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Implications of a metal-bearing chemical boundary layer in D" for mantle dynamics

Michael Manga
Department of Geological Sciences, University of Oregon

Raymond Jeanloz
Department of Geology and Geophysics, University of California, Berkeley

Abstract. At lower mantle conditions, the thermal conductivity of non-metallic crystals is an order of magnitude lower than that of dense metallic minerals such as FeO, FeSi and Fe, which may be present in a chemical boundary layer at the base of the mantle. Because the core-mantle boundary is nearly isothermal, variations in the thickness of a metal-bearing layer induce lateral temperature variations of several hundred kelvin, which in turn affect the pattern of mantle convection. Upwellings should occur where the layer is thickest; the resulting stability of the metal + silicate layer with respect to the flow may also stabilize the pattern of mantle convection, and reduce its time-dependence.

Introduction

The reduction of radial seismic velocity gradients in D" is commonly attributed to the temperature increase associated with a thermal boundary layer at the base of the convecting mantle [e.g., *Stacey and Loper*, 1983]. The large amplitude of lateral velocity heterogeneity at the base of the mantle in global seismic tomography models [e.g., *Su and Dziewonski*, 1994] is also compatible with the large lateral temperature gradients expected in a thermal boundary layer in an actively convecting mantle. However, more recent local and regional studies reveal the existence of complex radial and lateral structures [e.g., review by *Loper and Lay*, 1995] which suggest the presence of chemical heterogeneity.

The correct interpretation of velocity variations in D" is not simple; small-scale topography at the core-mantle boundary, anisotropy, and strong scattering can affect seismological inferences, and a limited and imprecise knowledge of high-pressure elastic properties limit geophysical interpretations of seismological observations. Nevertheless, based on the range of seismic observations, it seems likely that the D" region is a complex thermal and chemical boundary layer, and that the local and detailed seismic structure is due to a combination of thermal and compositional effects [e.g., *Lay*, 1989]. Here we consider the evidence for, and effects of, a metal-bearing chemical boundary layer at the base of the mantle.

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Existence and Thickness of Metal-rich Chemical Boundary Layers

Metallic minerals in a chemical boundary layer may have four sources: *i)* primordial differentiation, *ii)* ongoing differentiation in the outer core, *iii)* differentiation, and segregation of metals, in the mantle, and *iv)* reactions between iron in the outer core and silicates or oxides in the lower mantle. In all four cases, the metallic phases would probably be iron alloys.

For cases *i* and *ii*, a thin layer of iron alloy, e.g. FeO, between the mantle and core would have a P-wave velocity reduction $> 10\%$, using elastic properties cited by *Williams and Garnero* [1996]. This velocity anomaly is compatible with that required by *Garnero and Helmberger* [1996] to explain the thin (< 40 km thick) basal layer beneath the central Pacific. This layer has alternatively been attributed to partial melting [*Williams and Garnero*, 1996; see also *Kendall and Silver*, 1996]. Since the melting temperature of FeO is higher than that of Fe at lower mantle and core pressures [*Knittle and Jeanloz*, 1991b; *Boehler*, 1992; *Shen et al.*, 1993], the layer of FeO may be solid.

In cases *iii* and *iv*, the reaction between iron and perovskite at lower mantle pressures and temperatures produces metal alloys FeO and FeSi [e.g., *Knittle and Jeanloz*, 1991a]:

$$Mg_{0.9}Fe_{0.1}SiO_3 + 0.15Fe =$$
$$0.9MgSiO_3 + 0.2FeO + 0.05FeSi + 0.05SiO_2. \quad (1)$$

Calculations suggest that reaction (1) may proceed further [*Song and Ahrens*, 1994]:

$$Mg_{0.9}Fe_{0.1}SiO_3 + 0.3Fe =$$
$$0.9MgSiO_3 + 0.3FeO + 0.1FeSi. \quad (2)$$

The reduction of seismic velocities due to these reactions is in the range $\sim 2 - 5 \pm 2\%$, again using values for material properties cited by *Williams and Garnero* [1996], and thus compatible with the magnitude of velocity reduction beneath the Pacific [e.g., *Vinnik et al.*, 1995].

The density contrast between the lower mantle and either a layer of pure Fe-alloy (cases *i* and *ii* above) or a layer of metals and oxides (cases *iii* and *iv*, and reactions 1 and 2), is $\Delta\rho \sim 20\%$ and $\sim 3 - 7\%$, respectively. In both cases, the density contrast is sufficiently large that complete entrainment and dispersal of these phases into the convecting mantle is unlikely [e.g., *Sleep*, 1988; *Olson and Kincaid*, 1991; *Kellogg and King*, 1993]. The thickness of such layers will be determined by the pattern and strength of flow in the lower mantle; balancing viscous stresses by buoyancy gives an estimate of the magnitude of thickness variations, $h \sim 4\mu\dot{\epsilon}/\Delta\rho g$, where μ is the viscosity of the lower mantle, and $\dot{\epsilon}$ is the strain-rate. Reasonable choices of $\mu \sim 10^{23}$ Pa s and $\dot{\epsilon} \sim 10^{-15}$ s^{-1} imply that layers of pure Fe-alloy and of mixed (metal plus oxide) reaction products will vary in thickness by O(10) km and O(100) km, respectively. These estimates are consistent with more detailed numerical calculations [e.g., *Davies and Gurnis*, 1986; *Hansen and Yuen*, 1989]. The thicknesses of both types of metal-bearing chemical boundary layers are also compatible with the presence of both thin (0-40km thick) [*Garnero and Helmberger*, 1996] and thick (0-300 km thick) [e.g., *Vinnik et al.*, 1995] low-velocity layers found by seismologists.

In summary, seismological observations and mineral physics are compatible with the existence of two types of chemical boundary layers at the base of the mantle: a thin layer of metal alloy at the very base of the mantle, and a thicker layer containing ~10-20% by volume of iron alloy intermixed with non-metallic silicate and oxide (Fig. 1).

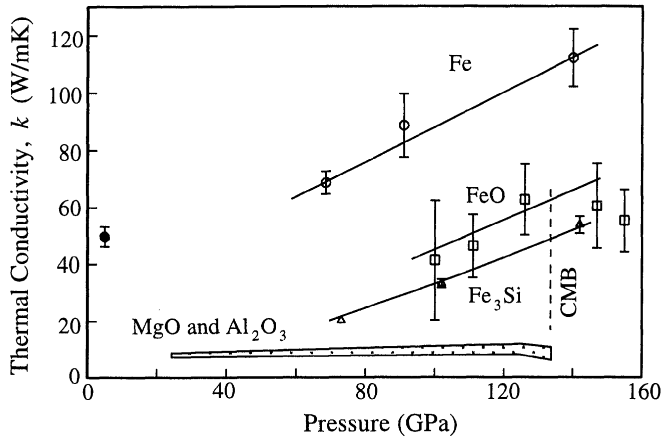

Figure 2. Thermal conductivity of iron, iron-alloys and non-metals at lower mantle pressures and temperatures. The thermal conductivity of metals is determined from electrical conductivity measurements made in shock experiments, and is expected to be independent of temperature [*Berman*, 1976]. The filled circle is based on piston cylinder experiments [*Secco and Schloessin*, 1989]; other references are cited in the text. The range of thermal conductivity of MgO and Al_2O_3 indicated by the stippled region accounts for the temperature-dependence of k, and uncertainties in the lower mantle temperature [*Manga and Jeanloz*, 1996]. The thermal conductivity of other non-metals is expected to be similar to that of MgO and Al_2O_3.

Geodynamic Implications

The electrical and thermal properties of metal alloys are significantly different from those of non-metals. Previous investigations have focussed on interactions with the geomagnetic field, as the presence of metal in the lowermost mantle influences electromagnetic coupling between the mantle and core [e.g., *Love and Bloxham*, 1994], generation of the Earth's magnetic field, and the propagation of magnetic field from the core through the mantle [e.g., *Busse and Wicht*, 1992; *Buffett*, 1996]. Here we consider the effects of a metal-bearing boundary layer on geodynamic processes in the mantle. In the discussion below, we focus on the thermal consequences of Fe-alloys being present in a layer composed of the reaction products in (1-2), and note that the results can be rescaled for a thinner layer of nearly pure Fe-alloy.

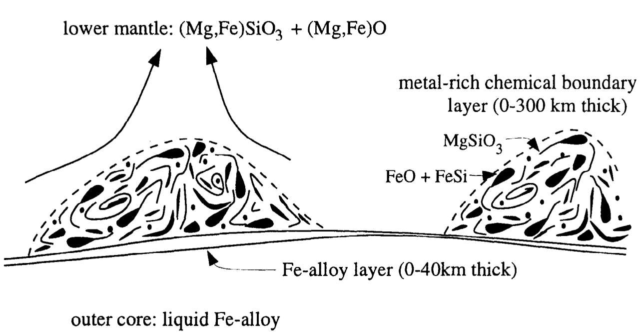

Figure 1. Schematic illustration of two possible types of metal-bearing boundary layers at the base of the mantle. The thick layer contains FeO and FeSi produced by a reaction between Fe in the core and (Mg,Fe)SiO_3 perovskite in the mantle. The thinner layer may be pure Fe-alloy. The thickness of such layers will be governed by the strength and pattern of flow in the lower mantle.

Thermal conductivity in the lower mantle

Because the core-mantle boundary is approximately isothermal [e.g., *Bloxham and Gubbins*, 1987], the high thermal conductivity of metal alloys (Fig. 2) affects the thermal structure in the lower mantle. The thermal conductivity of metals and alloys, $k \sim 30-100$ W/mK, was calculated from electrical conductivity measurements using the Wiedmann-Franz-Lorenz law,

$$k = L_o \sigma T \tag{3}$$

where σ is electrical conductivity, $L_o = 2.45 \times 10^{-8}$ WΩ/K^2 is the Lorenz number, and T is the temperature at which the measurement was made (the Hugoniot temperature for shock experiments). Measurements of σ used to generate Fig. 2 come from shock experiments by *Keeler and Mitchell* [1969], *Matassov* [1977] and *Knittle et al.* [1986] for Fe, Fe_3Si, and FeO, respectively. The corresponding Hugoniot temperatures are from *McQueen et al.* [1970], *Matassov* [1977] and *Jeanloz and Ahrens* [1980]. To leading order, k is independent of temperature for metals [e.g., *Berman*, 1976], so that values obtained from (3) using shock data should also apply at lower mantle temperatures.

The thermal conductivities of MgO and Al_2O_3 were determined from measurements made in the laser-heated diamond-anvil cell, and are expected to be similar to those of other non-metals in the lower mantle [*Manga and Jeanloz*, 1996]. Values of $k \sim 5 - 10$ W/mK at lower mantle conditions are compatible with theoretical predictions [e.g., *Brown*, 1986] and low-pressure experimental results [*Yukutake and Shimada*, 1978], as well as with geodynamic constraints [e.g., *Leitch*, 1995].

For convenience, we assume a thermal conductivity of 50 W/mK and a conservative 10 W/mK for Fe-alloys and non-metals, respectively. Using the *Hashin and Shtrikman* [1962] bounds for composites, the thermal conductivity in the metal + silicate boundary layer is increased $\sim 30-50\%$ compared with that in the mantle, assuming 15% by volume of Fe-alloys.

Implications for mantle dynamics

In Fig. 3 we illustrate schematically the effect of a high thermal-conductivity layer on the globally-averaged temperature distribution in the lower mantle. Relatively high temperatures are induced at the top of the variable-thickness metal-bearing layer, leading to lateral temperature variations. Taking a mean vertical temperature gradient of ~ 3 K/km at the base of the mantle, and assuming that the reduction of temperature gradients is proportional to the increase in the thermal conductivity, lateral temperature variations of up to 500 K should exist if the layer's thickness varies by 300 km. Although these approximations assume that the heat flux from the core to the mantle is not significantly affected by the layer, numerical calculations that allow k to increase with depth are consistent with the reduction of the average vertical temperature gradient [e.g., *Leitch et al.*, 1991, Fig. 12].

Mantle upwellings, induced above the thickest regions of the metal-bearing layer by the horizontal temperature gradients, in turn tend to further thicken the layer. Indeed, the spatial correlation of the long-wavelength, low-velocity anomaly beneath the central Pacific [*Su et al.*, 1994] with the location of the basal layer found by *Garnero and Helmberger* [1996] and a thicker layer with a velocity reduction of a few percent [e.g., *Vinnik et al.*, 1995], supports the dynamical picture proposed here. The Pacific low-velocity region is partially surrounded by high velocity D" [*Wysession et al.*, 1995] which may be related to cold subducted slabs. The absence of structure in D" and a sharp core-mantle boundary found in other locations [e.g., *Vidale and Benz*, 1992] could be caused by the presence of downwellings. Complex layered structures with high velocities overlying low velocities may be due to cold mantle interfingered with a metal-rich layer [*Wysession*, 1996].

An analogous situation occurs at the top of the mantle, where lateral temperature variations associated with the presence of continents induce upwellings that promote the breakup of continents [*Zhong and Gurnis*, 1993; *Lowman and Jarvis*, 1993; *Anderson*, 1994]. *Guillou and Jaupart* [1995] found that conductive lids (in their experiments, associated with continents) induce a large-scale flow even at very high Rayleigh numbers. In contrast to the dynamics at the top of the mantle, the upwelling induced in the lower mantle is stable with respect to the boundary layer, since upwellings are promoted where the layer is thickest. Thus, the interaction of a high thermal-conductivity boundary layer with a convecting mantle may partially suppress some of the characteristic features of high Rayleigh number convection, namely, short-wavelength structure and strong time-dependence. These inferences are compatible with the calculations of *Leitch et al.* [1991], who find that increasing k in the lower boundary layer results in a broader boundary layer and broader plumes, results that can be explained by the decrease of the local Rayleigh number in the boundary layer. Thus, the presence of a metal-bearing layer, in addition to a decrease in thermal expansivity and increase in viscosity with increasing depth in the mantle [e.g., *Hansen et al.*, 1993; *Bunge et al.*, 1996], may partially explain the dominance of long-wavelength structure in the lower mantle.

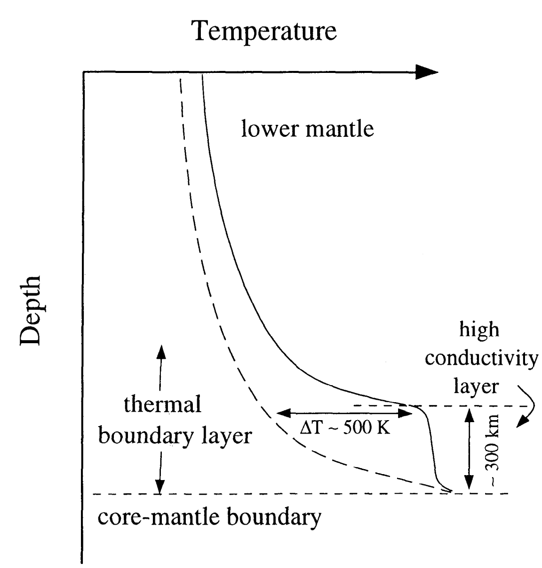

Figure 3. Schematic illustration of the globally-averaged temperature in the thermal boundary layer at the base of the mantle. Temperatures are shown for a homogeneous lower mantle (dashed curve, left) and a mantle with a high thermal conductivity, metal-bearing, basal layer (solid curve, right).

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