THERMAL EVOLUTION OF MERCURY IN THE CONDUCTIVE REGIME AND IMPLICATIONS FOR MAGNETISM.

C. C. Reese, P. E. Peterson, *Department of Science and Math, University of Minnesota Morris, Morris MN 56267*, V. S. Solomatov, *Department of Physics, New Mexico State University, Las Cruces NM 88003, USA.*

Introduction. Early theoretical estimates of convective instability and convective heat transport in the mantle of Mercury suggested that the mantle convects although not as vigorously as on Earth [1-3]. The strongest observational support for convection was the presence of a magnetic field on Mercury. A major difficulty with these models is that they predicted a marginally unstable mantle with very weak convection. Since the estimates were based mainly on studies of constant viscosity convection they could have substantially overestimated the vigor of convection [4]. It is unclear whether convection occurs at all with realistic rheologies [5]. Also, the magnitude of the magnetic field predicted by convective models was substantially larger than the observed field. We address the question of whether the thermal evolution of Mercury occurred in conductive or convective regime and whether the magnetic field of the required strength can be generated with a conductive mantle.

Convection or conduction? In the absence of plate tectonics, mantle convection, if it occurs at all, must take place in the stagnant lid convection regime [4-6]. It is driven by small temperature differences, on the order of 200 K, controlled by rheology rather than by the total temperature difference across the convective layer (which would be the case for constant viscosity convection). One of the key parameters is the viscosity. Laboratory data for viscous creep in olivine [7] can be used as reasonable constraints for mantle viscosity (if pyroxene is the dominant component then the mantle is likely to be stiffer). The choice between the two major mechanisms, diffusion creep and dislocation creep, depends on the grain size. As discussed in [8] a likely range for the grain size in the mantle of Mercury is 0.1-1 cm. Application of scaling relationships for temperature- and pressure- dependent viscosity convection [5] with rheologies corresponding to olivine suggests several conclusions (Fig. 1). Firstly, for grain sizes estimated for Mercury, dislocation creep is more efficient than diffusion creep. This basically means that for the typical viscous stresses associated with convection the dislocation viscosity is lower than diffusion viscosity. Secondly, convection occurs only at temperatures higher than 1800 K for “dry” mantle and higher than 1500-1600 K for “wet” mantle. Thirdly, the range of temperature (and the heat flux) in which convection occurs without melting (assuming dry solidus of peridotite) is very small: about 10 K (between 9 and 13 mW m^{-2}) for “dry” mantle and 100 K (between 6 and 22 mW m^{-2}) for “wet” mantle. Further insight into this problem can be obtained with the help of parameterized convection calculations of the entire thermal evolution [9]. We find that melting occurs during very early stages of evolution causing differentiation of nearly all radioactive isotopes into the crust. The rest of thermal evolution occurs in the conductive regime (Fig. 2).

Coupled evolution of the mantle and the core. Assuming that thermal evolution occurs mainly in the conductive regime, we can consider a very simple model of coupling between the mantle and the core in which most radioactive elements are in the crust. A finite difference approach is used to calculate the temperatures in a spherical mantle (considering only radial variations) while the temperature in the core is calculated using parameterization suggested in [1,2]. In the model shown in Figures 3-5 mantle temperature quickly reaches a quasi-equilibrium profile after which the temperature of the core changes very slowly. Convection in the core is very weak and is driven by compositional buoyancy due to release of sulfur during solidification of the solid core. The temperature of the core and the size of the inner solid core changed very little during planetary evolution.

Conclusion. Models of thermal evolution of Mercury with realistic temperature-dependent and pressure-dependent viscosity suggest that mantle convection ceased early, after a short period of intensive melting and differentiation. We also argue that conductive mantle can generate a sufficiently small magnetic field which can be reconciled with observations. The cooling rate of the planet, the change in the planetary radius and the magnitude of the magnetic field depend on the abundance of radioactive elements in the mantle, the initial temperature of the core and the mantle, and the amount of sulfur in the core.

References. [1] D. J. Stevenson et al., Icarus, 54, 466-489, 1983; [2] G. Schubert et al., in Mercury, Univ. Arizona, pp. 429-460, 1988. [3] T. Spohn, Icarus, 90, 222-236, 1991; [4] V. S. Solomatov, Phys. Fluids, 7, 266-274, 1995; [5] V. S. Solomatov and L.-N. Moresi, J. Geophys. Res., 105, 21795-21818, 2000; [6] V. S. Solomatov and L.-N. Moresi, J. Geophys. Res., 101, 4737-4753, 1996; [7] S.-I. Karato and P. Wu, Science, 260, 458-461, 1993. [8] V. S. Solomatov, LPSC 33rd, 2002, this volume; [9] C. C. Reese et al., Icarus, 139, 67-80, 1999.

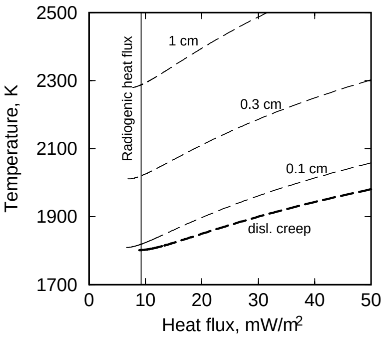

Figure 1: Mantle temperature as a function of the heat flux from the mantle (dry olivine). Dislocation creep is shown with a thick line. Diffusion creep is shown with thin lines (the grain size is indicated). The dashed line corresponds to a partially molten mantle.

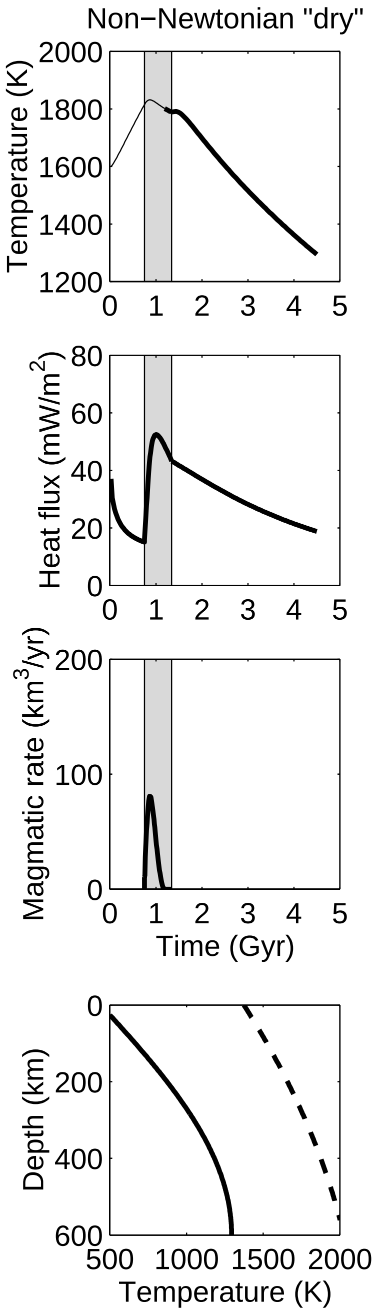

Figure 2: Thermal evolution models for Mercury. The viscosity is controlled by dislocation creep in “dry” olivine. The shaded region indicate the time interval when convection took place. Convective recycling of mantle material through supersolidus regions results in an extensive magmatism and fast depletion of the mantle in the abundances of radioactive elements. Thin line in top figure indicates supersolidus mantle. The bottom figure shows the present-day temperature distribution (solid line) and solidus (dashed line).

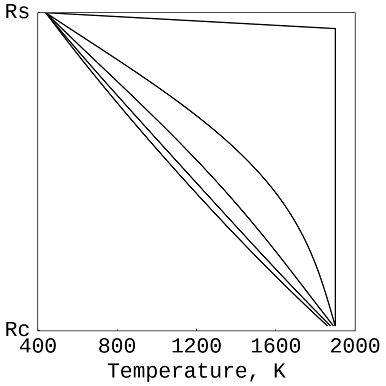

Figure 3: Temperature profiles at 0,1,2,3 and 4 Gyr for a completely depleted mantle and an initial temperature of 1900 K.

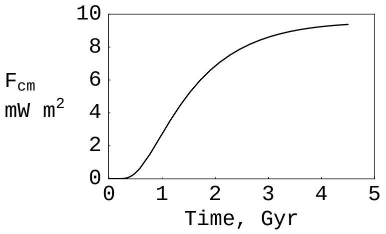

Figure 4: Heat flux from the core as a function of time.

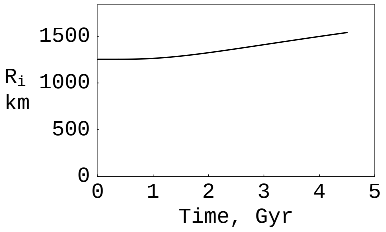

Figure 5: Radius of the inner core as a function of time.