

The Electrical Conductivity and Thermal Profile of the Earth's mid-Mantle.

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Abstract. Electrical conductivity in the Earth's mantle is sensitive to temperature and chemical environment. Recent laboratory measurements of electrical conductivity are combined with candidate mantle geotherms to produce synthetic electrical conductivity profiles. These profiles are used to forward model the Earth's geomagnetic response function C , results of which are compared with the observed globally averaged response function at periods of 3.5 days to 4 months. Candidate lower mantle geotherms, representing whole-mantle and layered convection end-members, are compared using published electrical conductivity measurements on alumina-bearing and alumina-free perovskite in the conductivity models. Comparison of the predicted response functions with the observed geomagnetic response of the Earth shows that a) if lower mantle alumina is incorporated into perovskite, then the lower mantle must be cool, and b) if the alumina is not incorporated in perovskite then the results are only consistent with a hot lower mantle. In addition, the maximum alumina content of lower mantle $MgSiO_3$ perovskite is constrained at 4%.

Introduction

Temperature is fundamental in controlling the physical properties of the Earth's mantle such as viscosity, density, electrical conductivity, convection and melting (Fowler, 1990; Ahrens, 1995). Thermal profiles of the upper mantle are well constrained from geothermometry (Mercier, 1980) and the temperature of the olivine-spinel-post spinel transitions (Ito and Katsura, 1989). Lower mantle temperatures are poorly known, however, with models of layered convection showing a mid-mantle thermal boundary, of 1000 K or more, which is not present in whole mantle models (Jeanloz and Knittle, 1989; Glatzmaier and Schubert, 1993). Such widely varying geotherms have vastly different implications for Earth processes and properties. Electrical conductivity could, in principle, be used to distinguish between different mantle geotherms, provided a) geophysically measured response functions have sufficient depth resolution and b) high quality laboratory conductivity measurements for the relevant minerals exist.

Fluctuations in the magnetic field external to the Earth induce electric currents within the Earth's mantle which give rise to an internal component to the geomagnetic field variations. These field variations can be measured and inverted to give an electrical conductivity profile of the Earth. Field variations with periods of a few months are sensitive to

lower mantle conductivity and have a relatively good signal to noise ratio (Constable, 1993), however, solutions to such inversions are non-unique (Parker, 1970), and we probably cannot say more than conductivity of the top of the lower mantle is around 1S/m (Egbert and Booker, 1992; Schultz et al, 1993; Lizaralde et al, 1995), which is consistent with measurements of perovskite conductivity. An alternative approach is to forward model the global geomagnetic response based on estimated geotherms and measured mineral electrical conductivities. The high sensitivity of electrical conductivity to temperature allows us to test the geotherms by comparing the predicted response functions with those measured for the Earth. We will show that, despite the present uncertainties in the laboratory measurements, the results of such forward modeling are robust and can discriminate between different lower mantle geotherms.

Geotherms

Upper mantle temperatures are well constrained by geothermometry of mantle xenoliths (Mercier, 1980) (to 150 km) and the temperatures of the olivine-wadsleyite (1700 K at 410 km) and the post-spinel transitions (1873 K at 670 km) (Ito and Katsura, 1989). Lower mantle geotherms are less well constrained. The geotherm of Brown & Shankland (1981) represents a lower limit for the lower mantle, being an adiabatic extrapolation of upper mantle temperatures. Spiliopoulos & Stacey (1984) derived a geotherm from their estimate of the core-mantle boundary temperature of 3800 K. Their adiabatic extrapolation upwards from a superadiabatic boundary of 800 K at the base of the lower mantle resulted in a thermal boundary of 430 K at 670 km. Both Brown and Shankland's and Spiliopoulos and Stacey's geotherms are consistent with whole mantle convection models since layered models require a large thermal boundary of about 1000 K at 670 km (e.g.; Glatzmaier and Schubert, 1993). In order to test the effect of a lower mantle geotherm consistent with layered convection we have considered a lower mantle geotherm which is 1000 K hotter than Brown and Shankland's geotherm. This hot geotherm is similar to that of Anderson (1982), but does not include the differences between thermodynamic parameters used in Anderson's and Brown and Shankland's adiabatic extrapolation. We are not suggesting this as a new potential lower mantle geotherm, but rather testing the effect of a large thermal boundary layer at 670 km depth using this simple, hot, lower mantle geotherm.

Laboratory Measurements of Electrical Conductivity

We have used values for electrical conductivity given in Table 1 in our calculation of mantle electrical conductivity profiles. Values for magnesio-wüstite (Dobson and Brodholt,

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Table 1. Electrical conductivity values used in the geomagnetic forward model.

Phase	Activation enthalpy ΔU (eV)	Activation volume ΔV (cm ³ /mol)	Preexponential $\log \sigma_0$ (eV)	Temperature K	Pressure GPa	Fe/Fe+Mg
Ol*	1.87(4)	0.0(5)	3.63(14)	1273-2000	1,5,10	0.1
	1.50 (3)		2.71(20)			
Wad	0.82(4)	-	1.90(20)	873-1673	15	0.1
Wad	0.79(2)	-	1.24(8)	1273-1573	17	0.1
Pv	0.70(5)	-	1.87(11)	1650-1890	25	0.08
	0.62(4)		1.12(12)			
Mw†	0.72(5)	0.0 (5)	1.93(25)	1000-2000	5,10,15	0.045
	0.51(5)		3.25(25)			0.196

* Present study, upper and lower bounds for f_{O_2} = Ni-NiO, Fe-FeO, C-CO.

† Multi anvil press. Values of σ_0 and E are approximate and for comparison, because the arrhenius relationship does not hold.

2000), olivine and wadsleyite come from our measurements in a Walker-type multi-anvil press using the technique described in Dobson et al (1997). Our olivine conductivity measurements at oxygen fugacities buffered by graphite-CO, iron-wüstite and nickel-nickel oxide (Dobson and Brodholt, 1999) are consistent with previous studies (Xu et al, 1998a) and cover the likely oxygen fugacity of the upper mantle. The upper mantle electrical conductivities we obtain from these experimental data agree with the range obtained for the upper mantle from short-period geomagnetic field variations (e.g.;

Campbell and Schiffmacher, 1988). Our wadsleyite measurements are identical within experimental error to those of Xu et al (1998a), who also found ringwoodite conductivity to be almost identical to wadsleyite. We have used our wadsleyite value and the significantly more resistive values from Omura (1991) for the upper and lower bounds of transition zone conductivity. Xu et al (1998b) have recently published results of conductivity measurements at 25 GPa and temperatures up to 2000 K, on perovskite with and without 4 at. % Al_2O_3 . Alumina is important for perovskite electrical conductivity because the most favorable charge balancing mechanism in magnesium silicate perovskite is by incorporation of ferric iron on the cation site (McCammon, 1997; Richmond & Brodholt, 1998). The presence of 4 % Al_2O_3 causes a three-fold increase in electrical conductivity and we have used the aluminous and alumina-free electrical conductivities to represent possible end-member lower mantle values. We will show later that this assumption is robust. Previous low temperature perovskite measurements can not be extrapolated to lower mantle temperatures because of a change in activation energy above 1000 K (Katsura et al, 1998; Xu et al, 1998b).

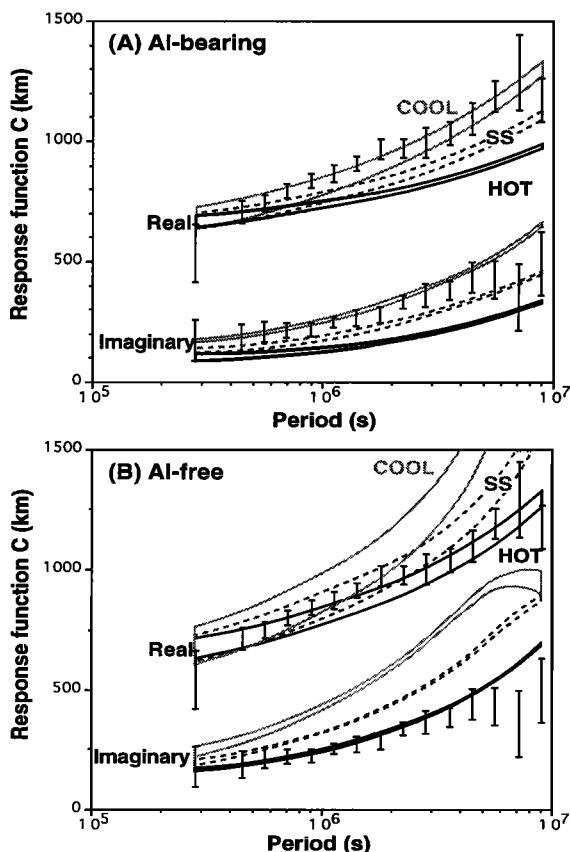


Figure 1. Predicted and global average response functions calculated for aluminous (A) and Al-free (B) perovskite. Upper curves are for the real component and lower curves denote the imaginary component of the response function. The three sets of curves were calculated for cool (Brown and Shankland, 1981; labeled COOL), intermediate (Spiliopoulos and Stacey, 1984; SS) and hot (see text; HOT) geotherms. Errors in the global average response are 95% confidence intervals.

Predicted Response Functions

The geomagnetic response given by the synthetic conductivity profiles was calculated using the 1-dimensional layered model method described in Parker (1970) and Constable (1993) (fig. 1). Measured global average response functions from Constable (1993) are plotted with two standard error bars. The upper and lower bounds for predicted response functions associated with each geotherm were generated using the most resistive and most conductive possible upper mantle and transition zone discussed above. In the case of aluminous perovskite (fig. 1a), only the cool geotherm of Brown and Shankland produces a response function compatible with the Earth's response. If, on the other hand, the alumina-free perovskite is used in generating the conductivity profiles (fig. 1b), the hot geotherm produces the only reasonable fit. In other words, the Earth's response function is only compatible with either an alumina-free perovskite in a hot lower mantle or an aluminous perovskite in a cool lower mantle. Figure 2 shows that the two possibilities are indistinguishable and that they fit the field measurements.

Discussion

The alumina and ferric iron contents of magnesium silicate perovskite have been the subject of considerable recent study. Phase equilibria experiments (Wood and Rubie, 1996) show

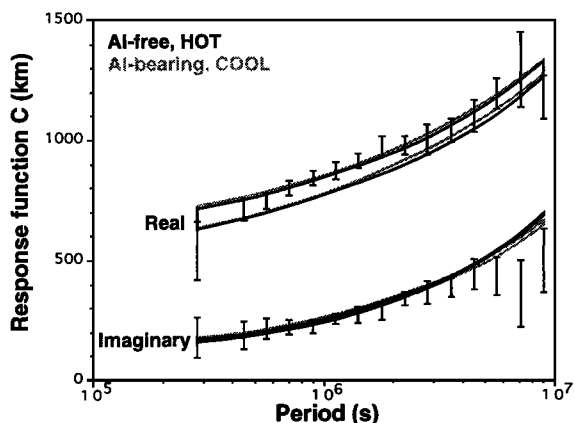


Figure 2. The best fit calculated response functions: light grey = aluminous perovskite, cool geotherm; black = Al-free perovskite, hot geotherm. The two calculated response functions are indistinguishable and fit the global average response function.

that alumina is soluble in perovskite above 25 GPa and that the presence of alumina increases iron solubility in perovskite. Richmond & Brodholt (1998) suggest that Al^{3+} is incorporated on the Si site and charge balanced by Fe^{3+} on the cation site, which is supported by Mössbauer studies of synthetic aluminous $(\text{Mg,Fe})\text{SiO}_3$ perovskite (McCammon, 1997). These studies are consistent with the high alumina and ferric iron contents of MgSiO_3 inclusions, occurring with magnesiowüstite, in deep-sourced diamonds, which probably represent a relict lower mantle assemblage (McCammon et al, 1997). Thus our current understanding of Al in the lower mantle suggests that alumina is incorporated into perovskite and that there is no aluminous phase above 25 GPa pressure. This therefore supports a cool lower mantle. Hot lower mantle models would require either a separate volumetrically minor aluminous phase, or a lower mantle with no alumina. A lower mantle with no alumina could only be achieved if it has a bulk chemistry significantly different from the upper mantle. Layered convection models lead to divergent histories and hence distinct compositions for the upper and lower mantle. The favoured chondritic lower mantle models, however, still have 3.5 % or more alumina and are rich in iron, leading to a relatively conductive perovskite-rich lower mantle. If the Earth truly has layered convection, the resulting hot geotherm would therefore require an alumina (and/or iron)-poor lower mantle composition, not previously suggested, or a new volumetrically minor aluminous phase.

There are a number of uncertainties associated with applying experimental results to predicting mantle electrical conductivity profiles (fig. 3), however these uncertainties act to increase lower mantle conductivity and thus preclude the hot geotherm.

1) Our upper mantle conductivity model includes an uncertainty of ± 0.4 log units. This encompasses the likely range in upper mantle conductivity due to uncertainties in oxygen fugacity and water content.

2) We have used the wadsleyite measurements of Omura (1991) to give a lower bound of transition zone electrical conductivity. However, loss of iron to the Pt electrodes mean that wadsleyite conductivity may have been severely underestimated in Omura's experiments (Duba and Wanamaker, 1994). Our measurements of wadsleyite conductivity provide

the upper bound. Figures 1 and 2 show that the response function at the periods considered here is relatively insensitive to upper mantle and transition zone conductivity variations.

3) The lower mantle conductivity profiles for the cool and hot geotherms with Al-bearing and Al-free perovskite are shown in figure 3. Repeat experiments showed reproducibility in measured perovskite conductivity of 0.1 log units, comparable to our inter laboratory agreement for olivine and wadsleyite. A 1000 K increase causes a much larger increase of 0.5 log units, comparable to the increase due to alumina.

It is also important to consider the uncertainties in the composition of the lower mantle phases:

4) The effect of magnesiowüstite depends on the equilibrium geometry and relative proportions of the phases. If the magnesiowüstite forms isolated regions within perovskite (Martinez et al, 1997) it would give negligible contribution to conductivity. At the other extreme, a totally interconnected network of 20 vol.% $\text{Mg}_{0.86}\text{Fe}_{0.14}\text{O}$ magnesiowüstite would double the conductivity of an Al-bearing lower mantle. The increase in conductivity over the Al-free perovskite, if the magnesiowüstite is interconnected, may be as much as 5 times (interpolated from Webman et al, 1976). Therefore the only likely effect of magnesiowüstite (if any) is to increase lower mantle conductivity sufficiently to necessitate a cool lower mantle geotherm. Most layered convection models, however, favour perovskite-rich lower mantle compositions, where shorting by a volumetrically minor magnesiowüstite would be unlikely.

5) The perovskite used in Xu et al (1998b) was synthetic $(\text{Mg}_{0.92}\text{Fe}_{0.08})\text{SiO}_3$, more iron-rich than the lower mantle perovskite composition estimated from Al-free perovskite-magnesiowüstite partitioning experiments (Katsura and Ito, 1996). Partitioning studies with aluminous perovskite (Wood and Rubie, 1996) lead to an equilibrium iron content

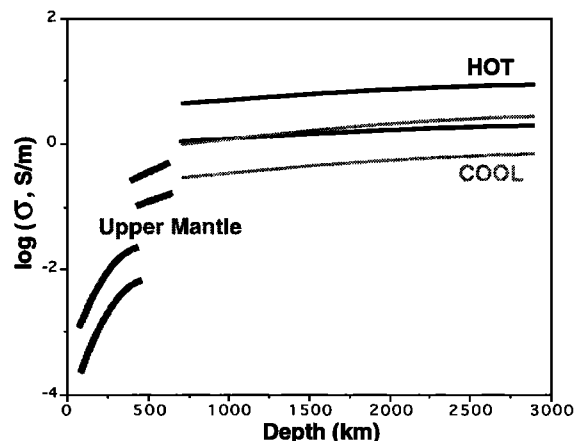


Figure 3. Electrical conductivity profiles used. All models have a common upper mantle profile. Olivine conductivity was measured at C-CO, Ni-NiO and Fe-FeO oxygen fugacities. Transition zone conductivities are based on the values for wadsleyite from Omura (1991), lower line, and new measurements which are consistent with Xu et al (1998a). Lower mantle conductivities are based on the cool geotherm (light grey lines) (Brown & Shankland, 1981) and a 1000 K hotter geotherm (black lines). The upper black and light grey lines are for aluminous perovskite and lower lines are for Al-free perovskite, both from Xu et al (1998b).

of 0.1 and a slightly more conductive lower mantle. The partitioning studies, therefore, reinforce the conclusion of a cool, aluminous or hot, Al-free lower mantle.

6) The effect of water on mineral conductivity is unknown. In general, the most likely effect of water in the major mantle silicates is to enhance ionic conduction, via proton mobility or the increased concentration of extrinsic cation vacancies (Shankland and Duba, 1997). This is unlikely to have a significant effect on the electronic conduction mechanisms which are dominant at the temperatures of interest (e.g.; Bai et al, 1995 in olivine: Iyengar and Alcock, 1970; Wood and Nell, 1991 in wadsleyite: Li and Jeanloz, 1994 in perovskite). If, however, proton conduction were a significant mechanism it might increase electrical conductivity, favouring cool lower mantle geotherms.

Conclusions

We have combined laboratory measurements of electrical conductivity with proposed lower mantle geotherms to generate synthetic Earth geomagnetic response functions. The Earth's geomagnetic response is compatible with an alumina free hot lower mantle perovskite, without shorting from magnesiowüstite, or an aluminous cool lower mantle perovskite. Recent EOS-based geotherms (Yagi and Funamori, 1996; Da Silva et al, 2000) show that a 'pyrolite-composition' lower mantle is consistent with cool geotherms, whereas 'perovskite' models require a lower mantle ≥ 500 K hotter, consistent with our conductivity constraints. Experimental and theoretical studies and natural samples all suggest that the equilibrium lower mantle magnesium silicate perovskite will contain significant alumina and ferric iron. This study constrains the maximum possible alumina content of lower mantle MgSiO_3 perovskite to 4 %, since a higher alumina content would render even the cold geotherm of Brown and Shankland too conductive. Although we can not rule out a hot lower mantle geotherm we note that it would require a novel alumina-free or iron-poor composition, or a new volumetrically minor aluminous phase, not suggested by other lines of investigation.

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