Global EM induction studies of deep Earth from ground and space. Progress status.

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SUMMARY

In the last decade, global electromagnetic (EM) induction studies of the Earth have found a new breath, particularly due to a growth in the amount and variety of data available, especially from recent geomagnetic satellite missions, like CHAMP and *Swarm*. The increased interest in such studies is also due to the development of new approaches for analyzing long-period (LP) EM data, including joint inversion of the signals of different (tidal, ionospheric, and magnetospheric) origins. Such interest has also resulted from the significant methodological progress made during the last few years that enabled accurate and efficient modelling of the LP EM responses in spherical Earth models with a three-dimensional distribution of electrical conductivity. In this abstract the current state of global EM induction studies is overviewed and possible directions for future work are outlined.

Ключевые слова: global EM induction, satellite and ground-based data, electrical conductivity of deep Earth

INTRODUCTION

Understanding 3-D physical properties of the Earth's mantle on a global scale is outstanding problem of modern geophysics. There exist only two direct methods that can deliver depth-dependent information about the structure of the Earth's mantle: seismic and electromagnetic (EM) sounding methods. Global seismic tomography provides today a variety of three-dimensional (3-D) velocity models that can be interpreted in terms of cratonic roots, mantle plumes and slab graveyards. While substantial progress in seismology models has been achieved, the interpretation of the seismic velocity anomalies in terms of thermo-dynamical and compositional parameters is often uncertain especially when it comes to constraints on hydrogen content.

The goal of EM sounding methods is to identify spatial variations of the electrical conductivity in the Earth's interior. Since the conductivity is sensitive to the temperature, chemical composition and hydrogen content, it helps understand the Earth's origin, past evolution and modern dynamics.

In this abstract I will overview EM methods/transfer functions used to probe deep Earth, discuss the recent results, and outline possible directions for future work.

TRANSFER FUNCTIONS USED FOR DEEP EM SOUNDING AND THE RESULTS BASED ON THEIR ANALYSIS

In the following three subsections, I will discuss TFs coming from the analysis of magnetic field variations - the main data source for the EM probing of the deep Earth.

Periods between a few days and a few months

At these periods magnetic field variations mostly originate from the magnetospheric ring current, which is usually described by a first zonal spherical harmonic (SH) $Y_1^0 = \cos\vartheta$. In this case, the conductivity can be obtained by inverting *local C-responses* (Banks 1969) that relate local radial and tangential magnetic fields

$$C_1(\overline{r}_a,\omega) = \frac{a}{2}\tan\vartheta\frac{B_r(\overline{r}_a,\omega)}{B_\vartheta(\overline{r}_a,\omega)},$$

(1)

where ω is the angular frequency, $\overline{r}_a = (a,\vartheta,\varphi)$, a = 6371.2 km is a mean Earth's radius, ϑ and φ are geomagnetic colatitude and longitude, respectively. There are numerous studies (Munch et al, 2018; Zhang et al., 2020; Yuan et al, 2020; among others) that perform site-by-site 1-D inversions of local *C*-responses estimated at mid-latitude geomagnetic observatories in an attempt to detect lateral variations in mantle conductivity. Moreover, a few semi-global (Utada et al., 2009; Shimizu et al, 2010; Koyama et al, 2014) and global (Kelbert et al, 2009; Semenov and Kuvshinov, 2012; Sun et al, 2015; Li et al., 2020) 3-D mantle conductivity models were published based on the inversion of these responses.

A few comments on the recovered 3-D models are relevant at this point. First, due to their frequency content, local *C*-responses, and thus models based on them have limited sensitivity in the crust and upper mantle. Second, spatial sparsity and irregularity of observatories' distribution (with high- and low-latitude observatories eliminated) precludes obtaining a truly global 3-D conductivity model of the Earth's deep interior. Note that the usage of solely mid-latitude observatories is dictated by the fact that variations due to large-scale magnetospheric source (ring current) are analysed. This, on the one hand, allows researchers to use local C-response concept, but on the other hand, impedes the usage of polar data where another (ionospheric) source dominates, as well as equatorial data where signal-to-noise ratio in radial magnetic field component is prohibitively low. Third, the family of 3-D models produced to date is yet to reach a consensus. The present discrepancy is mostly due to: (a) inherent strong non-uniqueness of the inverse problem arising from spatial sparsity and irregularity of data distribution and limited period range; b) inconsistency of external (magnetospheric) field model, which ignores the more complex spatial structure of ring current. Indeed, there has long been evidence for a more complex structure and asymmetry of the magnetospheric ring current (cf. Balasis and Egbert, 2006).

To overcome the latter problem, Püthe et al (2015a) introduced arrays of new transfer functions (TFs) that are capable of handling sources of arbitrary complexity. These TFs relate the expansion coefficients describing the source globally with a locally measured vertical magnetic field, hence these TFs are referred to as global-to-local (G2L) TFs. Assuming that the magnetic field (above the Earth, $r \geq a$, and below the source) is described as a gradient of the scalar potential, i.e. $\overline{B} = -\mathrm{grad}\, V$ and V is represented via SH expansion

$$V = \sum_{n,m} \left[\varepsilon_n^m(\omega) \left(\frac{r}{a} \right)^n + \iota_n^m(\omega) \left(\frac{a}{r} \right)^{n+1} \right] Y_n^m(\vartheta, \varphi), \quad (2)$$

these G2L TFs, $T_n^m(\overline{r}_a, \omega)$, are given as

$$Z(\overline{r}_a, \omega) = -\sum_{n,m} \varepsilon_n^m(\omega)\, T_n^m(\overline{r}_a, \omega) \quad .$$

(3) Here $\sum_{n,m} = \sum_{n=1}^{\infty} \sum_{m=-n}^{n}$, $Z = -B_r$, and ε_n^m and ι_n^m are external and induced (due to EM induction in conducting Earth) coefficients. The determination of these TFs includes two steps: first, time series of external coefficients are determined from observatory data, and, second, TFs are estimated using relation described by equation (3). Munch et al (2020) estimated and inverted G2L TFs at several continental geomagnetic observatories in terms of local 1-D conductivity distributions and revealed lateral variations in mantle water content.

Despite continuing efforts, spatial irregularity of observatories with only a few in oceanic regions precludes obtaining the cogent global 3-D model of mantle conductivity of uniform lateral resolution from observatory data. In contrast to ground-based data, satellite-borne measurements provide better spatio-temporal data coverage. With the *Swarm* satellite constellation mission, the possibility of obtaining global images of 3-D mantle heterogeneities, especially in oceanic regions, comes into reach. Bearing this in mind, mapping 3-D electrical conductivity of the Earth's mantle has been identified as one of the main scientific objectives of *Swarm.* The challenge here is that one cannot use – because satellites move in space – local or/and global-to-local transfer functions concepts described above. As part of the *Swarm* mission preparation, Püthe and Kuvshinov (2014) have developed a formalism that allows us to interpret the satellite magnetic data in terms of 3-D mantle conductivity distribution. It relies on an estimation and 3-D inversion of *matrix Q-responses* that relate external and induced coefficients

$$\iota_k^l(\omega) = \sum_{n,m} Q_{kn}^{lm}(\omega) \varepsilon_n^m(\omega) \quad .$$

(4)

Matrix Q-responses are global TFs, meaning that they do not depend on $\overline{r}$. We also note that if the Earth's conductivity model is assumed to be 1-D, equation (4) degenerates to

$$\iota_n^m(\omega) = Q_n(\omega) \varepsilon_n^m(\omega). \quad (5)$$

Taking in account that dominant term in the spatial representation of the signals of magnetospheric origin is Y_1^0, the best resolvable response is Q_1, which can be further converted into *global C-response* as (cf. Kuvshinov et al, 2021, Appendix A)

$$C_1(a,\omega) = \frac{a}{2}\frac{1-2Q_1(\omega)}{1+Q_1(\omega)}.$$

(6)

Püthe et al (2015b) estimated global *C*-responses from satellite and observatory data and obtained from them a global 1-D conductivity profile, which, however, poorly constrains conductivity in the crust and upper mantle (depths: 0 – 400 km). Grayver et al (2017) performed joint 1-D inversion of global *C*-responses and tidal signals and showed that joint inversion allows one to constrain conductivity in the whole depth column (0 – 1500 km).

Kuvshinov et al (2021) implemented matrix Q-responses concept to obtain global 3-D conductivity distribution in the mid mantle (depths: 400 – 1200 km) using as an input time series of external and induced SH coefficients of magnetospheric origin estimated from 6 years of *Swarm*, Cryosat-2 and observatory data. Time series of external and induced coefficients are estimated up to degrees 2 and 3, respectively, implying one can obtain only a low-resolution 3-D conductivity model.

Periods between four hours and one day

Tighter constraints on the electrical structure in the upper mantle (depths down to 400 km) would require considering magnetic field variations in a period range between a few hours and one day. At these periods the dominating source of the magnetic field variations is an ionospheric current system (Yamazaki and Maute 2017), which has a much more complex spatio-temporal structure than the magnetospheric ring current. Despite this, several studies analysed daily magnetic variations and utilized a variant of local *C*-response concept, which represents the source by using a single SH (cf. Simpson et al*.,* 2000). However, presently there exists a consensus that the description of the ionospheric source by a single SH is too simplistic. Alternatively, local *C*-responses can be estimated without prior assumptions about the source geometry (cf. Olsen, 1998). Such estimation requires local tangential gradients of the tangential magnetic field. Since direct measurement of the gradients is extremely challenging in practice, they are commonly computed with the use of tangential magnetic field measured at an array of observation sites located nearby. The prerequisite for the successful implementation of this approach is a relatively dense regional grid of observations in the region of interest. This significantly limits its wide practical adoption.

To overcome the discussed difficulties, Koch and Kuvshinov (2013) explored the approach proposed by Fainberg et al. (1990) for analysing daily magnetic variations. Rather than working with the responses, they work with field spectra directly. Despite promising results (Koch and Kuvshinov 2015), the analysis of spectra has one significant shortcoming: it requires an actual description of the source. In practice, however, one determines the source with an inevitable error, and this injects an undesired and uncontrolled uncertainty into the recovered conductivity models.

As was discussed earlier, Püthe et al (2015a) introduced a new type of transfer functions (TFs) that relate expansion coefficients describing the source globally with a locally measured vertical magnetic field. With the new TFs, one avoids complications of finding the actual description of the source. The approach only requires to specify a set of basis functions (in this work SH) which adequately describes the source geometry. Additionally, this approach enables us to estimate the statistical uncertainties of the TFs. Guzavina et al (2019) discussed the applicability of this approach to the solar quiet (Sq) variations, estimated Sq G2L TFs at a global net of observatories, and inverted estimated TFs at a few locations around the world in terms of 1-D mantle conductivity. Zehnhausern et al (2021) explore an alternative parametrization of the ionospheric source based on a principal component analysis (PCA) of the Fourier transformed output from the physics-based Thermosphere Ionosphere Electrodynamics General Circulation Model (TIE-GCM; Qian et al., 2014).

Periods between a few minutes and 3 hours

The structures at depths 0 – 200 km can be imaged using magnetic field variations with periods shorter than 3 hours. The source of these variations is approximated by a vertically incident plane wave. The plane wave assumption allows one to relate the vertical component with the horizontal components via magnetic MT responses - tippers (Berdichevsky and Dmitriev, 2008)

$$B_z = T_{zx}B_x + T_{zy}B_y, \qquad (7)$$

where all quantities depend on frequency and location. Because of the plane-wave excitation, vertical component, and thus tippers, are nonzero only above non-1-D conductivity structures. Wang et al (2014) performed the first 3-D inversion of tippers using AWAGS data (Chamalaun & Barton, 1993) to image the conductivity beneath the whole Australia. Morschhauser et al (2019) carried out quasi 1-D inversion of tippers to constrain the conductivity beneath two island observatories: TDC and GAN. Here the term "quasi" is used to stress the fact that during 1-D inversion the 3-D forward modeling operator is exploited to account for the ocean induction effect (cf. Parkinson and Jones, 1979).

OUTLOOK

Alternative approaches to estimate time series of inducing and induced coefficients

As shown by Kuvshinov et al (2021) many elements of the *Q*-matrix are poorly resolved. The most probable reason for this is an imperfect estimation of the inducing and induced coefficients. Note again that the induced coefficients are responsible for 3-D conductivity effects, and one 3-D effect that strongly influences the results but cannot be properly addressed by the Gauss method is the so-called ocean induction effect (Kuvshinov, 2008). Recall that the Gauss method is based on a simultaneous analysis of radial and horizontal magnetic field components. Given deficient spatio-temporal distribution of observatory and *Swarm* data, with Gauss method one is able to recover only low SH coefficients both in the inducing and induced parts. But induced radial magnetic field requires higher SH degrees for its proper description, since this component is strongly affected by the (localized) ocean effect.

There exists an alternative approach to isolate the inducing part of the signals from the induced part. It exploits the precomputed EM fields/responses induced by "elementary" extraneous currents in an a priori model of known conductivity; commonly this model includes oceans and continents of laterally variable conductance underlain by a layered (1-D) medium. This approach has been routinely used for the last two decades to retrieve the inducing part of the signals *in the frequency domain*. In particular, the concept was used in analyses of ground based (cf. Kuvshinov et al., 2007) and satellite (cf. Sabaka et al., 2015) magnetic data of ionospheric origin (assumed to be mostly periodic). It was also used for analysis of aperiodic ground-based data of magnetospheric origin (cf. Honkonen et al., 2018). Note that the latter studies aimed to explore the spatio-temporal evolution of the induced EM field, and time-domain results in these studies were obtained by converting the frequency-domain results into the time domain by means of a Fourier transform (FT).

Working with satellite data we are interested in isolating inducing (external) and induced parts of (aperiodic) signals of magnetospheric origin. This prompts a two-step procedure for retrieving time series of the corresponding SH coefficients. The procedure is based on the fact that the magnetic tangential components are much less influenced by 3-D effects compared to the radial component (Kuvshinov, 2008). Thereby, by analysing the tangential magnetic components and assuming an a priori Earth's conductivity model, one determines time series of inducing coefficients. Note, that to account for time-domain induction effects in satellite data the FT approach hardly works, since satellites move in space. But retrieving the inducing coefficients can be done directly in the time domain using a concept of impulse responses (Svetov, 1991). With the retrieved time series of inducing SH coefficients, one determines in a second step the induced coefficients by analysing the magnetic radial component only. An implementation of this two-step approach is discussed in Grayver et al (2021).

Exploiting data from non-dedicated satellite missions

In the context of 3-D EM induction studies we strive for a precise and detailed description of the spatiotemporal structure of inducing and induced signals. From this perspective, the ideal geomagnetic satellite mission would consist of a large number of low Earth orbiting (LEO) satellites (in polar, circular orbits) uniformly separated in local time (that is, in longitude). This configuration allows for detecting both latitudinal and longitudinal variability of the signals. The more satellites, the higher the resolution of the recovered inducing and induced signals, both in time and space. The existing *Swarm* constellation mission comprises two satellites with varying local time separation between zero and six hours, thus limiting detection of longitudinal variability of the signals. Olsen et al. (2020) and Stolle et al. (2020) used data collected by the CryoSat-2 and GRACE-FO satellites, and demonstrated that platform magnetometers do provide valuable geomagnetic field measurements, given that vector magnetic field can be properly calibrated. These platform magnetometers are present onboard many LEO satellites, as a part of their attitude control system. In (Kuvshinov et al., 2021) authors used calibrated CryoSat-2 data in addition to *Swarm* and observatory data, and observed that these data indeed improve the recovery of the aforementioned signals. However, improvement by adding data from one single satellite is limited. Obviously, more data from LEO satellites in orbits with different local times are necessary for a substantial improvement; for instance, usage of duly calibrated data from satellites like the Iridium-Next and Spire constellation could be promising.

Another opportunity are dedicated satellite constellations with low-inclined satellites (Hulot et al., 2021) which will enable researchers to better characterize the complex spatio-temporal nature of the ionospheric and magnetospheric signals.

Multi-response, multi-source and multi-resolution global 3-D imaging

Most of the results discussed above rely on an analysis of magnetic field variations in each period range separately. Joint analysis of variations in a period range as wide as practicable would provide an opportunity for imaging the 3-D conductivity structures of the Earth throughout its full depth range - from the crust down to the lower mantle. However, this is a challenging task, because it requires combining transfer functions from sources of different morphology, specifically, matrix Q-responses, G2L TFs and tippers. The latter two (local) responses are estimated from ground-based data, whereas the (global) matrix Q-responses are estimated from satellite (and observatory) data. As discussed above, the satellite-based matrix Q-responses allow for a global 3-D imaging of mid-mantle conductivity but at rather low (continental-scale) lateral resolution. Ground-based data allow for higher resolution 3-D imaging of the Earth's mantle in regions with a dense net of continuous observations (like in Europe and China), or temporary long-term (like in Australia, AWAGS) observations. However, bearing in mind an overall very irregular spatial distribution of the ground-based magnetic sites (with substantial gaps in oceanic regions), a proper determination of a global 3-D mantle conductivity model of (uniform) high-resolution is hardly feasible. The above considerations suggest a multi-resolution approach to global 3-D imaging. Specifically, at a first step a low-resolution baseline global 3-D conductivity model at mid-mantle depths is obtained by inverting matrix Q-responses. Wherever possible, large scale regional 3-D conductivity models in the full depth range are obtained by joint inversion of tippers and global-to-local transfer functions. As for oceanic regions, one can decipher local one-dimensional (1-D) conductivity profiles beneath island observatories. The final step of the discussed approach is a compilation of the retrieved global, regional and local models in a global multi-resolution model, for instance, as done by (Alekseev et al., 2015).

So far, we discussed the works utilizing magnetic field data, and confined to onshore observations as ground-based data. However, as discussed earlier, there is an overall deficiency of geomagnetic data in oceanic regions. Data from sea-bottom long-term (with measurement period from a few months to a few years), large-scale MT surveys (Baba et al., 2017; Suetsugu et al., 2012, among others) can fill, at least partly, this spatial gap. A rather exhaustive summary of available sea-bottom MT data sets is presented in (Guzavina, 2020). From these data one can estimate and invert both MT responses (impedances and tippers) and "daily band" global-to-local TFs. It is noteworthy that with sea-bottom long-term MT data it is possible to estimate not only "vertical magnetic" global-to-local TFs, but also TFs relating SH expansion coefficients with local horizontal electric and magnetic fields (Guzavina, 2020).

Besides, sea-bottom MT data generally comprise detectable tidal signals which also can be used for constraining conductivity distribution in the lithosphere and upper mantle beneath oceans (Zhang et al., 2019).

In addition, one can expand the database for EM sounding of the deep Earth with long-period MT data from a few continental-scale MT projects, such as EarthScope, AusLAMP and SinoProbe.

Note that MT data interpretation is routinely done by assuming Earth is flat, which may become unjustified for large-scale MT surveys. Grayver et al (2019) presented a methodology for calculating the MT impedance tensor in spherical Earth with a 3-D conductivity distribution; the authors show how the MT impedances in spherical coordinates can be modeled using three polarizations of a uniform external magnetic field.

As for modelling in spherical geometry magnetic MT responses – tippers, one needs another type of excitation because the uniform external magnetic field of any polarization contains a non-zero radial component. Kruglyakov and Kuvshinov (2021) elaborated a source model, which leads to valid tippers on a whole sphere or a part thereof. This opens an avenue for obtaining global/semi-global 3-D conductivity models at depths < 200 km. It is important to stress that in contrast to local C-responses or/and G2L TFs, tippers can be estimated in equatorial regions. Moreover, to obtain reliable estimates for tippers one can use data with rather short (a few weeks or so) measurement periods. This enables exploiting not only observatory data but an immense amount of magnetic data from numerous projects around the globe (Shimizu & Utada 1999; Gjerloev 2012; Denardini et al. 2018, among others).

Finally, multi-year observations of electric field (Fujii et al., 2015; Wang et al., 2020, among others) is another promising source of data to probe deep structures of the Earth.

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