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Study of the Structure of the Lower Lithosphere by Explosion Seismology in the USSR

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Abstract. Deep seismic sounding (DSS) conducted along long range profiles in various regions of the USSR yielded good seismic records of deep waves in a distance range of 400–600 km (on some profiles to 1000–1500 km). These experimental DSS data permitted to construct detailed velocity models of the upper mantle to a depth of 100–120 km or more. The analysis of kinematic and dynamic characteristics of the mantle waves P^M (P_n), which are everywhere recorded as first arrivals beginning from distance at 130–200 km on land and at 30–40 km at sea, shows that this group of waves has a complex structure and consists of several refracted and reflected waves. Travel times and other characteristics of P^M (P_n) waves are greatly variable from area to area. Velocity sections of the upper mantle were constructed by comparison of the experimental P^M (P_n) travel-time curves referred to the flattened base of the earth's crust with the theoretical travel-time curves computed for vertically and horizontally inhomogeneous velocity models. The principal results of the study are as follows:

1) Lateral velocity inhomogeneities have been established in the upper mantle, the linear dimensions of which vary from tens to hundreds of kilometers or more:

2) The uppermost mantle has been found to incorporate layers, 10–20 km thick of lower (7.7–7.9 km/s) and higher (8.4–8.6 km/s) compressional velocities;

3) A change in the structure of the lower lithosphere has been recorded in the transition from one large geological province to another. This change is marked by variation of velocities and velocity distribution with depth. The results demonstrate the effect of the upper mantle processes on the formation and evolution of the crustal structure.

Key words: Structure of the upper mantle – Deep seismic sounding – Long range seismic lithospheric profiles.

Introduction

According to the present day knowledge, many geological phenomena observed at the surface and in the upper layers of the earth's crust, the study of which is of great importance for scientific and practical applications, are in many respects caused by the processes operating in the upper mantle. In this connection, the solution of many urgent problems of geodynamic requires quantitative results on the distribution of lateral and vertical inhomogeneities in the upper mantle because these inhomogeneities are indicators of the mechanism and energetics of abyssal processes. Most of the information available to date about the upper mantle structure was obtained by seismic recording of earthquakes. The intensive development of explosion seismology for the last 10–15 years is of great importance. This method allowed investigations to be made not only in regions of seismicity but also in aseismic platform regions. The accuracy and resolution of investigations were increased by using higher frequencies and by determining the time and place of shots more precisely than origin and coordinates of earthquakes. Another advantage was the fact that in explosion seismological investigations, shot and detector arrangements could be varied depending on the aim and object of investigation. As a result, the reliability of field measurements and hence the accuracy of geological interpretation could be greatly improved.

In the early sixties crustal studies by deep seismic sounding (DSS) were made in some areas of central Asia (Ryaboy, 1966, 1969). These experiments proved that the DSS method could be effectively applied to not only crustal studies but also to investigations of the upper mantle to a depth of 100–120 km. Later this conclusion was proved in other areas on land (Alekseev et al., 1973; Kosminskaya et al., 1972) and also in the Pacific Ocean (Zverev, 1970) and in the Black Sea (Neprochnov and Rykunov, 1970).

At present there are two trends in the study of the upper mantle by explosion seismology. One of them is based on a technique developed in seismology (Lewis and Meyer, 1968; Archambeau et al., 1968; and others).

This method is characterized by large separations between recording stations (up to 100 km). Reversed and overlapping systems are commonly not employed. The accuracy of the results in this case is comparable to that of earthquake studies. The other trend derived its ideas and techniques from exploration geophysics and from the experience in deep seismic sounding. It employed a technique of correlating deeply travelling waves using reversed and overlapping shot and detector arrangements, and of detailed analysing kinematic and dynamic characteristics of waves in the distance range from 500–1000 km or more from an explosion (Ryaboy, 1966; Burmakov et al., 1975; Hirn et al., 1973; Bamford et al., 1976, etc). The second trend appears to yield more accurate and more detailed results. In the USSR the development of seismic study of the upper mantle followed the second trend.

To date more than 200 profiles have been observed in the USSR by the DSS nethod. About 10% of them are long-range profiles to a distance range of 400–600 km (in some regions up to 1000–1500 km). This paper is an account of the main results of the DSS study of the upper mantle structure obtained

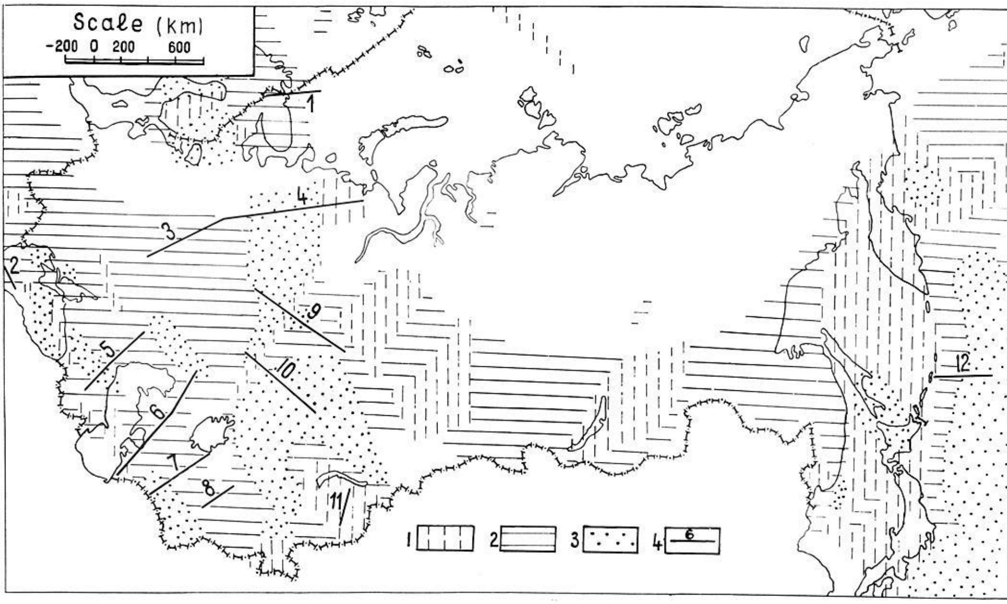

Fig. 1. Schematic map of boundary velocities of compressional waves measured along the Mohorovičić discontinuity (V_B^M) for the territory of the USSR and adjacent countries, compiled by N.A. Belyaevsky and V.Z. Ryaboy, V_B^M values: 1: $V_B^M \leqq 8.0$ km/s; 2: $V_B^M = 8.1$–8.2 km/s, 3: $V_B^M > 8.2$ km/s; 4: DSS profiles used in the construction of the lower lithosphere velocity sections in this paper

along the long-range profiles. The location map of the profiles is given in Figure 1.

Observational Technique

Field observations were made on DSS lines by a method of continuous longitudinal profiling with a correlation of reversed and overlapping travel-time curves of deep waves. A distinctive feature of the field procedure was that, apart from studying the crustal structure, it provided the possibility of investigating the upper mantle using the same types of reflected waves at less than and beyond the critical distance, refracted and converted waves. This was achieved by making measurements at distances greatly exceeding those of the conventional DSS systems, which on land commonly cover 200–250 km. Distances of 400–600 km were used to record the mantle phases. On some profiles, as for instance from Kineshma to Vorkuta (Profile 4 in Fig. 1), the distances reached 1000–1500 km (Burmakov et al., 1975).

Spacings between shotpoints varied from 50–80 km to 150–200 km or more.

The measurements were made with field equipment of oscillographic and magnetic recording in a frequency range of 5–8 to 15–20 cps. The stations were spaced 100–200 m apart. On those profiles where travel-time curves reached distances up to 1000–1500 km, recording was made with the equipment having a frequency band from 1–2 to 10–15 cps and with separations between recording stations of 5–15 km.

As a rule, field measurements were made with calibrated equipment, which made it possible to use not only kinematic but also dynamic characteristics of the recorded waves in the interpretation. It should be noted that the employed DSS systems of crustal profiles were sufficient to overcome the effect of lateral inhomogeneities within the crust on the kinematic and dynamic characteristics of mantle waves.

P-Waves from the Lower Lithosphere

It is the purpose of this section to present the P-wave travel-time and amplitude data observed at a distance (R) exceeding 130–150 km from the shotpoint.

Group of $P^M_{\text{refl.}}$ Waves

A group of waves reflected from the Moho discontinuity here termed $P^M_{\text{refl.}}$ (in other publications termed P_MP) yields the most energetic arrivals in a range of distances from 60–80 km to 250–300 km from the shotpoint. In some areas this group of reflections can be followed directly from the shotpoint to a distance of more than 300 km. Apparent velocities of these reflections regularly decrease with distance from about 9–10 km/s at R=60–80 km to 6.5–7.0 km/s at R= 250–300 km. The overlapping travel-time curves are not parallel but tend to diverge.

In most cases P^M reflections are most intensive in a distance range of 80–100 km to 180–200 km. Figures 2 and 3 show the examples of averaged travel-time curves, the graphs of amplitude vs. distance variation, $A(R)$, and the photographs of seismic records showing P^M reflections obtained on a profile from Kopet Dagh to the Areal Sea (Profile 7 in Fig. 1) (Ryaboy, 1966, 1969).

The times of early P^M reflections on the DSS profiles measured on land and shown in Figure 1 differ by about 2.5–3.0 s for the same distances from the shotpoint. The standard deviation of travel-time curves from the mean curve, which was smoothed with the help of a rectangular function, was ± 0.88 s (Fig. 4). The travel-time curves earlier of P^M refl. arrivals that have been used in the statistical analysis were obtained in areas of essentially differing crustal structure such as Baltic Shield, Russian plate, Turanian plate, North Tien-Shan, Fore-Caucasus, Central Kazakhstan, Ural folded area and others. Therefore, the above mentioned time differences of P^M reflections and standard deviations may be regarded as indications of heterogeneity of the continental crustal structure.

Group of P^M (P_n) Waves

These waves, designated in seismology as P_n, are everywhere recorded as first arrivals beginning from a distance of 130–200 km. The transition from the

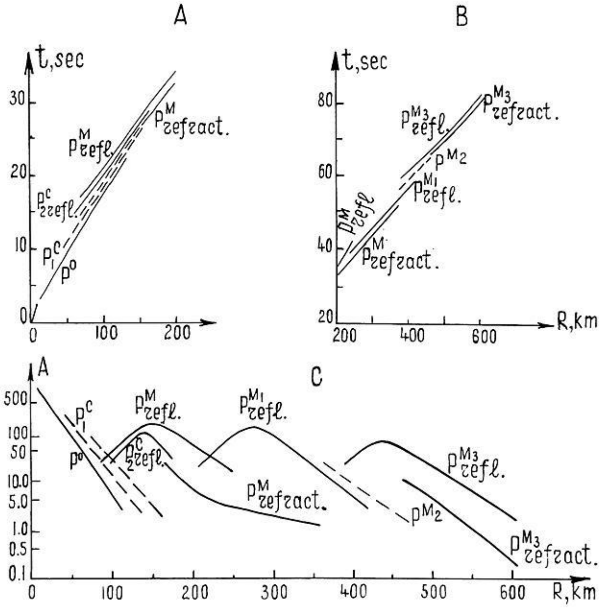

Fig. 2A–C. Schematic travel-time curves (**A, B** for distances of 0–200 and 200–615 km, respectively) and amplitude vs. distance graphs (**C**) for deep waves recorded along the profile Kopet Dagh-Aral Sea (Profile 7 in Fig. 1). Dotted lines designate low intensity and poorly correlatable waves

respective crustal waves to a group of P^M waves is commonly marked by a pronounced bend of the travel-time curve in first arrivals and by an increase of apparent velocities from 6.5–7.0 km/s to 7.7–8.2 km/s. The P^M travel-time curves are not strictly linear in the region of first arrivals. Apparent velocities tend to increase with distance from 7.7–8.2 km/s to 8.8–9.5 km/s at $R=$ 500–1500 km. The P^M-travel-time curves display rapid increases or decreases of apparent velocities against the background of their normal gradual increase. This is well illustrated by Figure 5A which shows first arrivals of $P^M(P_n)$ waves recorded on some of the reported long-ranges DSS profiles.

As a result of the detailed study of kinematic and dynamic characteristics of a P^M wave group, it has been found that the group has a complex structure consisting of several waves. This was particularly well seen in detailed observations. The most common case is a gradual change of waves, when waves of higher velocities enter the region of first arrivals as the distance from the shotpoint increases. In some cases this phenomenon is accompanied by an attenuation of the early waves, a break in a travel-time curve and a displacement of its retrograde branch by 0.5–1.5 s.

In some areas some of the P^M waves were reliably identified and traced in the reversed and overlapping systems in the regions of later arrivals. These waves are commonly most intensive. As compared to those recorded as first arrivals, they have higher and nearly identical apparent velocities which slightly decrease with greater distance from the shotpoint. Figures 2, 6, 7 and 8 show

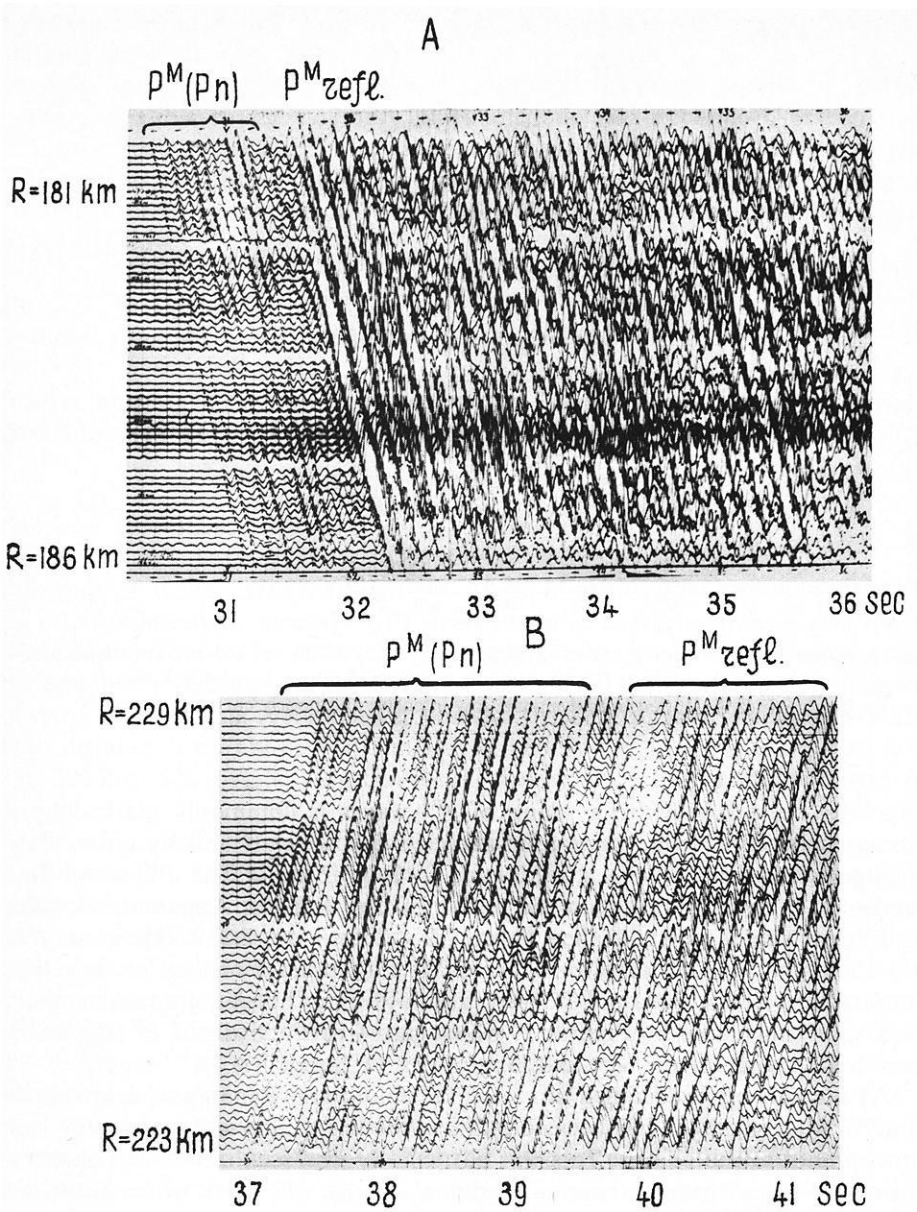

Fig. 3. Seismic records showing variation in intensity of $P^M(P_n)$ and $P^M_{\text{refl.}}$ with distance (R) from the shotpoint (profile Kopet Dagh-Aral Sea)

examples of travel-time curves, amplitude vs. distance graphs and photographs of seismic records showing some P^M waves recorded along the DSS profile from Kopet Dagh to the Aral Sea (Ryaboy, 1966, 1969). This profile was the first to yield experimental data revealing a complex structure of P^M waves.

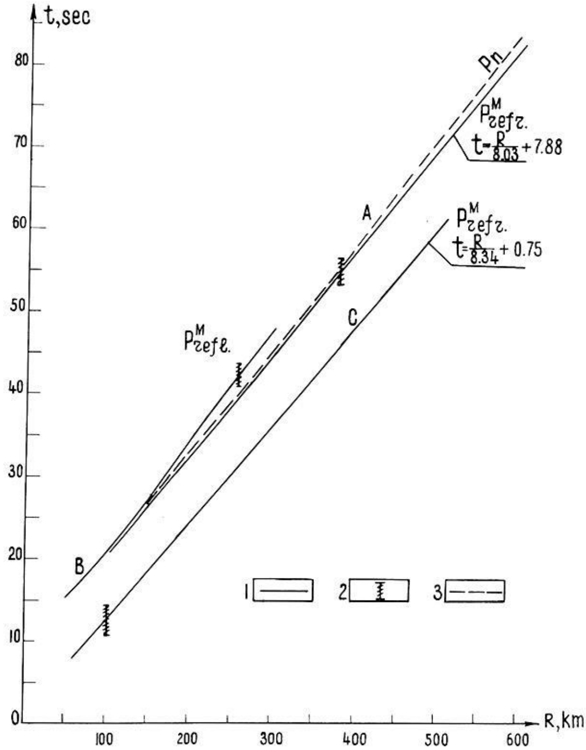

Fig. 4. The observed $t-t$ curves (A, B) and a t–t curve recomputed with reference to the flattened base of the crust (C). The curves average experimental travel times of first arrivals of $P^M(P_n)$ (A, C) and $P^M_{\text{refl.}}$ (B) on land. 1: averaged t–t curves from DSS data; 2: standard deviation from average t–t curves; 3: Jeffreys-Bullen reference P_n travel-time curve

This paper describes the results of reinterpretation of these data, which was made using the latest techniques of constructing two-dimensional velocity sections of the upper mantle.

In the region of first arrivals, $P^M(P_n)$ waves are usually less intensive than later P^M refl. waves reflected from the base of the crust. With increasing distance, however, the picture reverses, and from 250–300 km $P^M(P_n)$ waves become more intensive (Fig. 2C; Fig. 3). The amplitudes of $P^M(P_n)$ first arrivals rather rapidly decay with distance to 250–300 km, the attenuation being weaker at greater distances, where abnormal build-ups and decays of amplitudes are frequently observed.

The visible frequencies of $P^M(P_n)$ waves recorded in a frequency band of 5–8 to 15–20 cps vary from 6–8 to 10–12 cps, whereas those recorded at frequencies of 1–2 to 10–15 cps vary from 2–3 to 5–6 cps. The regions of abnormally increasing and decreasing visible frequencies of $P^M(P_n)$ waves are usually found to correspond to complications in the travel-time curves and to extremes in amplitude vs. distance graphs $A(R)$. As compared to the more intensive later events, the waves recorded as first arrivals are commonly 1–3 cps higher in frequency.

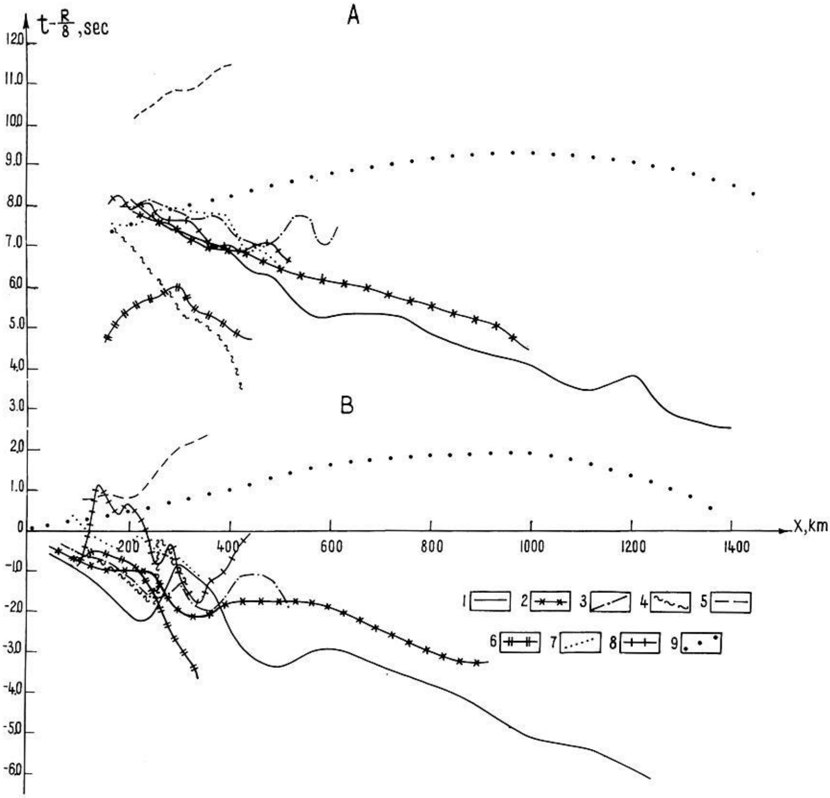

Fig. 5. The observed (**A**) and recomputed (**B**) travel-time curves of P^M (P_n) first arrivals from DSS results obtained in: 1: East European platform (Moscow syneclise) (4); 2: East European platform (Pechora syneclise) (4); 3: central region of Turanian platform (7); 4: Baltic shield (1); 5: North Tien-Shan and Balkhash depression (11); 6: Scythian platform (Fore-Caucasus) (5); 7: Central Kazakhstan folded area (10); 8: East Ural and western areas of West Siberian platform (9); 9: Jeffreys-Bullen reference P_n travel-time curve. Numbers in parentheses are numbers of profiles in Figure 1

The analysis of the experimental data obtained from the reversed and overlapping systems of observations has shown that the above mentioned features of kinematic and dynamic characteristics of P^M waves are, to a great extent, cuased by heterogeneities in the velocity structure of the upper mantle. However, in many cases it was found that kinematic and dynamic characteristics of P^M waves were also greatly affected by variations in the structure of the crust, particularly of its upper part. The number of waves in a P^M group, the intervals of their recognition and the main parameters were found to vary when we passed from one region to another, and in some cases even within one and the same profile. Similar conclusions on the structure of P^M (P_n) waves were made as a result of long-range profiles observed in France (Hirn et al., 1973).

A general idea of the distinctive features of a velocity model of the upper mantle can be obtained from comparison of the travel times of P^M first arrivals

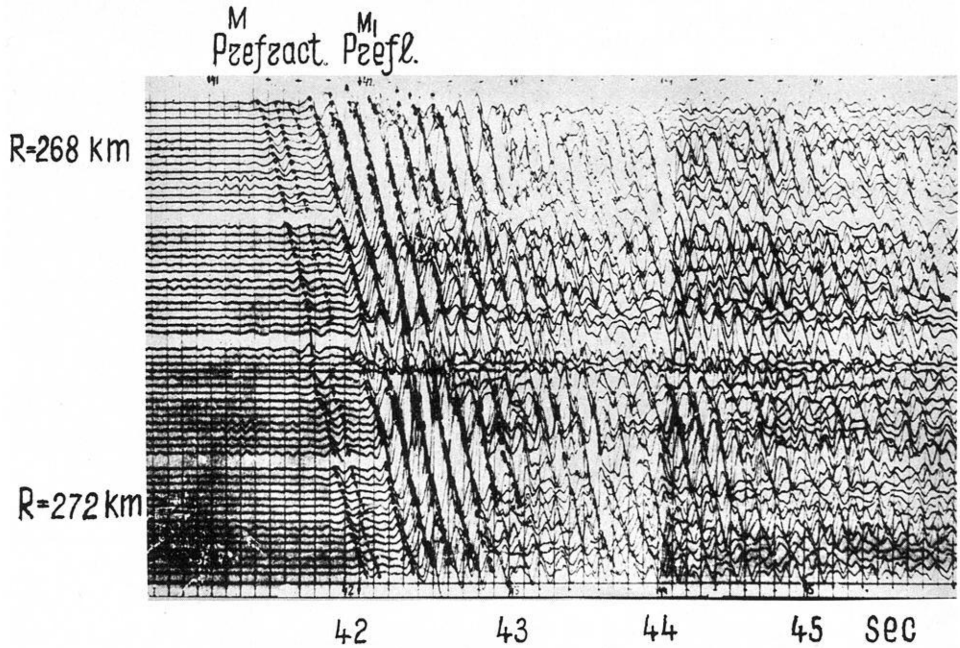

Fig. 6. Seismic records showing strong later arrivals of P^M wave reflected at an interface in the upper mantle ($P^{M_1}_{\text{refl.}}$) wave refracted ($P^{M}_{\text{refract.}}$) in a layer below the Mohorovičić discontinuity is seen in first arrivals (profile Kopet Dagh-Aral Sea)

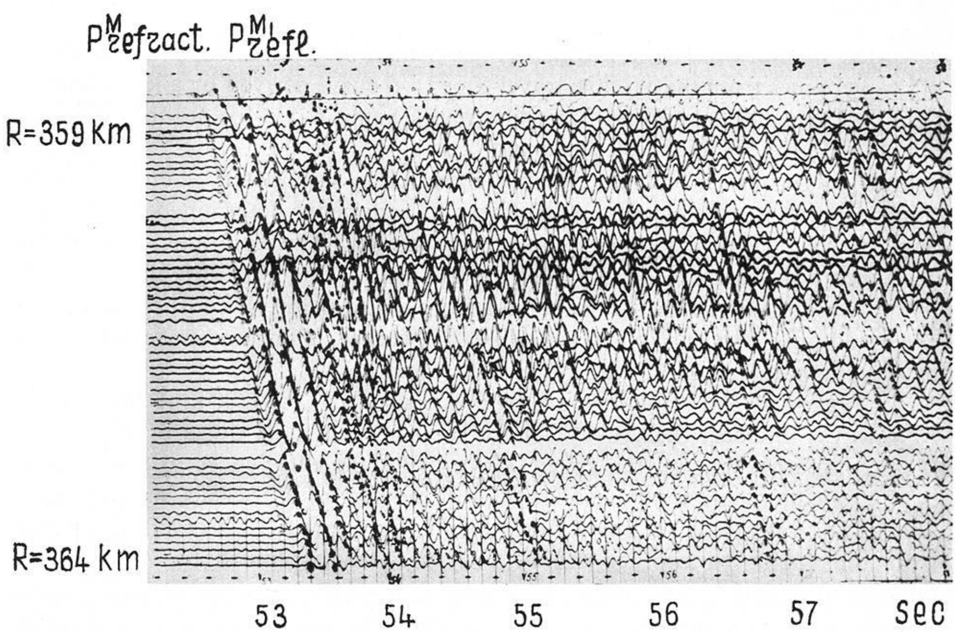

Fig. 7. Seismic records showing the interference of $P^{M}_{\text{refr.}}$ and $P^{M_1}_{\text{refl.}}$ in first arrivals (profile Kopet Dagh-Aral Sea)

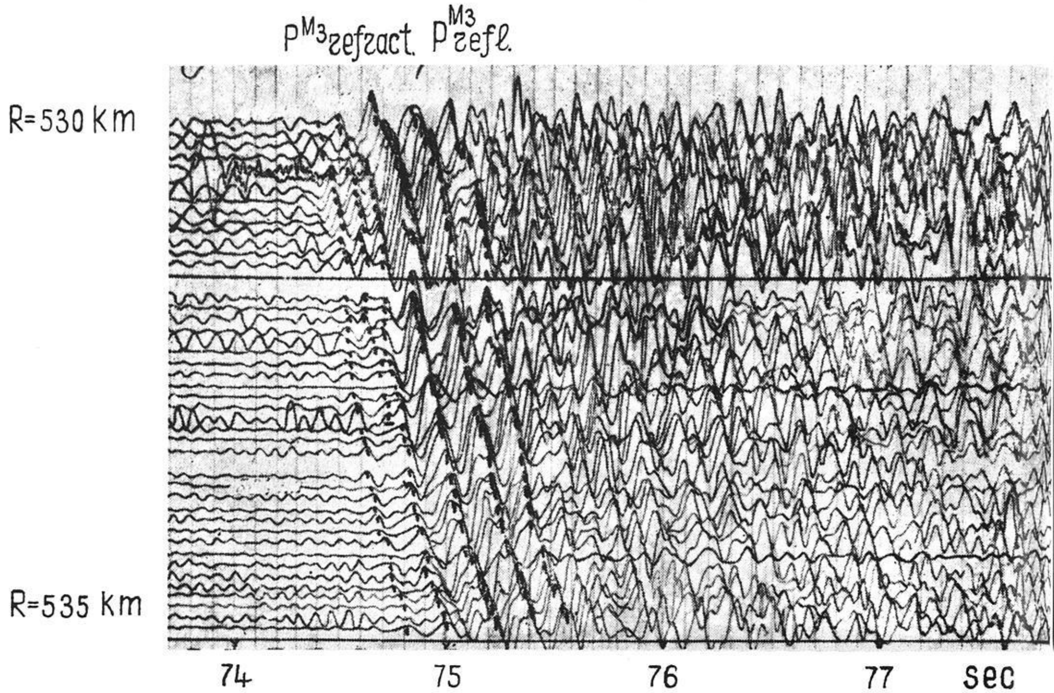

Fig. 8. Seismic records showing $P^{M_3}_{\mathrm{refr.}}$ and $P^{M_3}_{\mathrm{refl.}}$ waves at distances exceeding 500 km (profile Kopet Dagh-Aral Sea)

recorded on various DSS profiles (Figs. 4 and 5). A standard deviation of P^M travel times from the average travel-time (t–t) curve derived by a least-squares fit is ± 1.65 s, the maximum time differences being 6–8 c for stations on land and 12–13 s for combined offshore and land data. Variations in P^M travel times proved to be poorly correlated with the crustal parameters such as crustal thickness, average velocity and thickness of sediments determined from other data irrespective of P^M. Besides they vary greatly. The computed correlation coefficients do not exceed 0.35–0.45. This means that the differences in P^M travel times observed in the transition from one region to another cannot be explained only by variation in the crustal structure. The same is demonstrated by the comparison of P^M and $P^M_{\mathrm{refl.}}$ travel times.

In region of the critical point of P^M phases (at 100–200 km) the rays of P^M_{refl} waves intersect the crust at about the same angle as those of P^M waves. It is remarkable that on the same DSS profiles the differences in travel times of P^M reflections are much smaller than in the case of P^M waves, as mentioned above. So, variations in travel times of P^M first arrivals observed in passing from one region to another can be only partially explained by variations in the crustal structure, and seem to be indicative of the presence of large lateral velocity heterogeneities in the upper mantle. This conclusion is corroborated by the results presented below on the computation of the observed P^M travel-time curves with reference to the flattened base of the crust.

Interpretation Procedure

To eliminate the misleading effects of the crust, the P^M travel-time curves were recomputed with reference to a flattened crustal base (Ryaboy and Egorova, 1973). The procedure consisted in the computation of the angles between seismic rays and the ground surface. The computation was based on the Benndorf law using the original travel-time curve and seismic velocities in the upper layer as the initial data. Then, the paths of seismic rays were reconstructed to the intersection with the chosen reference level at a depth of 35–45 km near the Moho using the principles of geometrical optics and the crustal velocity data available.

Corrections were applied to the observed travel times to bring the source and the receiver to the obtained intersection points of the rays with the reference level. Theoretical modelling and analytical estimation of possible computational errors showed that the accuracy of P^M travel-time curves referred to the base of the crust varied from ± 0.3 s–± 0.5 s depending upon the chosen station separation and system of observation.

Examination of P^M first arrivals in the travel-time curves (Figs. 4, 5) shows that the differences between them are larger than the computational error, reaching 4–6 s (the standard deviation from the average travel-time curve is ± 1.46 s). This confirms the above conclusion of a horizontally inhomogeneous lower lithosphere. Velocity sections of the lower lithosphere were constructed as a result of comparison of the observed P^M travel time curves referred to the flattened base of the crust with the theoretical ones computed for vertically and horizontally heterogeneous velocity models (Alekseev et al., 1973; Burmakov and Ryaboy, 1973). A theoretical computation of travel-time curves, ray paths and isochrones was made using the Runge-Kutta method. Differential equations of seismic rays were solved by numerical methods. The procedure of computing rays and travel-time curves in horizontally heterogeneous media using computer was discussed in detail by Burmakov and Oblogina (1971).

The purpose of computation was to find a set of velocity sections in which the difference between theoretical and observed travel times did not exceed the accuracy of the latter (± 0.3–0.5 s) for a given region of seismic parameters with a minimum computer time. A detailed analysis of all possible solutions for two-dimensional models of interpretation is virtually impossible because it requires great computer time consumption. For this reason, a representative sample of all possible solutions was used to estimate the accuracy and stability of two-dimensional velocity sections of the upper mantle thus obtained with respect to probable errors of the observed travel-time curves.

Thus, the chosen procedure of inversion included a computerized comparison of theoretical and observed travel-time curves, estimation of the difference between them using the criteria of standard and maximum deviation, and correction of a theoretical model to minimize the discrepancy between experimental and theoretical data. A velocity section was constructed by an iteration method. At first, the velocity distribution in the upper layer was determined. The parameters thus found were then fixed, and the velocity distribution was calculated at greater depth. Possible solution of the problem were looked for among

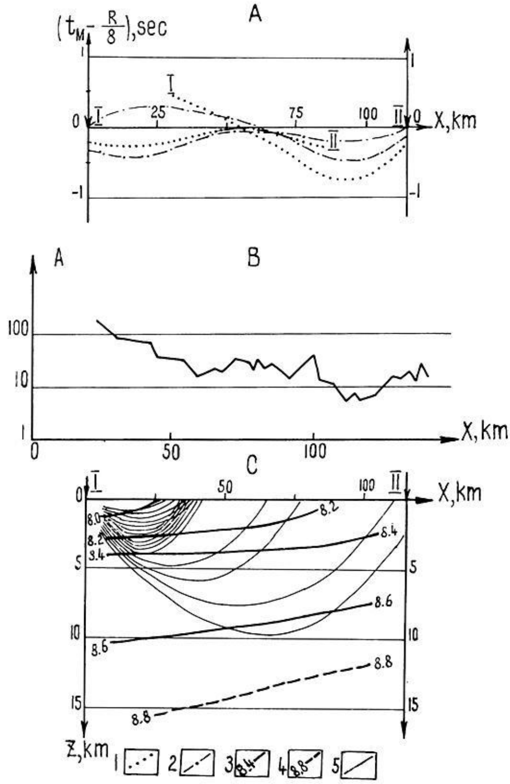

Fig. 9 A–C. Horizontally inhomogeneous velocity section of the uppermost mantle from DSS results obtained in the central part of the Black Sea (Neprochnov, Rykunov, 1970) (Profile 2 in Fig. 1)
A Comparison of reversed travel-time curves (*I* and *II*) (recomputed with reference to the base of the crust) of $P^M_{\mathrm{refr.}}$ first arrivals with theoretical travel-time curves of waves refracted in the upper mantle, calculated for velocity model $V(X, Z)$ of Figure 9C
B Amplitude vs. distance graph for $P^M_{\mathrm{refr.}}$
C Horizontally inhomogeneous velocity model of the uppermost mantle (from data of V.Z. Ryaboy and Y.A. Burmakov). The depths are given from the M-discontinuity. 1: observed travel-time curves; 2: theoretical travel-time curves; 3, 4: velocity contour lines from reliable (3) and unreliable (4) data (velocities are given in km/s); 5: paths of seismic rays

the simplest models of the medium structure. Dynamical characteristics were used for qualitative comparison of the observed amplitude vs. distance graphs, $A(R)$, of P^M waves with geometrical spreading of seismic rays whose paths were calculated at regularly spaced interval in the angle of radiation from the source.

When the computed reversed and overlapping travel-time curves of P^M waves differed by less than ± 0.3–0.5 s, or when single travel-time curves were obtained from field measurements, the interpretation was made for horizontally homogeneous models by a technique similar to the above mentioned one (Alekseev et al., 1973; Matveeva, Ryaboy, 1975).

Figure 9 shows an example of a two-dimensional velocity structure of the uppermost mantle for the central region of the Black Sea depression (Profile 2 in Fig. 1). The deviation of the theoretical travel-time curves of the waves refracted in the upper mantle, calculated for the model $V(X, Z)$, from the observed P^M first arrivals does not exceed ± 0.3 s (Fig. 9C). All velocity models, $V(X, Z)$, in which the theoretical travel-time curves are consistent with the observed ones, are qualitatively similar in vertical and lateral velocity variation. They differ in velocities by ± 0.1–0.3 km/s at the same depth. Figure 9C shows the velocity distribution in the upper mantle obtained by averaging all velocity sections consistent with the observed data.

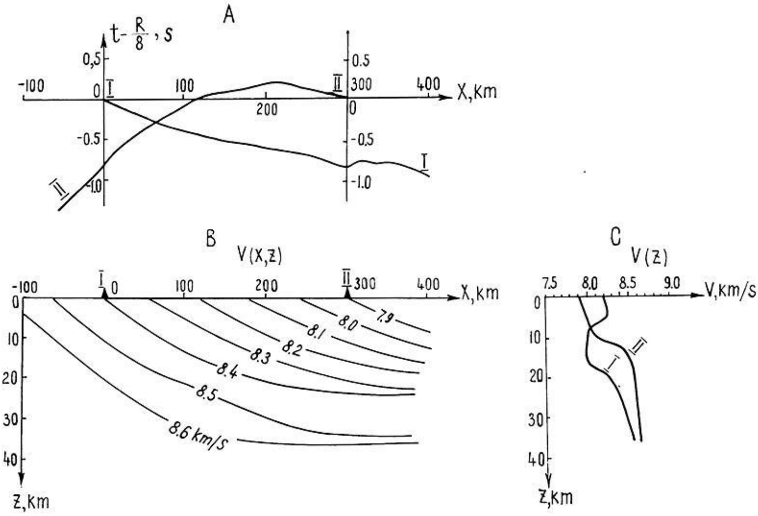

Fig. 10 A–C. The effect of horizontal inhomogeneities on the resulting upper mantle velocity sections. The depths measured with reference to the M discontinuity ($M=0$ km). **A** Theoretical travel-time curves of refracted waves for the horizontally inhomogeneous velocity model of the upper mantle given below in the same figure, **B** (*I*, *II*-shotpoints and respective reversed travel-time curves, reduced with a velocity of 8 km/s). **B** Horizontally inhomogeneous velocity model V (X, Z), of the upper mantle. **C** Velocity vs. depth variation, $V(Z)$, obtained by iteratic from the *I* and *II* travel-time curves in **A**

Variation in the density of seismic rays with distance (Fig. 9 C) agrees with the extreme values of the amplitude vs. distance graph $A(R)$. It should be noted that the velocity model of the upper mantle shown in Figure 9 C was constructed by starting at the Mohorovičić discontinuity, whose depth varies only slightly and averages 18 km. As a result, the extremes of the $A(R)$ graph are at a greater distance from the source than the respective regions of a larger and smaller density of seismic rays, the difference in the distance being equal to a horizontal projection of the waves refracted in the upper mantle to the crust obtained in migration. A horizontally homogeneous velocity section of the uppermost mantle previously constructed by Neprochnov and Rykunov (1970) is shown in Figure 11.

Thus, theoretical models helped to evaluate the accuracy of the observed data and to investigate the effect of horizontal velocity inhomogeneities unaccountable in interpretation. The computation has shown that the error of the determination of the upper mantle velocities from DSS results is about ± 0.1–0.2 km/s. The average velocity is determined with a higher accuracy (± 0.05–0.1 km/s). The parameters of layers of higher velocities are determined more accurately than those of lower velocities.

In the case of large horizontal velocity inhomogeneities an interpretation based on horizontally homogeneous models may result not only in substantial averaging of the results (lower resolution) but also in an erroneous understanding of the velocity structure (identification of non-existing interfaces, low-velocity layers, etc.). This is well seen in Figure 10, which shows horizontally homogeneous velocity models of the upper mantle constructed from the travel-time curves (Fig. 10A) calculated for a two-dimensional model (Fig. 10B). Velocity sections constructed from reversed travel-time curves (I and II in Fig. 10A) without allowance for horizontal inhomogeneities are seen to be substantially different both in velocity values and in velocity variation with depth (Fig. 10C) giving an erroneous idea of the real velocity distribution.

To conclude, the technique used by the author to interpret the P^M (P_n) data observed on DSS profiles allows the identification and determination of the parameters of large scale velocity inhomogeneities in the upper mantle extending for radii not less than several dozens of kilometers and amounting to 5–10 km or more in layer thickness.

The Results

In the seismic study of the upper mantle, the most reliable and accurate information is the determination of the boundary velocity of compressional waves along the Mohorovičić discontinuity (V_B^M). As a result of the analysis and systematization of the DSS and seismological data available, a schematic map of V_B^M values was compiled for the territory of the USSR and adjacent countries using about 400 determinations (Belyaevsky and Ryaboy, 1969). New results have been obtained since then for the same regions, which allowed the map to be improved and refined (Fig. 1).

The V_B^M values are known to vary greatly from 7.6–7.8 to 8.4–8.6 km/s or more, the most common values being 8.1–8.2 km/s. The lower values of 7.6–8.0 km/s are found in tectonically active regions, such as the neotectonic areas of the Tien-Shan and south Siberia, the Baikal rift zone, the transition zone from the continent to the Pacific Ocean, ets. The highest V_B^M values reaching 8.4–8.6 km/s or more have been determined in the Paleozoic folded areas of Central and North Kazakhstan and in the western Pacific. The zones classified as anomalous by their V_B^M values are confined to structural features distinguished in the crust by geological and geophysical data.

For some regions of the USSR indicated in Figure 1, horizontally homogeneous and two-dimensional velocity sections of the lower lithosphere were constructed from DSS results, shown in Figures 10 and 9, 12, respectively.

Before discussing the results, it should be noted that by using only first arrivals it is not possible to determine with certainty the number of layers in the really existing medium and to distinguish the velocity-depth function at the various interfaces (of first or second order). The information provided by first arrivals is capable only of providing velocity variations with depth and velocities in the separate layers. Intensive waves recorded as reflections from upper mantle interfaces, as, for instance, $P^{M_1}_{\text{refl.}}$ and $P^{M_3}_{\text{refl.}}$ recorded on the profile Kopet Dagh-Aral Sea, indicate an abrupt change of velocities at these interfaces (first-order boundaries).

The investigations reported here revealed the presence of large horiyzontal velocity inhomogeneities in the upper mantle found not only in tectonically

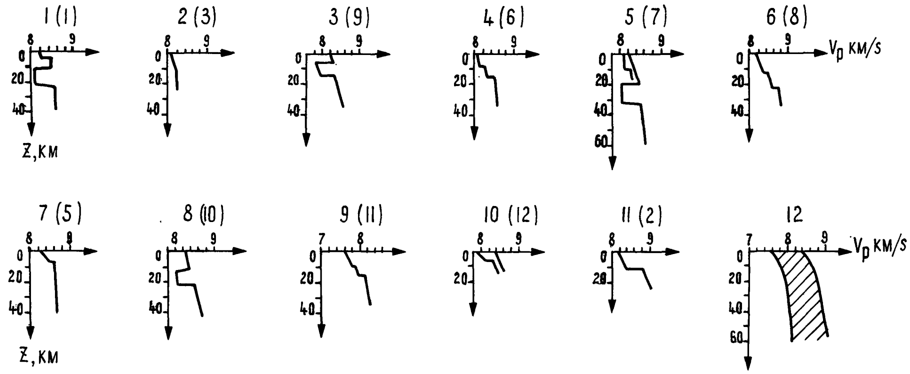

Fig. 11. Horizontally homogeneous velocity sections of the upper mantle from DSS data for different geologic structures of the USSR: 1 (1): Baltic shield; 2 (3): Voronezh massif; 3 (9): Russian platform (eastern areas) and Uralian folding (western areas); 4 (6), 5 (7), 6 (8): Turanian platform (western, central and eastern areas, respectively); 7 (5): Scythian platform (Pre-Caucasus); 8 (10): Central Kazakhstan folded area; 9 (11): North Tien-Shan and Balkhash depression; 10 (12): western part of the Pacific Ocean; 11 (2): central region of the Black Sea depression; 12: Shaded region indicates DSS velocity data for the upper-most part of the mantle; (1–10): from data of Ryaboy and Matveeva; 11: from data of Neprochnov and Rykunov. In parentheses are given the numbers of profiles given in Figure 1. The depth are measured with reference to the *M* discontinuity

active folded areas, as shown by seismologic studies (Alekseev et al., 1973; Nersesov et al., 1972; Vinnik, 1976, etc), but also in ancient platforms. The horizontal inhomogeneities in the upper mantle which would be found to extend from tens to hundreds of kilometers or more. The velocities in these depth ranges appear to be different from the average upper mantle velocities by 5–6%.

In the investigated depth intervals of up to 100–150 km compressional velocities in the upper mantle vary from 7.7–8.0 to 8.6–8.8 km/s or more. A remarkable feature of the upper mantle velocity structure in platform areas is the presence of low- and high-velocity layers about 10–20 km thick at a depth of 80–100 km. In the last years similar and even thinner layers were identified by explosion seismology in Western Europe (Hirn et al., 1973; Ansorge, Mueller, 1973; etc.). The upper mantle in platform areas, as compared to active regions, is characterized by a rather smooth and gradual lateral variation of velocities, the presence of layers with higher velocities and the absence of thick asthenospheric layers of lowered velocities, at least in the case of compressional waves.

It has been found that the transition from one large geological structure to another is marked by a change in the velocity structure of the upper mantle manifested by velocity variation and a different vertical and horizontal velocity distribution. For instance, velocities in the uppermost mantle were found to decrease at the transitions from the Moscow to the Pechora syneclise, from the eastern margin of the Russian platform to the Urals and to the western parts of the West Siberian platform, and from the Kopet Dagh fore-deep to the central areas of the Turanian platform (Fig. 12). Also, a strong decrease in velocities of the uppermost mantle was recorded at the transition to the tectonically active regions of North Tien-Shan (Fig. 11C). Thus, differences

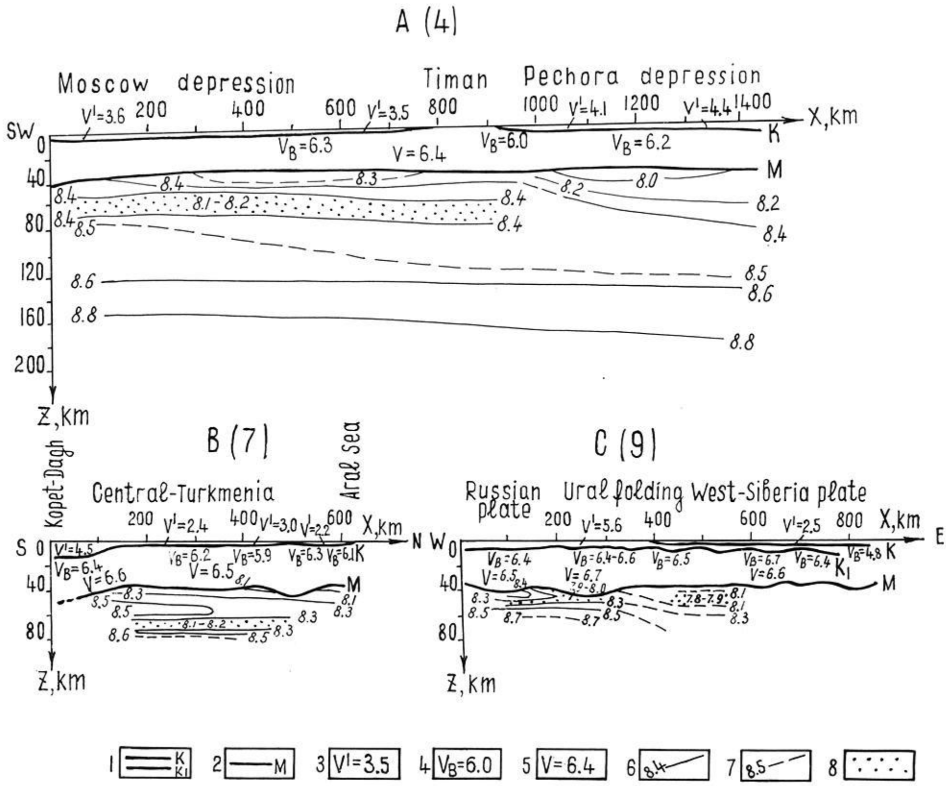

Fig. 12A–C. Velocity sections of the crust and upper mantle along DSS profiles: **A** (4) Moscow depression-Pechora depression (by Y.A. Burmakov, A.V. Egorkin, E.A. Popov and V.Z. Ryaboy); **B** (7) Kopet Dagh-Aral Sea (by V.Z. Ryaboy and Yu. A. Burmakov); **C** (9) Russian platform – West-Siberian platform (by V.Z. Ryaboy, L.N. Starobinets and V.S. Druzhinin). In parentheses are given the numbers of profiles as in Figure 1. 1: top of the consolidated crust (K) and top of the ancient (Archean) basement (K_1) on profile C (9) (in the east of the Russian platform $k_1 = K$); 2: Moho discontinuity; 3: average velocity in the sediments; 4: boundary velocity at the top of the consolidated crust; 5: average velocity in the consolidated crust; 6: contour lines of compressional velocities in the upper mantle from reliable data; 7: same from unreliable data; 8: low-velocity layer in the upper-most mantle. Velocities are given in km/s

between structural features can be traced not only in the crust but also in the upper mantle, which strongly indicates the influence of upper mantle processes upon the evolution of crustal structure.

It may be supposed that the velocity inhomogeneities revealed in the upper mantle of platform areas at depths of less than 80–100 km are caused by compositional inhomogeneities, i.e. for example, by an alternation of peridotites and eclogites, rather than by variations in thermodynamic conditions (Belyaevsky et al., 1975).

This seems to be likely because a temperature of 1000–1200° C, required for partial melting of the upper mantle material, may be found beneath the platforms at depths exceeding 100 km as evidenced by thermal and petrologic studies.

In conclusion it should be stated that the idea, which has been popular until recently, of a world-wide three-layer structure of the upper mantle consisting of approximately homogeneous layers-namely supraasthenosphere (lower lithosphere), asthenosphere and sub-asthenosphere, – is giving way to new concepts of a more complex three-dimensional distribution of upper mantle velocities, with the low-velocity asthenosphereic layer being found at depths less than 150–200 km and probably not everywhere.

A horizontally inhomogeneous model of the structure of the upper mantle opens the way for solving many urgent problems of geodynamics.

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