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Electrical conductivity of the Earth's mantle from the first Swarm magnetic field measurements

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Abstract We present a 1-D electrical conductivity profile of the Earth's mantle down to 2000 km derived from L1b Swarm satellite magnetic field measurements from November 2013 to September 2014. We first derive a model for the main magnetic field, correct the data for a lithospheric field model, and additionally select the data to reduce the contributions of the ionospheric field. We then model the primary and induced magnetospheric fields for periods between 2 and 256 days and perform a Bayesian inversion to obtain the probability density function for the electrical conductivity as function of depth. The conductivity increases by 3 orders of magnitude in the 400–900 km depth range. Assuming a pyrolitic mantle composition, this profile is interpreted in terms of temperature variations leading to a temperature gradient in the lower mantle that is close to adiabatic.

1. Introduction

On 22 November 2013, the European Space Agency successfully launched the Swarm satellite mission devoted to the study of the Earth's magnetic environment. The Swarm scientific mission consists of three identical satellites carrying vector and scalar magnetometers, two of which (A and C) flying side by side at about 450 km and a third (B) being at about 530 km with a phase shift increasing with time that will allow the constellation to survey all local times during its nominal lifetime. Swarm will thus permit the best ever separation of the internal and external magnetic field sources. This configuration, in particular, opens the possibility to better probe the conductivity of the Earth's mantle [Kuvshinov et al., 2006; Püthe and Kuvshinov, 2013a, 2013b; Velímský, 2013], which is one of the primary science objectives of the mission [e.g., Olsen et al., 2013]. Electromagnetic (EM) induction studies from space were carried out during the past two decades [Olsen, 1999a; Constable and Constable, 2004; Kuvshinov and Olsen, 2006; Velímský, 2010; Civet and Tarits, 2013]. They show reasonably good agreement for periods ranging from 1 day to a few months but differ more significantly at the shortest and longest periods. This difficulty arises because some internal and external magnetic field sources overlap in time and in space so that their separation is ambiguous over these time periods [e.g., Olsen, 1999b]. For example, the long-term external field variation can hardly be distinguished from the Earth's main field secular variation, and periods of about 1 day are smeared with various effects such as the ionospheric field daily variation [Tarits and Grammatica, 2000] or the field induced by ocean tides [Tyler et al., 2003]. A further complication is due to the heterogeneous distribution of the satellite magnetic field measurements in space and especially in time that introduces data gaps [Civet and Tarits, 2013]. The common way to circumvent this difficulty is to average the magnetic components over one or several orbits [e.g., Olsen, 1999a]. Another approach recently put forward by Civet and Tarits [2013, 2014] in the context of the planetary exploration is to fill gaps using a proxy for the variability of the external magnetic field.

Despite these limitations EM induction satellite-based studies generally agree with an electrical conductivity increase with depth from $\simeq 0.01$ S m⁻¹ near the surface to $\simeq 10$ S m⁻¹ at 2000 km depth. Such conductivity values are supported by laboratory mineral conductivity measurements [e.g., *Xu et al.*, 2000]. However, major mineralogical discontinuities arising in the mantle are not seen by these EM studies without prior information.

Electrical conductivity of mantle minerals depends on internal structure through pressure, temperature, oxygen fugacity, and composition. Composition includes not only the constituent mineral phase but also the chemistry of the phase such as iron content and minor phases such as partial melt and water. Recent laboratory measurements of mineral conductivity (see the reviews of *Yoshino* [2010] and *Karato* [2011]) have identified the sensitivity of the conductivity to all these parameters and therefore allow a precise modeling of the conductivity in terms of internal structure. For example, *Khan and Shankland* [2012] have recently

©2015. American Geophysical Union. All Rights Reserved. evaluated the water content in the Earth's mantle from Bayesian inversion of electromagnetic induction data recorded at geomagnetic observatories distributed across the globe using laboratory-based conductivity profiles.

In this paper we derive a 1-D electrical conductivity profile from electromagnetic induction theory based on 10 months of Swarm satellite measurements using a Bayesian approach. These profiles are interpreted in terms of temperature variations using laboratory-based electrical conductivity of minerals.

2. Method

The application of the EM induction theory to remote magnetic field measurements can be complex, and we introduce some classical simplifying assumptions (see, for instance, Kuvshinov, [2012], for a recent review). First, the low Earth-orbiting Swarm satellites are assumed to fly in source-free regions where the magnetic field **B** derives from the potential V through $\mathbf{B} = -\nabla V$. The magnetic potential V in space, being the solution of Laplace's equation, can be expanded in terms of internal and external spherical harmonic (SH) functions. The EM method for probing the mantle requires the contributions of the external magnetic field and of their induced counterparts to be isolated from other fields such as the core and the lithospheric fields. The magnetic field measurements have thus to be corrected for a model describing the static internal magnetic fields and their temporal variations. After these corrections, the magnetic field residuals are expected to contain only the externally inducing and the internally induced parts. Then the major source of external field is assumed to be produced by the ring current in the Earth's magnetosphere for periods longer than 1 day. This source is far enough from the measurements to be considered as large scale and mostly dipolar but close enough to the Earth to further assume that the electromagnetic wave is stationary for these periods. This 1 day lower bound for the period also allows us to mitigate the induction effects from the Earth's ionospheric field that is prominent at shorter periods and smaller spatial wavelengths [e.g., Friis-Christensen et al., 2006; Olsen et al., 2006]. Under these approximations the magnetospheric potential can be written in the space frequency domain as

$$V(\mathbf{r},\omega) = a \sum_{n=1}^{\infty} \sum_{m=-n}^{n} \left[\varepsilon_n^m(\omega) \left(\frac{r}{a}\right)^n + \iota_n^m(\omega) \left(\frac{a}{r}\right)^{n+1} \right] P_n^{|m|}(\cos\theta) e^{im\phi},\tag{1}$$

where ω is the angular frequency, a the Earth's reference radius equal to 6371.2 km, and ε and ι are the external (inducing) and internal (induced) Gauss coefficients, respectively. $P_n^m(\cos\theta)$ are the associated Schmidt-normalized Legendre function, of degree n and order m, and r the vector position where r, θ , and ϕ are radius, colatitude, and longitude, respectively. A classical procedure to infer the 1-D electrical conductivity of the mantle is to estimate geomagnetic response functions. These functions are defined as the ratio, in the frequency domain, between different observed electromagnetic components [e.g., Olsen, 1999b]. The response function depends only on the degree n, assuming that the Earth's mantle is spherically symmetric. We further consider that the geometry of the inducing source is dominated by the SH degree 1 and order 0 of the ring current. Once external and internal potentials are known for several frequencies, it becomes possible to construct a model in depth of the electrical conductivity in the Earth's interior [e.g., Schmucker, 1985]. We use a forward problem that estimates the internal response $l_{n, \text{model}}^{m}(\omega)$ of a conductive medium induced by a unitary source. We further use the linearity of the transfer function Q (i.e., $\iota = Q.\varepsilon$) to compare $l_{n, \text{ model}}^{m}(\omega)$ and $l_{n, \text{ observed}}^{m}(\omega)$ [Tarits and Mandea, 2010; Civet and Tarits, 2013, 2014]. The algorithm assumes a spherical semi-infinite medium of high electrical conductivity at the core mantle boundary and goes upward through successive conductive layers to estimate the induced response in terms of internal potentials for the considered frequencies ω .

3. Data Selection and Processing

3.1. Selection of the Swarm Measurements

The magnetic field measurements of the three Swarm satellites are considered from 26 November 2013 to 27 September 2014 (ESA L1 product, baseline 03). We select only the latest or the reprocessed Swarm data. We screen their quality flags defined in the Level1b Product Definition Document [*Tøffner-Clausen*, 2013]. We reject all Absolute Scalar Magnetometer (ASM) measurements corresponding to the flag quality code 255 and keep only the Vector Field Magnetometer (VFM) measurements identified as being in nominal mode. We also exclude the measurements made during satellite maneuvers which induce artificial time-varying magnetic fields. We carried out a preliminary comparison between the selected data and a candidate to the eleventh

generation of the International Geomagnetic Reference Field model [*Thébault et al.*, 2010]. This allowed us to identify and remove large outliers remaining in the data sets for specific days (25 March, 26 March, 8 April, 11 September, and 12 September).

3.2. Correction for the Main and Lithospheric Fields

We correct the raw vector measurements (\mathbf{B}_{raw}) with a dedicated main field model (\mathbf{B}_m) built from a subset of the entire Swarm magnetic measurements. This model is computed using the following approach. We first subsample the data set every 10 s and separate the scalar and vector data into midlatitudes (between -52° and 52° magnetic latitudes) and polar regions (for magnetic latitudes larger than $|52^{\circ}|$). Vector data in the polar regions are discarded. All scalar and vector data at midlatitudes are selected for 22:00–5:00 local time in order to minimize the contributions from the ionospheric field. In contrast the scalar data are used in the polar regions at all local times in the dark side (Sun at least 2° below the horizon). A further selection is based on the provisional *Dst* (Disturbance storm time) and *ap* (index of the auroral geomagnetic activity). The *Dst* index and its time variation over the three previous hours is requested to be lower than |5| nT, and the *ap* index, which measures the general magnetic activity at the planetary scale, is requested to satisfy $ap \le 27$ after having met the same requirement over the previous 3 h. All selected data correspond to $Kp \le 2^{\circ}$ (Kp is a 3 h planetary index of the geomagnetic activity).

The resulting subset of measurements is then inverted in terms of spherical harmonics up to degree 40 for the internal part with a linear secular variation up to degree 13. We coestimate the static external magnetic field up to degree 2 with the degree 1 parameterized by the provisional Dst index. The inversion is carried out using a robust iteratively reweighted least squares (IRLS) inversion scheme with Huber weighting to minimize the effect of outliers. This field model **B**_m is then used to correct the entire selected vector measurements. The scalar measurements are now no longer considered. Then the lithospheric field model (B_i) of Lesur et al. [2010] is used to correct for the lithospheric field at higher spatial resolution, between degrees 41 and 80. This step is important to reduce the leakage of the lithospheric field along the satellite orbits [Thébault et al., 2012]. A visual inspection of the corrected data \mathbf{B} (with $\mathbf{B} = \mathbf{B}_{raw} - \mathbf{B}_m - \mathbf{B}_l$) allows us to identify measurements with remaining suspicious behavior. We identify the measurements that differ by more than 3.5 times the standard deviation of the residual field at midlatitudes. When outliers are identified, the entire day is removed from our database, including in the polar areas. Figure 1 shows that before this processing, measurements from satellites A and C contained aberrant values that need to be filtered out. The ionospheric field and, in particular, the equatorial electrojet in the latitude range $[-20^\circ, 20^\circ]$, can be seen in the right part of Figure 1. Finally, only data for which the Sun is 6° above the horizon are selected. This limit is a good compromise between the global data coverage and the reduction of ionospheric disturbances. One side effect of selecting mostly sunlit data is to introduce gaps in the time series for each satellite data set lasting from a few hours to a few days.

3.3. Computation of the Electrical Response

In the selected and corrected measurements the magnetospheric field is assumed to be dominant over the ionospheric field. This is a simplification considering the complexity of separating ionospheric from magnetospheric parts. Nevertheless, we restrict ourselves to periods larger than 1 day to further reduce the effect of the ionospheric field at shorter periods [*Hutton*, 1972]. For each day we select the residual measurements of the three satellites and estimate the external magnetospheric field and its induced counterpart. We consider all three magnetic components of the magnetic field to better separate the external and the internal contributions.

The inversion is carried out in the geomagnetic dipole coordinate system. All vector data are weighted by $\sin \theta$, in order to account for the higher data density in the polar regions. The parameter estimation relies again on an IRLS algorithm using Huber weighting, and the solution was expanded to SH degree 2 to minimize the spectral leakage effects of smaller spatial scales on the degree 1 parameter. For each day we verify that the inverse problem is well conditioned (small covariance between Gauss coefficients). This guarantees that the separation between internal and external sources is numerically reliable and that the estimated Gauss coefficients are individually meaningful. We exploit this lack of correlation to directly estimate the standard deviation of the Gauss coefficients from the diagonal elements of the covariance matrix. In addition we apply a bootstrap estimation approach to investigate the fluctuation of the parameter estimation. Each day, the internal and external Gauss coefficients are estimated 1000 times from random subsets containing each 50% of the entire data set. The daily standard deviation on each Gauss coefficient estimated by this approach



Figure 1. North component of Swarm magnetic field measurements after reduction of the main and lithospheric fields, (left) before and (right) after rejection of outliers (see text for details).

is smaller than the standard deviation estimated via the parameter covariance matrix, confirming that the internal and external coefficients are robustly estimated. Finally, the effect of the mismatch between the ASM magnetic field intensity and the intensity field computed from the VFM magnetometers, which may reach a few nT peak to peak, is analyzed in the time and frequency domains (not shown). We verify that this mismatch does not significantly affect the estimation of the dipole Gauss coefficients for periods exceeding 1 day.

The daily estimations of the internal and external Gauss coefficients are hereafter noted $\tilde{t}_n^m(t)$ and $\tilde{\varepsilon}_n^m(t)$, with $\tilde{\sigma}_i(t)$ and $\tilde{\sigma}_e(t)$ their estimated standard deviation, respectively. Only the degree 1 and order 0 coefficient is used in this study. Nonetheless, we argue that more internal/external field coefficients will be resolved when longer measurement time series are available and when the satellites reach their definitive orbital configuration, thus providing robust magnetospheric field estimations at different local times.

One major advantage of the Swarm mission is that having three satellites flying reduces the longest time gap to 1 day. Five such gaps are found, when comparable studies using a single satellite may have gaps exceeding 15 days [*Civet and Tarits*, 2013]. For the internal and external coefficients of degree 1 and order 0, we fill these temporal gaps with the help of the provisional *Dst* index. The direct comparison between the *Dst* index and

the Gauss coefficients is not feasible, but a linear correlation between the Gauss coefficients and the *Dst* index can be assumed [*Langel and Estes*, 1985]. A regression analysis over the available time series leads to

$$\tilde{\iota}_{1}^{0}(t) = -2.12 - 0.14 \, Dst(t),$$

$$\tilde{\varepsilon}_{1}^{0}(t) = 12.33 - 0.53 \, Dst(t).$$
(2)

These linear relationships are used to convert the mean *Dst* index into $\tilde{\iota}_{1}^{0}(t)$ and $\tilde{\varepsilon}_{1}^{0}(t)$ for the five missing days, leading to a regular and complete time series over 305 consecutive days; for these five above mentioned days, the standard deviation of the dipole coefficients is set to 0.

The Fourier transform $\tilde{\tau}_1^0(\omega)$ and $\tilde{\varepsilon}_1^0(\omega)$ of the time series and, by linearity, their standard deviation $\tilde{\sigma}_i(\omega)$ and $\tilde{\sigma}_e(\omega)$ are computed. Each time series of 305 days was tapered to zero until the next power of 2 (512 days) in order to avoid spectral leakage and Gibbs effects. This defines a new space of frequencies ω_f . The signal-to-noise ratio for the external potential is 1–2 orders of magnitude larger than for the internal potential. From $\tilde{\varepsilon}_1^0(\omega_f)$ we then compute the estimated complex internal induced response $\tilde{\varepsilon}_1^0(\omega)$ for periods ranging between 2 and 256 days assuming that the inducing potential $\tilde{\varepsilon}_1^0(\omega_f)$ is error-free, thus considering in the following only the standard deviation $\tilde{\sigma}_i(\omega)$.

3.4. Estimation of the Vertical Mantle Electrical Conductivity Profile

Since our algorithm computes the internal response from an inducing source and because we can neglect the errors on the inducing source, we impose that conductivity parameters σ should minimize

$$\chi^{2} = \sum_{f=1}^{N_{f}} \left[\log \left(\left| \frac{\tilde{l}_{1}^{0}\left(\omega_{f}\right) - l_{1}^{0}\left(\omega_{f}\right)}{\tilde{\sigma}_{i}(\omega_{f})} \right| \right) \right]^{2} + \alpha \sum_{l=1}^{L} \log \left(\frac{\sigma_{l+1}}{\sigma_{l}} \right) \,. \tag{3}$$

where N_f is the number of frequencies used and L is the number of layers. This inverse problem is nonlinear and generally ill-posed, especially when dealing with noisy measurements that are only available for a restricted time period. We designed a Markov chain Monte Carlo (McMC) algorithm to solve equation (3) to circumvent some of the difficulties arising from such a nonlinear ill-posed inverse problem relying on a poorly conditioned inverse matrix. Within the Bayesian framework, the inverse problem consists in estimating marginal probability distributions for the conductivity. It is well known [e.g., Tarantola, 1987, chapter 2] that this requires an extensive exploration of the model space. An elegant way to overcome this difficulty is to construct a Markov chain during which model parameters are randomly updated at each iteration. In the McMC algorithm, parameter values decreasing the misfit function have better chance to be reselected at the next iteration than values increasing it. The second term of equation (3) is designed to smooth the conductivity contrasts between two consecutive layers with a damping parameter α . We choose a value of α so that the smoothing constraint does not exceed about 2% of the total cost function on average, i.e., the conductivity estimates are not entirely determined by the smoothing constraint. The posterior marginal probability distribution is explored through a Markov process based upon a Gibbs's sampler of the conductivity values (see, for instance, Schott et al. [1999], for a detailed algorithm). We consider mantle conductivity values lying between 10^{-4} and 10^3 S m⁻¹ and divide this interval into 50 cells equally spaced in a logarithmic scale. We do not a priori impose the number and thickness of the vertical layers, as Constable [1993] highlights the danger of the a priori layered approach that results in oscillatory solutions. Instead, the algorithm arbitrarily starts with four layers of 500 km thickness between the surface and 2000 km depth, and iteratively refines the vertical discretization of the conductivity profile in the following way. After convergence of the Markov chain for the initial problem with four layers, the algorithm identifies the maximum likelihood estimated values of the conductivity for each layer. When the difference of conductivity between two consecutive layers exceeds half an order of magnitude, the algorithm considers that there is a discontinuity. It, therefore, divides the lowermost layer into two thinner layers. The thickness of the upper one of these two is rounded to the nearest 50 km (maximum vertical resolution). As a result, the lower part of the divided layer is always as thick as or thinner than the upper one, and no layer is thinner than 50 km. A new Markov chain inversion is then run for the problem involving this new distribution of layers. This procedure is designed to obtain a conductivity model whose vertical complexity is determined by the data rather than by the a priori model of the mantle stratification. At the end of the iterative process, when no discontinuity is found for layers thicker than 50 km, the algorithm returns the probability density function (pdf) with the maximum likelihood of the electrical conductivity for 14 layers of varying thickness. The vertical resolution is better in the 500 km to 1000 km depth range thus



Figure 2. (left) Probability density function (pdf) of electrical conductivity profile obtained from McMC inversion of Swarm L1b data. The maximum likelihood (red) and the mean value (plain blue) with 1 standard deviation error bar (dashed blue) are also represented. (right) Maximum likelihood of the McMC pdf (red) compared to previous studies [Semenov and Jozwiak, 1999; Olsen, 1999a, 1999b; Constable and Constable, 2004; Kuvshinov and Olsen, 2006; Civet and Tarits, 2013].

showing the depths at which the conductivity values are better constrained by the measurements. The normalized root-mean-square between the observed and the maximum likelihood induced Fourier coefficients after convergence of the McMC is 1.43 and their correlation is 0.90. This gives us confidence that the model is statistically significant.

4. Results and Discussion

We show in Figure 2 the pdf of the electrical conductivity profile derived from the Bayesian inversion of internally induced potentials. The pdf maximum likelihood, its mean, and its associated 1 standard deviation interval, are also displayed, as well as profiles of previous studies. As the pdf is not symmetrically distributed, small discrepancies appear between the mean and the maximum likelihood. In the whole depth range, the one sigma uncertainty is of the order of 0.5 log unit with a somewhat larger uncertainty between 0 and 500 km depth, because we do not consider periods shorter than 2 days.

The pdf of the electrical conductivity profile is characterized by an increase from 0.001 S m⁻¹ at 400 km depth to \simeq 1 S m⁻¹ at 900 km depth. This 3 orders of magnitude increase may be related to the mineralogical transformations of upper mantle minerals into their lower mantle phases [e.g., *Xu et al.*, 2000]. For depths larger than 900 km, the electrical conductivity shows a small increase from \simeq 1 S m⁻¹ at 900 km depth to \simeq 4 S m⁻¹ at 2000 km depth. As no mineralogical transformation occurs in this depth range, the small conductivity increase can be directly related to pressure increase and temperature variation. Compared to previous studies by *Semenov and Jozwiak* [1999], *Olsen*, [1999a, 1999b], *Kuvshinov and Olsen* [2006], and *Civet and Tarits* [2013], we observe that our results are characterized by a much smaller conductivity in the upper mantle and the transition zone. For depths larger than 800 km, our results show a very good agreement with the results of *Olsen* [1999a, 1999b] and *Kuvshinov and Olsen* [2006]. Below 1000 km there is a disagreement with the results of *Constable and Constable* [2004], which is possibly due to the misuse of the multitaper approach as pointed by *Kuvshinov and Olsen* [2006].

We tentatively interpret this electrical conductivity profile in terms of temperature variations with depth. We follow the modeling of *Khan and Shankland* [2012] who computes the electrical conductivity of a rock aggregate as the effective medium mean of individual mineral phase conductivity σ_i of the form

$$\sigma_i = \sigma_0^i \exp\left(\frac{-H_i}{kT}\right) \tag{4}$$

where *T* is temperature and *k* is Boltzmann constant. The preexponential factor σ_0^i and the activation enthalpy H_i depend on the composition of mineral *i* and to a lesser extent on pressure and oxygen fugacity. Two



Figure 3. Temperature profiles associated with maximum likelihood and mean conductivity values of Figure 2. Temperature profiles in red and dark green are obtained with the electrical conductivity database of *Karato* [2011] (KD), whereas light green and orange profiles are obtained with the database of *Yoshino* [2010] (YK). Geotherm of *Stacey and Davis* [2008], along with the adiabatic temperature profile of *Katsura et al.* [2010] are shown for comparison.

different databases are used for σ_0^i and H_i parameters: YK (Yoshino, Katsura, and coworkers [Yoshino, 2010]) and KD (Karato, Dai, and coworkers [Karato, 2011]). The YK database is supplemented with the more recent results of Zhang et al. [2012] in order to model the conductivity of hydrous pyroxene along with corrected values for the conductivity of hydrous wadsleyite [Yoshino and Katsura, 2012]. These two databases are different for the upper mantle but agree for lower mantle minerals for which the measurements of Xu et al. [2000] are used in both cases.

Although oxygen fugacity correction is included in the two databases, assumptions on composition and pressure profile are required in order to interpret conductivity in terms of temperature only. The mantle mineralogy is assumed to be given by a pyrolitic model [*Irifune*, 1987]. The mineralogical transformations are computed at each pressure/temperature conditions according to phase diagrams discussed in *Vacher et al.* [1998] for dry mineralogy. Water content is assumed to

be equal to 0.01% in the upper mantle and 0.1% in the mantle transition zone, in agreement with both geochemical analysis of basalts and study of geophysically inferred electrical conductivity of Earth's interior [*Karato*, 2011]. Pressure is assumed to be identical to preliminary reference Earth model [*Dziewonski and Anderson*, 1981] values. These assumptions on composition and pressure allow us to isolate the temperature effect on conductivity and thus to derive a temperature profile from conductivity through standard numerical root search.

We show in Figure 3 the interpretation in terms of temperature variations of the maximum likelihood and mean estimators of electrical conductivity pdf of Figure 2 calculated using the KD and YK mineralogical databases sampled at the center of each layer. Two recent geotherms [*Stacey and Davis*, 2008; *Katsura et al.*, 2010] are also represented. Given the close agreement between the maximum likelihood and the mean electrical conductivity in the upper mantle, the variations of the temperature deduced from these estimators are nearly identical in this depth range for a given model. The choice of the database used to constrain the conductivity of minerals, however, induces large differences in the temperature profile in the upper mantle and the transition zone (Figure 3). This is largely due to the discrepancies that exist between laboratory measurements of hydrous minerals performed by the two research groups.

In the lower mantle both databases are identical and share the modeling of *Xu et al.* [2000] for lower mantle phases. In the 1000–1500 km depth range, the computed temperature profiles are close to the geotherms of *Stacey and Davis* [2008] and *Katsura et al.* [2010] and are characterized by an adiabatic gradient of 0.3 K/km. For depths greater than 1500 km, the mean and maximum likelihood estimators become significantly different, highlighting the decrease in resolution in the pdf at such depths.

Negative temperature gradients are observed in the upper mantle and transition zone. One is found 50 km below the 670 km deep transition zone. High-pressure mineral phases are more conductive than their low-pressure ones, and conductivity increase is expected. As the interpretation of conductivity is done in terms of temperature only, an increase in conductivity smaller than the one predicted by laboratory-based measurements and pyrolitic model is falsely interpreted as a temperature decrease. Instead, it should be interpreted as inaccurate estimate of the depth at which the conductivity increases or as erroneous assumptions on composition of the a priori pyrolitic model or the chosen water content. Two other negative gradients are observed in the uppermost and lowermost parts of the profile where uncertainties on the conductivity

model are larger (Figure 2). Joint inversion of Swarm data supplemented with complementary data (such as seismological data) are required to go beyond the simple conductivity to temperature interpretation in order to discriminate between temperature and composition effects.

5. Conclusions

In this study we use magnetic field measurements acquired during the first 10 months of the ESA Swarm mission to estimate the ratio of the externally inducing to the internally induced magnetic field and to derive a 1-D electrical conductivity profile to a depth of 2000 km. We start from the raw L1b Swarm magnetic field measurements of the three satellite vector and scalar magnetometers. We build a model for the main internal magnetic field and its secular variation to highlight the magnetospheric external field and its induced part and then estimate the electrical conductivity in the Earth's mantle. This, we think, warrants a good control of the entire processing scheme that allows detecting electrical conductivity discontinuities. The 1-D conductivity profile is then interpreted in terms of temperature variations, and we obtain a temperature gradient in the lower mantle which is close to an adiabatic one.

We note that these results have previously been obtained by *Khan and Shankland* [2012] from observatory data and relying on thermodynamical modeling to identify the phase transitions. Using less than 1 year of satellite measurements is not sufficient to derive a definitive temperature profile for the Earth's mantle. We, nevertheless, show that the Swarm mission already provides consistent and promising results in the lower mantle and has also the ability to challenge our current view of temperature and composition in the upper mantle and transition zone. This study will be pursued as more Swarm measurements become available. We also have to acknowledge that discrepancies between laboratory measurements of hydrous iron-bearing minerals have to be reconciled too. In order to use the whole information contained in the pdf of the conductivity, a further step will be to perform a Bayesian inversion of the Swarm data directly in terms of temperature, along with composition if complementary data or priors are used [e.g., *Verhoeven et al.*, 2009; *Khan and Shankland*, 2012]. It is also hoped that additional measurements acquired during the next years of the Swarm mission will considerably improve these results. First, the number of available measurements will provide narrower estimates of the electrical conductivity pdf function. Second, longer time periods will allow deeper conductivity estimates. The final orbital configuration will also permit a better estimation of the magnetospheric fields at shorter periods and thus to investigate the conductivity at shallower depths.

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