ELECTROMAGNETIC STUDIES OF GLOBAL GEODYNAMIC PROCESSES

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Abstract. The deep electromagnetic sounding (DES) technique is one of the few geophysical methods, along with seismology, gravity, heat flow, which may be use to probe the structure of the Earth's mantle directly. The interpretation of the DESs may provide electrical conductivity profiles down to the upper part of the lower mantle. The electrical conductivity is extremely sensitive to most of the thermodynamic processes we believe are acting in the Earth's mantle (temperature increases, partial melting, phase transition and to a lesser extent pressure). Therefore, in principle, results from DES along with laboratory measurements could be used to constrain models of these processes.

The DES technique is reviewed in the light of recent results obtained in a variety of domains: data acquisition and analysis, global induction modeling and data inversion and interpretation.

The mechanisms and the importance of surface distortions of the DES data are reviewed and techniques to model them are discussed. The recent results in terms of the conductivity distribution in the mantle from local and global DES are presented and a tentative synthesis is proposed.

The geodynamic interpretations of the deep conductivity structures are reviewed. The existence of mantle lateral heterogeneities in conductivity at all scales and depths for which electromagnetic data are available is now well documented. A comparison with global results from seismology is presented.

1. Introduction

The knowledge of the structure of the Earth's mantle provides a major clue to the understanding of the mechanisms of the evolution of our planet. The evolution seems to be the result of thermal processes acting between the core and the surface. Our present understanding of these processes is the result of studies which combine the outcome from petrology, geochemistry, laboratory experiments on the physics and chemistry of the Earth materials and from the numerical simulation of the thermal and mechanical evolution of the planet. One of the major limitations is the lack of direct observations of the actual structure of the Earth's mantle. Very few techniques provide such data. Seismology and electromagnetic induction are the most prominent together with gravity and astronomy.

Since the pioneering work in the electromagnetic induction study of the Earth (e.g. Schuster 1889, Chapman 1919, Chapman and Price 1930, Banks 1969), the field has not provided much more quantitative additions to the present knowledge of the global structure of the Earth's mantle. This is due mainly to the difficulty of gathering the data to probe the mantle structure to great depth and until recently the lack of high speed/large capacity computers. These data consist of long period variations (from a few hours to several years) of the geomagnetic field

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recorded mainly at magnetic observatories. In the period range 6 hours – 100 days, these fluctuations are principally caused by the magnetic tides (the daily variation and its harmonics) and by the magnetospheric ring current. Geoelectric data are very rarely used for deep investigations because of their extreme sensitivity to local distortion.

Until recently, deep electromagnetic investigations were restricted to the determination of the electrical conductivity in a spherically symmetric Earth, averaging out possible lateral variations. In the late seventies and the early eighties however, it became clear that the geomagnetic data could be used to infer the conductivity structure to a higher degree than the degree zero global mean (e.g. Berdichevsky *et al.* 1976, Roberts 1983).

Nevertheless, as a result of the limited amount of data available (both in the spatial and the frequency domain) as well as of the fundamental limitation of the diffusion equation which governs the electromagnetic field, the knowledge of the conductivity structure of the Earth's mantle is still in its infancy.

This does not mean that the other techniques have been so successful in helping to understand the inner structure of the Earth. The results obtained from seismologic tomography for instance, though spectacular, have brought more questions than solutions. Among the reasons for these difficulties, are the still limited resolution of the images obtained and the major problem of relating the observations to the thermodynamics of the mantle.

It becomes mandatory to get independent sets of data sensitive to geophysical parameters other than the density, the elastic or the anelastic properties of the mantle. Without this new information, it is possible that our steady increase in knowledge of the structure of the Earth's mantle could stall. It is quite clear to people interested in the Earth's mantle that knowledge of the 3-D conductivity structure of the earth, even limited to depths less than or equal to 1,000 to 1,500 km, the present maximum depth of investigation of deep electromagnetic soundings (DES), might be the next breakthrough.

In this paper, I intend to take stock of our progress in constraining the conductivity distribution in the Earth. I restrict the discussion to the mid-mantle (about 400 to 1500 km). This is the critical depth range to be investigated for understanding the global processes acting in the whole mantle because it scans the limit between the upper and the lower mantle, whose nature is still not understood, and is of fundamental importance for a number of important issues, namely the scale of mantle convection, the origin of hot spots and the recycling of subducted slabs.

Recent Results from Deep Electromagnetic Soundings

The most important result obtained by the deep electromagnetic studies during the last decade is the demonstration that the geomagnetic data are sensitive to the presence of internal inhomogeneities. In order to study thoroughly the possible existence of those deep heterogeneities that are able to be resolved by electromagnetic data, authors have tried to interpret locally or semi-locally the long period data sets available at magnetic observatories. Several studies have been carried out with this approach, using different or identical data sets, sometimes including long period electric data, with the help of sophisticated data analysis techniques (e.g. Roberts 1984, Counil *et al.* 1987, Schultz and Larsen 1987, Schultz *et al.* 1987, Campbell and Schiffmacher 1988)). Figure 1 is an example of the scatter of the data sets, here the transfer functions deduced from Dst geomagnetic data from Schultz and Larsen (1987) and Semenov (1989).

It has not been obvious and it is not definitively clear for all data sets, that the differences observed in the transfer functions from several observatories could have a mantle origin. One possible explanation for instance, is that these differences have their origin in a poor description of the source field, whose geometry must be accounted for in deep electromagnetic investigations. However, early studies (e.g. Banks 1969) had partially ruled out this hypothesis. A very critical selection of the data sets based on a uniform field hypothesis (P_0^1) for periods of more than a few days was made, and the results demonstrated that the Dst data sets which passed the test for being relatively free of external contamination, showed significant scatter. The Sq data however may not be as source free as the (P_0^1) data because the geometry of the source field is far more complex than the Dst's (e.g. Roberts 1986b).

Some of the observed differences may also be due to techniques used for the determination of the transfer functions. Nevertheless as we shall see, the comparison between models obtained by different authors at a same site is reasonably satisfactory.

The alternate hypothesis proposed to explain the variability of the transfer functions is that it may be due to the effect of local surficial heterogeneities, namely the coast effect (e.g. Fainberg 1980, Roberts 1986a).

Tri-Dimensional Distortion of Long Period Electromagnetic Data

The coast effect

The coast effect has long been considered to be a limitation on the investigations of the deep Earth by electromagnetic methods.

The progressive availability of algorithms to simulate the effect of realistic oceans, the coast lines, and the bathymetry by a heterogeneous thin shell, has motivated studies of this effect at a global scale (e.g. Rikitake 1961, Bullard and Parker 1970, Hobbs and Dawes 1979, Fainberg and Singer 1980, Beamish *et al.* 1983).

During the last years, several new or improved algorithms have been proposed

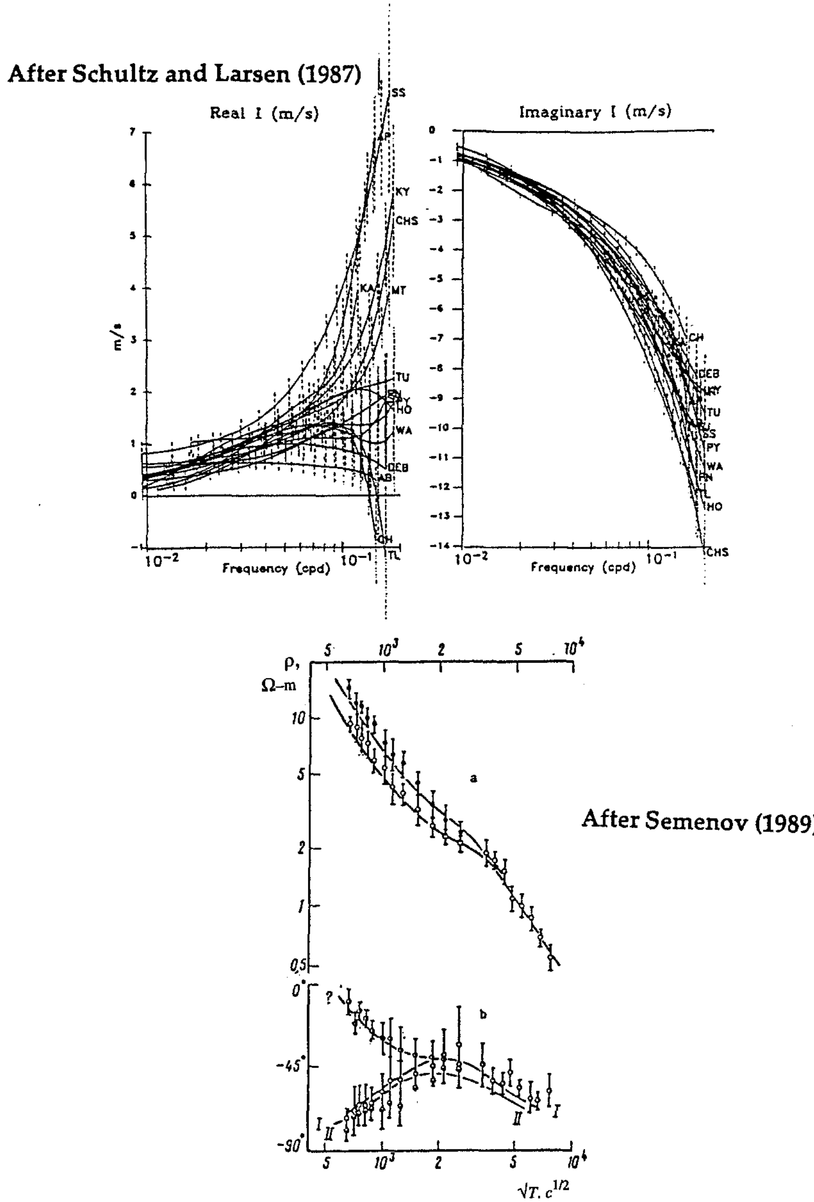

Fig. 1. Examples of geomagnetic transfer functions. Upper part: real and imaginary parts of the equivalent MT impedance. Lower part: Equivalent MT impedance represented in terms of apparent resistivity and phase. It is quite clear that the scatter is particularly strong on the phase.

(Winch 1989, Takeda 1991, Fainberg *et al.* 1991a, Zhang and Schultz 1992) and detailed discussions of the ocean and the coast effect on the geomagnetic data have been undertaken (Kuvshinov *et al.* 1990, Takeda 1989, 1991, Fainberg *et al.* 1991b).

The algorithm used by Fainberg *et al.* (1991b) and Kuvshinov *et al.* (1990) is presented by Fainberg *et al.* (1991a). This algorithm is bi-modal and hence it makes possible the use of realistic radial distributions of conductivity conversely to most previous algorithms in which the toroidal magnetic mode was cancelled because of the insulated layer beneath the thin heterogeneous shell. Fainberg *et al.* (1991b) used this capability to address the problem of the determination of the resistivity-thickness product of the medium underlaying the outer surface shell and have proposed a global value of $3.10^9\ \Omega\text{-m}^2$ in good agreement with Mackie *et al.*'s estimation, $1.10^9\ \Omega\text{-m}^2$ for the Pacific ocean (1988).

Kuvshinov *et al.* (1990) have investigated the coast effect for a P_0^1 geometry at periods from 1 to 15 days. Their results (only displayed for 1 day and 15 days period) show that the coast effect is quite small (9%) at period of 15 days. They also showed that lateral variations of the resistivity-thickness product may have a significant effect on the apparent resistivity, calculated from the transfer functions between the magnetic components of the electromagnetic field.

They ran different models with various resistivity-thickness products in order to test the sensitivity of the geomagnetic data to surficial and sub-surface heterogeneities. They concluded that their models do not explain the long period geomagnetic data and that additional effects should be taken into account.

Winch (1989) extended Rikitake's theory to include a realistic land-ocean distribution. The model was used by Takeda (1991) by studying the effect of the electric currents induced in the oceans by a Sq source field, to determine the mantle conductivity. Using the representation of the source field determined by Takeda (1985, 1989), Takeda (1991) determined the bias introduced by the presence of a heterogeneous shell (30% for the determination of the depth and conductivity) and so confirmed earlier results that showed a substantial effect due to the presence of the ocean (see for example Parkinson and Hutton 1989).

The most recent published algorithm that is able to solve for the induction in a spherical heterogeneous Earth is proposed by Zhang and Schultz (1992). This algorithm may address the full 3-D induction problem. The solution is found by a perturbation method. The model includes both poloidal and toroidal modes. This algorithm does not allow for strong lateral contrasts of conductivity. Its application is in the study of the influence of deep contrasts of conductivity on the geomagnetic data and allows one to address the inverse problem in a very promising way.

In order to complete this discussion, I have reproduced some of the aforementioned results as well as additional calculations in order to better describe the very long period coast effect in both the spatial and frequency domains.

The calculations were carried out with a new algorithm developed by Wahr *et al.* (1992) and Tarits and Wahr (1992), which solves the 3-D induction problem in a spherical heterogeneous Earth. The model consists of a stack of layers which may be either homogeneous or lateraly inhomogeneous. The Maxwell equations are expanded in generalized spherical harmonics (Phinney and Burridge 1959) and the radial derivative of the expanded electromagnetic field components are numerically integrated using a 4th-order Runge-Kutta integration technique. We use an effective direct and inverse generalized Legendre transform algorithm (Lognonne and Romanowicz 1990) to match the boundary conditions in the spatial domain and to return to the spectral domain to propagate the field through the next layer. The algorithm converges with large contrasts of conductivity at any period of interest. It has been checked satisfactorily with various published results.

Figure 2 presents a case similar to the one considered by Kuvshinov *et al.* (1990). The model consists of a realistic distribution of continents and bathymetry (the database used is ETOPO5 from NOAA) overlying a 1-D medium similar to Kuvshinov *et al.*'s. The source has a P_0^1 geometry. The results are presented for a 10 days period (Figure 2a).

I have represented the square of the radial magnetic component B_r over the north magnetic component B_θ for the model including the coast effect, divided by the same quantity for a uniform ocean. The ratio $(B_r/B_\theta)^2$ is proportional to the classical magnetotelluric apparent resistivity for a symmetric spherical case and may be interpreted as an equivalent apparent resistivity (EAR) for heterogeneous situations. This means that the values presented in Figure 2 may be interpreted in term of percentage of EAR.

Figure 2b presents the results for the same model for a 5 days period. The effect is more important at 5 days than at 10 days (25% against 9%) and increases sharply at 2 days (not shown here). I have also run the same model at a 1 day period with the Sq source field from Malin and Gupta (1977). The effect is dramatic, as may be seen from Figure 2c.

I have not presented the results for the phase-lags of the distorted electromagnetic field because the conclusions are identical to the conclusions drawn from the study of the EAR. The coast effect may be responsible for a significant distortion of the magnetic field at periods less than about 2–3 days but it becomes vanishingly small at mid-latitude as the period increases. At 10 days period (Figure 2a) or 15 days period (Kuvshinov *et al.* 1990), the effect is very small everywhere.

Mantle heterogeneities

Roberts (1986a), Semenov (1989), Schultz and Larsen (1990) have shown that the analysis of their data sets in term of conductivity profiles gives an indication of the presence of shallow (less than 400 km) heterogeneities. They also suggested that for some places such as Japan, the coast effect could not be effective enough to explain these features. This conclusion seems to be confirmed in our preceeding discussion.

However, the influence on the geomagnetic data of upper mantle heterogeneities is still an issue. So far, only Kuvshinov *et al.* (1990) have published models including both a realistic upper heterogeneous shell and an inhomogeneous resistive underlying medium. Nevertheless, they have shown that the introduction

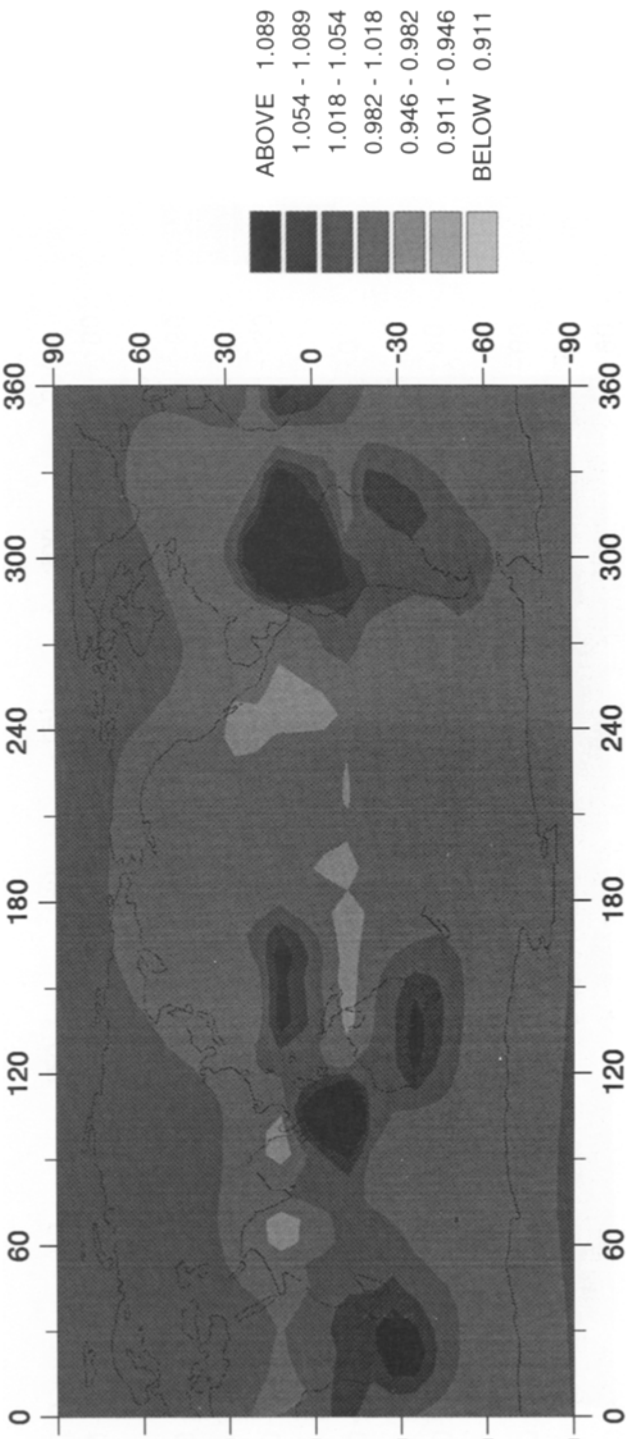
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Fig. 2a.

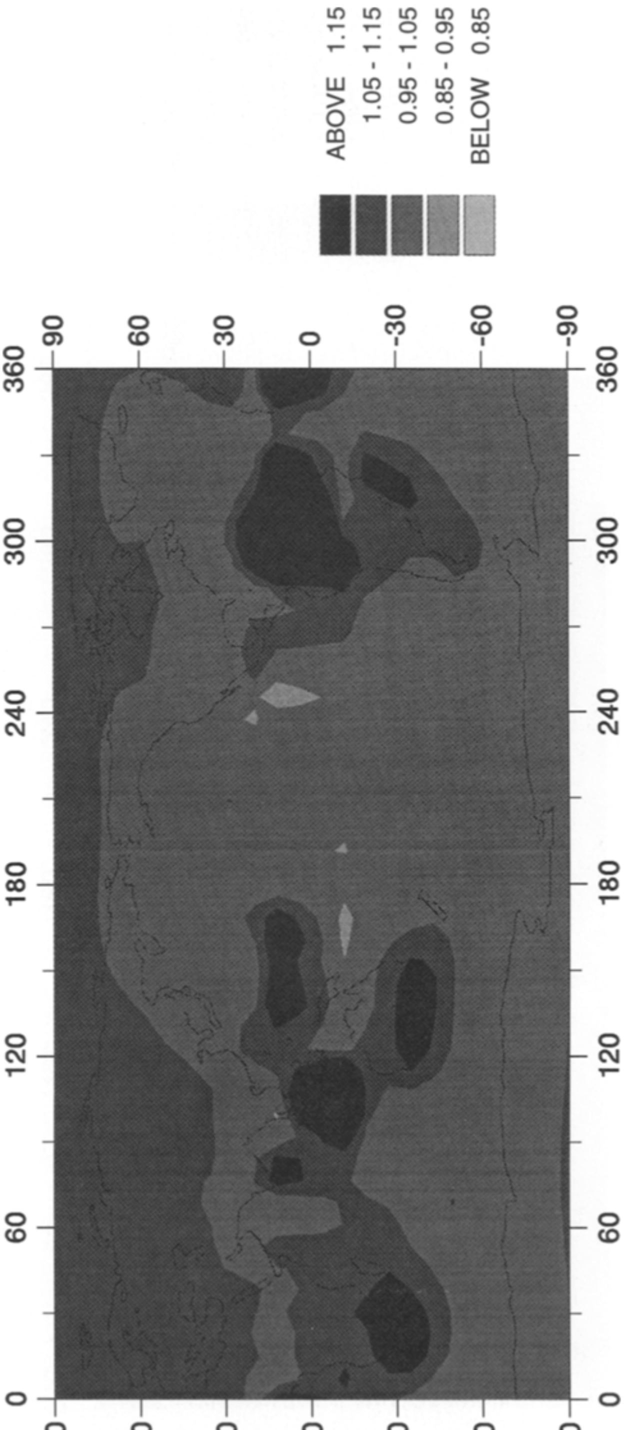
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Fig. 2a.

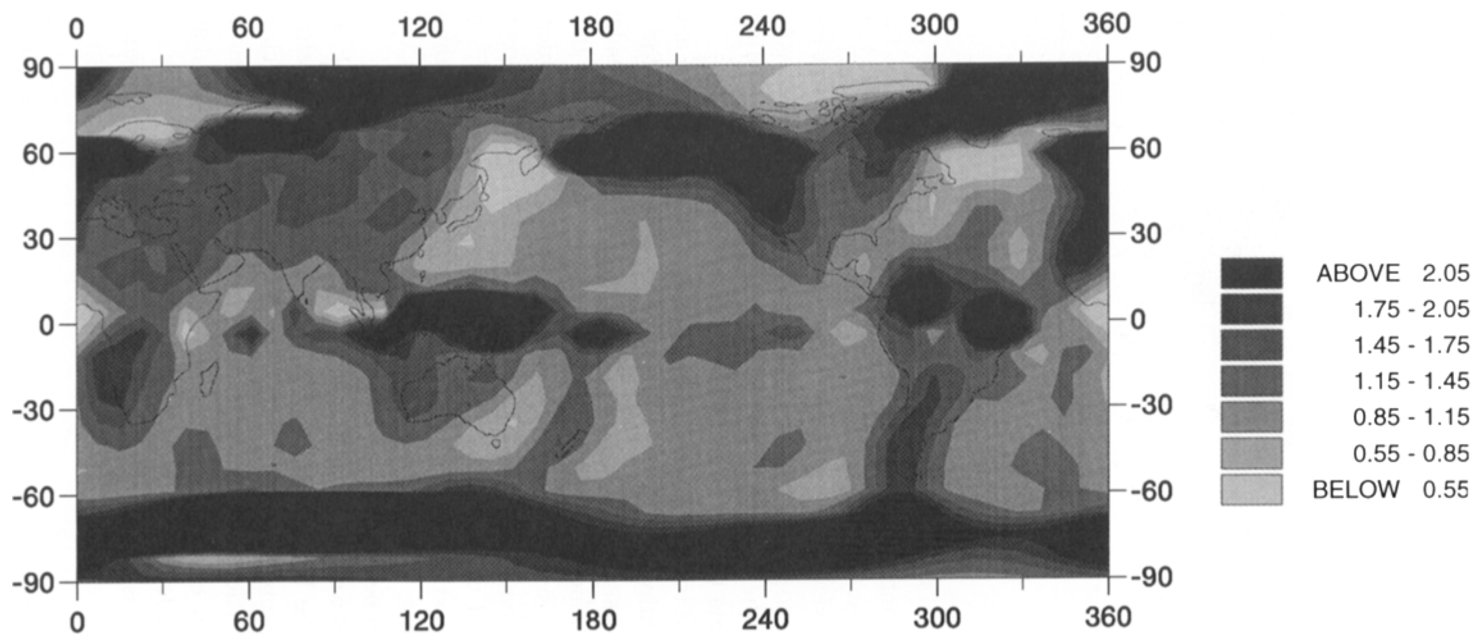

Fig. 2b.

Fig. 2. Coast effect on the equivalent apparent resistivity (EAR) with respect to the apparent resistivity for a uniform ocean. (a) The source is a P_0^1, the period is 10 days. (b) Same source field as (a) but the period is 5 days. (c) The source is the Sq field (period 1 day) from Malin and Gupta (1977).

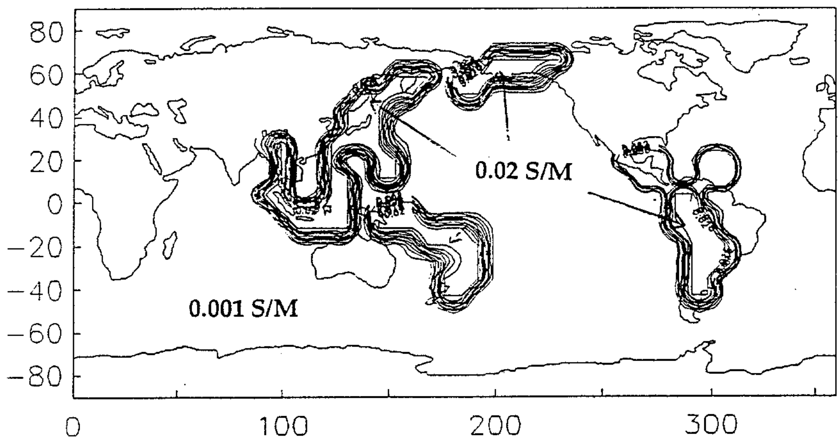

Fig. 3a.

of resistive contrasts below the outer shell could not explain the observations, in particular those at the Japanese observatories.

I have considered the effect of large conductive bodies in the upper mantle. I have chosen to model the subduction zones for which structural information at the scale of the upper mantle, is available from a combination of seismological and geoid data (Hager and Clayton 1989).

In addition to the upper shell representing the continents and the oceans with a realistic bathymetry, I added the global distribution of the subduction zones from Hagger and Clayton (1989). The dipping slabs are considered vertical down to 300 km (which is a very approximate way to represent the subduction) and expanded into surface spherical harmonics up to degree and order 20 (Figure 3a). At this scale, I considered that the subduction zones could be represented by conductive patches. I used $0.02\,\Omega^{-1}\,\mathrm{m}^{-1}$ for the slabs and $0.001\,\Omega^{-1}\,\mathrm{m}^{-1}$ for the rest of the upper mantle. I checked that a lower conductivity for the upper mantle did not change the resulting geomagnetic field. Also, a higher conductivity for the subduction zones would not be very realistic at this scale.

The model was run at periods of 2, 5 and 10 days with a P_0^1 source field and at 1 day with the Malin and Gupta's Sq external harmonic coefficients (1977). The deeper shells are homogeneous and their characteristics are close to Kuvshinov *et al.*'s model. The heterogeneous structures in the induction model were described with harmonic expansion to degree and order 15 (I checked that the use of higher degree and order did not change the results significantly). The same representation of the results as before is used. The EAR for the model including both the coast effect and the slabs are now divided by the EAR for the model with the coast effect only.

I have reported the results for 5 and 2 days only, on Figure 3b and 3c respec-

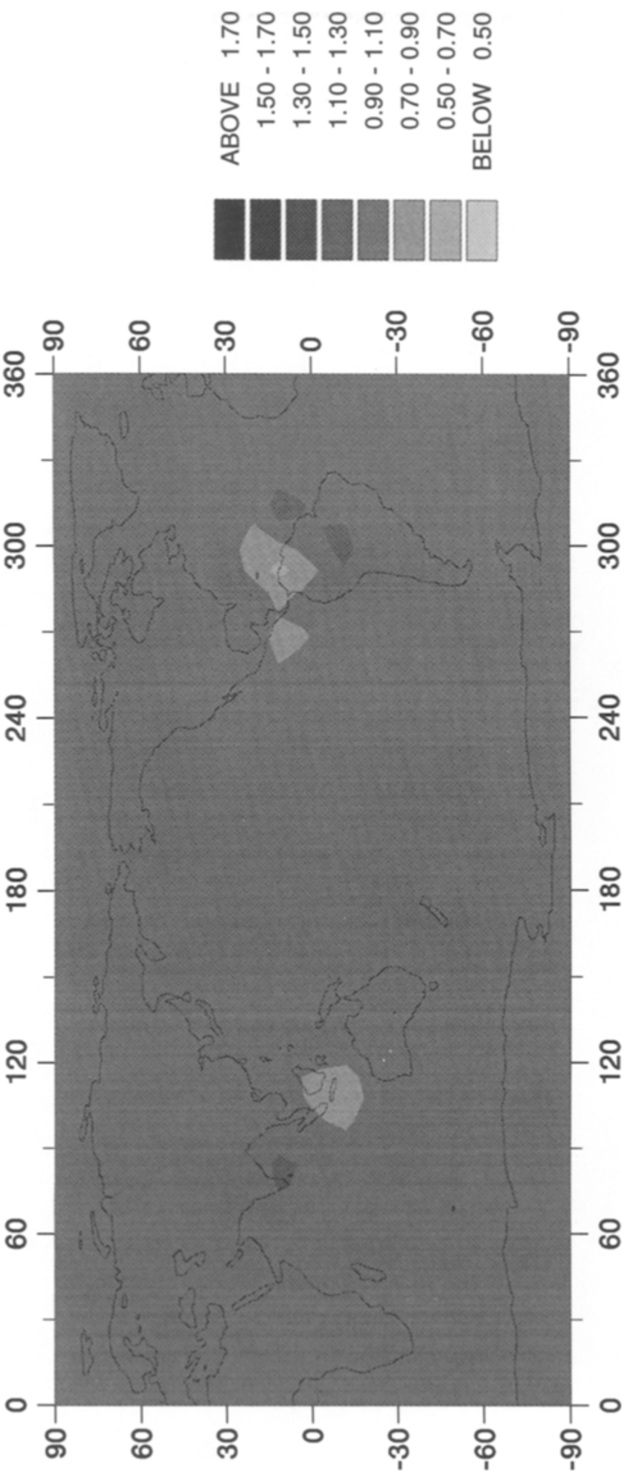
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Fig. 3b.

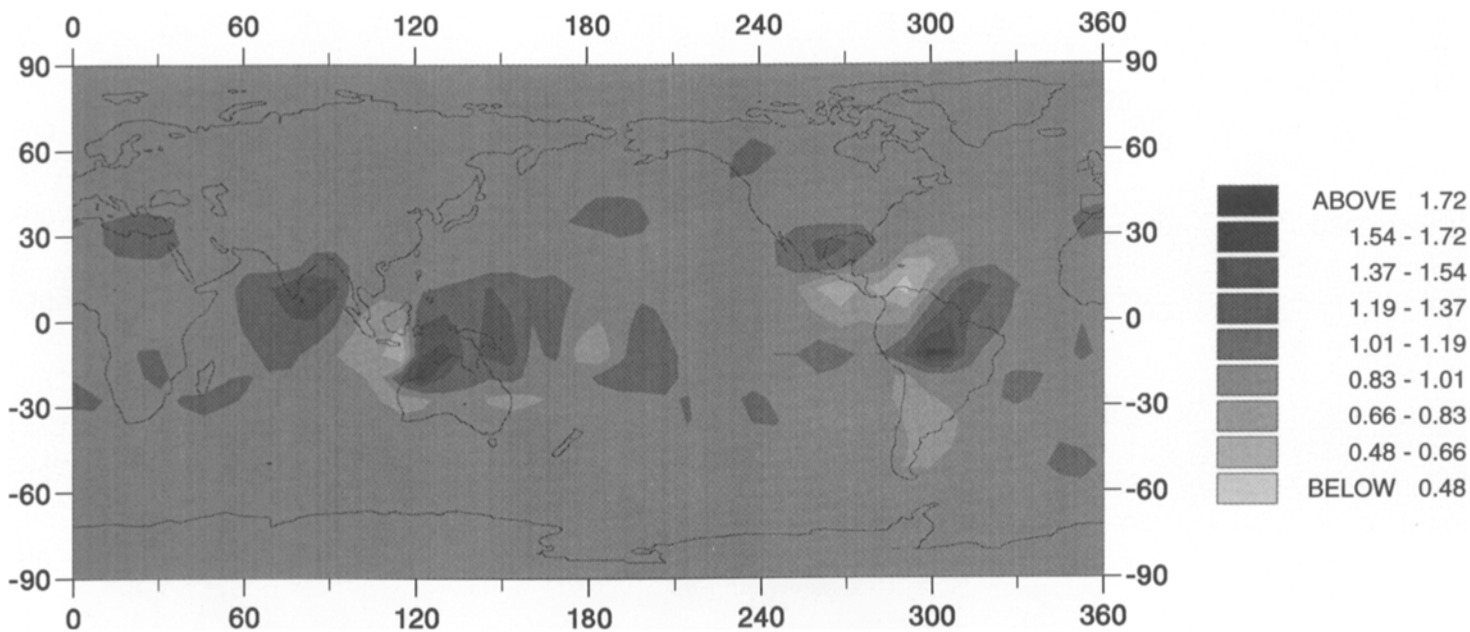


Fig. 3c.

Fig. 3. (a) Regional model of the subduction zones after Hager and Clayton (1989), converted into a conductivity model (see text for details). (b) EAR at 5 and (c) EAR at 2 days periods for a P_0^1 source field.

tively. At 2 days period, significant EARs are visible in South and Central America and in South Asia and Indonesia. At 5 days period, the effect has completely vanished almost everywhere.

The weakness of the resulting perturbation at periods more than 2 days could be related to the limited harmonic expansion used. Nevertheless, with the same expansion, we found the correct order of magnitude for the coast effect (see Figure 2).

As an illustration of the consequence of this result, consider the case of the Japan Anomaly. This anomaly has been described by several authors (e.g. Rikitake 1966, Honkura 1978) over a period range roughly from a few hours to a few days. The coast effect does not seem to be sufficient to explain the anomalous electromagnetic field over this period range (e.g. Menvielle *et al.* 1982, Schultz and Larsen 1990) and it has been suggested that deeper anomalous structures, likely related to the active subduction, could account for the observations. According to our calculations, this hypothesis only stands to periods less than 4–5 days. At longer periods, crustal or lithospheric heterogeneities do not seem to be sufficient to influence the data. Therefore the observed anomaly might be linked to deeper processes.

The results of the modeling of the coast effect indicate that only the geomagnetic data at the shortest periods (a few days) are significantly distorted. According to the models, the effect is fairly large for Sq at 1 day and is therefore expected to be larger at shorter periods. The addition of resistive contrasts (Kuvshinov *et al.* 1990) or of large conductive bodies in the first 300 km below the surface does not explain the features observed at the smallest periods in the data sets. These tests are very preliminary and more thorough studies are necessary, which should include more realistic crustal and lithospheric structures.

Interpretation of the Deep Electromagnetic Soundings

Conductivity profiles

I shall only present the results obtained since R. Roberts wrote his review paper about global induction studies in 1986 (1986b). The data sets used for these interpretations are geomagnetic transfer functions at different periods (the Sq and its harmonics and the Dst from about 3 to 100 days) from magnetic observatories all over the world.

The most important issue discussed by Roberts (1986b) was the hypothesis that the long period geomagnetic data could provide information about lateral heterogeneities in the mantle. Since then, most of the published works have successfully strengthened this hypothesis. A lot of attention has been given to the local or semi-local processing of the induction data in order to get transfer functions representative of the conductivity distribution locally rather than globally. Hence,

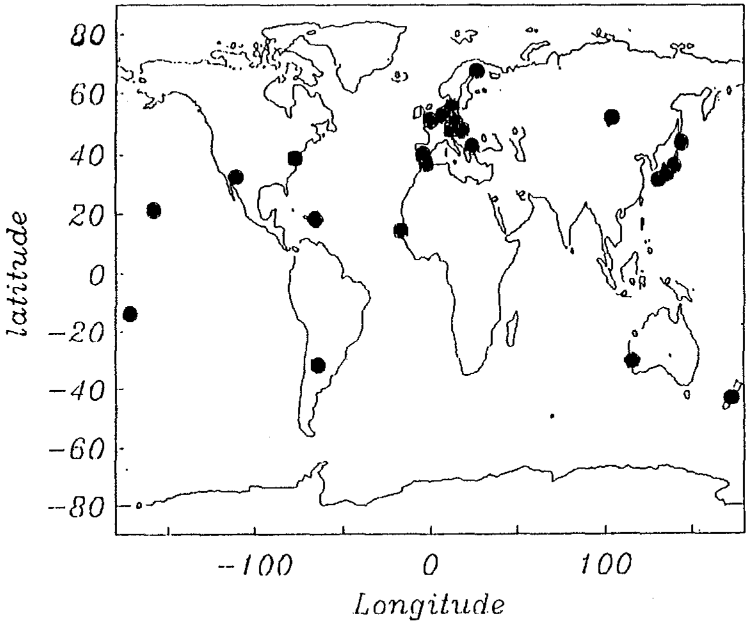


Fig. 4. Distribution of the observatories whose data passed the tests of uniform source field and of one-dimensionality.

the resulting conductivity profiles derived from those data are the horizontal average over distances from one thousand km or so (e.g. Schultz and Larsen 1990) to regional scales (continents, oceanic basins) (Campbell and Schiffmacher 1988).

All the authors were particularly careful to process the data so the residual source field bias should be minimal. They were therefore quite confident that the scatter observed in their sets of transfer function was, to a large extent, of internal origin (e.g. Figure 1).

Despite the large amount of work carried out to get data at as many sites as possible, the restrictions imposed by the authors to keep only geomagnetic data with pure source field and good 1-D characteristics, led to a limited set of data (Figure 4).

In the course of his global analysis of geomagnetic induction data, Semenov (1989) has shown that the Dst data at some sites could not be explained by a single conductivity model, though such a model could explain the transfer functions at all the other sites (Figure 5). This single model is characterized by a conductive slab at 600–700 km depth, in agreement with the zero degree models obtained by taking the geomagnetic data globally (e.g. Hobbs 1987). To explain these anomalies at the Asia-Pacific limit, Semenov (1989) assumed the presence of significant heterogeneities in the upper mantle as well as in the conductive slab.

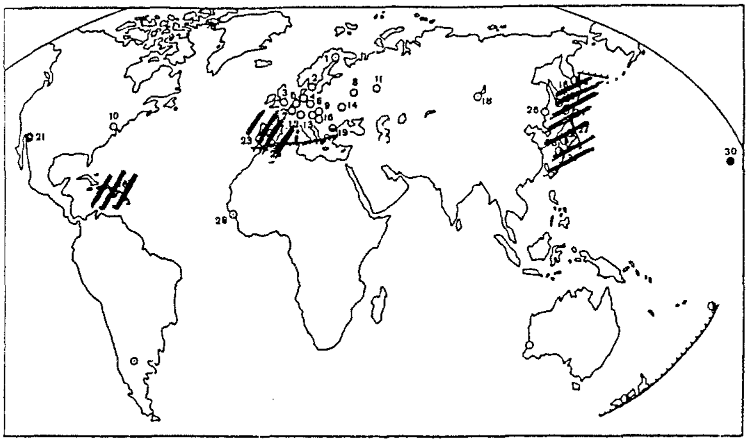


Fig. 5. After Semenov (1989, copyright by the American Geophysical Union); the hatched areas correspond to anomalous regions characterized by high apparent resistivities and phases.

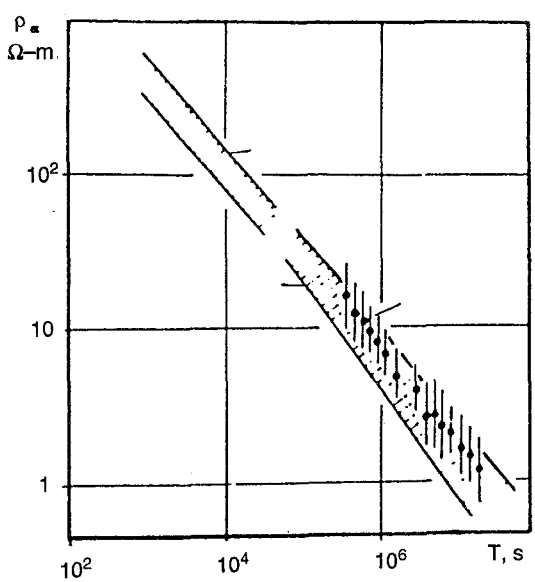


Fig. 6. After Kharin and Semenov (1985). The straight lines correspond to the envelope of the global transfer function. The dots are from Kharin and Semenov for the Honolulu observatory.

Conversely, Kharin and Semenov (1986) have studied Dst data from Honolulu with a continuum spectrum method and have shown that the corresponding transfer function lies within the error bars of global transfer functions (Figure 6).

Campbell and Schiffmacher (1988) determined conductivity profiles averaged

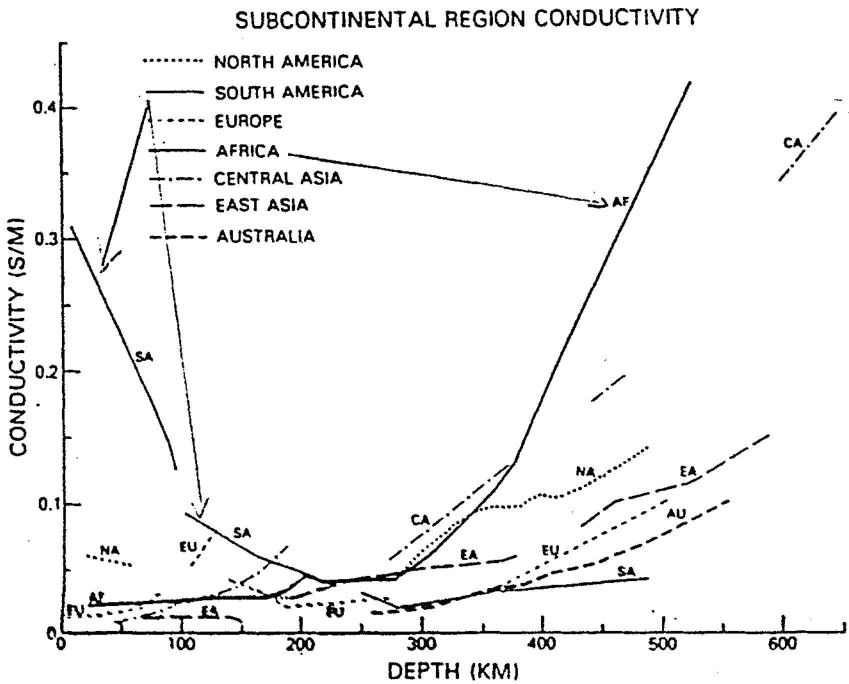


Fig. 7. After Campbell and Schiffmacher (1988, Fig. 13). Conductivity profiles beneath 7 continental areas.

over seven continental regions using a regionalization technique of the Sq field to determine the transfer functions (Campbell and Schiffmacher 1985). They inverted these data with the well known equivalent-conductor method (see for instance Schmucker 1987). Their technique however led to response functions that were sensitive to the position and intensity of the Sq vortexes. They tried to overcome this difficulty using a locally weighted regression fit. The authors show that the details of the conductivity profiles change significantly from one site to the other, a demonstration of the probable different structures below these regions (Figure 7).

We may notice however that the conductivity profiles below Africa and South America, indicated by the arrows (Figure 7), are widely different from the other sites. These results are in disagreement with the local and regional studies of Ritz and Robineau (1986) in Africa and Duhau and Favetto (1990) in both Africa and South America. The techniques used are very different however. Ritz and Robineau used deep MT soundings and Duhau and Favetto used the equatorial electrojet as an induction source.

Given the difficulty of using the Sq data for a number of reason (see Roberts 1986b for a discussion) and the limited frequency range available, most local analyses of long period geomagnetic data have involved data associated with the ring current field which is dominated by a P_0^1 geometry. The P_0^1 approximation can be used at periods not lower than about 3 days however, which prevents obtaining information at depths less than 300 to 400 km.

Roberts (1986b), Schultz and Larsen (1990), Schultz (1990), Petersons and Anderssen (1990), Chen and Fung (1991) have published local interpretations of P_0^1 data in terms conductivity profiles or of conductivity-analog. All the results suggest that the conductivity distribution varies significantly from site to site, an indication of mantle heterogeneities, either from shallow or from deep origin (or both).

We may wonder if the scatter of the conductivity models obtained by different authors for a set of sites was more important than the scatter among the conductivity models obtained by several authors at a single site. It is of course very difficult to compare at a glance the published results, since they are presented in a variety of ways with pictures as tiny as possible with log-log or semi-log scaling. Fortunately some authors provide tables (but others do not).

Figure 8 presents an example of conductivity profiles at three sites obtained from 3 different analyses. I also present at the same scale the whole set of models obtained by Schultz (1990) for comparison. On the whole, the agreement is reasonable between the authors and the scatter is much less than the scatter observed between the 16 models of Schultz (1990). This also suggests that the different interpretations proposed in the literature provide reasonable estimates of the conductivity, at least for the model space explored by the authors and in spite of the various processing techniques used to calculate the transfer functions.

Schultz (1990) paid special attention to the nature of the conductivity transition between the upper and the lower mantle. He demonstrated with the use of a smooth inversion technique that this transition should be sharp at most sites. Both Roberts (1986a, b) and Schultz (1990) reported that their analysis of the inversion of the data suggested a possible decrease of conductivity at some sites, at depths greater than 800 to 1,000 km. The very different approach and selection criteria used by these authors to select their data sets support the proposition that this feature is indeed required by the data and is not a bias introduced at some level of the analysis. Nevertheless, as mentioned by the authors themselves, the resolution of the data for this depth range is not sufficient to confirm this result.

Interpretation

The aforementioned variability of the conductivity profiles from one site to the other has been attributed to various causes. The preferred interpretations include the effect of surficial and uppermost mantle heterogeneities and/or inhomogeneities in the deep conductive layers.

Some authors have noticed the correlation between land-continent or major tectonic features and certain characteristics of the conductivity profiles (e.g. Roberts 1986a, b, Semenov 1989, Schultz and Larsen 1990, Schultz 1990).

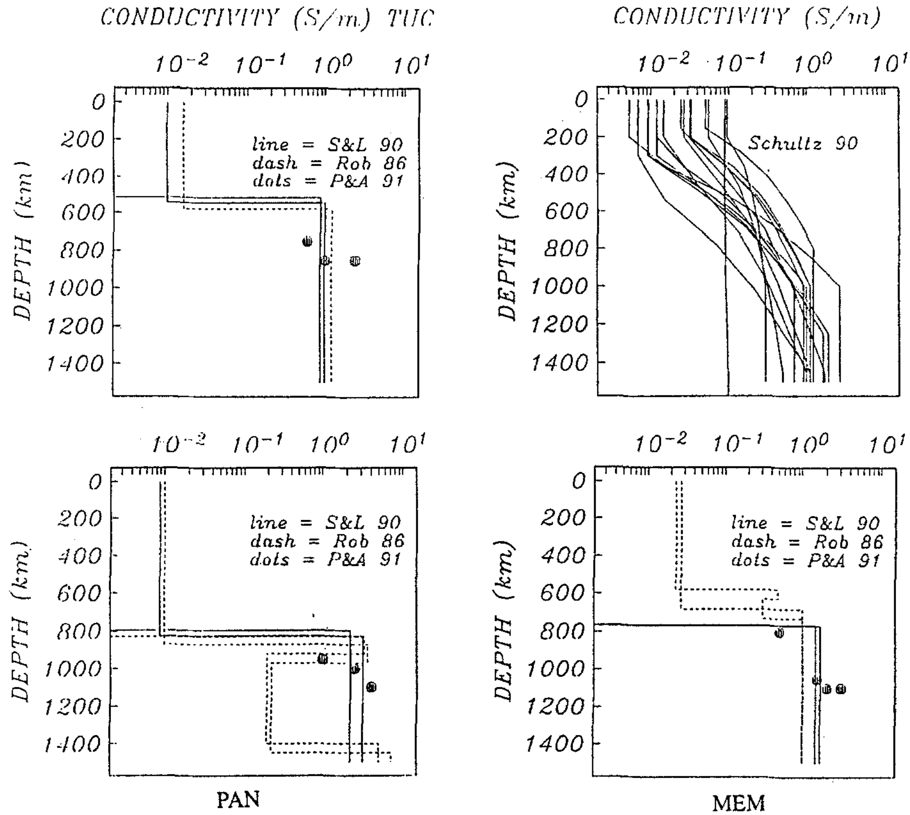


Fig. 8. Conductivity versus depth at 3 sites (Tucson, Panagurishe, and Memambetsu) studied independently by Roberts (1986), Schultz and Larsen (1990) and Petersons and Anderssen (1991). I also present at the same scale the whole set of models obtained by Schultz (1990) for comparison.

Schultz and Larsen (1990) distinguish three sets of data with respect to the presence or absence of shallow structures required by the geomagnetic data. The data sets which definitively require shallow structures (less than 400 km) are situated near or on known major tectonic features (eg Hawaii, Japan, New Zealand, Baikal area). Semenov (1989) also concludes that the data from the 3 stations along the Asia-Pacific limit have anomalously high phases and high apparent resistivity with respect to the majority of the data studied (at 24 sites). The remaining anomalous stations (4 in Europe and 1 in the Antilla Arc) only present anomalously high apparent resistivities.

These features appear only at short periods (1 to 10 days). In this period range, numerous other sites have failed to pass the test of one-dimensionality (e.g. Schultz and Larsen 1987) which suggests that the internal origin of the effect is shallow with respect to the penetration depth, which is within the first 400 km. Nonetheless, both Semenov (1989) and Schultz (1990) propose to reject the possibility of ocean perturbations as the sole source of the perturbations at the Japanese sites. They roughly estimated the conductance necessary to concentrate the amount of electric currents required to explain the anomaly, and they found excessive values with respect to the ocean conductance (30,000 Ω^{-1} against 15,000 Ω^{-1}). They suggested correlating the anomalous behavior of their data to the Japan Conductivity Anomaly (Rikitake 1966, Honkura 1978). On the other hand, there is no explanation (that I know of) for the other stations (in the Iberian Peninsula for instance), though Semenov suggested that the vicinity of plate boundaries may play a role.

The variability in the deeper features of the conductivity profiles could also be due to possible heterogeneities in the deeper layers in the mantle. This hypothesis is much less documented that the former. On the basis of the scale length of the observed conductivity-depth profiles in Europe, Roberts (1986a) suggested that the origin of the deep features in the conductivity profiles could be in the mid-mantle rather than at lithospheric depths. However, Semenov (1989) concluded that the anomalously high apparent resistivities in the Iberian Peninsula and in the Antilla would be incompatible with deep inhomogeneities for reasons which are not discussed.

Roberts (1986b) proposed that the conductivity contrast observed in the depth range 500–1000 km could be correlated with the 670 km seismic discontinuity rather than with the olivine-spinel transition, 400 to 500 km in depth.

To mention a 500–1000 km interval does not mean that the resolution of the depths is that bad. This actually corresponds to several sites with different depths at which the conductivity increases significantly either at relatively short scale (as between Kakioka and Memanbutsu in Japan) or at larger scales in Europe (Roberts 1986b) or in China (Chen and Fung 1991).

Not many studies have been carried out however on the precision of the determination of the conductivity model parameters. Schultz (1990) has discussed the resolving kernels of the smooth inversion of his data sets and has concluded that the resolution is maximum in the depth range 600–700 km. The consistency of the conductivity distribution found by different authors at the same sites supports this conclusion and suggests that the expected values of the model parameters are reasonably reliable despite the fact that their variances are poorly known. This means that the relative differences between the conductivity profiles are meaningful (Roberts 1986a, Dmitriev *et al.* 1987, Schultz and Larsen 1990). This is the most important point in the discussion of the interpretation of the conductivity-depths results in terms of Earth structure.

However, it is difficult to match the important variability observed between the conductivity profiles to known mantle characteristics such as the phase transitions or the 670 km discontinuity. Variations of these features over a range 500 to 1,000 km seem to be excluded because their topography does not appear to exceed 30 km (e.g. Shearer and Masters 1992). Furthermore, laboratory experiments of the thermodynamics of the transition phases indicate that the Clapeyron slope is such that great changes of depths of the transition would lead to unrealistic lateral changes of temperature (500 to 1000 degrees).

It is interesting to note however, that recent results concerning the global estimation of the mean conductivity distribution for a spherical symmetric Earth (Hobbs 1987, Dmitriev *et al.* 1987, Semenov 1989) have confirmed that the mean global conductivity (the model of degree zero) has a rapid increase at a depth of about 600 to 700 km. The conductance of this transition zone could range from 125,000 to 500,000 S, an estimation in reasonable agreement with earlier results. As an illustration, Figure 9 shows the global conductivity models obtained by Hobbs (1987) and Dmitriev *et al.* (1987). Although the authors used two representations as extreme as possible (a smooth continuous model, and a two layer model), it is still possible to recognize the rapid increase of conductivity (or decrease of resistivity) at a depth of about 600 to 700 km.

If we believed that this is not coincidental, it would mean that part of the variability of the conductivity profiles could be correlated (through a still unknown mechanism) to the 670 km seismic (and probably chemical) discontinuity rather than to the transition zone, despite the fact that laboratory experiments on the possible minerals of the transition zone suggest that the conductivity should increase significantly (e.g. Omura 1991).

This result suggests that conductivity and seismic velocity could be correlated to some extent. In addition, if any correlation exists, this could mean that the conductivity profiles, obtained from various analyses, have an internal consistency, and that the uncorrelated part could likely be a real feature to which the seismic data would not be sensitive.

Electrical conductivity and seismic velocity

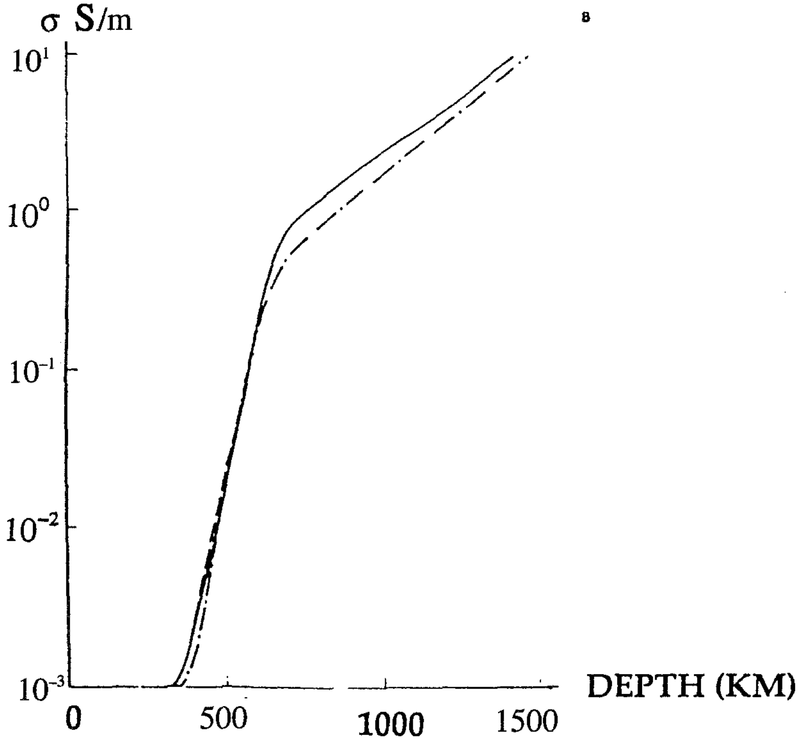
Conductivity data

The examination of the available distributions of conductivity within the depth range 400 to 1,500 km provided by various authors shows that one class of models is easily comparable from all authors.

This class consists of layered conductivity models obtained by Schmucker's inversion or some similar substitute. I used 33 sites (Figure 4) where Roberts (1986a) and Schultz and Larsen (1990) published such models and I selected at each site, the depth of the main reflector because it is the most important feature in the conductivity profiles and it is reasonably well resolved (Schultz 1990).

One may argue that the number of data points is ridiculously small. However,

Fig. 9. This figure presents the global conductivity models obtained by Hobbs (1987, Fig. 11) (lower part) and Dmitriev *et al.* (1987, Fig. 4b) (upper part). Although the authors used 2 representations as extreme as possible, it is still possible to recognize the rapid increase of conductivity (or decrease of resistivity) at a depth of about 600 to 700 km.


σ S/m
B
10¹
10⁰
10⁻¹
10⁻²
10⁻³
0
500
1000
1500
DEPTH (KM)

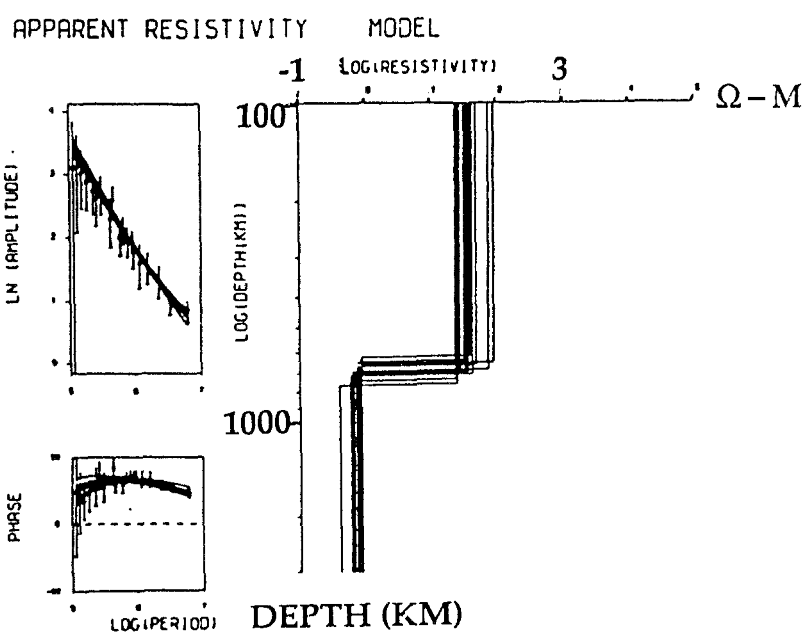

APPARENT RESISTIVITY
MODEL
-1
LOG(RESISTIVITY)
3
Ω – M
LN (AMPLITUDE)
PHASE
LOG(PERIOD)
LOG(DEPTH(KM))
100
1000
DEPTH (KM)

TABLE I
Correlation between the 670 km discontinuity topography and the depth of the main electrical conductor. Degrees of freedom are 14 for depths less than 700 km and 19 for depths more than 700 km.

	Correlation	Probability of significance
Corr. with depths <700 km	0.48	0.92
Corr. with depths >700 km	0.61	0.99

the electromagnetic data correspond to conductivities averaged over quite large areas, which minimizes somehow the aliasing problem encountered in other types of data such as seismic data.

Therefore the depths of the conductor determined from the geomagnetic data are averaged over an area with a size that depends on the whole process to determine the local transfer functions. Schultz and Larsen (1987) estimated the radius of the area to be about 800 km. Roberts's processing combines a global estimate of the horizontal geomagnetic field with the local estimate of the radial component. I attributed a radius of 1,600 km to his estimates of the conductor depths.

Seismic data

I have considered two sets of results obtained from the global analysis of the seismic data. First I considered the volumetric distribution of velocity anomalies provided by tomographic analysis. I used the model of relative S-wave velocity anomalies $\Delta\beta/\beta$ with respect to a PREM model published by Tanimoto (1990). In this model, the Earth is divided into a stack of inhomogeneous layers.

The second set of results I considered use the topography of the major seismic discontinuity at 670 km depth, determined by the global analysis of seismic waves reflected from the discontinuity (Shearer and Masters 1992).

The seismic models come as coefficients of a spherical harmonic expansion. It is straightforward to integrate these parameters over the area covered by a given electromagnetic parameter. The integrated seismic parameter may then be directly compared to its electromagnetic counterpart.

Correlation between electromagnetic and seismic models

I have split the set of 33 conductor depths into 2 sets, one with 14 depths less than 700 km and one with 19 depths greater than 700 km. I ran several statistical test to estimate the correlation between the electromagnetic and the seismic parameters. Finally I selected the non-parametric Spearman test (or rank correlation) for its robustness to undefined probability distribution of the data.

Table I presents the correlation between the 670 km discontinuity topography and the conductor depths at all sites. We observe that the correlation is significant at a high probability level. Both the shallow and the deep conductor depths are

TABLE II

Correlation between the 2 sets of conductor depths (less than 700 km and more than 700 km) and the 4 first layers (down to 1022 km) of Tanimoto's model. Degrees of freedom are 14 for depths less than 700 km and 19 for depths more than 700 km.

Layer	Depths <700 km Corr, Proba	Depths >700 km Corr, Proba
0–220 (1)	No Correlation	No Correlation
220–400 (2)	0.51 at 0.94	0.33 at 0.83
400–670 (3)	0.46 at 0.90	0.42 at 0.92
670–1022 (4)	No Correlation	0.35 at 0.87

correlated to the seismic topography, with a maximum for the set of deeper conductors. The positive correlation indicates that the topography of the conductor depth varies in the same direction as the topography of the seismic discontinuity.

Consider now the correlation between the depth of the electrical conductor and the 3-D distribution of the S-wave velocity variation determined by Tanimoto (1990). Table II presents the correlation between the 2 sets of conductor depths and the 4 first layers (down to 1022 km) of Tanimoto's model. Most of the correlation is found with layers 2 and 3. Shallow conductor depths do not correlate with layer 4 but deep conductor depths do.

We observe a positive correlation between the topography of the conductor depth and the structure of the S-wave velocity variation. In other words, the fast seismic zones correlate with the most resistive zones of the mantle in the depths range considered. This is a reasonable result and in accordance with the correlation we found for the 670 km discontinuity topography.

The interpretation of such correlations is not easy. At this stage we may say that both conductivity and S-wave velocity tend to vary in the same direction under the influence of an unknown process.

It is however possible to test one hypothesis, i.e. that the topography of the 670 km discontinuity could be directly responsible, for the variability of the conductivity. The test was carried out by taking the 670 km discontinuity topography, converting it into a slab of variable conductivity, and calculating the effect on the geomagnetic field of such a heterogeneous structure.

I considered a model of the Earth made of an upper heterogeneous shell which represents the continents and the ocean, overlying two homogeneous layers down to 630 km. Between 640 and 670 km, I converted the topography of the discontinuity published by Shearer and Masters into an inhomogeneous slab of variable conductivity.

Let $H(\theta, \phi)$ be the topography of the discontinuity. I calculated the effective conductivity $\sigma^*(\theta, \phi)$ of the slab as follow:

$$\sigma^*(\theta, \phi) = (\sigma_{MS}[H(\theta, \phi) - 640] + \sigma_{MI}[670 - H(\theta, \phi)])/30 \qquad (1)$$

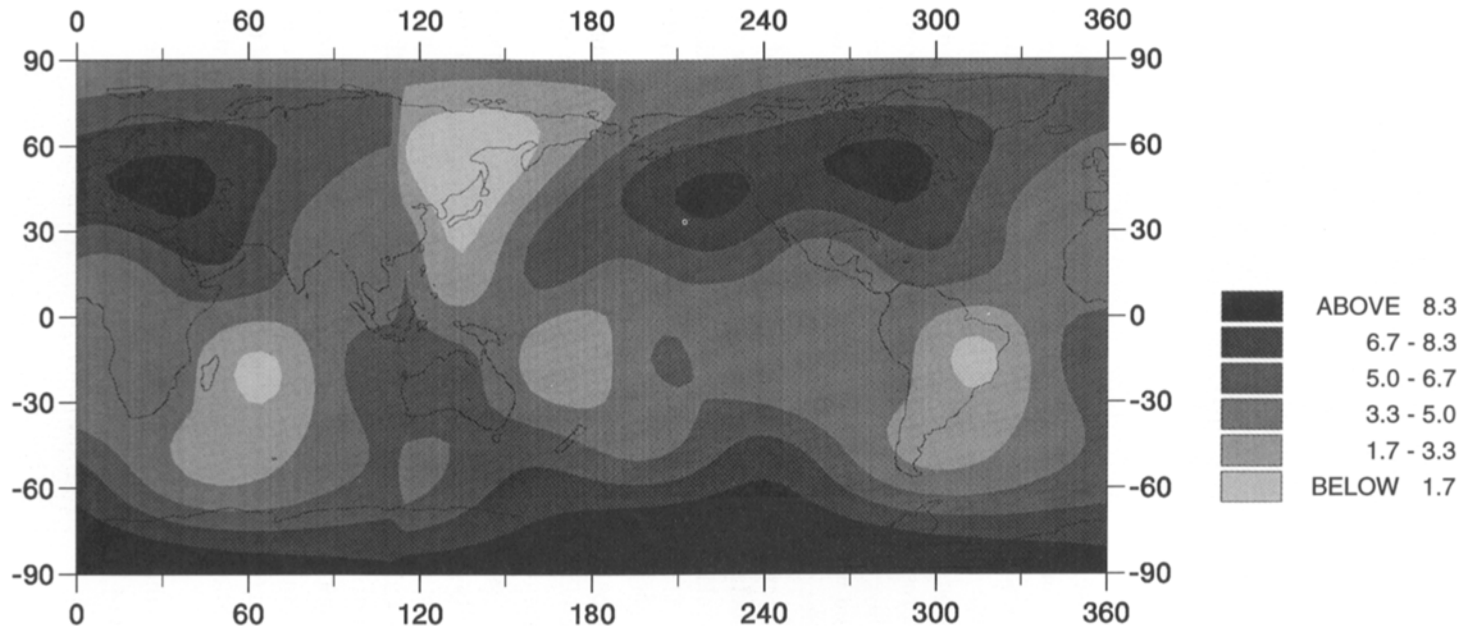


Fig. 10. Topography of the 670 km seismic discontinuity converted into a slab 30 km thick with a variable electrical conductivity. Values are in S/m. (see text for details).

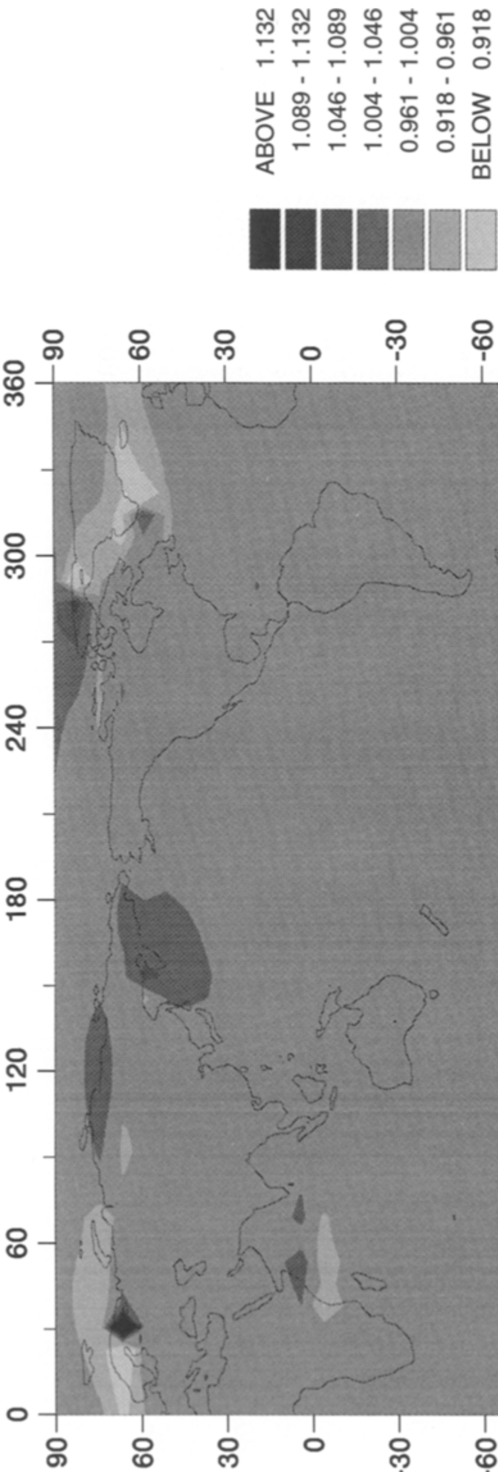

ABOVE 1.132
1.089 - 1.132
1.046 - 1.089
1.004 - 1.046
0.961 - 1.004
0.918 - 0.961
BELOW 0.918
0
60
120
180
240
300
360
90
60
30
0
-30
-60
-90

Fig. 11a.

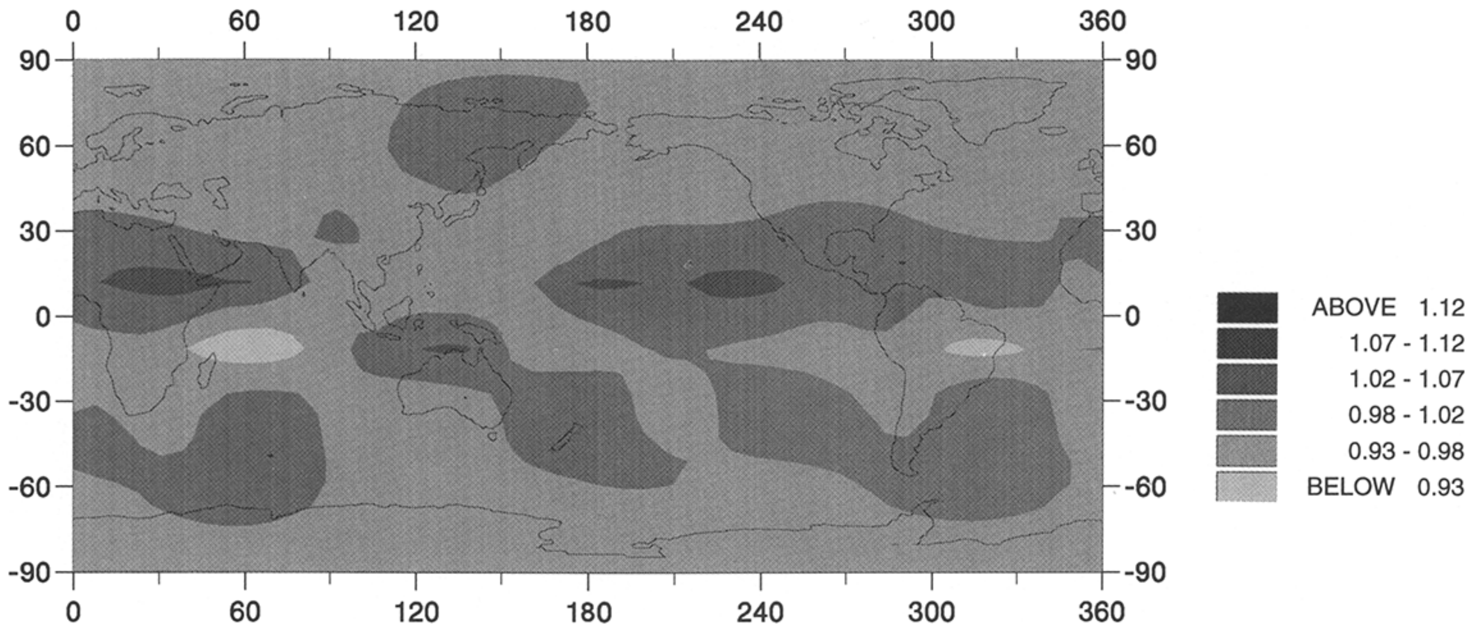


Fig. 11b.

Fig. 11. Ratio of the EAR for the model including the coast effect and the 670 km deep heterogeneous slab divided by the same quantity for the model with the coast effect only. (a) at 1 day period and a Sq source field. (b) at 5 days period for a P_0^1 source field.

σ_{MS} is the upper mantle conductivity taken as $0.1\,\Omega^{-1}\,m^{-1}$ and σ_{MI} is the lower mantle conductivity taken as $10\,\Omega^{-1}\,m^{-1}$. I used this value to maximize the effect of the conductivity contrasts. The conductivity distribution obtained is displayed Figure 10 (in S/m). I ran the model at 1, 5 and 10 days. Figure 11 presents the EAR for the total model divided by the same quantity calculated for the model without the inner heterogeneous layer. The results are presented at 1 and 5 days. The effect is small at all periods (less than 5%).

This suggests then, that the correlation observed between the conductor depths and the 670-km discontinuity might be the indirect manifestation of a process acting differently on both parameters. This could explain why we obtain a good correlation with the 670-km discontinuity topography as well as with the volumic anomalies of velocity.

This is certainly not a definitive demonstration that the spatial variations of a major seismic discontinuity is far too small to affect the surface geomagnetic data, because we did not take into account the structures visualized by the tomographic imaging which seems to be also correlated with lateral contrasts of conductivity and therefore might contribute to the observed variability of the mantle conductivity structure.

Nevertheless, if the correlations observed between the electromagnetic and the seismic parameters were not an artifact, we have a clear demonstration that both parameters are linked by a common process which is sampled very differently by seismic and electromagnetic data. Of course, we have not yet reached a situation where we may provide 3-D images of mantle conductivity. Much work is necessary. It is in particular mandatory to increase the data sets and therefore to progress in the processing and interpretation of geomagnetic data in areas where tests both for a homogeneous source field and for underlying 1-D structures fail.

In addition, I would like to emphasize that we now understand reasonably well the mechanisms of electromagnetic distorsion by local heterogeneities and therefore we should be able to properly use long period electric data as invaluable complementary sets of information in classical geomagnetic analyses.

The knowledge of the 3-D conductivity structure of the mantle is one of the important keys to understanding the inner mechanisms of our planet.

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