A semi-global reference model for electrical conductivity in the mid-mantle beneath the north Pacific region

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[1] One-dimensional electrical conductivity structure in the mid-mantle of the one-fourth of the Earth beneath the north Pacific Ocean was obtained by a semi-global electromagnetic induction study. Electromagnetic response functions estimated from electric field variations measured by submarine cables and geomagnetic field variations obtained by magnetic observatories and long-term observations sites were inverted into radially symmetric conductivity distribution by taking the distribution of land and ocean at the surface into account. As a most preferred model, a smooth conductivity-depth profile was obtained with two abrupt increases that possibly correspond to the seismic discontinuities at 410 and 660 km. *INDEX TERMS:* 1515 Geomagnetism and Paleomagnetism: Geomagnetic induction; 3005 Marine Geology and Geophysics: Geomagnetism (1550); 3210 Mathematical Geophysics: Modeling; 5109 Physical Properties of Rocks: Magnetic and electrical properties; 9355 Information Related to Geographic Region: Pacific Ocean. **Citation:** Utada, H., T. Koyama, H. Shimizu, and A. D. Chave, A semi-global reference model for electrical conductivity in the mid-mantle beneath the north Pacific region, *Geophys. Res. Lett.*, *30*(4), 1194, doi:10.1029/2002GL016092, 2003.

1. Introduction

[2] Electrical conductivity is a material property that is strongly dependent on physical, chemical, and thermal conditions. All of these conditions in the Earth are strongly depth dependent and therefore Earth can be regarded as horizontally stratified (or radially symmetric) to the first order approximation. There exist horizontally stratified models of the Earth that can be used as a reference in seismology [e.g., *Dziewonsky and Anderson*, 1981]. Electrical conductivity, on the other hand, is known to be very heterogeneous especially at crustal and upper mantle depths. However, attempts to estimate such a reference model for the deeper mantle might well be justified, as its conductivity is high compared to its possible spatial variability.

[3] Radially symmetric conductivity models through the mid-mantle have been obtained from local to regional electromagnetic response functions [e.g., *Neal et al.*, 2000; *Ichiki et al.*, 2001] when they are shown to be consistent with a one-dimensional (1-D) model using the D+ criterion [*Parker*, 1980]. Lateral variation of these local 1-D structures then serve as a first approximation to the real lateral variability of the Earth. However, quantitative understanding of the heterogeneity through such an approach is rather difficult because lateral and radial (or vertical) variations of conductivity are not properly separated [*Schultz and Larsen*, 1990].

[4] There also have been a number of global induction studies to determine a 1-D mid-mantle conductivity model by using globally distributed observations. However, the 1-D global models obtained in past works [e.g., *Banks*, 1969; *Achache et al.*, 1981; *Olsen*, 1998] show differences of conductivity, which are too large to be accounted for only by different combinations of sites. Proper modeling of the large surficial ocean-land conductivity contrast is also essential to obtain a reliable 1-D model, as shown below.

[5] We try to separate the real 3-D conductivity distribution into a radially symmetric part and lateral variability around this profile, as has been done in seismic tomography. This paper is intended to provide the first step in this approach by using a recently obtained dataset from the Pacific region and its surroundings to obtain reliable estimate of the semi-global 1-D electrical conductivity structure in the mid-mantle that can be used as a reference.

2. Data and Inversion Scheme

[6] We used three component geomagnetic field variations observed at standard observatories and long term observation sites operated by *INTERMAGNET* [1994], the Japan Meteorological Agency or the Ocean Hemisphere Project [OHP; *Shimizu and Utada*, 1999], together with electric potential variations observed by using abandoned commercial submarine telecommunications cables obtained by Bell Laboratories [e.g. *Lanzerotti et al.*, 1992; *Fujii et al.*, 1995] and OHP. The locations of the geomagnetic observatories and cables are shown in Figure 1. These observation sites cover nearly one fourth of the Earth's surface, although their distribution may be too sparse for the quantitative study of lateral heterogeneity in the upper mantle.

[7] These geomagnetic and geoelectric data were transformed into either geomagnetic depth sounding (GDS) or magnetotelluric (MT) response functions in the frequency domain at each station or cable by applying a bounded influence method [*Chave and Thomson*, 2002]. In addition to response functions thus obtained, we used GDS response functions in the study area recently estimated by *Fujii and Schultz* [2002].

[8] Response functions at periods longer than a day should be incorporated in order to explore deeper into the mantle. At such long periods, the observed responses are functions not only of the electrical conductivity structure and the frequency but also of the morphology of the external source field producing them. We found that the source field of geomagnetic variations can be well approxi-

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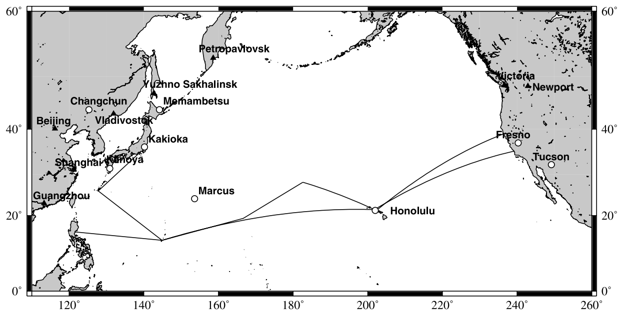

Figure 1. Locations of geomagnetic observatories and submarine cables. Circles and triangles denote the locations of the geomagnetic stations whose data are analyzed in this study and by *Fujii and Schultz* [2002], respectively. Lines denote the locations of submarine cables.

mated by a P_1^0 spherical harmonic over this period band [*Banks and Ainsworth*, 1992; *Fujii and Schultz*, 2002]. Considering also the accuracy of response function estimates, we selected GDS responses for the period band 5–27 days and MT responses for 1–8 days, both at 8 discrete periods at each station or for each cable.

[9] An optimum 1-D conductivity model that accounts for all response functions was then sought using an inverse method that minimizes

$$\Phi(m) = \mathrm{S}\left(Zobs\left(f_i, r_\mathrm{j}\right) - Zcal\left(f_i, r_\mathrm{j}; m\right)\right)^2 / var\left(f_i, r_\mathrm{j}\right) + \lambda g(m), \quad (1)$$

where f_i and r_j denote the frequency and the location of observation site, m is a set of model parameters (the electrical conductivity in each layer), *Zobs* and *Zcal* are observed and calculated response functions, *var* is the variance of the estimated response, and λ and g are a hyper-parameter and mathematical constraint on the model parameters, respectively.

[10] The procedure shown above, however, is not as simple as what is used to interpret local or global EM responses. Previous works assumed that the entire earth is radially symmetric including its surface where there is a large lateral conductivity contrast at the land-ocean interface. The effect of this heterogeneity is quite strong, particularly at the coastal geomagnetic stations, and cannot be ignored [*Kuvshinov et al.*, 1999; 2002]. The lateral variations of water depth are well known and dominant as seawater conductivity is much higher than in the rest of the structure. In this paper, the actual distribution of land and ocean is taken into account to estimate subsurface 1-D structure. Thus, the problem becomes essentially 3-D and requires a solver for the EM induction equation to obtain Zcal in a laterally heterogeneous sphere. We used a forward solver developed by *Koyama et al.* [2002] which is based on the modified iterative dissipative method (MIDM) in the frequency domain [*Singer*, 1995]. Water depth was given on a surface grid by using the ETOPO5 database with a grid spacing of 2 degrees. A uniform seawater conductivity of 3.0 S/m was assumed.

[11] In the inversion, the earth material except seawater was assumed to be stratified with a series of 50 km thick layers down to a depth of 1000 km, and the lowermost layer was assumed to be a uniform geocentric sphere. The conductivity of each layer is the parameter inverted for so that the number of unknown parameters is 21, while the total number of complex response functions to be inverted is 317.

[12] Inversion to find appropriate m in equation (1) poses convergence problems starting from a set of arbitrary initial values. In this study, an initial guess for a set of model parameters was estimated as follows. (1) The effect of surface heterogeneity at each station or cable is corrected for. The effect was approximated by

$$\delta Z\left(f_i, r_j\right) = Z_{3D}\left(f_i, r_j\right) - Z_{1D}\left(f_i\right) \quad (2)$$

where Z_{3D} and Z_{1D} are response functions calculated for laterally heterogeneous (where the lateral heterogeneity is confined to the surface) and homogeneous models, respectively. The effect can be corrected for each observed response function as,

$$Zcor\left(f_i, r_j\right) = Zobs\left(f_i, r_j\right) - \delta Z\left(f_i, r_j\right) \quad (3)$$

where *Zcor* is a corrected response function. (2) All corrected response functions are spatially averaged to have an equivalent single station magnetotelluric impedance. (3) This averaged dataset was inverted for a 1-D conductivity model by the Occam method [e.g. *Constable et al.*, 1987]. Constraints on spatial smoothness of the structure and the possible locations of conductivity jumps were applied during the inversion. (4) The 1-D model is used to improve the correction of the effect of surface heterogeneity. Procedure (1)–(4) was iterated until the model parameters converged. The resulting model was used as an initial guess to obtain the final result by minimizing $\Phi(m)$ in equation (1).

[13] We used an iterative inversion scheme (a combination of the steepest descent and the quasi-Newton methods), in which the total misfit was minimized between the observed responses and those calculated for all available

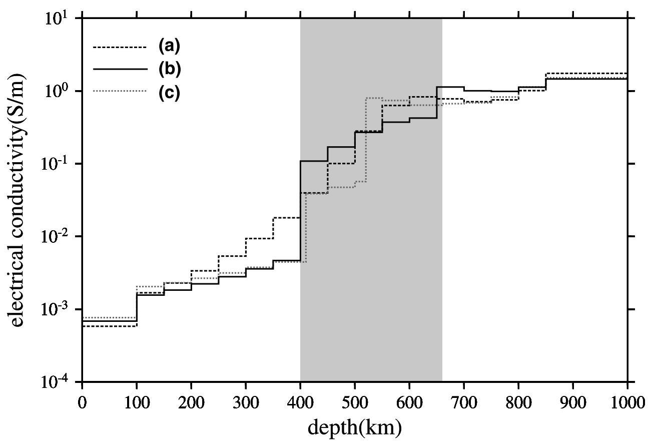

Figure 2. Conductivity profiles obtained from the inversion algorithm described in the text. (a) unconstrained, (b) constrained to have two jumps at 400 and 660 km depths, (c) constrained to have three jumps at 400, 550 and 650km depths.

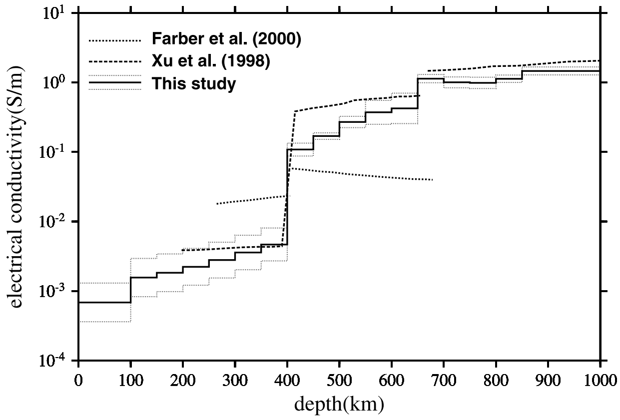

Figure 3. Comparison of the two-jump model and the results of laboratory experiments by *Farber et al.* [2000] and *Xu et al.* [1998].

magnetic stations and submarine cables jointly, to obtain final 1-D models.

3. Result

[14] First, we carried out the inversion by using an initial guess with a smoothness constraint throughout the depth range of interest. This resulted in a conductivity model smoothly increasing with depth between 200 and 600 km (Figure 2a). Conductivity increases the most rapidly around the transition zone (shown by the shaded area), but becomes almost flat at about 1 S/m in the lower mantle. The structure also suggests that the upper mantle is much less conducting, although the present dataset covers a period band of 1–27 days and hence does not have much resolution over this depth range.

[15] In the mantle transition zone, mineral phase changes occur from the upper-mantle olivine (α-phase) to wadsleyite (β-phase) at about 410km depth and from wadsleyite to ringwoodite (γ-phase) at about 520km depth [*Xu et al.*, 1998]. In the lower mantle below 660km depth, silicate perovskite or magnesiowüstite are considered to be the major constituents. Therefore there could be discontinuities in any material properties across these boundaries.

[16] Although seismological methods can detect these discontinuities in terms of elastic constants, electromagnetic methods intrinsically do not have sufficient resolving ability to detect the position of such boundaries and conductivity contrasts across them at the same time as accurately as seismological methods. Therefore we tried to invert the EM responses by giving a priori information about the depths of possible discontinuities. As a priori information, we reduced the smoothness constraint in the inversion over certain depths ranges by 30% so that it allows the presence of discontinuities in conductivity. Two cases were examined: (1) boundaries at 410 and 660km depths only are considered (Figure 2b), and (2) all three boundaries are taken into account (Figure 2c). Error bar of the conductivity of each layer was estimated for the inverted model by the jack-knife method [*Efron and Tibshirani*, 1993]. Note that these error bars are linear approximations with concomitant limitations.

4. Discussion

[17] The solutions shown in Figure 2 are statistically equivalent with the same misfit level (χ^2 = 1328.8, 1332.4, 1348.7, respectively). However, considering the presence of phase transitions in the actual Earth, we should reject Figure 2a which gives smooth variation of the electrical conductivity throughout the depth range of present interest (the mid mantle depth). A little more detailed

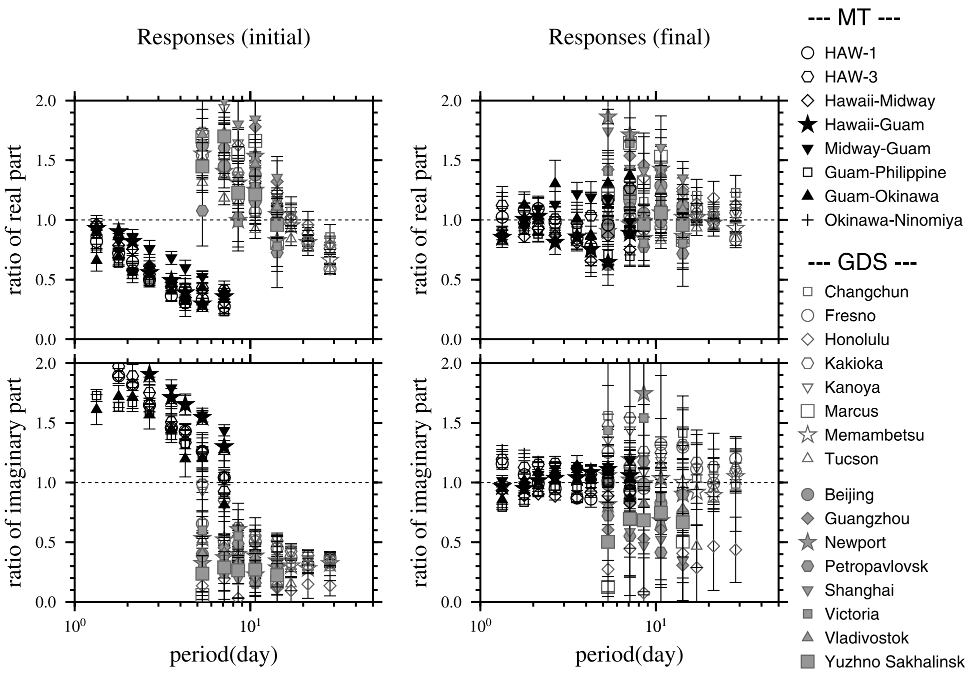

Figure 4. Real and imaginary parts of the observed response functions normalized by responses calculated for the initial (left) and final (right) conductivity model. MT (upper) and GDS (lower) responses are presented separately.

discussion is necessary to choose a physically plausible model among these models.

[18] Recent studies [*Shankland et al.*, 1993; *Xu et al.*, 1998; *Katsura et al.*, 1998; *Farber et al.*, 2000; *Dobson and Brodholt*, 2000] point out that the electrical conductivity at depths within the mantle transition zone or the lower mantle are mostly governed by the phase change of mineral constituents, temperature, and chemical composition. Most of these studies suggest the presence of abrupt increases in electrical conductivity at 400 km depth, corresponding to the phase transition from olivine to wadsleyite, and at 660 km depth, corresponding to the boundary of upper and lower mantle materials. In contrast to these two structural changes, these works predict only a small conductivity jump at 520 km depth corresponding to the phase transition from wadsleyite to ringwoodite. In Figure 3, two recent laboratory results are plotted for comparison. There is remarkable agreement between the conductivity profiles from our second solution and that given by *Xu et al.* [1998], especially in the size of the conductivity jumps at the discontinuities. If the transition at 520 km does not cause a significant change in conductivity, we may conclude that our second solution shown in Figure 3 is geophysically the most plausible.

[19] Figure 4 exhibits the fit of real and imaginary parts of EM responses calculated from the initial (uniform mantle with 0.1 S/m) and final one-dimensional conductivity models to the observed ones. Variance reduction is obvious, but there remains a scatter around the final model responses. The scatter may be due to a breakdown of the P_1^0 assumption for the source field geometry at periods of 1–5 days, but the presence of systematic spatial variation in the response functions rather than random scatter probably indicates the effect of lateral heterogeneity. In fact, the D+ test on EM responses used in this study strongly suggests the presence of significant lateral heterogeneity: i.e., although responses of each station or cable are individually consistent with a 1-D model, the inability to jointly invert all response functions indicates inconsistency with a 1-D model. Therefore we further aim to delineate the lateral heterogeneity so that observation-calculation misfit is minimized [*Koyama et al.*, 2002, in preparation].

[20] This paper introduced a new approach to the study of electrical conductivity in the mid-mantle in which a 1-D model was obtained by analyzing semi-global electromagnetic induction in the presence of the ocean. The resulting 1-D model can be used as a reference electrical conductivity model for one fourth of the Earth's mantle beneath the north Pacific region. Further efforts to increase observation sites and improvement of both data quality and quantity are still strongly required.

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