,, \quad t_0 = 0.255\,\mathrm{s}\,. \tag{5}$$

Equation (4) can then be considered as one of the equations for determining the unknown layer parameters v_1 and d_1. The velocity v_1 could, in principle, be determined from the travel–time curve of the direct wave. However, considering all the measured travel time data as belonging to the head wave, we must proceed in another way.

According to the paper by (Horálek et al., 1995), the seismograms of the micro–earthquakes, recorded by the West–Bohemian local seismic network, display on their vertical components a distinctive P–polarized wave that precedes the S wave by up to 0.25 s. This wave was interpreted as an S–to–P converted wave. Assume that the conversion occurs at the interface of the layer and the half–space of our model when the incident S wave propagates from bellow. Denote this converted wave as the Sp wave, and by t_{dif} the time difference between the S and Sp arrival times.

To simplify the problem, let us consider only the vertically incident S wave (in spite of the fact that no conversion occurs in this limiting case). Now we have

$$t_{\mathrm{dif}} = d_1/v_1^* - d_1/v_1\,, \tag{6}$$

where v_1^* is the S–wave velocity in the layer. Assume the ratio of the P– to S–wave velocities to be $v_1/v_1^* = \sqrt{3}$. Then, for a given value of t_{dif}, Eqs (4) to (6) can be used for determining v_1 and d_1.

Let us choose three values of t_{dif} for the vertical incidence within the stated range, namely 0.1 s, 0.15 s and 0.2 s. In this way we arrive at three structural models denoted as C, D and E, respectively, the parameters of which are given in Tab. 3. The travel–time curves of the head waves in these models begin at epicentral distances of 0.217, 2.324 and 5.272 km (critical points), and intersect the travel–time curves of the direct wave at epicentral distances of 0.822, 5.292 and 11.234 km, respectively. However, only the intermediate model D of these three models will be considered in further discussions. Model C should be refused since it has a very low velocity in the first layer (Tab. 3). Model E also leads to some problems since the first arrival at the epicentral distance of 6.34 km should belong to the direct wave (this distance is less than the value of 11.234 km given above) and not to the head

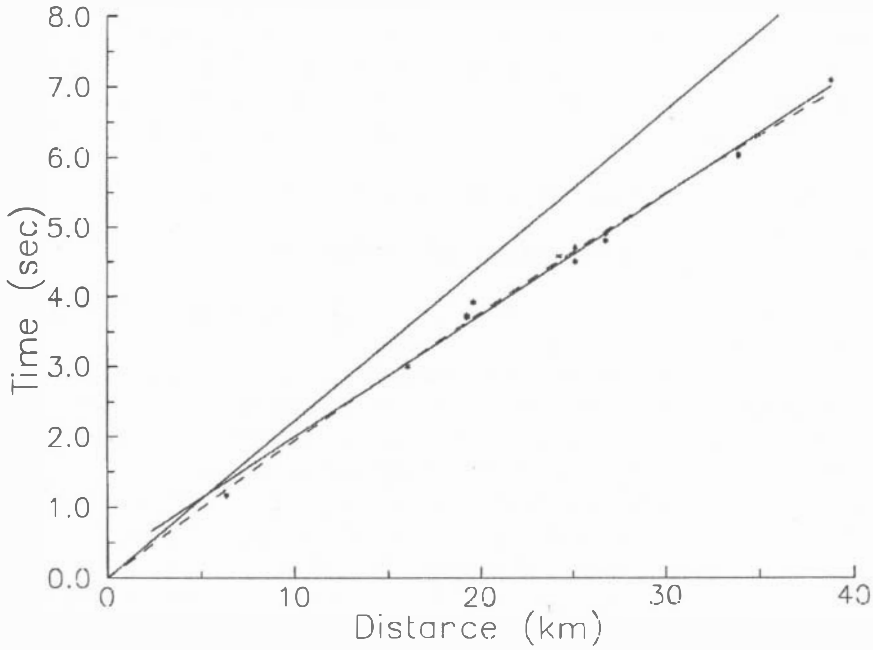

FIG. 3. Observed travel times (points) and their approximation by the direct and head wave travel times for model D (solid lines) and by the second order polynomial (dashed line)

wave, which we have assumed. Nevertheless, further investigations of the Sp wave are necessary to solve these problems.

The travel–time curves of the direct and head waves for model D (for $t_{\text{dif}} = 0.15$ s) are shown in Fig. 3, and the corresponding velocity–depth distribution is shown in Fig. 4. Another important seismic wave in this model, the reflected wave, has not been considered here, as its arrival times are later.

4. HALF–SPACE WITH A VERTICAL VELOCITY GRADIENT

Looking at the distribution of the measured travel time points (Fig. 2 or 3) it seems that a curved line could approximate these points better than a straight line. It provides us with another way to solve the problems discussed in Section 2. Any approximation of the data by a curved line will lead to a model with a vertical velocity gradient. For the sake of brevity, let us consider only parabolic approximations, $t = aD^2 + bD + c$. Two cases will again be distinguished, namely the case with the absolute term c retained and the case without this term.

The best–fitting parabola with the absolute term, representing the travel time data in Tab. 2, has the form

$$t = -0.00058 \cdot D^2 + 0.2007 \cdot D - 0.0196 . \tag{7}$$

It can be seen that the absolute term here is in its absolute value more than ten times smaller than the same term in (1). It means that the parabola passes fairly close to the coordinate origin although the data do not contain points from small epicentral distances. The linear coefficient in (7) yields the surface velocity $v = 4.984\,\mathrm{km/s}$. The corresponding sum of squares of residuals is now equal to $0.175\,\mathrm{s}^2$ (a more precise value is $0.1751\,\mathrm{s}^2$).

The best-fitting parabola without the absolute term has the form

$$t = -0.00055 \cdot D^2 + 0.1991 \cdot D\,. \tag{8}$$

It yields the surface velocity $v = 5.024\,\mathrm{km/s}$, and the sum of squares of residuals is again equal to $0.175\,\mathrm{s}^2$ (a more precise value is $0.1753\,\mathrm{s}^2$).

We should emphasize different behaviour of the linear and quadratic approximations when their absolute terms are omitted. On passing from the straight line (1) to (2), the sum of squares of residuals increased from $0.212\,\mathrm{s}^2$ to $0.299\,\mathrm{s}^2$, i.e. by about 40 per cent. On the other hand, on passing from parabola (7) to (8), the increase of this sum is quite negligible (about 0.1 per cent). Of course, this surprising similarity of the quadratic approximations may be accidental to a certain extent. Nevertheless, it indicates that these approximations of the data should be preferred over the linear approximations. In other words, we arrive at an analogous conclusion as in Section 3 that a vertical inhomogeneity is an important structural feature of the region under consideration.

Consider the approximation of the measured travel time data by the quadratic function (8) with the zero intercept time. This function is shown in Fig. 3 by the dashed line. The prescribed form of the travel–time curve implies, of course, a certain form of the corresponding velocity–depth distribution, $v(z)$. This distribution for the travel–time curve (8), obtained by the Wiechert–Herglotz method (Janský, 1977), is shown in Fig. 4 by the dashed line. Denote this model as model F.

Model F should again be considered as a very rough approximation to the real structure. Let us add only two comments to this model. First, the quadratic approximation has a tendency to diminish the slope of the travel–time curve at its end. Consequently, the velocities in model F are probably too high at greater depths. Second, the smooth velocity profile of model F cannot explain the generation of the converted Sp wave. Therefore, for example, a discontinuity should be added to the upper part of model F. However, also the S to P reflection at a discontinuity located slightly bellow the source can be taken into consideration (Zollo et al., 1995).

5. Location Tests

Consider the seismic network composed of seven stations in Fig. 1. Their Cartesian coordinates are given in Tab. 1 but, to simplify the problem, let us assume that all the stations are situated at the same height above sea level. The vertical z–coordinate (depth) is put to zero at this height, $z = 0$. Since no detailed structural model of the region is known, we shall solve the forward problems for three different models, namely for the continuous model F, layered model D and layered model WB95 proposed by (Novotný, 1996). The first five layers of the WB95 model

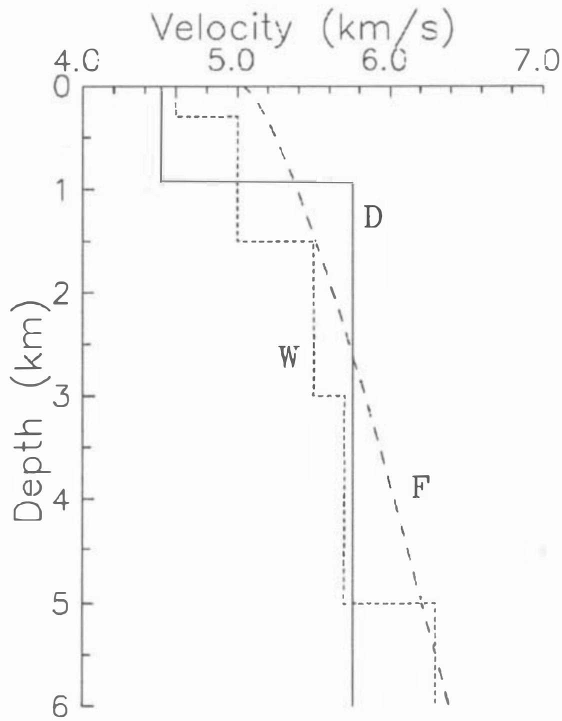

FIG. 4. P–wave velocity–depth distribution for model D (solid line), model F (dashed line), and model WB95 (dotted line)

have P–wave velocities of 4.6, 5.0, 5.5, 5.7 and 6.3 km/s, and thicknesses of 0.3, 1.2, 1.5, 2.0 and 9.5 km, respectively; see the dotted line in Fig. 4. For the purposes of this paper we shall abbreviate the WB95 model as model W (Tab. 4).

Many tests have been computed for an epicentre with the coordinates $x_s = 33$ km and $y_s = 24$ km in the neighbourhood of the Nový Kostel station (see tests 1 to 6 in Tab. 4). This epicentre simulates the position of the microearthquake epicentral area No. 1 according to the classification by (Horálek et al., 1995). Two source depths z_s have been considered for each of the models: 0 and 6 km for model F, and 6 and 10 km for models D and W. Program ZESY of Červený and Janský (1985) was used for computing the theoretical travel times of P waves propagating from these sources to the seismic stations. The origin time was put to zero in all cases, $t_s = 0$. These travel times, which we do not present in this paper, were then used as input data for the location of these hypocentres (as unknowns) using homogeneous structural models. We have used the following models in the location programs: first, model A with the fixed P–wave velocity of 5.757 km/s; second, model B with the fixed velocity of 5.461 km/s; third, a homogeneous half–space with an unknown

TABLE 4. Results of the location tests: the first column contains the number of the test; M_S, x_s, y_s and z_s denote the model and the source coordinates used for solving the forward problem; M denotes the homogeneous model used for the location and the next columns give the results of the location. For details see the text

No	M_s	x_s	y_s	z_s	M	x	y	z	t	v	σ
1	F	33	24	0	A	32.8	24.3	0.0	0.25	5.757	0.08
					B	33.0	24.1	0.0	0.12	5.461	0.03
					V	33.0	24.1	0.0	0.06	5.326	0.02
2	F	33	24	6	A	33.0	24.0	6.4	−0.06	5.757	0.01
					B	33.0	24.1	7.5	−0.29	5.461	0.02
					V	33.0	24.0	5.9	0.06	5.939	0.00
3	D	33	24	6	A	33.0	24.0	5.6	0.11	5.757	0.00
					B	33.0	24.0	6.8	−0.13	5.461	0.01
					V	33.0	24.0	5.9	0.05	5.679	0.00
4	D	33	24	10	A	33.0	24.0	9.7	0.10	5.757	0.00
					B	33.0	24.0	10.9	−0.20	5.461	0.01
					V	33.0	23.9	10.0	0.03	5.681	0.00
5	W	33	24	6	A	33.0	24.0	6.1	0.04	5.757	0.01
					B	33.1	24.0	7.2	−0.20	5.461	0.02
					V	33.0	24.0	5.7	0.12	5.861	0.01
6	W	33	24	10	A	33.0	24.0	10.7	−0.13	5.757	0.01
					B	33.1	23.9	11.9	−0.44	5.461	0.01
					V	33.0	24.0	10.0	0.05	5.949	0.00
7	W	18	30	6	A	17.6	30.0	7.8	−0.17	5.757	0.03
					B	17.0	29.9	10.5	−0.69	5.461	0.05
					V	17.8	30.0	5.7	0.17	6.045	0.02
8	W	18	30	10	A	17.7	30.0	11.9	−0.31	5.757	0.02
					B	17.3	29.9	14.4	−0.86	5.461	0.03
					V	17.9	30.0	9.9	0.09	6.051	0.01

velocity. In the latter case, not only the hypocentre parameters but also the half-space velocity may vary in the course of the location. This model will be denoted as model V.

Table 4 gives some results of these location tests. Let us describe this table in detail for the first example. In this case, the forward problem has been solved for model F (the continuous velocity–depth distribution) and the surface source, $z_s = 0$, with the epicenter mentioned above (see the first five columns of the first row in Tab. 4). Column M denotes the homogeneous models used for the location, i.e. models A, B and V, respectively. The next columns contain results of the location obtained by the gradient method. Here we present the results which we have looked for in a vicinity of the true values (the true values of x_s, y_s, z_s and

t_s have often been chosen as the starting parameters and up to 200 iterations have been carried out). It can be seen that, using model A, the calculated epicentral coordinates, $x = 32.8$ km and $y = 24.3$ km, are very close to the true values of 33 and 24 km, respectively. The best results have also been obtained for the source depths z close to the true value $z_s = 0$. Only the calculated origin time $t = 0.25$ s differs a little from the true value $t_s = 0$. The last but one column contains the half-space velocity. The last column contains a mean time deviation calculated by the formula $\sigma = \sqrt{s/n}$ where s is the sum of squares of the time residuals, and n is the number of the stations. The second row contains the location result for model B. It can be seen that this model yields a better result than model A. It can easily be explained since the low velocities in the upper part of model F are better approximated by model B with a lower velocity than by model A. Still a better result has been obtained for model V with the final velocity of 5.326 km/s (the third row in Tab. 4).

In example 2 a similar problem is solved for the source depth of 6 km. The determination of the epicentre is again very accurate, but the computed source depths differ from the true value (by 0.4 km for model A, by 1.5 km for model B and by 0.1 km for model V). Contrary to example 1, model A now yields a better result than model B. Such a situation will repeat in all remaining examples in Tab. 4.

In examples 3 and 4 the forward problem has been solved for model D. Since model D has been derived from model A, it is not surprising that model A gives good results.

In examples 5 and 6, five layers of model W in solving the forward problem have been replaced by homogeneous half-spaces in solving the inverse problems. The accuracy of the epicentre determination is again high, but the errors in the depth reach up to 0.7 km for model A, 1.9 km for model B and 0.3 km for model V.

The last two examples in Tab. 4, examples 7 and 8, consider again model W in solving the forward problem, but the epicentre is situated in the western margin of the seismic network. This epicentre simulates the position of the microearthquake epicentral area No. 3 (Horálek et al., 1995). The errors in the epicentre location now reach values of 0.4, 1.0 and 0.2 km for models A, B and V, respectively. The corresponding depth errors reach greater values of 1.9, 4.5 and 0.3 km, respectively. We could expect the lower accuracy of location as the epicentre is not surrounded by the seismic stations uniformly.

The examples given in Tab. 4 indicate that model A is relatively suitable for locating earthquakes in the region of the West–Bohemian earthquake swarms. In using this model the epicentre locations can be considered as sufficiently accurate, but the depth determinations may be in error of hundreds of meters up to a few kilometers. If the real structure of the region is close to model W then the calculated depths using model A are a little higher than the true values, i.e. the true hypocentres are probably a little shallower than the calculated ones.

Very promising results have been obtained for the homogeneous model V with a variable velocity. Although the models F, D and W, used for solving the forward problems, are vertically inhomogeneous, the locations using model V have yielded surprisingly accurate positions of the hypocentres. The maximum errors in Tab. 4 for model V are as follows: only 0.2 km in the epicentre coordinates, 0.3 km in the depth, and 0.17 s in the origin time. Naturally, the optimum velocity in model V varies with the source depth and epicentral distance.

It should also be noted that the computed depths for model V in Tab. 4 are always less or equal to the true values. It seems that it might be a general property of this method for a broad category of structural models. If it holds true generally (which should be tested in a greater detail or proven), in this manner we could obtain suitable lower estimates to the source depths.

In addition to the gradient method, also other methods were used in the location tests. Many numerical values were different from those in Tab. 4, but the general properties remained the same.

6. Conclusions

It happens very often that only a few values of measured travel times are available. It seems to be quite natural to approximate such points by a straight line and to adopt the reciprocal value of its slope as the velocity in the corresponding homogeneous half-space model. However, such a simple solution is acceptable only in case of the intercept time being close to zero. If the intercept time deviates from zero, a more complicated model must be considered. Several approaches of solving such a problem have been discussed in this paper.

The analysis of the travel times of seismic waves generated by quarry blasts in the region of the West–Bohemian earthquake swarms leads to the following conclusions:

a) The experimental data used by (Horálek et al., 1987) are very incomplete especially at short epicentral distances (distances less than about 6 km).

b) The straight–line approximation (1) of the experimental data yields the intercept time of 0.255 s, which is much higher than the accuracy of the measurements. We interpret it as an indication of a pronounced vertical inhomogeneity in the structure of the region. We are convinced that mean vertically inhomogeneous models could represent an intermediate step between initial homogeneous models and final 3–D models. Consequently, we recommend to replace the homogeneous model A in routine applications by a better model, e.g., by a model of type D (Tab. 3). More complicated layered models, such as model W (Novotný, 1996), or block models (Nehybka et al., 1993; Dvořák, Sýkorová, 1993) should also be tested.

c) Sometimes it may be difficult to recognize whether the first–arrival travel times should be attributed to the direct wave in a homogeneous half–space or if a more complicated model is required. The non–zero intercept time, mentioned above, may be an indication of a vertical inhomogeneity, but it may also be a mere consequence of a great scatter of the observed data. Therefore, as a better quantitative test, we recommend to approximate the observed data by the straight lines $t = k \cdot D + q$ and $t = k \cdot D$ (see lines (1) and (2)), and to compare the residuals. If the former approximation yields substantially smaller residuals than the latter one, a homogeneous half–space is not acceptable. For a detailed discussion see Section 4.

The following conclusions can be drawn from the location tests:

a) Despite the objections to the homogeneous model A with the velocity of 5.757 km/s, this model seems to give satisfactory results when used for locating the West–Bohemian earthquakes. The errors in the epicentre coordinates are small, the depth errors may vary between hundreds of meters to a few kilometers.
b) Homogeneous models with a variable velocity should also be used in the location algorithms (see the discussion in the preceding section).

We have arrived at somewhat contradictory conclusions concerning the homogeneous model A. However, they can easily be explained as follows. The observed travel times in Tab. 2, from which model A has been derived, do not contain data from epicentral distances smaller than 6 km. Therefore, we can hardly expect model A to describe well the shallow crustal structure of the region. On the other hand, model A characterizes well the structure at a little greater depths (say, at depths greater than about 3 km). And, by chance, most earthquakes of the region occur below this depth. Consequently, on the one hand, model A is an oversimplified structural model (which should be replaced by better models) but, on the other hand, this model seems to give satisfactory results when used for locating earthquakes.

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