The lithosphere of the Ukraine

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The study of deep crustal and upper mantle zones is one of the major problems of both theoretical and practical character in modern geology. Principal methods of lithospheric analysis are geophysical, mainly seismic ones, as they provide the most reliable data on lithospheric structure and its physical properties.

In the past twenty years the methods largely employed in the Ukraine have been those of deep seismic sounding (DSS) and common depth point. This article is based on the results of that work. As a rule, seismic exploration has been carried out along the geotraverses crossing the Ukraine’s main tectonic structures (Fig. 1).

Continuous in-line profiling was the dominant method for the field investigations, while detailed systems of correlated travel-time curves were obtained for all the main types of waves (refracted, reflected, converted waves, etc) from the intracrustal boundaries, its bottom and, if possible, the upper mantle (Sollogub et al. 1973). During the field investigations the geophone spacing was 1–2 km, while the distance between the shot-points was 15–20 km for upper crust studies and 30–50 km for deep horizon analysis. The length of the travel-time curves was, respectively, 70–100 km and 180–220 km; to study the upper mantle they were extended to 400–600 km (Sollogub et al. 1980a). Interpretation included the use of all wave types, which increased the amount of information obtained.

Numerous reflecting and refracting boundaries are to be found in the lithosphere. The following are stable: (1) surface of the Preriphean basement and the younger basement (C1); (2) surface of the ancient protobase (C2); (3) the Mohorovičić discontinuity (M). In addition, a layer of crust–mantle mixture (C–M) is found in tectonically active regions of lower crust and upper mantle. In some mantle regions another layer has been noted with a lower velocity of elastic wave propagation, which is characteristic of asthenosphere or mantle inhomogeneity.

The aforementioned seismic boundaries can be briefly described as follows:

(1) The surface of the Preriphean basement (C1) is traced everywhere, its depth of occurrence varying from zero (in the Ukrainian Shield) to 15–25 km (in the Dnieper–Donets palaeorift, the Near-Black Sea depression) (Chekunov 1972). Crystalline rocks of the basement are of Archaean and Proterozoic age. The regions of less ancient consolidation possess a younger folded-metamorphosed basement of sedimentary rocks. Head waves (\(V_B = 5.8–6.2 \text{ km sec}^{-1}\)) and, in a number of regions with a thick sedimentary cover, reflected waves are recorded from the surfaces of the ancient and the young basements.

(2) The ancient protobase (C2) can be observed everywhere within the Ukrainian Shield and its flanks where it occurs at depths of 5–20 km, going down to 20–25 km under the depressions that surround the shield.

The structure of crystalline complexes in the Ukrainian Shield above horizon C2 is very complicated (Sollogub 1982). Seismic exploration by reflected waves has revealed here numerous reflecting planes dipping at different angles, which commonly form intricate combinations of synclinal- and anticlinal-like structures, lenses, cappeaks, etc (Fig. 2). These structures are composed of metamorphosed and granitised sedimentary volcanoanogenic formations of Archaean and Proterozoic age (Sollogub 1982). Seismic reflectors below horizon C2 are subhorizontal.

The total data obtained makes it possible to consider the seismic horizon C2 as the surface of the ancient protobase which corresponds to the lunar stage of the Earth’s evolution. We believe that originally the horizon C2 had a basal structure, which allowed the elastic wave propagation velocity of about 7 km sec\(^{-1}\). Then it was overlapped with sedimentary formations of Archaean and Proterozoic age which later underwent tectonic deformations, metamorphism and granitisation. As a result of those processes, the original and sedimentary layers have changed their physical properties arriving at their present state (\(V = 5.5–6.0 \text{ km sec}^{-1}\)). Due to granitisation, there have been changes in the upper part of the ancient basement (horizon C2) and some deeper complexes as well, where the velocity of elastic oscillations has decreased from its original values (7.0 km sec\(^{-1}\)) to 6.4–6.5 km sec\(^{-1}\) and less. The velocity increases gradually with depth, reaching 7.0 km sec\(^{-1}\) at a depth of 30–35 km. No sharp boundary with \(V_B = 7.0 \text{ km sec}^{-1}\) has been observed.

The situation is somewhat different in the depressions with thick sedimentary Phanerozoic formations. With the sediment depth of 10 km and more, horizons with \(V = 7.0 \text{ km sec}^{-1}\) have been noted in some sections of consolidated crust. It is known that when such depressions are formed on account of tension and concomitant plutonic transformations, the reduction of the crust’s ‘granitic’ complex occurs along with the general thinning of the crust. The rocks of the upper mantle uplift and the crust becomes more basic, as a result of which the horizon C2 ‘regenerates’.

Let us give a more detailed analysis of several seismic
sections which characterise the horizon $C_2$. Figure 2 presents a cross-section of the upper consolidated crust (up to 12–13 km) taken in the Ukrainian Shield’s central part. The section has been constructed from the reflected wave method and DSS data (Sollogub 1982).

The region is characterised by very thin sediments of latest times (up to 1 km) which cover the surface of crystalline rocks. It is known from geological data that the crystalline rock mass in the east and west of the analysed profile features two synclines filled by sedimentary volcanogenic rocks of early Proterozoic age. Granites of Archaean age crop out to the surface between them.

At a depth of 10–12 km DSS data establishes a horizontal key horizon $C_2$ occasionally disturbed by faults. Numerous reflecting planes found under the Early Proterozoic synclines corroborate the existence of the aforementioned structures. The conventional horizon $C_1''$ characterises the bottom of the synclines which have, in their axial parts, a depth of about 4.0–5.0 km. Beneath the horizon $C_1''$, another conventional horizon $C_1'$ is observed, occurring in the interval from 4 to 8 km deep and forming an asymmetrical syncline. The latter is not revealed either geologically, i.e. on the basement surface, or in the gravity field. Most probably, the horizon $C_1'$ divides the rock masses of Archaean age
into two strata, the upper of which crops out to the present-day surface between the Early Proterozoic structures, while the lower one lies immediately on the surface of $C_3$.

The analysed seismic section clearly demonstrates the complex structure of the rock masses that superpose the horizon $C_3$ and consist of three tectonic beds: the first, Early Proterozoic, forms synclines; the second, Upper Archaean, occasionally comes up to the present-day surface and is confined by the horizon $C_4'$ at depth; the third, Lower Archaean, lies between the horizon $C_4'$ and the key horizon $C_5$. The latter is the basement of the complete complex of sedimentary–volcanogenic formations, now metamorphosed.

Figure 3 presents a total section of the Earth’s crust near the town of Kirovograd. Here at a depth of 10–15 km the surface $C_2$ is observed, which in many sections is faulted with vertical displacements of up to 3–4 km. Below this boundary there are numerous reflecting planes lying almost perfectly horizontal. The M-discontinuity, also broken by faults, is found at a depth of 40–42 km. Isometric lines show that the velocity at the horizon $C_3$ equals 6.3–6.4 km sec$^{-1}$. Velocity gradually increases with depth, reaching 7.0 km sec$^{-1}$ at 38 km deep. Therefore, the complex of rock masses with $V = 7.0$ km sec$^{-1}$ and more can be only 3 km thick and occurs between 38 and 41 km deep.

The structural details of the horizon $C_2$ are essential for determining the character of the movement of separate blocks against each other in the Ukrainian Shield.

(3) The M-discontinuity is observed everywhere. The boundary velocity along the M-discontinuity is 8.0–8.5 km sec$^{-1}$. This discontinuity is very rarely a simple seismic boundary of the first type, with spasmodic increases of velocity.

As a rule, the structure of transition from the crust to the mantle is more complicated and multiform (Guterch et al. 1980). By way of generalising, we have distinguished five principal types of transitions which, in their turn, can be subdivided into different kinds (Sollogub 1980). In some regions, two or more M-discontinuities are observed. This is probably caused by spasmodic vertical displacement of the M-discontinuity due to phase transformations, while the boundary in the section (Ancient Moho) retains its earlier position. Appearance of double M-discontinuities is typical of the regions that have undergone tectonic transformations. At the M-discontinuity level deep transformation takes place, forming a new structural plane, while relics of the former plane are preserved too.

In the Ukraine the M-discontinuity is found at depths of from 30 to 60 km. Far from being chaotic, the depth changes have the following regularities: the crust thickens under mountain formations, i.e. orogens of all ages, whether existing at present or denuded; the crust grows thinner under troughs and depressions. The connection between the near-surface structures and the deep ones is evident both from the previously mentioned changes in the crust thickness and from the strike congruence. In the sections where the tectonic plane of the structures strike is sharply transformed, the M-discontinuities have the interferential character, reflecting, to a certain extent, each of the interstratified tectonic planes. This phenomenon makes it possible to determine the age of any structural plane of the M-discontinuity by comparing it with the near-surface geological structures, i.e. stratifying it (Sollogub and Chekunov 1980). A diagram of an M-discontinuity relief is presented in Fig. 5. The isolines reflect primarily the tectonics of the Early Proterozoic plane. This can be applied to the Ukrainian Shield and, partially, to some younger tectonic regions like the Dnieper–Donets palaeorift, Mountainous Crimea and so on. The latter regions features two M-discontinuities in the sections.
Fig. 3. Seismic crustal section of the Ukrainian Shield near the town of Kirovograd. 1. Pretriplane basement surface (C1). 2. Protobasement surface (C2). 3. M-discontinuity (M). 4. Reflecting planes. 5. Faults. 6. Diffraction points. 7. Isolines of real velocities from refracted wave travel-time curves (in km sec \(^{-1}\)). 8. Isolines of real velocities from reflected wave travel-time curves (in km sec \(^{-1}\)). 9. ’Granite’ complex of the Earth’s crust. 10. ’Basaltic’ complex.

where structures of different ages cross: the ancient one—Early Proterozoic, and the younger one—Baikal–Hercynian (the Dnieper–Donets palaeorift), Kimerian–Alpine (Mountainous Crimea) and others.

Submeridional structures are distinct in the diagram as zones with the thickest crust (up to 55 km) being, in fact, ancient mountain ‘roots’ of the Early Proterozoic age. These structures are traced not only throughout the East European platform, but to the south of it as well, i.e. through a part of the Scythian plate, Kimerian mountain formations of the Crimea, and even the young Black Sea depression of the Alpine age. In the latter case the relative amplitude of the ‘roots’ is far less than in the platform.

Along with submeridional structures, there are sublattitudinal structures of the M-discontinuity (the Dnieper–Donets palaeorift, the Carpathians, etc). The Dnieper–Donets palaeorift generally has a thinned crust (35 km). In the Donbass region this regularity breaks up, and the crust here is slightly thickened. Great thickness of the ‘basalt’ complex and the absence of the ‘granite’ layer resemble the structure of the crust’s geosynclinal part at the initial stages of its development. Apparently, the scientists who excluded Donbass from the East European platform, were right (Stille 1929).

Apart from the Dnieper–Donets rift, the thinned crust is found in the Transcarpathian region (town of Mukachevo, 25–30 km), Black Sea depression (25–30 km) and under a section of the Ukrainian Shield, near the town of Dniepropetrovsk (30–32 km) (Sollogub 1982).

The foregoing sections and the structural diagram of the M-discontinuity clearly demonstrate that the Earth’s crust is broken by numerous faults. They can be subdivided, according to their size and significance, into the following categories: (i) the lineaments separating large segments of the Central and East European lithosphere with thick and thin crust; (ii) deep fault-sutures limiting the principal tectonic regions of Phanerozoic age; (iii) deep fault-sutures limiting the principal tectonic regions prior to the Phanerozoic; (iv) inner block faults.

The majority of deep faults are not vertical but dipping and every tectonic region has its particular features. The Dnieper–Donets palaeorift, for example, is confined by faults that dip toward the centre of the depression, which attests to the crustal extension at the time of the structure’s formation. In the Donbass region, the faults that limit the structure in the south dip gently to the north, which may suggest the movement of the Near-Azov massif down to Donbass.

In the Early Proterozoic geosyncline region, the faults that bound the thick part of the crust are inclined to both sides of the geosynclinal centre.

(4) In some lower regions of the consolidated crust, a layer has been found with \( V_1 = 7.4–7.7 \text{ km sec}^{-1} \) (crust–mantle ‘mixture’), whose surface reflects a characteristic overcritical wave which can be dynamically compared with the wave from the M-discontinuity.

The existence of the crust–mantle ‘mixture’ with \( V = 7.4–7.7 \text{ km sec}^{-1} \), i.e. the velocity which is too great for the consolidated crust but too little for the mantle, is most probably connected with the phase transformation at the crust–mantle boundary. Apparently, the upper horizon of this layer corresponds to a new M-discontinuity whose formation is not yet complete. A similar structure in the crust’s lower part is observed in many continental and oceanic regions of the Earth and is typical of such tectonically active structures as rifts (Baikal, Rhine graben, etc). Thus, the presence of the horizon C–M with \( V = 7.4–7.7 \text{ km sec}^{-1} \) may serve as an indicator for detecting rift structures.

Figure 4 presents a crustal section along the profile which crosses the strike of the Dnieper–Donets palaeorift near the city of Kharkov. The structure of the sedimentary cover has been studied by drilling and seismic prospecting. The sedimentary thickness changes along the profile from 2–3 km at the south of the section (the Ukrainian Shield) to 15–18 km in the centre of the depression, decreasing then to 5 km within the Voronezh massif. The sedimentary layer consists of two strata: the lower one is composed of Riphean sediments (possibly Lower Proterozoic), the upper one of younger rocks. An extended horizon with \( V = 7.0–7.2 \text{ km sec}^{-1} \) is found only under the depression below the Preriphcean basement (\( C_1 \)), at a depth of 25–28 km.

In all probability, this seismic boundary is the proto-basement—horizon \( C_2 \). At a depth of 32–35 km in the depression’s central part there is a layer with \( V_1 = 7.6–7.8 \text{ km sec}^{-1} \) (crust–mantle mixture), limited from the south and the north by deep faults which, in the section’s upper part, are the corresponding boundaries of the Riphean graben. The thickness of the crust–mantle layer does not exceed 10 km. At a depth of 40 km and more, numerous reflecting planes are observed, which is characteristic of the M-discontinuity. The existence of the crust–mantle layer under the Dnieper–Donets depression and Donbass gives a reason to attribute these structures to a palaeorift. Apart from the Dnieper–Donets palaeorift, the crust–mantle layer has also been found in the Ukraine under the Near-Black Sea depression and the Precarpathian trough. In Poland it exists under the Dano-Polish furrow. The thickness of the crust–mantle layer in each region is about 10–12 km.

The regions where the described layer has been found are rift structures. In some cases they are ancient palaeoriffs (the Dnieper–Donets one), in other cases younger ones, bordering on the ancient East European platform.

In some parts of the Ukraine, a layer with low velocity of elastic wave propagation has been found at a depth of 50–100 km in the upper mantle (Belyaevsky et al. 1977, Sollogub et al. 1980a). A cross-section of the lithosphere along the line Kerch–Kilia–Vrancea (Fig. 6) provides a characteristic of this layer and of all the crustal and
upper mantle structure (Sollogub et al. 1982a). The section crosses the seismically active Vrancea zone with its deep foci (up to a depth of 160–170 km) which are related to the plane that goes beneath the Carpathians, resembling thus the Zavarianky–Benioff zone.

The cross-section of features: surface of the young (Kimmerian) basement (C_i), surface of the Preriphean basement (C_u), crust–mantle layer (C–M), M-discontinuity, seismic horizons P_1 and P_2 in the upper mantle, conducting horizon from magneto-telluric sounding, MTS (Stanica and Stanica 1981) and asthenosphere cover from geothermal exploration. The surfaces C_u and C_i form in the section's central and eastern part a gently sloping depression, extended laterally along the southern boundary of the East European platform.

There are several regions showing the M-discontinuity with the crustal thickness reaching 55 km and more. They correspond to the Early Proterozoic geosynclinal zones traced from the Ukrainian Shield to the Plain of Crimea region. In the section's western part the thick-crust zone characterises the 'roots' of the Carpathians.

Two seismic boundaries P_1 and P_2 have been detected below the M-discontinuity, at a depth of 85–100 km. The layer between these horizons is a low-velocity one. A similar layer has been noted in several regions of the Ukraine (Belyaevsky et al. 1977, Sollogub et al. 1980a). A structural diagram constructed by the mentioned layer's cover points to the presence of a lengthy swell extending to the north-east in the upper mantle. According to the diagram, the layer's cover forms an antithetic bend, indicated in the cross-section with a dotted line (Fig. 6).

The data on heat flow and magneto-telluric sounding provide important information about the deep structure of the lithosphere. The cross-section shows the temperature distribution in the lithosphere. Also outlined is the asthenosphere cover whose position has been determined by geothermal data (Kutas 1982). The temperature in the asthenosphere cover changes from 1400 to 1700°C. The highest temperatures within the analysed section correspond to the zones with relatively recent tectonic and magnetic activity. Among them is, in the first place, a narrow zone of the volcanic Carpathians where the isotherm of 1000°C occurs at a depth of about 50 km, whereas at a depth of 75–80 km the temperature reaches the point of basic rock melting (1300–1400°C) (Kutas 1982). The second high-temperature zone coincides with the depression on the Preriphean basement's surface. In this section the isotherm of 1000°C corresponds to a depth of 55–65 km.

An example of a relatively cold block is the section confined by the mantle faults in the region of Vrancea D’D”–Odessa E”. Here at a depth of 50–55 km the temperature does not exceed 800°C. It is worth mentioning that the isotherm of 700°C, or rather the interval between 600 and 800°C, coincides with the M-discontinuity, which can serve to justify the idea of the isothermal character of the boundary between the Earth's crust and the upper mantle.

According to geothermal data, the asthenosphere cover along the profile changes from 110 to 200 km deep. Magneto-telluric sounding shows that a conducting layer exists in some points at the same depth. Thus, it is possible to conclude that geothermal and magneto-telluric data characterise the same section corresponding to the asthenosphere cover. As to the low-velocity layer (P_1–P_2), it coincides with the asthenosphere cover only at some points, while being independent in other sections and characterising the upper mantle inhomogeneities.

The cross-section (Fig. 6) shows that the depth differences prior to the asthenosphere cover reach 50–60 km. The sections with sharp depth changes are evidently correlated with mantle faults. The most western fault (D’D”) occurs in the Vrancea seismically active zone, and all the foci of the zone are associated with it. The foci are absent in the asthenosphere layer, being concentrated in the lithosphere.

The mantle fault E'E” also exhibits a substantial depth difference prior to the asthenosphere cover, confining from the east a block with extremely thick lithosphere (up to 200 km). This fault occurs slightly to the south of Odessa and coincides with the deep fault that limits the Early Proterozoic geosynclinal zone from the west. There the asthenosphere cover assumes minimal values (about 100–110 km). This lithospheric elevated block is quite wide and ends in the vicinity of the deep fault F’F”.

The fault F’F” stretches out to the north-east, crossing the articulation of the Ukrainian Shield's central part with the Near-Azov massif, and later the zone of articulation of the Dnieper graben with Donbass.

In all probability, the aforementioned mantle faults can be traced not only in the north-east direction, but in the south-west direction as well. A number of abnormal manifestations are correlated with them in the crust. The faults are supposed to be ancient, although they manifest themselves at present, too.

The block of very thick lithosphere is confined by inhomogeneities from all sides. Mantle faults D and E are situated in the west and in the east; a low-velocity layer is found in the upper mantle, slightly below the M-discontinuity; and below the lithosphere there is the asthenosphere layer which also has low velocities and other physical properties different from the lithosphere.

In general, this subsiding block of the lithosphere forms a sort of a 'conducting' channel for the seismic waves generated by deep-focus Vrancea earthquakes. The seismic energy is concentrated in the channel and then travels along it, mainly in the north-eastern direction. This is clearly demonstrated by the position of the isoseismic lines which regularly form a north-east orientated ellipse after each deep-focus earthquake.
References


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