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The Intercept-Time Method – Possibilities and Limitations

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Abstract. In deep-seismic sounding the reflected and refracted waves recorded form a complex system of traveltime curves. The inverse seismic problem consists in determining the function $v(x, z)$ along the profile for a series of velocity levely $v(x, z) = \text{const}$ and discontinuities as well. This paper described a method which allows the transformation of observed traveltime curves into an intercept-time section with isolines of velocity. Such a section permits discrimination of the various types of waves and gives a qualitative picture of the crustal structure, and is then converted into a depth section.

The application of this method is demonstrated by a number of examples from the Ukrainian Shield, the Tien Shan and the Caucasus showing both possibilities and restrictions.

Key words: Deep-seismic sounding – Inversion of DSS data – Intercept-time sections – Cross sections – Ukrainian Shield – Tien Shan – Caucasus

Introduction

Seismic measurements aim to determine the subsurface velocity distribution of elastic waves. In seismic reflection work there exist well developed methods and computer programs which permit detailed interpretation of the recorded data and the elaboration of structural models. In seismic refraction work the inverse problem is more difficult, due to the complicated paths of refracted rays. Especially in deep-seismic sounding (DSS), it is still difficult to determine the nature of the recorded waves and to invert the traveltimes into a structural model.

Two-dimensional seismic problems in DSS can be formulated as follows: Along a profile there is observed a system of direct, reversed, and overlapping traveltime curves of continously refracted, reflected, and head waves. The task is to determine a velocity model $v(x, z)$ which can reproduce the observed wave field kinematically and dynamically. The velocity function $v(x, z)$ is assumed as being of arbitrary shape; it may include sharp discontinuities, layers with constant velocities and gradient zones, layers with high and low velocities. The medium should be isotropic.

There is no direct solution for such a general inverse seismic problem. There are only some methods which consider separately the travel-time curves of some known types of waves, e.g. head weaves, or reflections, or refractions, and permit the determination of some elements of the two-dimensional model. As a result a multistage process is necessary in order to construct a complete seismic cross-section from the observed wave fields. The iteration method is applied at the final stage of the process.

The methods used at the first stage may be divided into three main groups. The first one includes one-dimensional solutions: the Wiechert-Herglotz method, τ-method (Bessanova et al., 1974), the synthetic seismogram methods developed in the last decade (Fuchs and Müller, 1971), optimization procedures (e.g. Matveeva, 1968; Orcutt, 1980). They treat different types of waves and permit the determination of detailed velocity distributions with depth for laterally homogeneous parts of the profiles. The bounds of possible solutions may also be determined (Bessanova et al., 1974).

The second group of methods is mainly directed to the construction of seismic boundaries. Reflections and head waves are used here. In DSS studies in the USSR the wavefront method proposed by Riznichenko (1946) is usually applied. In the last decade a numerical version of the method was developed (Pilipenko, 1979). Although wavefront idea was used in other countries (Thornburgh, 1930; Haagedorn, 1959; Hubral, 1977), the methods based on the concept of delay time became more popular (Gardner, 1939, 1967; Barthelmes, 1946; Wyrobek, 1956; Barry, 1967; Willmore and Bancroft, 1960; Bamford, 1973). The latter is applied to head waves, and it is very useful for the investigation of sharp boundaries. Puzirev et al. (1975) proposed and applied a method based on measurements at fixed distances, comparable with the fan-shooting technique. Corresponding special interpretation procedures have been elaborated for the correlation of refracted and reflected waves.

The third group of methods deals with the numerical solutions of the two-dimensional problem which have been proposed and developed by Alekseev et al. (1969), Oblogina (1965) and Romanov et al. (1980). They treat mainly refracted waves propagating in ve-

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locity fields with monotonically increasing values. Thus they are not applicable to media with sharp boundaries.

All these methods require that the wave types are known and that the relations between the head wave branches of different shotpoints and between the travel-time curves of reflections and refractions are determined. In deep-seismic sounding it is still difficult to distinguish the nature of the recorded waves and how these waves are related to each other. The correlation of the waves along the profile is also a difficult problem.

In this paper a procedure, a generalized intercept-time method, is proposed which permits analysis of the wave field along the profile and the simultaneous derivation of an approximate two-dimensional solution from all recorded waves. The solution describes the principal elements of the cross-section: shape of the seismic boundaries and velocity isolines, the velocity variation along the boundaries, low-velocity zones and so on, but the derived parameters of the model should be corrected by ray tracing and synthetic seismogram calculation. The possibilities and also the restrictions of this method are demonstrated by means of some theoretical and practical examples.

Intercept-Time Method

In seismic reflection work t_0-sections are quite common within the interpretation procedure. In seismic refraction work the intercept-time t_i corresponds to the quantity t_0 (Gamburtsev et al., 1952). The well-known τ-method (Bessanova et al., 1974) may be also considered as a version of intercept-time method as far as τ means t_i in the solution.

The intercept-time method proposed by Pavlenkova (1973) was especially developed for laterally inhomogeneous media. Other names for this method are reduced-traveltime-curve method or transformed-travel-time-curve method. A developed version of this method is described in this paper.

In detail this version serves the following goals:

1. All travel-time data of all recording lines along a profile are displayed in a comprehensive form as intercept-time section with the velocity as parameter.
2. This display facilitates the identification of the nature of travel-time segments, e.g. the distinction between subcritical and overcritical reflections and head waves.
3. The intercept-time section gives a qualitative picture of the structure under study. Such a section is comparable with the t_0-section in reflection seismics.
4. Finally this intercept-time section serves as base for the construction of a depth cross section.

The concept of the proposed method starts from the following travel-time relation valid for a horizontally homogeneous medium, but with an arbitrary vertical velocity distribution $v(z)$:

$$t(x-x_0)=\frac{x-x_0}{v_a}+2\int_0^{z_m}\sqrt{1/v^2(z)-p^2}\,\mathrm{d}z \tag{1}$$

$x-x_0$: distance between recording- (x) and shot (x_0) point,
v_a: apparent velocity at the point $(x-x_0)$,
$p=1/v_a$: ray parameter,
z_m: maximum depth of penetration.

The first term of relation (1) determines the slope of a straight line being the tangent of the travel-time curve at the point $x-x_0$, while the second term is the intercept t_i of this line with the time axis.

The travel-time curve reduced with the velocity v_r is obtained from the relation:

$$t_r(x-x_0,v_r)=\frac{x-x_0}{v_a}+2\int_0^{z_m}\sqrt{1/v^2(z)-p^2}\,\mathrm{d}z-\frac{x-x_0}{v_r}. \tag{2}$$

This curve $t_r(x-x_0,v_t)$ has a maximum at $v_a=v_r$. Here the reduced travel-time t_r is equal to the intercept-time $t_i(v_r)$ for the velocity level $v=v_r$ (Fig. 1a).

The concept is now extended to laterally inhomogeneous media. If the depth of the velocity level v_r varies along the profile, the upper limit of the reduced travel-time integral varies also. In order to determine the lateral variation of the velocity level v_r, the direct and reversed travel-time curves are reduced with the same velocity v_r and plotted along the profile at the half distance $x_0+(x-x_0)/2$. Due to the fact that in crustal studies a great part of the recorded waves are continuously refracted waves, this half-distance display refers in a first approximation to the horizontal distance of the point of deepest penetration of the ray under consideration. In the next step the reduced travel-time curves are enveloped by a curve now outlining approximately the course of the velocity isoline $v=v_r$ in the intercept-time distance domain $t_i(x,v_r)$. Using different reduction velocities a system of velocity isolines can be constructed (Fig. 1b, c).

The depth z at the level $v=v_r$ can be obtained with the approximation formula:

$$z(x,v_r)=\frac{t_i(x,v_r)\cdot\bar{v}}{2\sqrt{1-(\bar{v}/v_r)^2}} \tag{3}$$

where $\bar{v}$ is the average velocity between the surface and the isoline $v=v_r$. $\bar{v}$ is determined from travel-time curves by the formula (Kondratiev 1963):

$$\bar{v}(x,v_r)=\tfrac{1}{2}\left[(x-x_0)/t+\sqrt{\frac{x-x_0}{t}\cdot v_r}\right].$$

In principle, the travel-time curve transformation can be applied to all types of travel-time branches. Within these time sections the wave types are in different relation to each other and to the envelope. The reversed and overlapping travel-time curves of head waves from a plane boundary with velocity v_b coincide with the envelope $t_r(x,v_r=v_b)$, independent of the slope of the boundary. In the more general case of smoothly curved prograde travel-time curves which are generated by continuously refracted waves the reduction velocities of the corresponding envelopes can be regarded as material velocities.

This relation between reduced travel-time curves $t_r(x-x_0,v_r)$ and their envelopes $t_i(x,v_r)$ enables us to de-

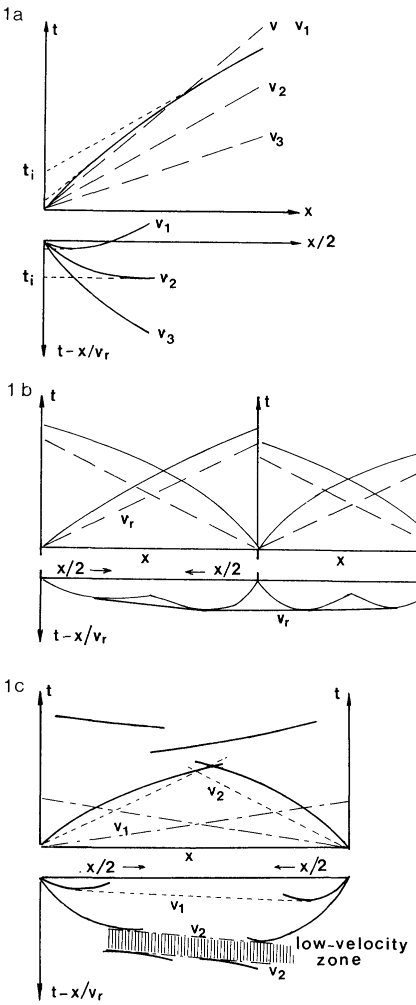


Fig. 1. a Reduced and transformed travel-time curves $t-x/v_r$ with different reduction velocities v_1, v_2, v_3 for one shot point. **b** Construction of a velocity-isoline v_r based on travel-time curves from a number of shot points. **c** Detection of a low-velocity layer in the reduced time section

termine the nature of the waves and to determine the velocity level $v=v_r$ for head waves.

The case is more complicated for retrograde travel-time branches. They can be caused by a wave reflected at a first-order discontinuity or by a wave penetrating into a zone having a strong velocity gradient. Just as in the normal time display, in the reduced presentation the retrograde branch must coincide at the critical point with the envelope of the corresponding prograde branch. Thus a discrimination between the sub- and overcritical parts of a retrograde branch becomes possible. The envelope derived from the distant part of the retrograde curves also represents a true velocity. The retrograde curves between the critical point and the endslope may represent true velocities in the case of continuously refracted waves in a strong gradient zone. For waves reflected at a sharp discontinuity the resulting velocity has the meaning of an apparent velocity. In the intercept-time section a discrimination of both types is not feasible, the answer can be obtained only by a detailed depth inversion of the time section (Giese, 1976).

One of the main difficulties in the intercept-time method and also in the inversion problem arises from inhomogeneities in the uppermost crust, e.g. sedimentary basins. This problem is discussed here by means of a model. This model presents a graben structure (Fig. 2a). In order to study the influence of the sedimentary fill on the travel-time of the waves penetrating deeper into the middle and lower crust, the graben model has been somewhat simplified. In the lower crust the isolines are not updomed, as is typical for a depression, but they run in parallel. Fig. 2d shows the calculated and reduced travel-time curves for the three velocities 6, 7, and 8 km/s and the enveloping velocity isolines.

As expected the enveloping curves show roughly the shape as the corresponding velocity isolines in the model section. Only the intercept-time curve $v_r=6.0$ km/s, however, reflects correctly the velocity isoline $v=6.0$ km/s of the model. The two other curves are not straight and horizontal but bent down because of the influence of the depression on the travel-times of the deeper penetrating waves.

Using the intercept-time formula (3) a first approximation to the original model was obtained (dotted lines in Fig. 2d).

For comparison, in Fig. 2d, the result is shown of a calculation based on the one-dimensional solution for individual travel-time curves. It became evident that the upper part of the original model could be determined in a good approximation, but in the lower part of the crust a strong scatter of the isolines results. Thus it is difficult to decide whether there is a simple structure or not. If starting from the intercept-time section, the depth transformation gives isolines 7.0 and 8.0 km/s which are approximately horizontal, as in the original model. However, the results illustrate simultaneously the very poor resolving power of the method, especially for small variations of velocity isolines in the lower crust, if a heterogeneous and complicated upper crust exists.

In order to improve the resolving power, special travel-time corrections can be applied already to the intercept-time sections. In Figure 2b the downbending of the isolines 7.0 and 8.0 km/s is caused by the sediments in the depression. If the sediments are replaced by basement rocks with $v=6.0\,km/s$ the delay can be eliminated with a sufficient approximation. This correction can be obtained from the intercept-time curve of the level $v=6$ km/s:

$$\Delta t=\tfrac{1}{2}[t_i(x_0, v_r=6)+t_i(x, v_r=6)]+t_d. \tag{5}$$

The small additional correction t_d takes into account the travel-time difference resulting from rays travelling through the sedimentary fill under different angles due to different penetration depths (top of basement and deeper crust in Fig. 2). In ordinary crustal studies t_d may be neglected.

Thus, in a general case the travel-time curves $t(x, x_0)$ may be corrected for the influence of lateral inhomogeneities above any velocity level $v=v_n$ with the formula

$$\Delta t_n(x)=\tfrac{1}{2}[t_i(x_0, v_{rn})+t_i(x, v_{rn})] \tag{6}$$

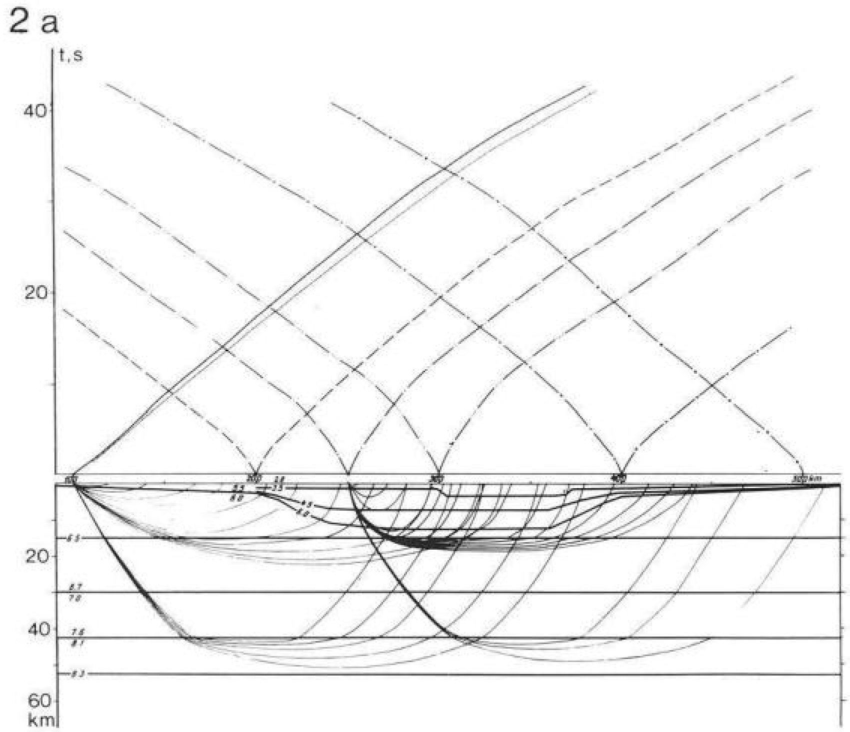
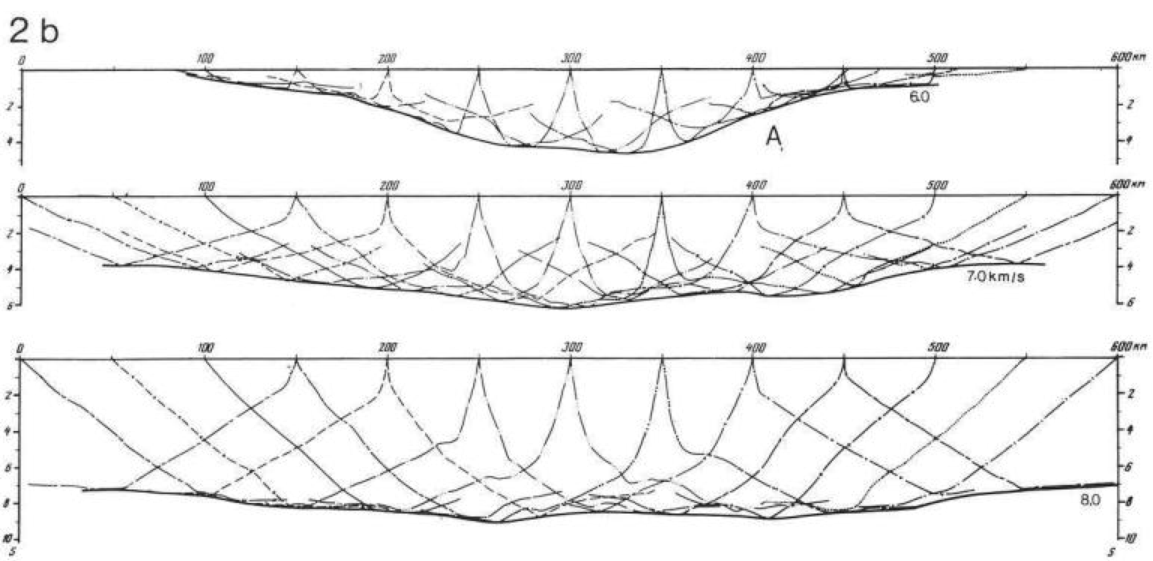
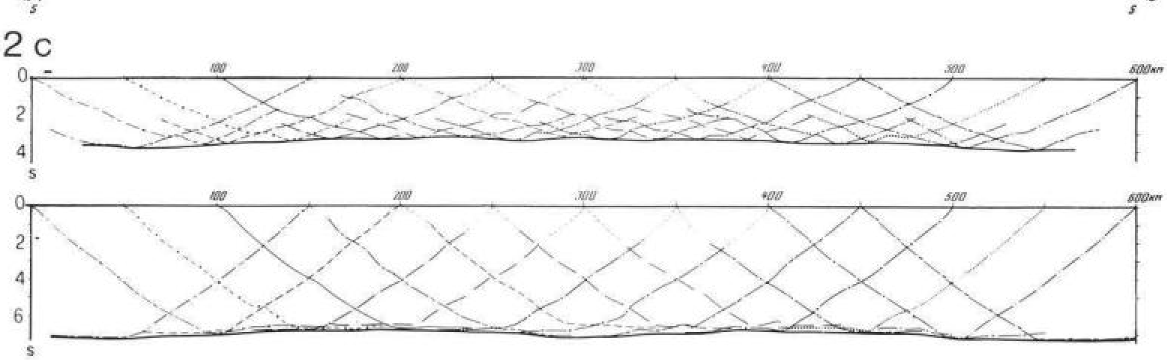


Fig. 2. a Crustal model with a graben-like depression and the calculated travel-time curves from a number of shot points. **b** Reduced travel-time curves and their envelopes (6.0, 7.0, 8.0 km/s). **c** Reduced travel-time curves and their envelopes (7.0 and 8.0 km/s) after having applied the time correction based on the topography of the isoline $v = 6$ km/s. **d** Comparison of different interpolation results with the original model. 1. boundaries based on one-dimensional solutions for each shot point; 2. derived from the system of reduced travel-time curves



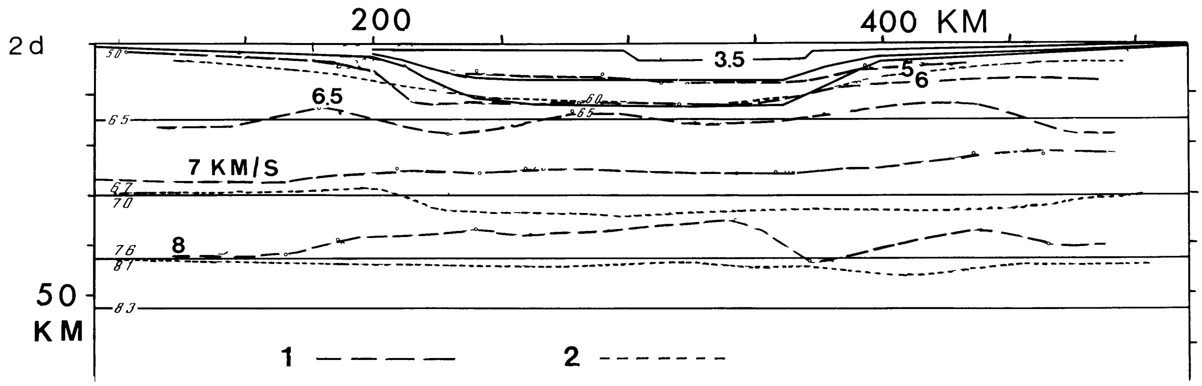


where $t_i(x, v_{rn})$ denotes the intercept-time taken from the envelope of the reduced and transformed travel-time curves with the reduction velocity v_{rn}.

Then the corrected travel-time curves are reduced with velocity v_{n+1}:

$$t_{r(n+1)} = t - \tfrac{1}{2}[t_i(x_0, v_{rn}) + t_i(x, v_{rn})] - \frac{x - x_0}{v_{r(n+1)}}. \qquad (7)$$

The form of the new envelope reflects more correctly the isoline $v = v_{r(n+1)}$ than the envelope of the uncor-



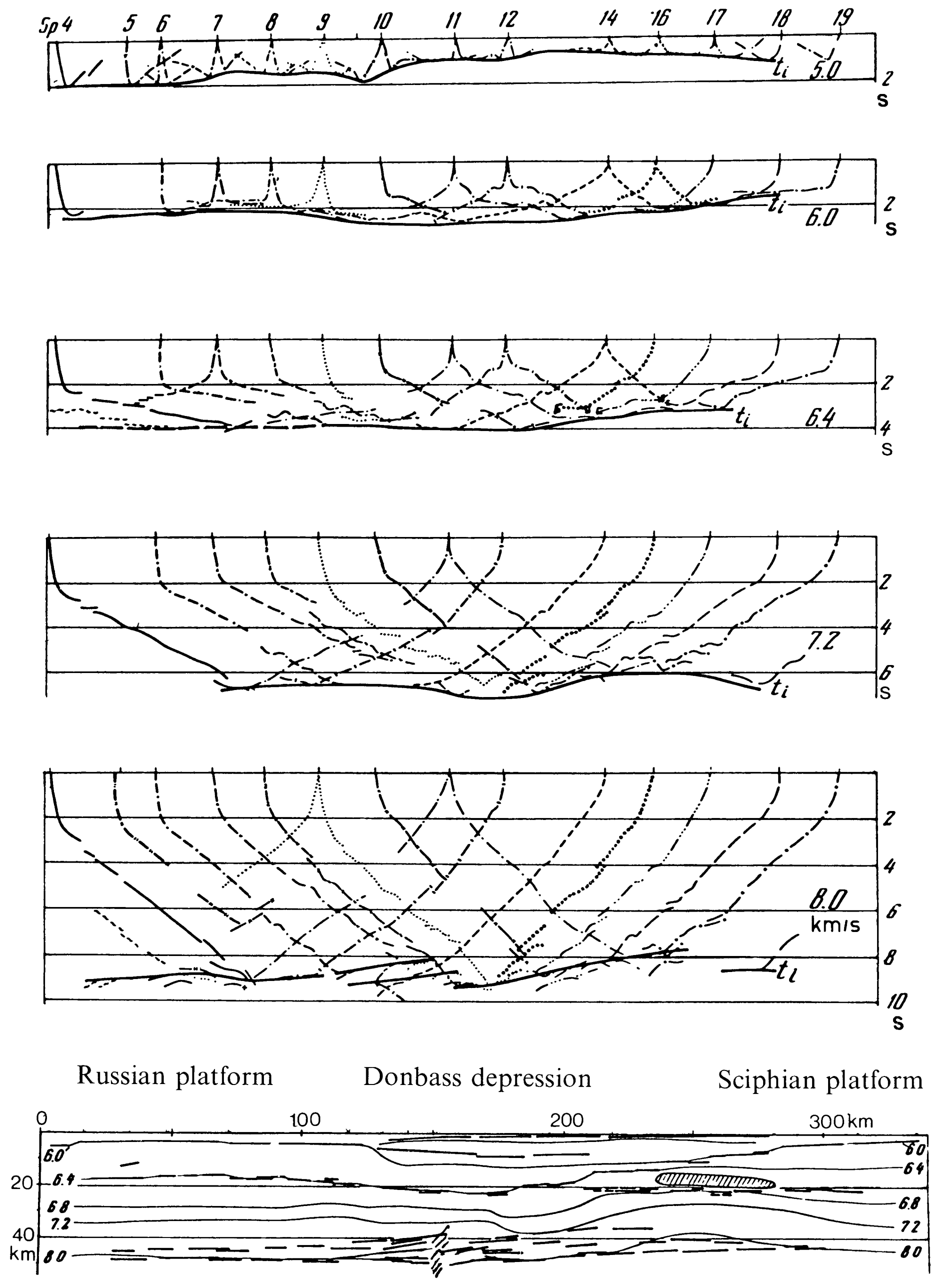


Fig. 3. Reduced travel-time curves for the Donbas profile with different reduction velocities and the resulting velocity isolines (5.0, 6.0, 6.4, 7.2, 8.0 km/s). At the bottom the resulting cross section is shown. *Hatched zones* indicate low-velocity layers (data from Konovalzev et al. 1980

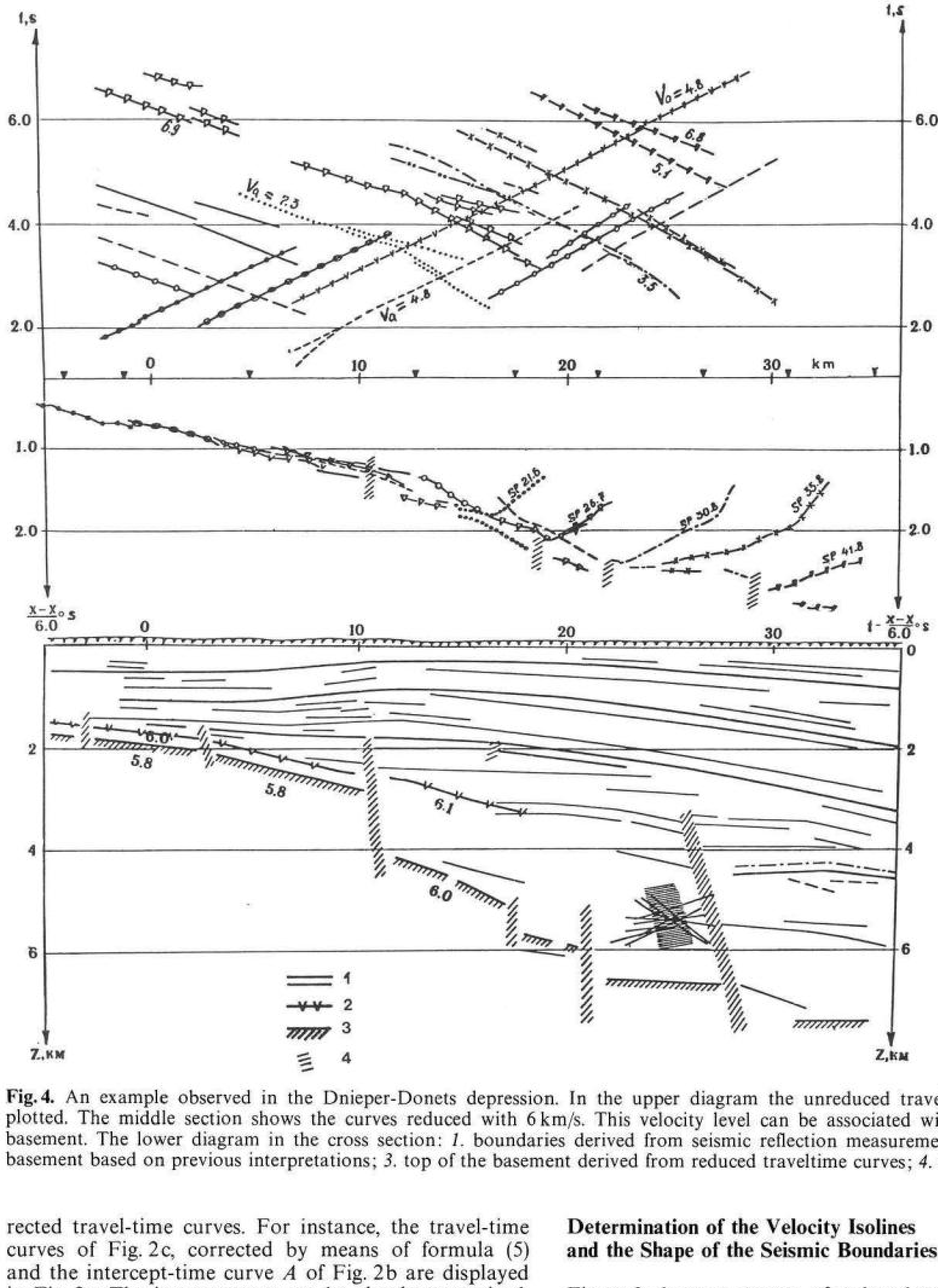


Fig. 4. An example observed in the Dnieper-Donets depression. In the upper diagram the unreduced travel-time curves are plotted. The middle section shows the curves reduced with 6 km/s. This velocity level can be associated with the top of the basement. The lower diagram in the cross section: *1.* boundaries derived from seismic reflection measurements; *2.* top of the basement based on previous interpretations; *3.* top of the basement derived from reduced traveltime curves; *4.* fault zones

rected travel-time curves. For instance, the travel-time curves of Fig. 2c, corrected by means of formula (5) and the intercept-time curve *A* of Fig. 2b are displayed in Fig. 2c. The improvement can be clearly recognized, the deeper isolines are now straight lines like the velocity lines in the primary model.

A correction for the level $v = 6$ km/s is necessary for studies of the lower crust whereas for the upper mantle the level $v = 8$ km/s must be corrected.

Some experimental examples of the intercept-time method application are considered in what follows.

Determination of the Velocity Isolines and the Shape of the Seismic Boundaries

Figure 3 shows a system of reduced travel-time curves of a deep-seismic sounding profile crossing the Donbas region, the eastern part of the Dnieper-Donets depression (data published by Konovalzev et al. 1980). The very dense system of recording lines enables a detailed study of the shape of the velocity isolines. The pattern of these lines reveals the quite different structures of the upper and lower crust along the pro-

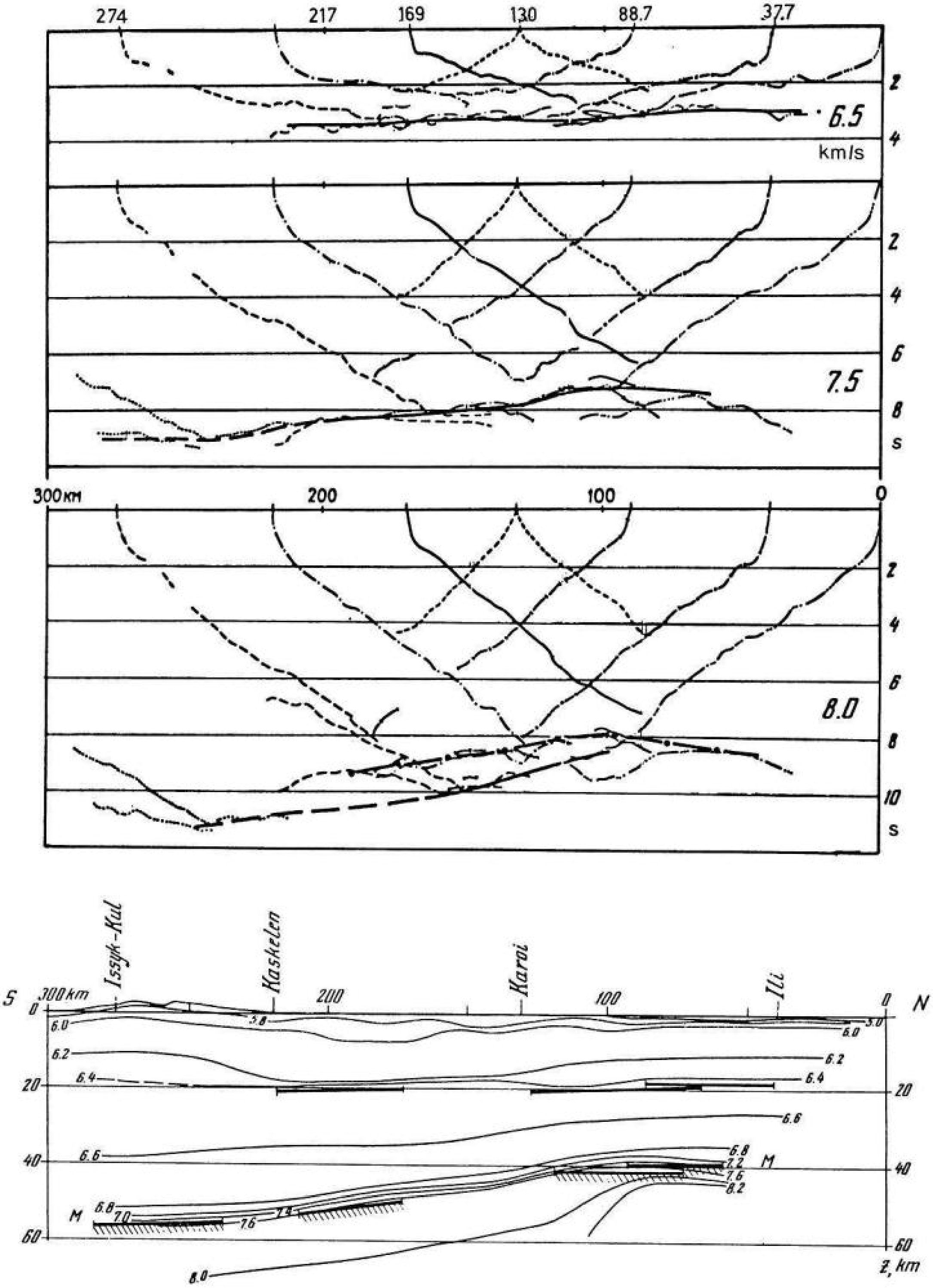


Fig. 5. Reduced traveltime curves for the Kaskelen profile (Tien Shan) with different reduction velocities (6.5, 7.5, 8.0 km/s). At the bottom the corresponding depth section is displayed

file. The isoline $v=5.0$ km/s outlines the principal uplift of the Carboniferous strata in the centre of the Donbas. The basement of the Dnieper-Donets depression can be associated with the isoline $v=6.0$ kmIs. The velocity range 6.8 to 7.2 km/s characterizes the lower crust. Along the profile this velocity interval range distinctly changes its depth. On the Russian platform (left side of the profile) it shows a slightly higher velocity gradient than on the younger Scyphian platform (right side of the profile). The Donbas fault zone is characterized by a downwarp of the isolines, that means by somewhat smaller velocities. A complicated alternation of low and high velocities could be detected in the transition zone between the Russian platform and the

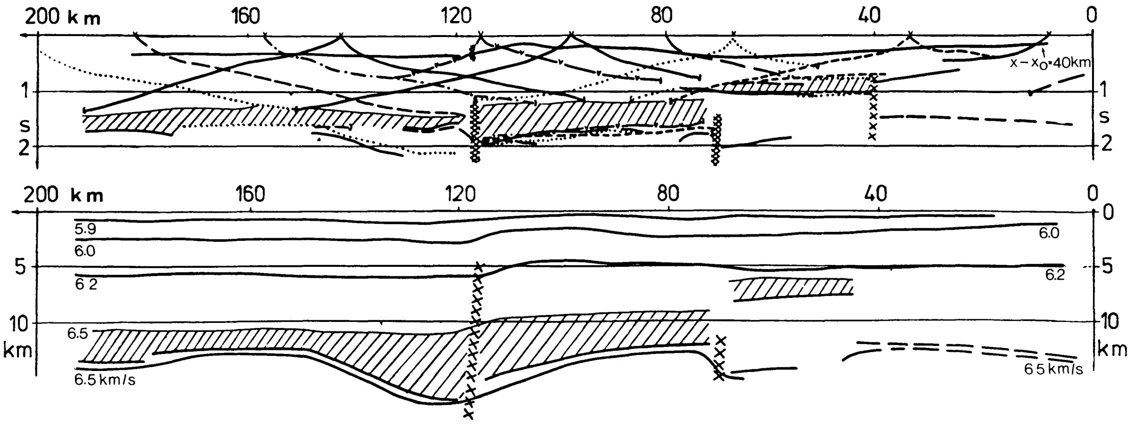


Fig. 6. The upper part shows the reduced traveltime curves of a seismic refraction line observed in the Ukrainian shield. The reduction velocity is 6.5 km/s. The *lower part* shows the resulting depth section. The *hatched areas* indicate low-velocity zones

Donbas depression. A very complex relief can be seen for the velocity level $v = 8.0$ km/s (Mohorovičić discontinuity). It forms a sequence of faults with over- and underthrusts, typical of the Donbas region. This detailed picture of the Mohorovičić discontinuity could be revealed mainly by the evaluation of reflected waves. The display as reduced travel-time segments allowed their identification and correlation with the line $t_i(x, v_r = 8\,\text{km/s})$ along that portion of the profile that traverses the fracture zone.

Another example of a complicated structure within a fracture zone is shown in Fig. 4. This profile was measured in the southern fracture zone of the Dnieper-Donets depression. From first arrivals, there is no evidence of a fracture zone. Only the evaluation of later arrivals as reduced travel-time curves proves the existence of a fracture zone. Later on such a structure could be confirmed by reflection seismics.

It should be emphasized that the main limitation of the method is the smoothing of sharp steps at refracting boundaries. It is very difficult to determine the correct position of the steps from the described head wave presentation because the travel-time is related to the point midway between the shot and recording points. To improve the results it is useful for such cases to combine the intercept-time method with the time-term presentation.

Study of the Velocity Variation Along a Seismic Boundary

The next example (Fig. 5) is taken from measurements carried out in the Tien Shan region. A first interpretation by Shazilov (1980) is based on the one-dimensional solution using the Wiecher-Herglotz method.

Characteristic features of the wave field in this area are the intensive second arrivals P^K and P_1^m, showing velocities near 6.8 km/s and 8.0 km/s. Their transformation into reduced travel-time curves proves the existence of two discontinuities. The first one has the stable velocity 6.5 km/s, the second discontinuity is characterized by a varying velocity. It might be derived from comparison of the envelope of the reduced travel-time curves at $v_r = 8$ km/s and position of P_1^m branches in Fig. 5. The P_1^m boundary coincides with the velocity level $v = 8$ km/s in the northern part of the profile and departs from this level in the southern part. This result indicates the existence of a mantle in the Tien Shan region, having in its upper part a lower velocity than normal. This new model has been checked by ray tracing yielding a good agreement between calculated and observed travel-time.

The Detection of Low-Velocity Layers

The existence of low-velocity layers is indicated in the time-distance plot by dying out of first arrivals and the appearance of secondary arrivals, forming a delayed travel-time curve which is parallel to that of the first arrivals. The method of reduced travel-time display can clarify these pecularities in the time cross-section (Fig. 1c).

Figure 6 shows an example taken from measurements in the Ukrainian shield. The travel-time curves are displayed with a reduction velocity of 6.5 km/s. The envelope of the first arrivals bounds the top of the low-velocity layer, whereas the corresponding envelope of later arrivals determines its bottom. The resulting model, shown in the lower part of Fig. 6 presents the low-velocity zone as a hatched area. The velocity inside this low-velocity zone has been assumed to be 5.8 km/s.

Mathematical Modelling

Mathematical modelling is the final stage of all interpretation methods. It aims to achieve an optimum agreement between the observed and the calculated wave field. Furthermore mathematical modelling permits determination of the parameters of low-velocity zones as well as the study of the bounds of possible solutions. Besides the kinematic data the dynamic characteristics should be included. Dynamical criteria can help to reduce the number of possible solutions. The calculation of wave amplitudes for horizontally inhomogeneous media can be performed by an extended ray method (Červený et al. 1977; Psenčik 1979).

Figure 7a shows a system of traveltime curves of a profile recorded along the Kura depression in the southern Caucasus. Three types of waves are recognizable, the branches of the waves travelling only through the

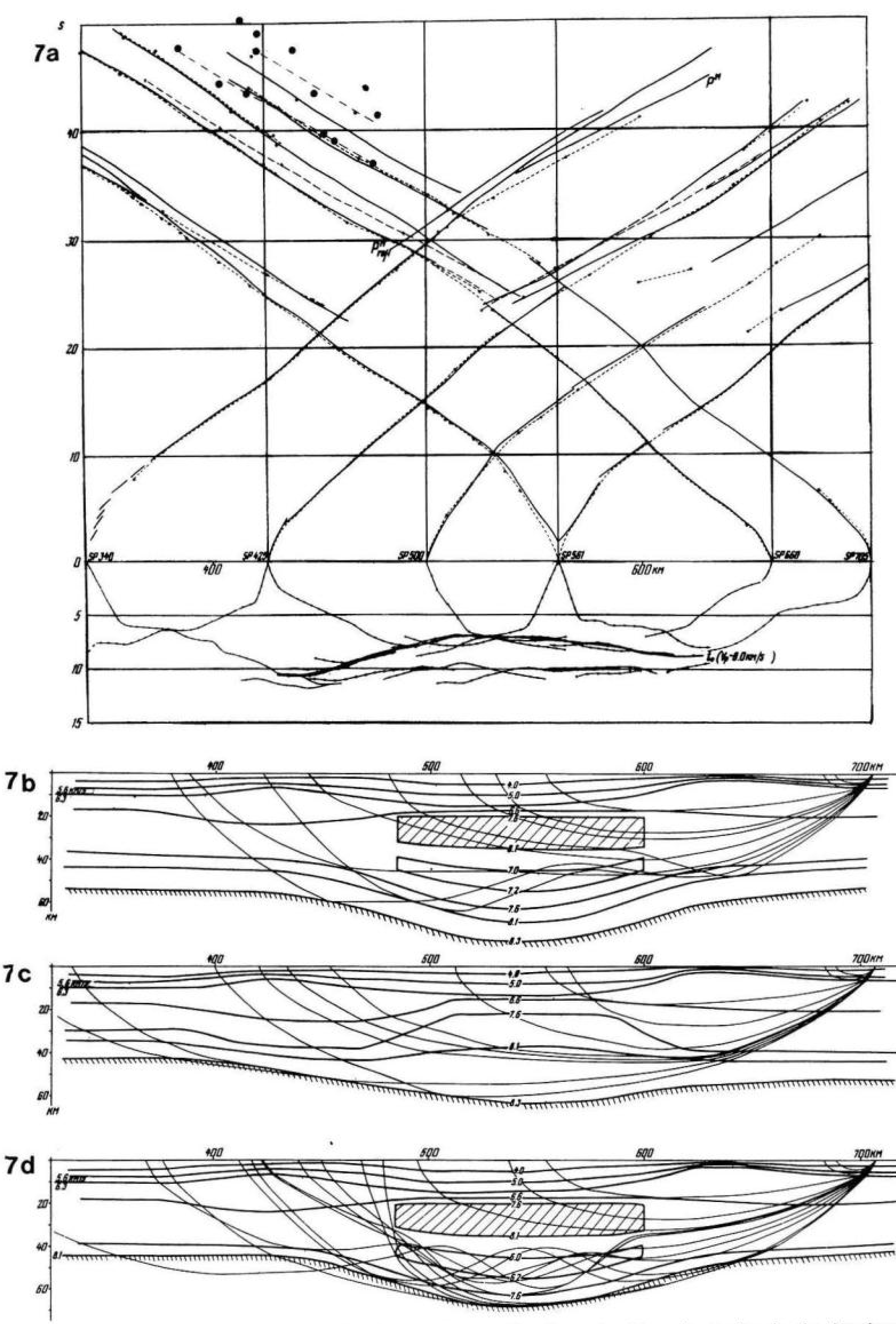


Fig. 7a–d. Traveltime curves of a deep-seismic sounding profile along the Kura depression in the Southern Caucasus. The observed branches are shown by *full lines*, the calculated by *dashed lines*. Three different models have been studied. Short dashes denote the traveltimes of the model without a low-velocity layer (Fig. 7c), the long dashes refer to the model of Fig. 7b and the circles to that of Fig. 7d, both with a low-velocity layer in the deeper crust. The model of Fig. 7b has a minimum velocity in this layer of 7.0 km/s and that of Fig. 7d a velocity of 6.0 km/s. The model of Fig. 7b agrees best with the observed data. At the bottom of Fig. 7a the observed traveltime curves in reduced form (8.0 km/s) are displayed. The *thick line* denotes the envelope of the reduced traveltime curves of the waves refracted at the top of a shallow high-velocity body. The later arrivals must be associated with a boundary situated at a depth of about 60–70 km

sedimentary cover with apparent velocities of 3.0–6.0 km/s, the anomalous phases with velocities 7.0–8.0 km/s and later arrivals forming retrograde time-distance curves. The characteristic features of this last group are large amplitudes, many phases, and a complicated variation of the wave shape. The apparent velocity of this group decreases with increasing distance from 8.5 to 6.5 km/s. An analysis of their travel-time curves revealed that they cannot be interpreted as normal reflections. When applying the wavefront method, no intersection of the wavefronts should be achieved. Mathematical modelling was used, including the calculation of travel-times as well as amplitudes of the most important waves. An example of these calculations is presented in Fig. 7b–d.

Two models differing in principle were considered, both being deducible from the reduced travel-time curves. The first presents an uplift of the material with velocities from 7.0 to 8.0 km/s near to the basement surface. In the second model this material is separated from the mantle by an inversion zone. The computation of the most important waves reveals that both models are equivalent with respect to the first arrivals. As regards later arrivals quite different results were obtained for each model. Only for the separated body with an underlying low-velocity channel does the form of calculated travel-time curves of refracted waves, propagating through this channel, correspond to the form of experimental curves satisfactorily.

The example presented demonstrates the advantages of mathematical modelling for the solution of two-dimensional seismic problems. The main advantage consists in its capacity to deal with very complicated models including laterally inhomogeneous media with curved interfaces. This cannot be done by any other direct method. Another advantage of modelling is the possibility of using the dynamic properties of waves for interpretation. Unfortunately, this may at present be done only in ray approximation limits (Červený et al. 1977). Nevertheless, the general laws of amplitude distribution in the wave field may be taken into account during the interpretation. In the above case the amplitude consideration helps to determine the nature of the later arrivals. As already mentioned, these waves have considerable amplitudes in comparison with the first arrivals and are formed of many phases. These properties could not be explained by means of simple refracted or reflected waves. The channel waves possess similar properties. They have rather large amplitudes, caused by weak divergence of the ray tube, and their energy is propagated along different paths inside the channel producing many phases of the arrivals (Pavlenkova et al. 1977).

The next advantage of mathematical modelling is an estimation of the stability and uniqueness of the resulting model. For instance, on the basis of a comparison of different models, it was found that the 10 km thick upper part of the crust was determined uniquely in the model discussed above. On the other hand the detailed velocity distribution inside the low-velocity channel cannot be calculated, only the average velocity can be determined. Two models have been checked, one with a channel velocity of 6.0 km/s the other one with 7.0 km/s. The second one fits the observed data satisfactorily. The most unstable part concerns the determination of the lower boundary of the high-velocity body and the Mohorovičić-discontinuity.

Conclusion

The solution of the two-dimensional seismic problem needs several stages of data evaluation. At first the main features of the model and the nature of the recorded waves must be determined. For this purpose, approximation methods, such as the intercept-time method as described above, the special time-field method of Puzirev and the time-term analysis may be used, In the next step a detailed analysis of reflections, head waves and refracted waves follows, aiming to determine correctly the elements of the model, i.e., the seismic interfaces and the velocity distribution within the layers. It should be noted that several methods may serve to elaborate the parameters of the model. In the last stage the model is checked by ray tracing and amplitude calculation. If necessary, the model must be improved stepwise in order to get a better agreement between the observed and calculated data. To develop and improve this method it is necessary to work out algorithms for automatic modelling of media with complex structures.

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