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Influence of an insulating megaregolith on heat flow and crustal temperature structure of Mercury

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ABSTRACT

Mercury is covered by a megaregolith layer, which constitutes a poor thermally conducting layer that must have an influence on the thermal state and evolution of the planet, although most thermal modeling or heat flow studies have overlooked it. In this work we have calculated surface heat flows and subsurface temperatures from the depth of thrust faults associated with several prominent lobate scarps on Mercury, valid for the time of the formation of these scarps, by solving the heat equation and taking into account the insulating effects of a megaregolith layer. We conclude that megaregolith insulation could have been an important factor limiting heat loss and therefore interior cooling and contraction of Mercury. As mercurian megaregolith properties are not very well known, we also analyze the influence of these properties on the results, and discuss the consequences of imposing the condition that the total radioactive heat production must be lower than the total surface heat loss (this is, the Urey ratio, *Ur*, must be lower than 1) in a cooling and thermally contracting planet such as Mercury at the time of scarp emplacement. Our results show that satisfying the condition of $Ur < 1$ implies that the average abundances of heat-producing elements silicate layer is 0.4 times or less the average surface value, placing an upper bound on the bulk content of heat producing elements in Mercury's interior.

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1. Introduction

In Mercury, as in other bodies that lack a substantial atmosphere, impact processes may have resulted in the production of a megaregolith layer, which is a porous, fragmentary layer formed by large compact and coherent blocks with regolith material filling the gaps between them. This rubble has impact and ejecta origin, and covers the outer few meters–kilometers of a planet (e.g., Warren and Rasmussen, 1987; Ziethe et al., 2009). Because megaregolith has a thermal conductivity much lower than that of equivalent solid rock, it insulates the hot interior and slows down cooling. Therefore, megaregolith has an influence on the thermal state and evolution of the planet.

Previous works have estimated paleo-heat flows for Mercury from the depth of large thrust faults associated with lobate scarps, interpreted to reach the crustal brittle–ductile transition (BDT) depth (Watters et al., 2002; Nimmo and Watters, 2004; Egea-González et al., 2012), or from the effective elastic thickness of the lithosphere (Ruiz et al., 2013); the so-obtained paleo-heat flow values refer to the time of deformation (i.e., the time of faulting or loading). Works calculating heat flows from proxies of lithospheric strength relate the mechanical state of the lithosphere to their thermal structure through procedures which involve the surface temperature. These works disregard the effects of an insulating megaregolith, although such a thermally insulating blanket would reduce the interior heat loss, which in turn would imply lower surface heat flows derived from the BDT depth (Ruiz and Tejero, 2000), the raising of near-surface crustal temperatures and the reduction of thermal gradients below the megaregolith.

Another way to analyze the thermal evolution of Mercury is through thermal history models (e.g., Hauck et al., 2004; Grott et al., 2011; Williams et al., 2011; Tosi et al., in press). Most of them neglect the presence of a megaregolith layer, resulting in enhanced heat dissipation, and hence in substantial planetary cooling and contraction, which is hard to reconcile with the relatively limited observed global contraction deduced from shortening measurements in compressional structures (for an updated estimate of radial contraction see Watters et al., 2013). On the other hand, Grott et al. (2011) demonstrated that the inclusion of an insulating megaregolith layer has important implications on the thermal evolution of Mercury, since predicted planetary cooling and radial contraction are notably reduced, relaxing the severe constraints imposed by the observed contraction.

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In this work, we consider the effect of an insulating megaregolith layer in the calculation of surface heat flow and crustal thermal structure from the depth of thrust faults associated with several prominent lobate scarps on Mercury. We first derive upper limits for the surface heat flow by neglecting the megaregolith layer. Then, we solve the heat equation by taking into account the properties of a typical lunar megaregolith, in accordance with previous studies which considered the lunar megaregolith as a good analog for Mercury (Grott et al., 2011). Furthermore, as mercurian megaregolith properties are not very well known, we also analyze the influence on the results of higher thermal conductivities, and the implications of the condition that the heat loss through the surface must be higher than the heat generated internally by radioactive decay in a cooling planet.

2. Study areas

Lobate scarps are the most prominent tectonic features on Mercury, and are interpreted to be the surface expressions of thrust faults related to global thermal contraction of the planet (e.g., Strom et al., 1975; Watters et al., 2009). In this work we have calculated surface heat flows for three different regions of Mercury by using published values of the BDT depth estimated from the analysis of thrust faults associated with lobate scarps (coordinates for the studied lobate scarps are given in Table 1). Watters et al. (2002) used a mechanical dislocation model to obtain a BDT depth of 35–40 km beneath the Discovery Rupes lobate scarp (hereafter Region A). Ritzer et al. (2010) employed a similar method to analyze the geometries of two faults associated with two lobate scarps located at the equator (Region B), and they obtained a BDT depth of 35 km for both structures. Egea-González et al. (2012) studied the depth of faulting for three lobate scarps located in the Kuiper region of Mercury (Region C); the depth of faulting estimated by these authors ranges from 30 to 39 km.

3. Temperature at the brittle–ductile transition

In order to obtain surface heat flows, we solve the heat equation. To establish the integration constants that are involved in this calculation, we take into account the temperature at the brittle–ductile transition (T_{BDT}) and the surface temperature (T_S). As the brittle strength is a function of depth and the ductile strength depends on temperature, we can equate both strength expressions at the brittle–ductile transition with the purpose of working out the value of the temperature at the BDT depth.

Smith et al. (2012) have shown that there are large variations in crustal thickness on Mercury. Our studied regions have crustal thicknesses ranging from similar (Region B) to clearly higher (Region C) than the mean value of 50 km assumed by these authors. For these regions, the crustal thicknesses values in Smith et al. (2012) are certainly thicker than the local BDT depth. Although there are some uncertainties in the absolute values of the crustal thickness model, local variations in this kind of models are robust. Thus we only consider the possibility of a BDT depth in the crust and restrict our calculations to crustal mechanical and thermal properties.

The critical stress difference necessary to cause faulting is given by (e.g., Ranalli, 1997):

$$(\sigma_1 - \sigma_3)_b = \alpha \rho g z, \tag{1}$$

where ρ is the density, g is the acceleration due to the gravity ($3.7\ \mathrm{m\,s^{-2}}$), z is the depth and α is a coefficient that depends on the friction coefficient and on the tectonic regime ($\alpha = 3$ for thrust faulting). In this expression we have assumed zero pore pressure, which is appropriated for the mercurian crust. In the lithosphere, ductile deformation takes place mainly by dislocation creep. In such a case the ductile strength is (e.g., Turcotte and Schubert, 2002);

$$(\sigma_1 - \sigma_3)_d = \left(\frac{\dot{\varepsilon}}{A}\right)^{\frac{1}{n}} Exp\left(\frac{Q}{nRT}\right), \tag{2}$$

where $\dot{\varepsilon}$ is the strain rate, A and n are laboratory-determined constants, Q is the activation energy of creep, R is the gas constant ($8.31\ \mathrm{J\,mol^{-1}\,K^{-1}}$) , and T is the absolute temperature. The brittle and ductile strengths are equal at z_{BDT}, so we can find the value of T_{BDT}:

$$T_{BDT} = \frac{Q}{nR\,\mathrm{Ln}\left[3\rho g z_{BDT}\left(\frac{\dot{\varepsilon}}{A}\right)^{-\frac{1}{n}}\right]} \tag{3}$$

This expression can be adapted to consider two different layers in case that megaregolith is assumed:

$$T_{BDT} = \frac{Q}{nR\,\mathrm{Ln}\left[3[\rho_1 z_1 + \rho_2(z_{BDT} - z_1)]g\left(\frac{\dot{\varepsilon}}{A}\right)^{-\frac{1}{n}}\right]} \tag{4}$$

We have used subscript 1 to refer to the megaregolith layer and subscript 2 to apply to the deeper layer. z_1 is the megaregolith thickness.

At the surface we assume temperature values provided by the present-day surface temperature model of Vasavada et al. (1999), which takes into account the insolation dependence on latitude and longitude. Table 1 shows temperatures on the surface that are representative for the location of the three regions studied here (see also Williams et al. (2011) and Egea-González et al. (2012)).

4. The case without insulating megaregolith

For the case without a megaregolith layer, surface heat flows can be easily calculated by solving the steady-state, 1-D heat conduction equation with radiogenic heat production for a layer of z_{BDT} thickness. We have assumed that thermal conductivity, density, heat capacity and volumetric heat production rate are constant values that represent the average properties of the crust, so the surface heat flow is expressed as:

$$F_S = \frac{k(T_{BDT} - T_S)}{z_{BDT}} + \frac{z_{BDT} H}{2}, \tag{5}$$

where k is the thermal conductivity of the crust and H is the volumetric heat production rate. We use a thermal conductivity of

Table 1
Relevant parameters and surface heat flows calculated by neglecting the megaregolith layer. Surface temperature depends on both latitude and longitude due to the coupled spin–orbit resonance and the relatively high eccentricity (Vasavada et al., 1999; Williams et al., 2011).

	Lobate scarps	z_{BDT} (km)	T_{BDT} (K)	T_S (K)	F_S ($\mathrm{mW\,m^{-2}}$)
Region A	Discovery Rupes (56°S, 40°W)	35–40 (Watters et al., 2002)	731–808	365	20–31
Region B	Western scarp (0°, 59.3°E) Eastern scarp (0°, 64.7°E)	35 (Ritzer et al., 2010)	737–808	350	24–32
Region C	Santa Maria Rupes (3.5°N, 19°W) S_K4 scarp (4°N, 15°W) S_K3 scarp (10.3°N, 13°W)	30–39 (Egea-González et al., 2012)	732–816	435	17–30

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