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Lateral variations of the mid-mantle conductance beneath Europe

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Abstract

Europe is a region with the largest density of geomagnetic observatories and several authors have used these data to estimate local geomagnetic response functions for various period ranges, typically of the width of 1.5 to 2.5 decades. By collecting the local response functions from 35 European observatories, and by their precise selection and subsequent combination, the independent regional geomagnetic induction data set could be extended to a period range of 4.5 decades. The initial local responses that were estimated by two magnetovariation methods, with two different external source fields employed, have been further supplemented by continental and global 11-year data, providing thus a data set extending over a period range from the harmonics of the daily variations up to 11 years. The combined responses have been inverted individually for each observatory by two techniques, by an Occam procedure and a stochastic 1D inversion for spherically symmetric Earth. The integrated mantle conductance has revealed rather regular lateral changes that have been used to design a mantle conductance image down to a depth of about 770 km. The presented conductance image can be correlated with major European tectonic units like the Baltic Shield and the Trans-European Suture Zone.

To examine possible distortions to the inferred mantle conductance models due to large-scale near-surface heterogeneities, specifically those caused by the oceans, seas and large sedimentary basins, a spherical forward modeling was carried out for a radially symmetric conductor coated by an inhomogeneous thin shell with the variable surface conductance. The model responses for the 35 observatory positions were inverted in the same way as previously the experimental data. The results for 28 observatories have shown that the depth down to a pre-defined conductance level could be retrieved with a high accuracy of a few percents, but for seven southernmost observatories the recovery error increased up to 9%. With these seven observatories removed from the analysis, the effect of the seas and oceans on the upper and mid-mantle conductance estimates beneath Europe can be considered negligible.

Keywords: Mid-mantle; Mantle conductivity; Europe; Induction methods

1. Introduction

Investigations of the Earth's mantle conductivity are based on the analysis and interpretation of the long period magnetic field variations induced by primary sources located in the ionosphere or/and magnetosphere. The natural magnetic variations have been recorded continuously by a world-wide net of geomagnetic observatories for tens of years. Analyses of these records have resulted in designing various global conductivity depth profiles within the Earth's mantle (e.g., Banks, 1969; Berdichevsky et al., 1969; Rokityansky, 1982; Schmucker, 1985). A mid-mantle conductive

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layer at depths of about 700 to 900 km has been detected by local magnetovariation soundings (e.g., Roberts, 1984; Schultz and Larsen, 1987; Semenov, 1989). Note, that already Zharkov (1983) theoretically predicted the existence of this layer as a global conductivity feature, and the analysis and interpretation of the MAGSAT satellite data (Oraevsky et al., 1993) confirmed its global nature experimentally. Later, lateral variations of the mid-mantle conductive layer on a global scale have been reported by Schultz and Semenov (1993) and Schultz and Pritchard (1999).

A particularly dense net of the geomagnetic observatories in Europe provided a good opportunity to detect some trends in the spatial distribution of the mantle conductivity on a regional scale. By interpreting data from various period bands, several different trends in the deep conductivity distribution, limited to specific depth ranges, in the mantle beneath Europe have been reported so far. By analyzing the amplitude and phase characteristics of the 27, 13, and 9-day geomagnetic variations and by studying their distribution over the European region Pecova et al. (1980) concluded that '. . . the northern part of Europe might be a region with increased electrical conductivity at depths of about 1,000 km'. Another conclusion was obtained by Roberts (1986) that 'the depth to the conducting zone seems to vary from about 600 km in north-western Europe to about 800 km in south-eastern Europe'. Recent investigations by Olsen (1998), based on the analysis of the C-responses, have shown that 'there is a general increase of Re{C} to the north for periods between 6 and 300 hr, which probably indicates lower conductivity in the north'. In this paper an attempt is made to specify the spatial distribution of the mantle geoelectrical structure beneath Europe more accurately. This work is based on processing a collection of the geomagnetic induction responses obtained by various authors for various period ranges various subsets of observatories. A thorough selection and re-combination of the original response functions from 35 observatories located across Europe (Fig. 1) has been carried out in order to provide a maximum number of reliable response estimates over a period range as broad as possible, specifically from 6 h up to 11 years. As a next step, the individual observatory transfer functions have been inverted for observatory-specific radially symmetric conductivity models of the Earth. Eventually, all the individual 1D models have been stitched together to provide a picture of the lateral variations of the mantle conductivity structure beneath Europe.

In order to investigate a possible impact of the conducting oceans and seas on the induction responses analyzed we perform an additional model study on a quasi-3D model that consists of a thin spherical shell of a variable conductance on the Earth's surface that overlays a radially symmetric layered conductivity of the deeper crust and mantle.

2. Selection and combination of the response collection

The collected data for the European observatories include the response functions estimated by Roberts

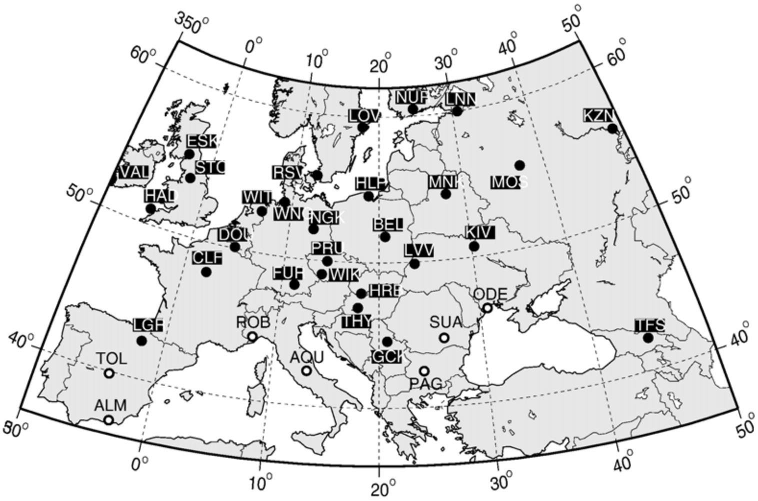

Fig. 1. Positions of the European geomagnetic observatories that are involved in this study. The international codes of observatories ignored in the conductance image are shown without the black background.

(1984; 9 observatories, period range of 3 to 73 days), Schultz and Larsen (1987; 7 observatories, period range of 5 to 100 days), Semenov (1998; for 17 observatories, period range of 2 to 750 days) and Olsen (1998; 33 observatories, period range of 0.125 to 30 days). The responses by Roberts (1984), Schultz and Larsen (1987) and Semenov (1998) have been obtained by using the Z/H method (Banks, 1969) while the responses presented by Olsen (1998) have been estimated by applying the Z/Y method (Schmucker, 1985). This collection was augmented by regional responses obtained for the Eurasia region for periods from 100 days up to about 2 years, as well as by selected global response functions for three time harmonics of the 11-year variations (Semenov and Jozwiak, 1999).

All the data have been converted into complex apparent resistivity values ρ, which are defined as $\rho = \pm\zeta^2/i \cdot \mu \cdot \omega$, where μ—the magnetic permeability, ω—the cycle frequency, i—the imaginary unit, ζ—the complex impedance, and the sign depends on the sign within the time-harmonic factor exp $(\pm i\omega t)$. The physical meaning of such a conversion is simple: this complex apparent resistivity is real and equal to the true resistivity for a homogeneous isotropic medium. The above apparent resistivity is defined as a complex-valued parameter, contrary to the commonly used EM transfer functions that considers a real apparent resistivity (as absolute value) and phase of the impedance or C-response. A mean apparent resistivity computed via averaging the real and imaginary parts of the C-responses on the one hand and an average value of the above introduced complex apparent resistivities on the other can be generally different values because of using the different average procedures (Semenov, 1998). This is significant for our analysis to prepare the homogeneous data. The module and phase are physically measurable parameters in contrast with the purely mathematical entities, the real and imaginary parts. Thus, we consequently use the terms module (or absolute value) of the complex apparent resistivity and phase of the complex apparent resistivity (from $-90°$ up to $90°$) everywhere to avoid confusion with the commonly used apparent resistivities and phases of the impedance.

In the selection step of our analysis, those responses have been eliminated from the collection for which the coherence between the geomagnetic components Z and H was less than 0.7 or the phase value of the complex apparent resistivity was out of the interval $\pm 90°$, the latter indicating inconsistency with a 1D conductivity model of the Earth. Finally, the responses corresponding to the daily variation harmonics with periods shorter than 6 h have been removed from the ensemble because of their unacceptably large scatter for different observatories.

Contributors of the primary data gave estimates of the confidence limits for the original responses. Those statistical characteristics were converted to confidence limits of the complex apparent resistivity if necessary. The response functions for the observatory PAG had been estimated independently by all the authors mentioned above, and the accuracy of those responses was evaluated by considering all the response estimates in a common period range. Two standard deviations of the mean apparent resistivity were found to correspond to 15% relative error for the complex resistivity modules and $7°$ for its phases (Semenov, 1998). The analogous estimates were obtained for responses at the observatories MEM, KAK and HON where the above parameters were found to vary between 8% and 14% and $6°$ and $11°$ for the modules and phases of the complex resistivity, respectively (Semenov, 1998). Finally, the 68% confidence limits of the collected responses were assumed, but not less than $\pm 10\%$ for the modules and $\pm 10°$ for the phases of the complex apparent resistivity. Unless a lower bound on the confidence limits was introduced, the 1D spherical and Occam inversion could not fit the apparent resistivity values for several observatories.

In total, the response functions for 35 European observatories have been composed for the period range of 6 h up to 11 years. Fig. 2A and B show the combined experimental responses for all the observatories. Along with the experimental responses, results of 1D Occam inversions for each observatory are also shown in this figure.

It is clearly seen from Fig. 2 that difficulties with fitting the data occurred particularly for the responses at shorter periods for the observatories ESK, LOV, NUR, STO, VAL, LNN and RSV, all with geomagnetic latitudes above 55°. This effect, discussed by Olsen (1998) and studied recently by Fujii and Schultz (2002), is most probably due to the influence of the auroral electrojet. Modules and phases of the experimental complex apparent resistivities for the above observatories cannot be both simultaneously in agreement with a 1D inverse model. Judging from the agreement between the experimental phases and 1D model phase data for the northern observatories, the apparent resistivity modules may be disturbed more seriously by the effects of the auroral zone than the phases (see Appendix). Actually, no 1D model can fit the course of the module curves in the period range

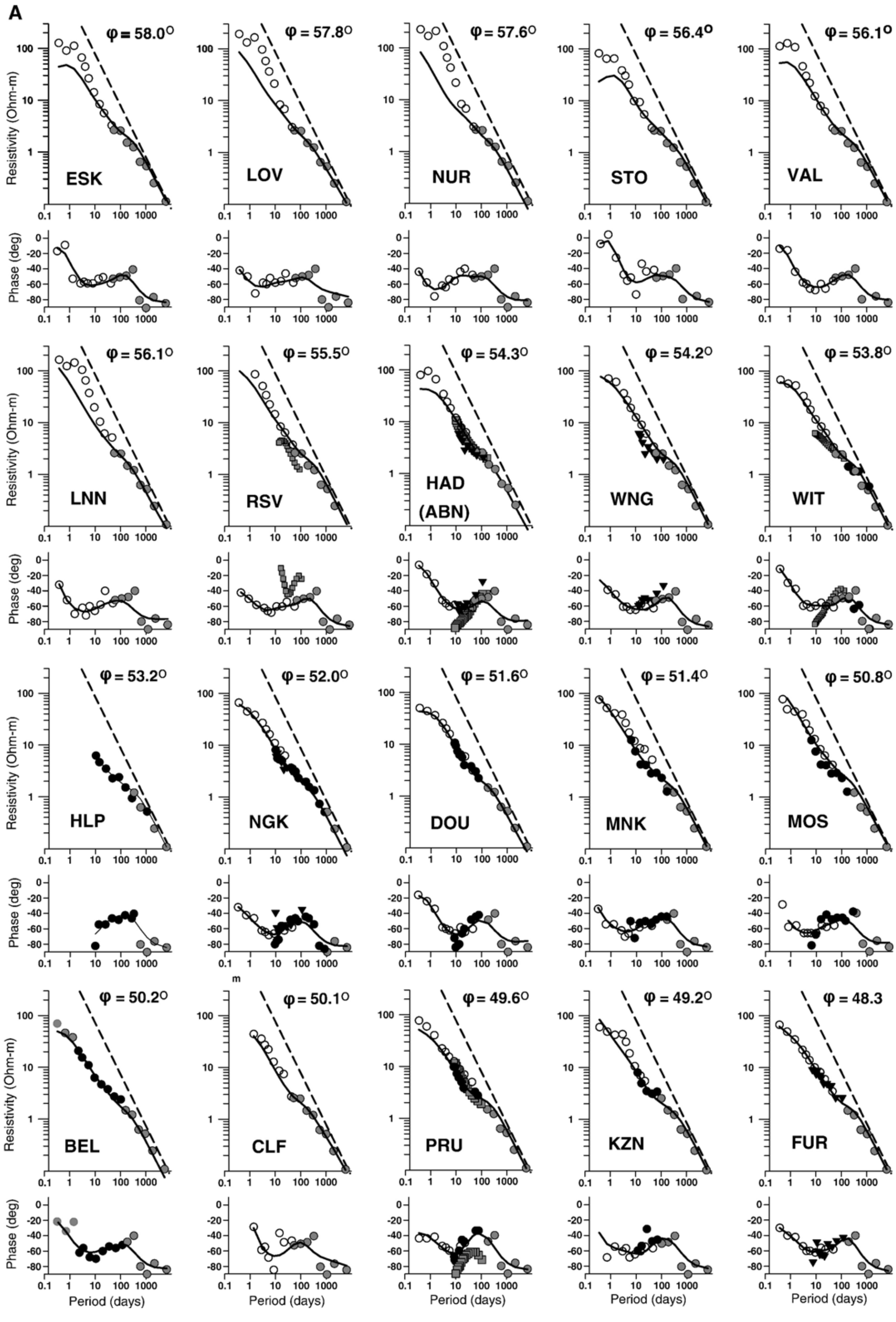

Fig. 2. (A and B) The collected responses for the European observatories are shown with their geomagnetic latitudes φ and marked by symbols: the local responses by Olsen (1998)—empty circles, by Semenov (1998)—dark circles, by Schultz and Larsen (1987)—grey squares, and by Roberts (1984)—dark triangles; the Eurasia continental and 11-year global (last 3 points) responses by Semenov and Jozwiak (1999)—grey circles. The 1D Occam inversion responses and the forward spherical modeling line of the weakly conductive Earth with the metallic core are marked by the solid and dashed lines, respectively. The phases of the dashed lines equal to -90°.

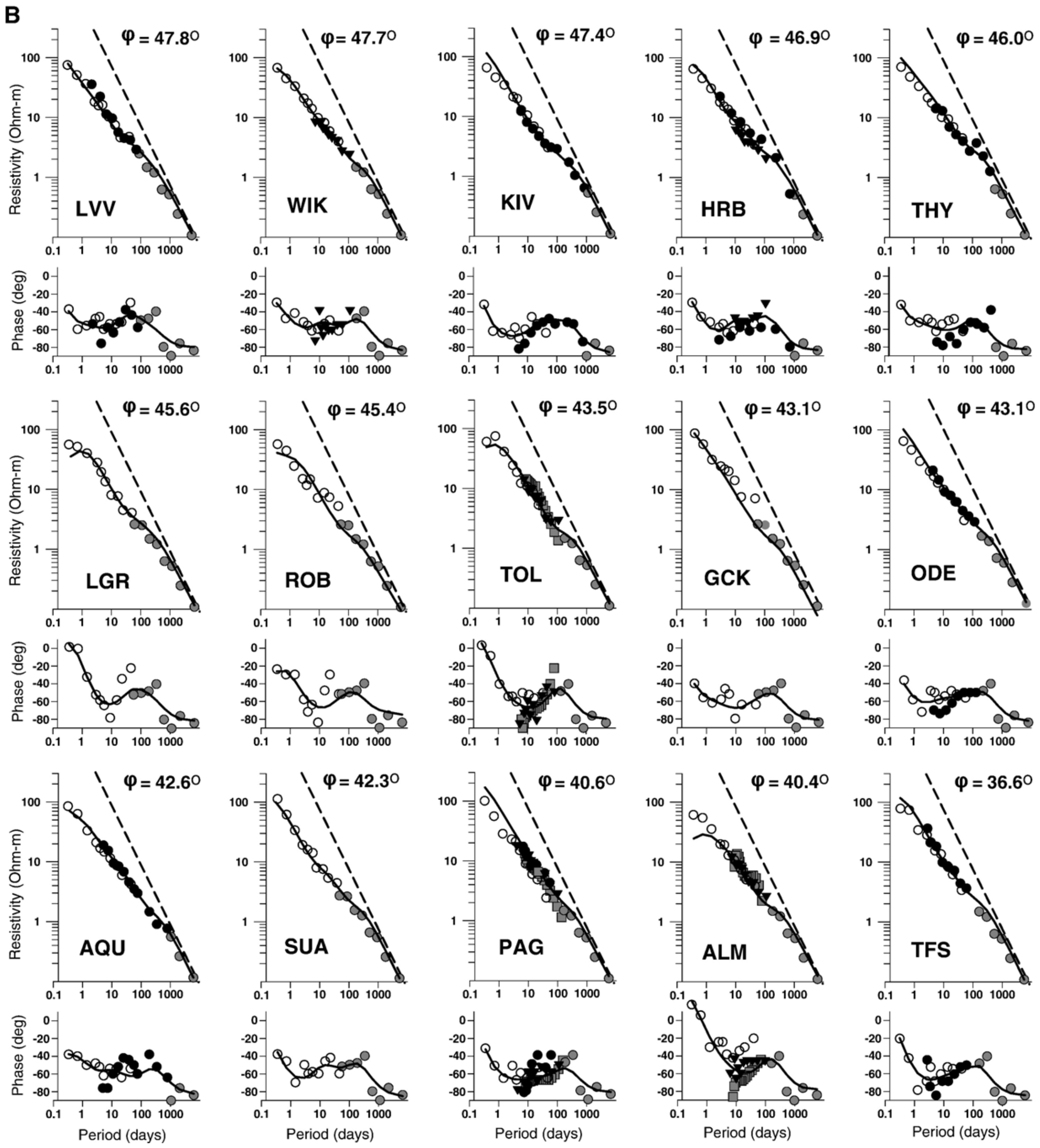

Fig. 2 (*continued*).

close to 1 day for some of the observatories (NUR, LNN, LOV, ESK), while 1D modeling can be done for their phases. Besides, the phases are independent of the geomagnetic latitudes for Dst variations in contrast to their absolute values (Anderssen et al., 1979). The source effect for the high-latitude observatories has been also discussed recently in connection with the 'BEAR' project (Varentsov et al., 2004).

3. Inversions

Two 1D different inverse techniques were applied to the collection of the experimental data. First, the combined responses were inverted individually for each observatory by applying a 1D Occam routine with a roughness penalty employed (Constable et al., 1987). The obtained inverse layered models were further converted into spherical geometry by applying Weidelt's (1972) transformation. As a result, the conductivity versus depth profiles have been obtained for each individual observatory.

Second, a 1D inversion scheme based on a stochastic method (Jozwiak, 2001) applied directly to spherical Earth models (1D spherical inversion) has been used to invert the combined data individually for each observatory. The main feature of this inverse algorithm is that it inverts directly for a conductance section within the model, rather than for the conductivity distribution. The conductance of a layer is the product of the conductivity of that layer with its thickness, and an integrated conductance is introduced as a sum of the conductivity times thickness products over a pile of layers. Often, the terms conductance and integrated conductance are considered synonymous, in the sense of the latter one, for multi-layered media, and we are using them as such in what follows. Berdichevsky and Dmitriev (2002) have proved that 1D inversion of electromagnetic data for the conductance in the Earth is a well-posed problem, i.e. it possesses a stable solution, contrary to the inversion for the conductivity which can vary largely under the same conductance. The confidence limits for the conductance in the spherical inversion show to be several times smaller as compared to those for the conventional conductivity, provided the same data, with the same accuracy, are used (Wieladek and Nowozynski, 1989). Note that for the 1D inversion to be feasible for our data from the northern observatories the confidence limits for the responses at the shortest periods had to be changed to substantially larger values with respect to their original estimates. Eventually, by employing the 1D spherical inversion technique, the conductance distribution down to the lower mantle beneath each observatory has been estimated.

Conversion of the conductivity versus depth profiles recovered by the 1D Occam routine into conductance distributions, on the one hand, and results obtained directly by the 1D spherical inversion, on the other, revealed fairly similar lateral changes of the mantle electrical properties. If the two inversion results were found consistent with each other and they both showed sufficiently regular behavior, only the results from the 1D spherical inversion were considered for the subsequent analysis. An image of the interpolated mantle conductance above the depth of 770 km was constructed beneath the European observatories and presented as an automatically smoothed scheme. The particular depth level of 770 km was chosen as an average depth of the mid-mantle conductive layer as it follows from the model of the resistivity distribution for the Eurasia region (Fig. 3). It is obvious that for the conductive layer elevated to a smaller depth, we will observe higher conductance values in the conductance map. If this layer is located deeper, the values of the conductance become smaller, as shown in Fig. 4, with the southernmost observatories omitted.

4. Modeling of the ocean effect

In order to estimate a possible influence of the ocean and seas on the induction responses, as well as to check the accuracy of the method for the conductance interpretation suggested above, we analyze results of model simulations performed on a quasi-3D model that consists of a radially symmetric conductor covered by a surface spherical shell of variable conductance approximated on a $1^\circ \times 1^\circ$ grid. The shell conductance is taken as realistic as possible, and it considers a joint contribution from the sea water and from sediments (see Kuvshinov et al., 2005, for details of the construction of the conductance map). For the underlying spherical conductor, a layered Earth model consisting of a crust, mantle and a metallic core is adopted with parameters

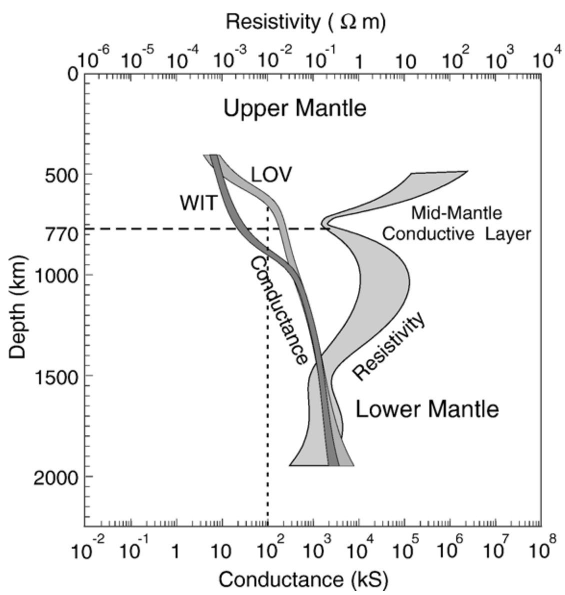

Fig. 3. Regional cross-section of the mantle resistivity beneath the Eurasia continent (Semenov and Jozwiak, 1999) and conductance's beneath the observatories LOV and WIT with their 68% corresponding confidence limits. Two last cross-sections show approximately maximal and minimal conductance beneath European observatories. The position of the 770-km depth and 100 kS conductance are marked by the dashed lines in the mid-mantle.

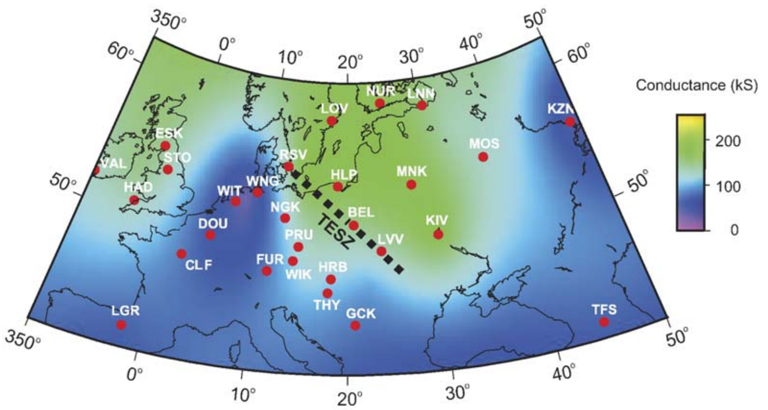

Fig. 4. Smoothed image of the integrated conductance, in kilosiemens, from the surface down to the depth of 770 km in the mantle beneath Europe. The image is presented without the southern littoral observatories indicated in Fig. 1.

shown in Table 1. The marked reference conductance of 100 kS has been chosen as shown in Fig. 3 (and by Semenov et al., 2003) for testing the accuracy of its depth estimates from the spherical modeling results for the individual observatory places. This value is not directly connected with the experimental data presented here.

We are also aware of a possible distorting effect of crustal conductors on the observatory data. These shallow conductors, however, distort the induction responses at shorter periods as compared to those we consider here and, moreover, positions of the crustal conductors are not known precisely in Europe.

The model is excited by a Dst-type source that is described by the first zonal spherical harmonic in the geomagnetic coordinate system. Calculations were carried out for the period range of 0.25 up to 4096 days with the numerical algorithm and code by Kuvshinov et al. (2002, 2005).

The model responses were obtained for the positions of all the European observatories considered in this work. These model responses, with 10% of a relative error added to the modules of the apparent resistivity and 10° of absolute error added to their phases, have been inverted by using the 1D spherical inversion in exactly the same way as it was applied earlier to the experimental responses. The inverse results for the model responses have shown that for the majority of the observatories (28 from 35) the integrated conductance of 100 kS is reached at a mean depth of 864 ± 9 km (two standard deviations). This depth coincides well with the depth range of 866 to 861 km (the particular value depends on the mean conductance over the surface shell) calculated directly from the model radial conductivity distribution (see Table 1).

However, for the observatories ALM, TOL, ROB, AQU, PAG, SUA and ODE, the conductance of 100 kS is reached at depths exceeding the mean depth value significantly, with the maximum deviation being as much as 9% (see Fig. 5). It means that we can expect non-negligible influence of the seas on the interpreted integrated conductance estimates beneath those coastal observatories. A strong effect of the Mediterranean Sea and quite a weak influence of the Atlantic Ocean are perhaps connected with the directions of the electric currents induced by the Dst variations. Actually, the induced currents flow along the geomagnetic parallels, i.e. along the direction from the Mediterranean to the

Table 1
Parameters of the spherically symmetric conductor used in the quasi-3D spherical modeling

Conductivity σ in S/m	Thickness h in kilometers	Depth down bottom, km	Conductance $S=\sigma \cdot h$, kS	Integrated conductance, kS
Shell	1	1	0.03–10	0.03–10
0.001	69	70	0.069	0.1–10.07
0.01	750	820	7.5	7.6–17.6
2.0	46.2–41.2	866 – 861	92.4 – 82.4	100
2.0	130	950	260	267.6–277.6
1.0	1050	2000	1050	1318–1328
1000	900	2900	900,000	901,318–901,328
3,000,000	3470	6370	–	–

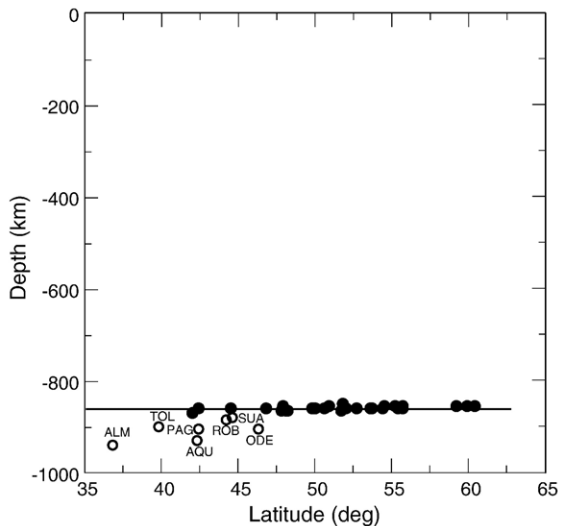

Fig. 5. Depths up to 100 kS conductance, estimated from the described forward modeling by the same way as for experimental data, as a function of the latitudes of the observatory positions. The design depth of the 100 kS conductance is shown by the solid line. The estimated depths differ from that modeling one more than 3% are shown by the empty circles with observatory codes.

Black Sea. They are, however, approximately orthogonal to the British Islands.

5. Conclusions and discussion

A new approach is presented for estimating the geoelectrical structure of the Earth's mantle beneath Europe based on the analysis of collected, properly selected and recombined magnetovariation sounding results obtained by several authors during the recent decades. The distinctive feature of this approach consists in our applying consequently the concept of the conductance instead of the traditional parameters of conductivity or resistivity. This allows us to increase the reliability of the inversion results substantially at the expense of loss of a layered structure of the overlying cross-section, which is not of vital importance. Actually, the inversions of responses generated by a synthetic model which consists of a surface spherical shell and a radially symmetric conductor underneath have shown that the error in determining the depth to a fixed conductance level of 100 kS is less than ± 10 km, if observatories evidently affected by the ocean effect are removed from the analysis. However, the experimental data inverted in the same way detect the depths to the 100 kS conductance level in the mid-mantle to vary with amplitude of about 200 km (Semenov et al., 2003), which exceeds the uncertainty in determining this depth due to inhomogeneities in the surface layer by a factor of twenty.

The integrated conductance of the Earth's mantle has revealed rather regular lateral changes. This fact allows us to present a smoothed image of the integrated conductance beneath Europe at a fixed depth of 770 km in the mid-mantle (Fig. 4). The image shows that the Trans European Suture Zone (TESZ) coincides with a gradient zone of the conductance between Western Europe and the East European plate, with variations of depths down to the 100 kS conductance level from about 600 down to 900 km beneath Europe (Semenov et al., 2003). We have to remark here that a depth increase down to the mid-mantle layer in approximately a southward direction in Europe has been already detected by Roberts (1986). However, our results do not distinguish between models with the variations in the integrated conductance caused either by a varying depth or by a varying conductivity of the mantle conductive layer — both alternatives are possible. The variations of the integrated conductance in our image reach as much as 150 kS. This value is connected mainly with the mid-mantle conductive layer because the conductance of the upper mantle is about 10–20 kS beneath Europe in accordance with estimates for different regions (Semenov, 1998). These values are roughly comparable with ocean conductance. Consequently, variations of the integrated conductance of the lithosphere and upper mantle may reach as much as 10–15 kS, which is by one order of magnitude less than the observed value of about 150 kS.

Additionally, the integrated conductance beneath the Baltic Shield is characterized by higher values as compared with the rest of Europe which shows a relatively smaller conductance down to the depth of 770 km. This fact is in agreement with the conclusion made by Pecova et al. (1980) for the depth of 1000 km, and it seems to be in contradiction with the result obtained by Olsen (1998) who used a part of those data, '... probably indicates lower conductivity in the north'. However, Olsen's conclusion was obtained for the periods between 6 and 300 h while the integrated conductance considered here was mainly connected with the mid-mantle and resulted from the analysis of longer periods, between 150 and 3000 h, as mentioned above. Thus the above conclusions relate to different depths. A phenomenon of a reversal of anomalous physical properties is known in the Earth from the satellite gravimetric data: 'the sign of anomalous density in radial direction is changed to the opposite at the 700 km boundary' (Martinec and Pec, 1989).

Thus we have studied the integrated conductance of the Earth beneath Europe down to the depth of the mid-mantle conductive layer and have shown that its reliably estimated spatial variations are mainly related to variations of the conductivity or to the depth of this essential mid-mantle feature of the Earth. A question arises in this respect: Can the method suggested for deep induction studies be used to analyze the data for greater depths, on the one hand, and especially also for shallower depths, on the other? Deeper studies can be carried out by using the available data, but from some depth down, the conductances beneath the individual observatories are approximately approaching the same values that correspond to the common regional and global data at the longest periods.

For the investigations into shallower depths, with less conductive structures, we have to exclude the high-conductive mid-mantle layer, and also supplement additional data for shorter periods to the analysis. Recently, magnetotelluric (MT) data were obtained in the frame of the project CEMES for the region of Central Europe starting from the period of about 300 s (Semenov et al., 2003), and now the Euro-Array project has been suggested with the experimental part intending carrying out regular MT measurements together with the seismic study (www.euroarray.org). The advantage of the suggested combination of the seismic and deep MT observations has been considered in detail by Jones (1999). Because the MT method is most sensitive to conductive layers, the main objective is to detect a conductive asthenosphere beneath Europe with a low velocity seismic characteristic (Thybo and Perchuc, 1997; Nielsen et al., 2002). The analysis considered in this paper can be helpful in this case for detecting depths down to a fixed conductance level in the upper mantle and for inferring the lateral variations of this depth which can be interpreted as a change in the conductance or as a variable depth of the asthenosphere (Gung et al., 2003).

Acknowledgments

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Appendix A

The magnetovariation sounding, using the ratio H_r/H_θ of Dst field, requires that we know the geomagnetic co-latitude of the observatory position to estimate response functions (Roberts, 1984). However, the observatory geomagnetic position undergoes changes during a day. One of possible reasons of this effect is the lack of coincidence between the Earth's rotational axis and the geomagnetic one if the ring current source has a stable position in the Earth–Sun as system. The angle between these axes can reach 10–15° shown in Fig. A. Actually, locates being at a geomagnetic longitude of the geomagnetic pole the observatory situates at a minimal distance from the pole, but through 12 h this distance becomes maximal (see Fig. A). The difference between those positions is formed by double distances between the Earth's geomagnetic and spin poles. For example, the observatory Belsk (BEL) has geomagnetic latitude $\approx 63°$ near the geomagnetic pole and $\approx 41°$ far away from it for epoch of 2002 year. Thus, the geomagnetic latitude depends on geomagnetic longitude of the observatory or on the local time as it has been established by the CHAMP satellite (Balasis et al., 2004). These daily oscillations will form a part of daily variations induced by the Sq source, i.e. disturb them in comparison with the pure Sq source. Additionally, the direction to the geomagnetic pole (the geomagnetic declination) will vary during a day depending on the local time.

It is a worth noting that only absolute values of responses estimated through H_r/H_θ are depending on the geomagnetic co-latitude of the observatory while

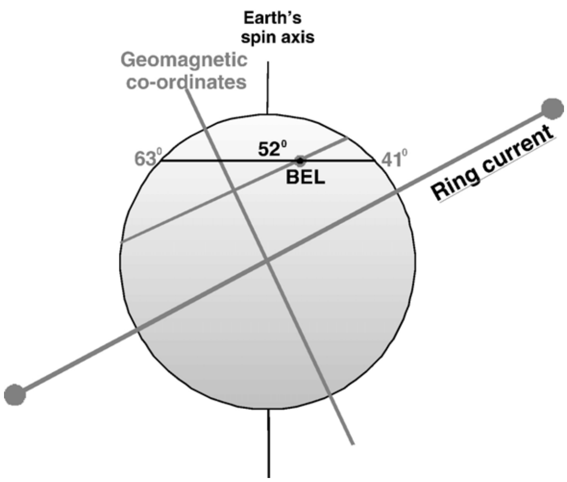

Fig. A. Scheme of the possible daily changing of the observatory geomagnetic coordinates. These oscillations for BEL can vary from 41° up to 63° around its geographic latitude of 52°.

their phases are independent of it (Anderssen et al., 1979). Thus, the sounding can be more reliable when only phase values of impedance are used.

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