Thermal Evolution of Mercury with a Volcanic Heat Pipe Flux G. A. Peterson[1], C. L. Johnson[1,2], A. M. Jellinek[1], [1]Departmentof Earth, Ocean and Atmospheric Sciences, University of British Columbia, Vancouver, BC, V6T1Z4 (gpeterso@ubc.eosc.ca), Canada, [2]Planetary Science Institute, Tucson, AZ 86819, USA.

Introduction: Thermal evolution models for Mercury are challenged by three aspects of the early history of the innermost planet [1–3]. First, the planet was marked by widespread effusive volcanism and crustal production [4]. Second, Mercury's surface is characterized by extensive tectonic structures indicative of crustal shortening, resulting from 5-10 km of radial contraction during secular cooling of the planet [5]. Cross-cutting relationships with craters suggest that these shortening structures began to form at ~3.9 Ga, at a rate that then monotonically declined to the present day [6]. Third, in addition to a present-day internally-generated magnetic field, crustal magnetic fields suggest a core dynamo was also active at ~4-3.5 Ga [7]. How the present-day or past core dynamos have been driven are still unknown [1,2,7].

Previous studies examining Mercury's thermal evolution have used solid-state models, ignoring crustal production [1] or used methods that produce negligible rates of volcanic heat loss to examine Mercury's phase of crustal production [2,3]. Here, we revisit the thermal history of Mercury, and incorporate an improved parameterization of the production and advection to the surface of buoyant partial melt. This picture is consistent with extensive crustal resurfacing of Mercury to produce a crust ~20 – 50 km thick by 4 Ga, forming the Intercrater Plains [4]. Inclusion of this volcanic heat flux better explains the timing and extent of shortening structures at Mercury's surface, the ancient magnetic field, and satisfies the constraint of extensive early crustal resurfacing[4–7]. This mode of heat transfer has been found to be an important process on Io and for the Hadean Earth, and is often referred to as the Heat Pipe mechanism [8].

Stagnant Lid Model Set up: Mercury comprises a metallic core (radius ~2020 km) and thin (~420 km thick) silicate shell [2]. Initial solid-state mantle convection, driven in response to intense early radioactive heat production is plausible in a stagnant lid regime [2,9]. Assuming quasi-steady state (SS) thermal conditions, parameterizations for the surface heat flux [9]enable thermal histories for the mantle and core to be computed by integrating the coupled equations [2,3]:

$$(1)\ \rho_m C_m V_m \frac{dT_m}{dt} = -q_m A_m + H_m V_m - q_{vol} A_m$$

$$(2)\ \rho_c C_c V_c \frac{dT_c}{dt} = -q_c A_c$$

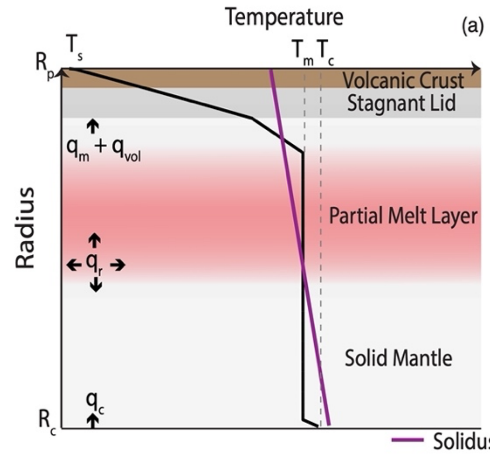

Figure 1: Diagram of the temperature profile (Black solid line) used for the parameterization of mantle convection with melt included. The partial melt zone (shaded red) is defined by the intersection with solidus (purple). Heat fluxes as a function of depth are shown as well.

Here, T_m is the mean internal temperature of the convecting mantle, T_c is the temperature of the core. ρ_m and ρ_c are the mantle and core density, C_m and C_c are mantle and core specific heat capacities. A_m, V_m and A_c are the surface area and volume of the mantle, and the surface area of the core. H_m is the heat production rate in the mantle with an enstatite chondrite composition [1,3]. q_m is a SS surface heat flux that scales as the Rayleigh number, based on the mantle layer depth, D, to the 1/3 power (Fig. 1).

Calculating the volcanic heat flux q_{vol}: q_{vol} is the advective heat flux associated with extracting partial melt out of the mantle to form crust [8]. Depending on the mantle solidus, convectively ascending mantle with temperature T_m can exceed the melting temperature to produce a partial melt layer (Fig. 1). If the difference between radioactive heating and power leaving the mantle by convection is positive, this excess power is consumed by melting to produce a vertically-averaged melt fraction ø and total melt volume V_{melt}.

Buoyant partial melt will rise through a combination of permeability-controlled porous media flow and mantle convective upwelling [10]. In the limit that porous media flow is much faster than mantle convective overturning, partial melt rises through the cold overlying lithosphere and erupts at the surface. The temperature profile within the partial melt zone is that of the solidus and we establish an upper bound for this contribution to the surface heat flow (Fig. 1).

Models without q_{vol}: Fig. 2 shows the thermal evolutions of the mantle and core and the associated

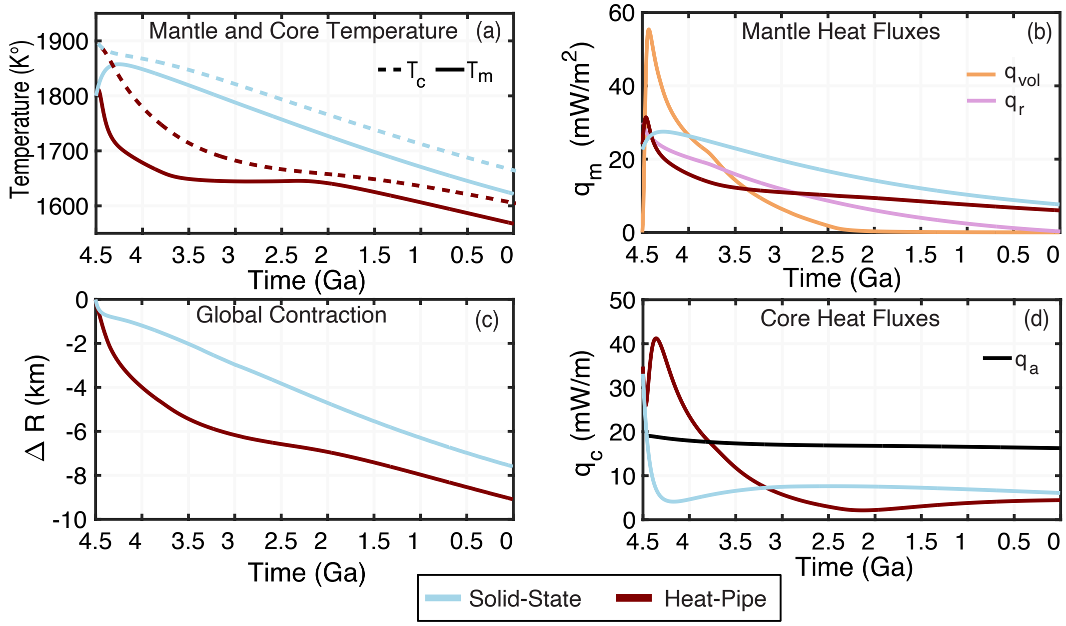

Figure 2: Thermal evolution for representative heat-pipe (red) and solid-state models (blue) with the same initial conditions (T_m = 1800 K°, T_c = 1900 K°, initial radioactive heating of 3 x 10^{11} W/kg, mantle thermal conductivity of 3 MW/K). a) T_m (solid line) and T_c (dashed line) for each model. b) Mantle heat flux, q_m for heat-pipe (red) and solid state (blue), q_{vol} (orange) and q_r (purple) c) The evolution of radial contraction (ΔR) d) The evolution of q_c for each model. The black line is the minimum core heat flux (q_a) that permits a dynamo through considerations of the core entropy balance.

radial thermal contraction ΔR for a typical model run. The results are similar to those in [1–3] where insufficient heat loss in solid-state models causes the mantle to initially warm when radioactive heating is high in Mercury's mantle early in the evolution (Fig. 2a,b). The initial mantle warming observed cannot explain the following observations:

1) An accelerated rate of shortening structure formation early in the planet's evolution. Instead solid-state models are characterized by a relatively constant formation rate of contraction (Fig. 2c).

2) The presence of a global magnetic field at ~3.9 Ga [7], if the dynamo is driven by thermal convection alone. To have a magnetic field, q_c must be at least as large as the core's adiabatic heat flux (~15 – 20 mWm^2 [2], and q_c falls below this by ~4.25 Ga (Figure 1e).

Model with q_{vol}: We find that mantle melt production and the resultant heat-pipe-controlled crustal production profoundly affects the early thermal evolution of the planet. Heat-pipe models are characterized by rapid mantle cooling as q_{vol} is the dominant mode of heat transport when melt production and extraction is high (Fig. 2a,b). This crucial difference between heat-pipe and solid-state models has important implications for the early geological record of Mercury:

1) The enhanced core and mantle cooling can explain shortening structures forming at 3.9 Ga (Fig. 2c) and producing less than 10 km of radial contraction over Mercury's evolution [5,6].
2) Early rapid mantle cooling sustains high core heat loss rates, which significantly increases the probability that an ancient dynamo is thermally driven (Fig. 2d).
3) Extensive early volcanism is predicted, consistent with formation of the Intercrater Plains.

Conclusions: Our approach and results reconcile key aspects of the contractional, magnetic and volcanic histories of Mercury's evolution.

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