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Low velocity and low electrical resistivity layers in the middle crust

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Abstract
Some Deep Seismic Sounding (DSS) revealed low velocity layers in the upper and middle crust of old platforms. The layers are often characterised by a lower electrical resistivity. It is not clear, however, how reliable the layers recognized from DSS data are, if they are regular or occasional events and how they correlate with other geophysical parameters. To answer these questions the experimental DSS data obtained in the Baltic and Ukranian shields by different institutions were reinterpreted by the author with unified methods. The shield areas are well studied using both the DSS and high-frequency magnetotelluric sounding (MTS) methods. As a result a marked velocity inversion (waveguide) was observed in a 10 to 20 km depth range in the majority of the DSS profiles. An increase in the electrical conductivity is typical for the waveguide. A comparison of the results with the data from other platform regions allow the conclusion that this low velocity and high electrical conductivity layer has a global significance. In the continental crust, the layer is characterised by changes in the reflectivity pattern, earthquakes number and changes in velocity pattern where the block structure is transformed into a subhorizontal layering. These structural features suggest that the layers separate brittle and weak parts of the crust. Usually they play the role of detachment zones at crustal block moving. A possible factor responsible for this phenomenon is an increase in porosity and in the salinity of the waveguide pore water compared with the upper crust. This suggestion is confirmed by the Kola superdeep borehole data. Porosity increasing in the middle crust is explained by the change in rock mechanical properties with depth, by fracturing porosity and by dilatancy effect, at a depth of 10-20 km.

Key words *seismic and electromagnetic studies – Earth's crust – Baltic and Ukranian shields*

1. Introduction

At first the low velocity layers were determined in the crust from seismological data in tectonic active regions. The main indications of the velocity inversion with depth were the «shadow zones» (sharp attenuation of the first arrivals at some epicentre distances) and recording of the channel waves, which travel along the low velocity layer (waveguide) when an earthquake appears in the layer. The first Deep Seismic Sounding (DSS) confirmed the existence of the low velocity layers in the crust of tectonic active regions by the observed «shadow zones». These layer were interpreted as a result of high temperatures and partly melting.

Later the low velocity layers were determined in the cold crust of shield areas at a depth of 10-15 km. For example, they were determined on the Ukrainian Shield, where detailed seismic observations, so called continuous profiling, were carried out (Pavlenkova, 1979). It was the most detailed form of DSS: multichannel seismic stations moved along the profile with seismograph spacing of 100 m. Explosions from several direct and overlapping shotpoints with interval of 40-60 km were recorded. These

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results, however, were considered unrealistic because it was difficult to explain the nature of this layer: the velocity inversion in the upper crust of the cold shield might not be explained by temperature influence. Actually, the temperature at a 10 km depth under the Ukrainian Shield is not higher than 150-200°C, which can decrease the velocity gradient with depth but cannot form an inversion zone. Apparently, the geophysical studies revealed a «cold» type of weakening of rocks, which has another nature than the layers in tectonically active regions.

The velocity inversion zones were determined in the upper and middle crust in many platform regions: in the Baltic Shield (Grad and Luosto, 1987), on the Russian plate (EUROBRIDGE Seismic Working Group, 1999) and in the Siberian craton (Pavlenkova *et al.*, 2002). Often these layers are characterised by higher electrical conductivity (Kovtun *et al.*, 1994; Beljavsky *et al.*, 2001). It became clear that such layers may have a global significance, but up to now the nature of the layer has been debated (Jones, 1992; Karakin and Kambarova, 1997; Astapenko and Frainberg, 1999; Vannjan and Pavlenkova, 2001; Berzin *et al.*, 2002).

The goal of this paper is to answer the following questions.

How reliable are the low velocity layers recognized from DSS data and are they a regular or occasional feature of the continental crust? How do they correlate with electrical conductivity model? Are there any correlations between the layer structure, tectonic features and other geophysical parameters of the crust and what is the origin of these layers?

To answer these questions all experimental DSS data obtained in the Baltic and Ukranian shields during the last 40 years were reinterpreted with a unified methodology and the obtained velocity models were compared with the magnetotelluric and other geophysical and geological data. The shields were chosen for this research because they are the most investigated regions with detailed seismic and high frequency electromagnetic methods and their results may be compared with Kola superdeep borehole data.

Then the obtained data on the low velocity and high conductivity layers in the upper and middle crust of the shields were compared with the data on other platform regions. Finally problems of the layer origin are discussed.

2. Low velocity layers in the crusts of the Baltic and Ukranian shields

More than twenty seismic profiles crossing all tectonic structures of the Baltic Shield have been processed in this region during the last 40 years (fig. 1c). They were carried out by different institutions with different instrumentations, techniques and interpretation methods. As DSS is developed the old data were reinterpreted many times and now there are several velocity models for each profile. The low velocity layers (velocity inversion zones or waveguides) were distinguished in the middle crust at depths of 10-15 km along the seismic profile Sveka (Grad and Luosto, 1987). The inversion is observed not only in the *P*-wave velocities but in the *S*-velocities as well.

The inversion zones, however, were not distinguished along other profiles. It was difficult to understand if the layer is local and does not exist in other regions or they were missed during the data interpretation.

In order to obtain comparable data for the whole Baltic Shield, a comparative analysis of experimental records was made by the author for all DSS profiles. The wave fields were analysed and interpreted using unified methods. As a result the low velocity layers were determined in the upper crust along the most profiles. Some results of the old seismic data reinterpretation were published in Berzin *et al.* (2001). In this paper, new models are presented for the profiles Quartz and Fennia. The Quartz profile was carried out by Centre GEON (Moscow) as a part of the Russian superlong seismic profile net made for the lithosphere studies (Egorkin, 1991). The Fennia profile was made by the Finnish Institute of Seismology with distances between the seismic stations of 2-3 km and between shots of 80 km.

As mentioned above, the main feature indicating a velocity inversion is a regular attenuation of the first arrivals at some distances from the source (appearance of so-called «shadow zone»). Figure 1a-c shows the traveltime curves and seismic rays for the inversion zone determined from SP 9 along the Quartz profile. Seismic rays penetrating the crust and arriving at the surface at source-receiver distances of about 100 km (*Pg*-wave) form a traveltime branch having an apparent velocity of about 6.4. km/s. The rays entering a velocity inversion zone are deflected downward rather than upward, intersect the zone, and come back, being reflected from its base (*K*1 interface) or refracted in the underlying layer. The related reflection and refraction traveltime branches have velocities not higher than 6.5 km/s. The *K*1 and *Pg* traveltime plots are nearly parallel and are separated in time by Δt. The time delay depends on the waveguide thickness (Δh) and a value of the velocity inversion (Δv) and provides major constraints on these parameters.

The average velocity in the waveguide and its thickness are not uniquely constrained by refraction traveltime data. A set of the crustal velocity models having different waveguide parameters Δv and Δh may be determined. The velocity decrease could not be ambiguously determined even with a sufficiently long traveltime curve of the wave reflected from the waveguide base. For instance, in the case shown in fig. 1a-c, the velocity decrease can vary from 0 to 0.4 km/s, with the base depth varying from 15 to 11 km.

To obtain comparable velocity models for all profiles, the data from the Kola superdeep borehole were used. The borehole determined the velocity at depths of the waveguide of 7-12 km as 6.1 to 6.2. km/s (Kozlovskii, 1984; Pavlenkova, 1991). These values were used for the all velocity inversion zones in the Baltic Shield. Other parameters of the waveguide and

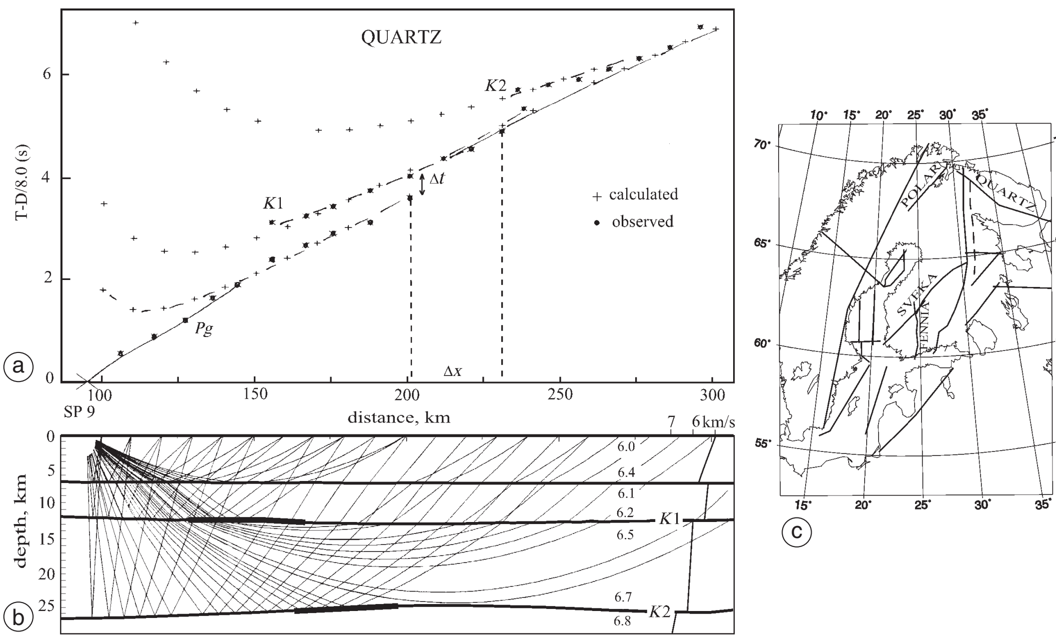

Fig. 1a-c. a) Traveltime plots and (b) seismic rays for the crustal velocity model constructed along the Quartz profile (Baltic Shield) with an inversion zone at depths of 7 to 12 km. *Pg* – refracted wave in the upper crust (5.8-6.4 km/s); *K*1 – wave reflected from the base of the inversion zone (6.4-6.7 km/s); *K*2 – wave reflected from the top of the lower crust (6.8-7.2 km/s); Δt is the time delay between *Pg* and *K*1 branches. Thick lines in the cross section are reflectors. c) A schematic map showing the position of DSS (solid lines) and MTS (dotted line) profiles in the Kola Peninsula and Karelia region used in this work.

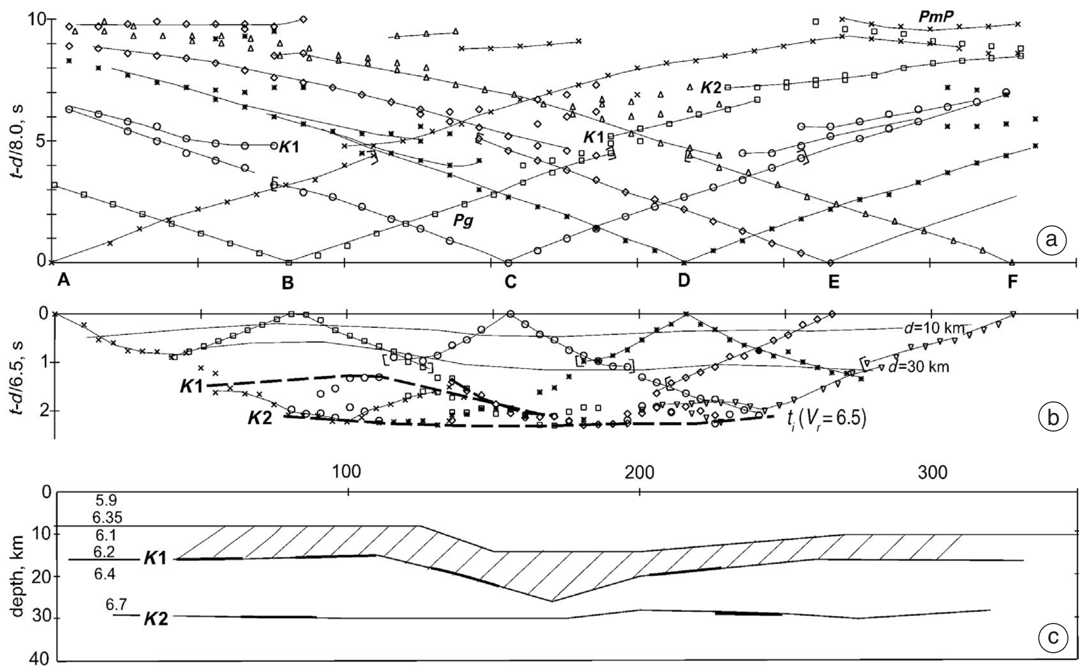

Fig. 2a-c. Observed travel times and the velocity model of the upper and middle crust along the Fennia profile. a) Travel times in a traditional form, the reduction velocity of 8.0 km/s; b) a time section: the observed travel times reduced with the reduction velocity 6.5 km/s and plotted at source-receiver midpoints with downwards time axis; the lines d = 10 and 30 km coincide the travel times at the offsets of 10 and 30 km, the dotted lines are intercept time curves for $K1$ and $K2$ boundaries; c) velocity model, thick lines are reflectors.

underlying layer were determined from 2D mathematical modelling (ray tracing).

The analysis has shown that the observed wave pattern involved in the recognition of low velocity layers is not always clear. It is often complicated by the fact that the shadow zone is not observed due to diffraction phenomena, and the waves reflected from the waveguide base are not recorded.

Wave patterns, which look like the «shadow zones», may also be observed at strong lateral inhomogeneity of the crust. To discriminate them from effects of low velocity layers, it is necessary to analyse the whole system of reversed and overlapping observations. The «shadow zones» connected with a low velocity layer are to be observed at the same distances (offsets) from all shots and it is the main criteria to trace the inversion zone along the profile. For this a special method of the travel-time analysis was applied and they were presented in form of the intercept time cross-sections (Pavlenkova, 1982).

Figure 2a-c presents an example of the time section for the Fennia profile. Figure 2a gives the system of the traveltime curves in a traditional form. From all shots the time curves show a regular picture. The waves *Pg* are recorded with apparent velocities around 6.0-6.3 km/s at offsets of 0-100 km. At larger offsets Pg attenuate and the $K1$ reflections became the first arrivals with velocities about 6.3 km/s. A time delay of about 0.5 s is observed between the *Pg*- and $K1$-waves. At offsets of 110-170 km the $K2$-waves are recorded with velocities of 6.8-7.0 km/s without any time delays regarding the $K1$-waves.

To see how this picture changes along the profile, the observed travel-times are shown in fig. 2b in another form: they are reduced with velocity 6.5 km/s (typical velocity at the inversion zone bottom) and are plotted at source-receiver midpoints. The lines coincided the travel times of first arrivals at constant offsets (d = =10 and 30 km) show the uppermost crust structure. The line t (d = 30 km) at which the first arrivals attenuate (1.0-1.2 s) outlines the inversion zone top. An envelope of the reflection $K1$ times may be used to determine a form of the inversion zone bottom. This line is intercept time (t_i) line for the $K1$ boundary and they may be recalculated in the depths to the boundary (the boundary reflections tough the t_i-line in the critical points). The $K1$ reflection envelope shows strong inclination of the boundary between the Shot Points (SP) B and C. In contrary the travel-times of refracted waves from the $K2$ boundary form a near horizontal t_i-line at the reduction velocity of 6.5 km/s.

The time section was used as a starting model for the following ray tracing. Resulting velocity model is given in fig. 2c. The low velocity layer is determined at the same depth as along the Quartz profile. But the layer structure is more complicate along the Fennia profile. Beneath SP-C the layer depth increases from 8 to 14 km and its thickness increases as well. It seams that it is influence of a crustal fault.

Similar velocity models were obtained for the Ukranian Shield. In the shield area, several DSS profiles were made in the 60-70's. That was the continuous profiling with analogue multichannel seismic stations. The data were

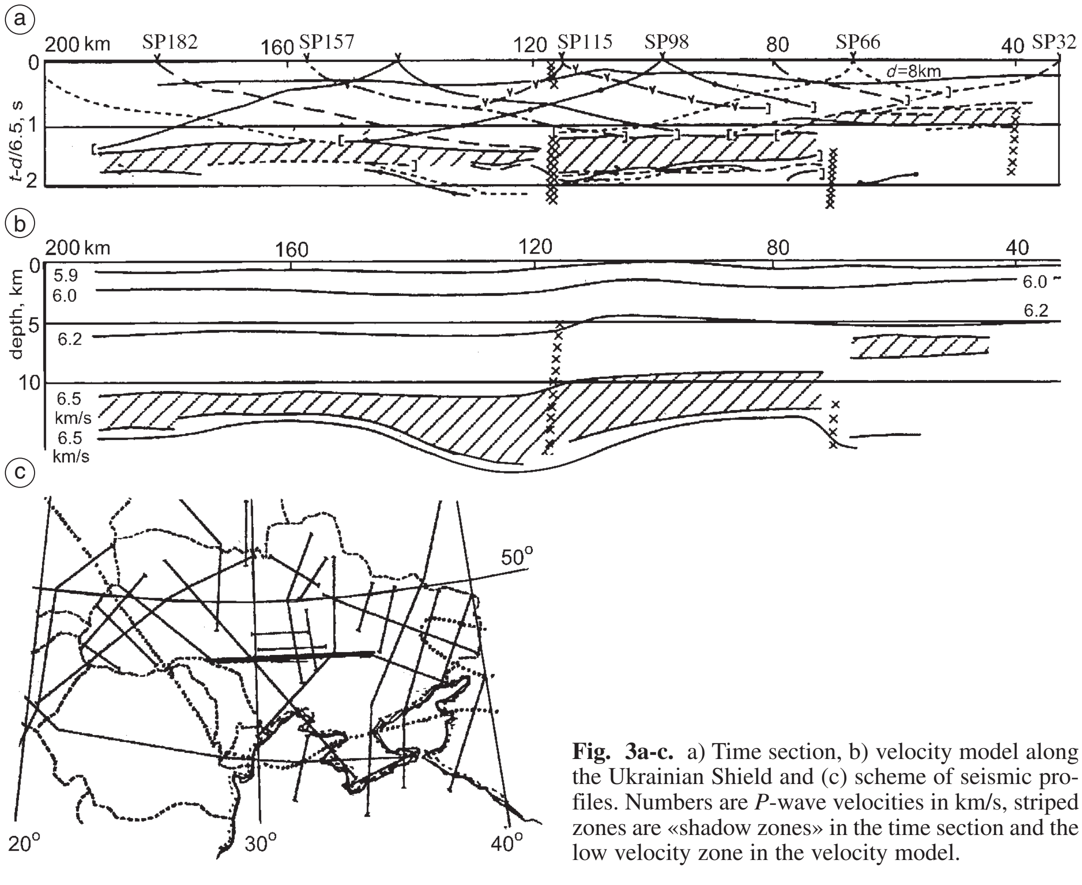

Fig. 3a-c. a) Time section, b) velocity model along the Ukrainian Shield and (c) scheme of seismic profiles. Numbers are P-wave velocities in km/s, striped zones are «shadow zones» in the time section and the low velocity zone in the velocity model.

reinterpreted with the same methodology as the Baltic Shield data. Below the results of the reinterpretation are presented for the profile along the shield (fig. 3a-c). Figure 4 presents a record section that is typical for many profiles in the Ukrainian Shield. It shows a clear «shadow zone» at offsets of 75 km. The observed travel times transformed to the midpoints and the corresponding time section (fig. 3a) show that the «shadow zone» may be traced along the whole profile. The time sections determine an increase in the waveguide thickness in the central part of the profile and a local anomaly in the profile interval of 50-70 km. Figure 3a-c also shows that the time section along the Ukrainian Shield is similar to that in the Baltic Shield and the low velocity layer was determined at the same depth of 10-15 km.

Thus, the main result of the time section comparison is that the velocity inversion exists

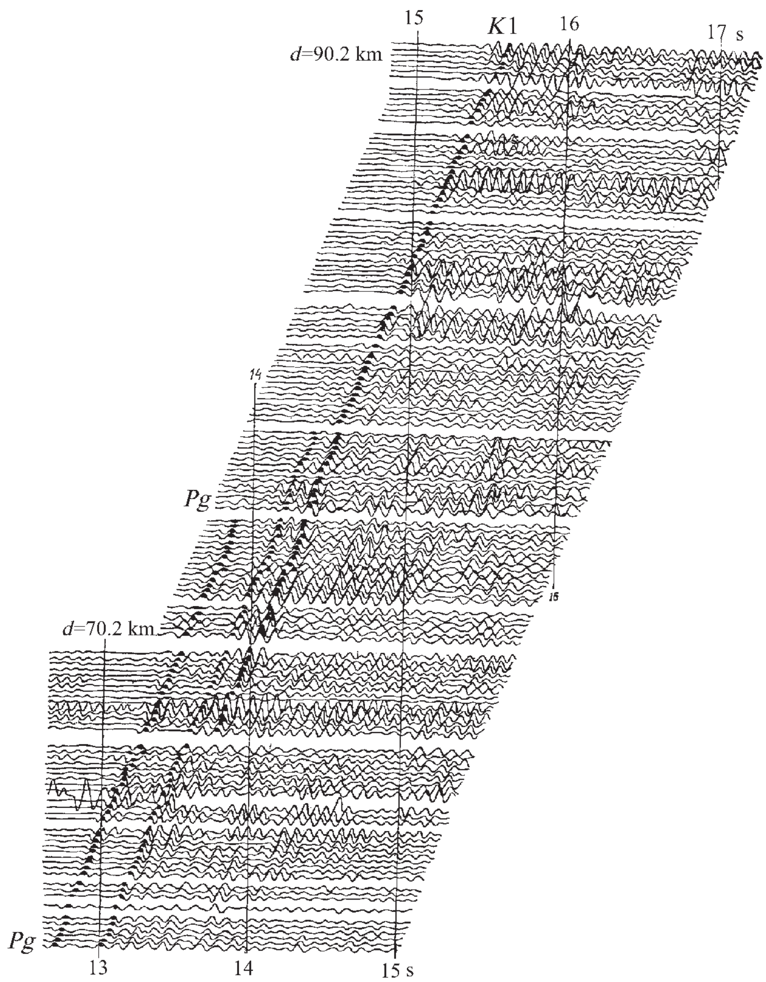

Fig. 4. Record section from DSS profile in the Ukrainian Shield which shows the «shadow zone» (attenuation of the *Pg*-wave) at distances from the source d=75 km. Legend in fig. 1a-c.

in the shield crust along the most seismic profiles. Thickness of the inversion zone varies but there is no doubt that it is a regular feature of the shield crust.

3. Magnetotelluric data

Electrical conductivity constraints on the upper crust of the Karelia region and Kola Peninsula were gained from high-frequency (10^{-2}-10^{-4} s variation period range) magnetotelluric soundings (MTS) (Kovtun *et al.*, 1994). Relatively dense coverage of the Baltic Shield by DSS and MTS profiles provides a good basis for examining in detail the upper crust anomalous layers and gaining constraints on their origin.

The most important result of these observations was the detection of an abrupt drop in the apparent resistivity in a variation period range of 10^{-2}-1 s, which is direct evidence of a decrease in the electrical resistivity at depths of 8-12 km. The apparent resistivity reaches a minimum near a 1 s period and increases by a few times with the further increase in the variation period. Such behaviour of the apparent resistivity indicates the presence of a conductive layer in the area studied.

As seen from fig. 5, the conductive layer is traced along the NS profile about 1000 km long from the Barents Sea coast to Lake Ladoga (fig. 1c).

In the Ukrainian Shield many magneto-telluric soundings were also carried out (Beljavsky *et al.*, 2001). For the most of them the high electrical conductivity layers are determined at a depth of 10-20 km.

A combined analysis of seismic and electromagnetic data makes their physical and geological interpretation more reliable, because these methods differ in their constraints on crustal properties. The seismic wave velocity is mostly controlled by the rock composition and partly by fluid concentration in pores and fractures. The electrical conductivity of fluid-bearing rocks is nearly independent on the solid phase composition and is determined by the fluid content and its salinity. If the depth and thickness of crustal conductive layers correlate with the parameters of the low *P*-velocity layers they determine the layers as zones of higher porosity and higher concentration of fluids.

The MTS data may be used for determination of the porosity in the high conductivity layer which is difficult to retrieve from seismic data. For the Baltic Shield such determinations show the following (Vannjan and Pavlenkova, 2002). The upper 10 km of the shield crust are characterised by low porosity. However, even negligible amounts of pore water dramatically decrease the electrical resistivity. This makes it

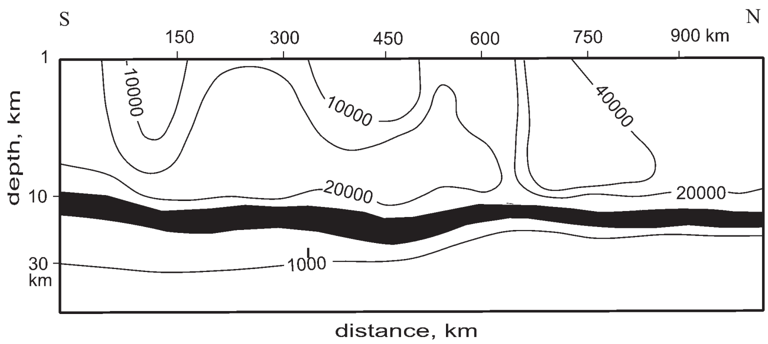

Fig. 5. Geoelectrical structure of the crust along the combined NS Teriberka-Loimola (TL) MTS profile (Kovtun *et al.*, 1994). The electrical resistivity contours in the crust are results of the 1D interpretation. A conductive layer is discovered in 8 to 15 km depth interval (black strip). The position of the profile is shown in fig. 1c.

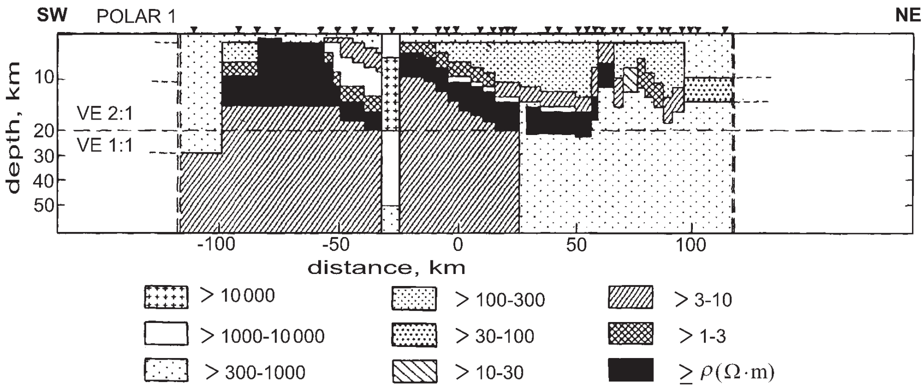

Fig. 6. A 2D geoelectrical model along the Polar profile (Korja *et al.*, 1989). The MT sites are indicated by the inverted triangles. The position of the profile is shown in fig. 1c.

possible to estimate the porosity ρ using the Archie Law, connecting the porosity with the electrical resistivities of the water-saturated rock ρ and pore water ρ_w: a high mineralization of the pore water can be expected in the static zone of the crystalline basement. Various direct effects indicate the presence of brines whose resistivity drops to 0.03 Ω·m at the upper crust temperature. Using this value and an average resistivity of the crust of 10^4 Ω·m, it was found $\rho = 0.1\%$. This value coincides with the generally accepted porosity of granite observed in samples tested in laboratory. This estimate can somewhat increase, if pores and fractures are filled with a fresher solution. Thus, the sensitivity of geoelectric data to the water saturation of rocks is high enough for detecting a porosity of a few tenths of a percent.

The depth and thickness of the Karelia and Kola Peninsula crustal conductive layers correlate well with the parameters of the low *P*-velocity layers. These layers are characterised with lower values of the *S*-wave velocities as well. The *P*- and *S*-velocities ratio do not differ from the ratio in other crustal layers. It is natural to conclude that the velocity and electrical resistivity decreases have a common origin. The increase in the porosity to about 1%, determined from electrical resistivity is also correlated with the seismic data. The 0.2-km/s decreases in the *P*-wave velocity results in the same value porosity (1%). Both approaches yield evidence that a 5 km layer containing up to 1% of water is present at a depth of about 10 km. A similar value (1.19%) was obtained from data of the Kola overdeep borehole (Kozlovskii, 1984).

The MTS data show another important feature of the layer discussed. The inclined high conductivity zones which correspond to the fault zones, flattened out to the low velocity layer. This example is presented in fig. 6 for the Polar profile. The latter was determined as a result of the data reinterpretation at depth of 10-20 km.

The same relation between the faults and the inversion zones are also observed from CDP data. Most faults, identified from geological data are traced as clearly visible reflectors usually within the upper crust. Their shape regularly changes with depth, so that near the surface they are practically vertical and at depth of about 10 km they flatten out. But it does not mean that there are no deep faults that cross the whole crust. The deep faults flatted out to the Moho are observed from the MTS data along the Sveka profile (Corja *et al.*, 1993) and from DSS and CDP data along the Kem-Uchta profile, Karelia (Berzin *et al.*, 2001).

The low velocity layers were determined in other regions of the East-European platform as well. Their depths are 10-20 km in the south and central parts of the platform (Pavlenkova, 1996; Baranova and Pavlenkova, 2003) and along EU-

ROBRIDGE profile in the western part of the platform (EUROBRIDGE Seismic Working Group, 1999). The inversion is observed in the *S*-wave fields as well. The *P*- and *S*-velocity ratio is determined as 1.71, the same as in the upper crust. The same results were obtained from the Sveka profile in the Baltic Shield (Grad and Luosto, 1987).

The low velocity and high conductivity layers were also discovered in the middle crust in other platform regions: in the Indian and Canadian shields, in the American ancient platforms (Berdichevsky *et al.*, 1984; Jones, 1992; Pavlenkova, 1996; Padilha *et al.*, 2000). In the young West-European plates the correlation between the low velocity layers and the high conductivity layers is also the same as in East-European platform (Aichruth *et al.*, 1992).

Such correlation allows the conclusion that this low velocity and high electrical conductivity layer in the middle crust has a global significance. What might be the origin of this layer?

4. Nature of the low velocity and high conductivity layer in the middle crust

The low velocity and high electrical conductivity layer in the middle crust is characterised by some structural peculiarities which are important for understanding the origin of the layer. The data presented above and the results of other data analysis show that in the crust at depths of 10-20 km there are a number of properties which identify the low velocity layer as a principal structural element of the continental crust. Among these properties the most typical are the following:

– The velocity inversion zones divide the crust in two parts with different structural pattern: the inclined reflectors which may be interpreted as faults, flatten out at the low velocity zone and the steeply dipping structures of the upper crust are replaced by sub-horizontal structures in the lower crust.

– A distinct seismic boundary (the reflector *K*1) often underlies the low velocity layer.

– Detailed research by seismic reflection method (CDP) shows the greater sub-horizontal stratification of the crust (increasing reflectivity) at depths below 10-15 km. This was also proposed from the detailed DSS data (Pavlenkova, 1996).

– The lower edges of the crustal bodies that give rise to gravity and magnetic anomalies are usually found at depths not lower than 10-15 km.

– A good correlation is observed between the velocity inversion and the distribution of the earthquakes: they occurred above the inversion zone. In the Baltic Shield crustal earthquakes are normally located above depth of 12 km (Korhonen and Porkka, 1981). It means that the mid-crustal layer is a weakened zone.

The main structural features of the crust are shown in a schematic form in fig. 7.

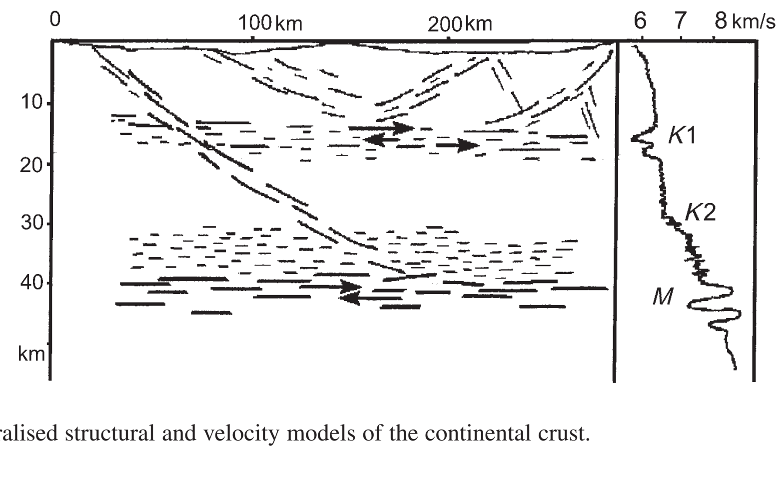

Fig. 7. Generalised structural and velocity models of the continental crust.

It should be noted, that these structural characteristics of the crustal waveguide are observed not only in the old platform areas. In the Baikal Rift Zone the low velocity layer, determined in the middle crust, has the same characteristics (fig. 8): higher reflectivity, flattening out of faults and lower earthquake number (Krilov *et al.*, 1990). It suggests that the crust of the Baikal Rift Zone has not been transformed by higher temperature regime and up to now maintains the platform crust features.

According to the current notions of crust rheology presented over the last fifteen years, the main factor responsible for the weakening of rocks and for the velocity inversion is the temperature increase with depth. The depth of the brittle/plastic transition zone increases as the heat flow in the studied region decreases. This depth exceeds 20-25 km under shields. New studies of the Baltic Shield rheology (Kukkonen and Peltonen, 1999; Kaikkonen *et al.*, 2000) confirmed that the depth of the brittle/plastic transition zone in the archean crust of Karelia is about 30 km. Thus, the heating-induced weakening of the crust is expected to occur at a depth three times greater than its value constrained by seismic and geoelectric data. Actually, the temperature at a 10 km depth under the Baltic Shield is not higher than 150°C, which can decrease the velocity gradient with depth but cannot form the inversion zone.

Apparently, the geophysical studies revealed a «cold» type of weakening of rocks at a depth of about 10 km, which cannot be accounted for by the temperature factor effective in tectonically active regions.

To explain the crustal high conductivity zones, the problem of the fluids origin is mainly discussed (Jones, 1983). Dehydration is considered as a more possible factor favouring the formation of lower velocity and higher electrical conductivity zones in the middle crust. This effect is supported by data from the Kola superdeep borehole. Here, a 0.2-0.3 km/s decrease in seismic velocities, discovered in a homogeneous sequence at a 4.5 km depth, correlates with a decrease in the bound water concentration and an increase in the amount of free water (Kozlovskii, 1984).

New studies of the Earth's fluid regime show that the Earth degasation is the main source of the mantle and crustal fluids (Letnikov, 2000). So the main problem is not the fluid origin but an origin of higher porosity in the middle crust at a depth of 10-20 km that enables concentration of the fluids.

The most reasonable explanation of the higher porosity and other properties of the low velocity crustal layer follows from the theory of dilatancy cracking of the crust (Nikolaevsky, 1985; Karakin and Kambarova, 1997). The theory is based on laboratory data on the mechanical properties of the

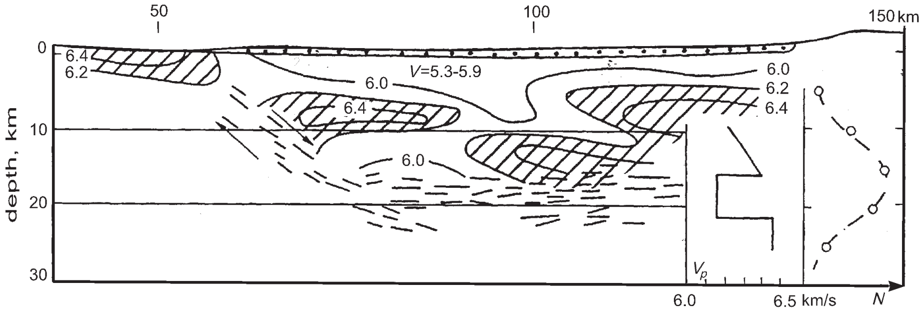

Fig. 8. Seismic cross-section and velocity model of the Baikal Rift Zone (Krilov *et al.*, 1990). V_p is seismic velocity, N is earthquake number.

rocks under high pressures and temperatures and on theoretical considerations of the rock mechanical properties at the PT condition typical for the cold crust (below 500-700ºC). In the upper crust normal vertical faults are formed as a result of horizontal displacement stresses; tilted shear faults are located below, and at a depth of 7-10 km they degenerate into completely sheared rocks. The same characteristics of faults in the uppermost crust have been described in Jackson and McKenzie (1983). The development of extremely fine cracking at depths of more than 10 km results in the saturation of rocks by fluids, the appearance of the low-velocity and high-conductivity layers and a corresponding increase in their plasticity.

Thus, the upper crust is a region of brittle deformations, the features of the low velocity and high conductivity layer appear to indicate an increase in the fracturing porosity in the middle continental crust. In other words, it is a weakened water-saturated layer.

This conclusion is supported by the Kola supperdeep borehole data. At a depth of 7-12 km the hole reaches a zone of weakened rocks and higher fluid flow (Pavlenkova, 1991).

The suggestion that weakened zones in the upper and middle crust are developed due to rock fracture and dilatancy effect can also explain all the structural peculiarities of the upper and middle crusts that were pointed out above and that are not considered by other hypotheses. The suggestion agrees with the observed structural peculiarities of the crust and with geological data. According to these data, the weakened layers can be associated with zones of relative horizontal motions of the upper *versus* lower crust (detachment zones, fig. 7). These are zones of fractured rocks similar to subhorizontal faults.

The horizontal displacement of separate layers, the formation of nappes, and plastic flow of the crustal matter can be associated processes causing the formation of a sub-horizontal seismic boundary $K1$ at the bottom of the crustal waveguide. It is difficult to explain this boundary nature in another way.

The weak layers can play an important role in any tectonic processes not only as the detachment zone at crustal block moving. Together with deep faults they form a channel system for the mantle fluids and matter transportation. It means that the low velocity and high conductivity crustal layers have a number of uses for identifying zones of mineralization, which are of economic importance for detecting fluids and determining the depth of weak zones. The latter is also important for studying seismicity.

4. Conclusions

The comprehensive analysis and reinterpretation of the seismic experimental data obtained in the Baltic and Ukrainian shields show that a layer of lower *P*- and *S*-wave velocities exists in a depth range of 10-20 km in their crust. The layer is reliably recognised from refraction and wide angle reflection data if systems of reversed and overlapping profiles are analyzed. The low velocities correlate with a lower electrical resistivity.

The comparison of the data with other geophysical and geological information from different regions suggests this low velocity and high conductivity layer have a global significance. Several structural features are typical for the layer: a change in velocity inhomogeneity where the block structure is transformed into a subhorizontal layering, changes in the reflectivity pattern and earthquake number. These properties of the middle crust layer mean that it can be associated with a weakened zone and it suggests a rheological stratification of the crust.

A possible factor responsible for this phenomenon is an increase in porosity and in the salinity of the waveguide pore water as compared with the upper crust. This suggestion is confirmed by the Kola superdeep borehole data. Porosity increasing in the middle crust is explained by a change in rock mechanical properties with depth and by their fracturing porosity and by delatancy effect at depth of 10-20 km.

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