Influence of Megaregolith on the thermal evolution of Mercury’s silicate shell

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Abstract A so-called Megaregolith layer that is considered to be produced by continuous impacts in Mercury’s early stages is integrated into the thermal evolution models of Mercury to study its influence on the thermal evolution of Mercury’s silicate shell. This research first implements a one-dimensional parametric global thermal evolution model. Our results indicate that Megaregolith directly affects the thermal evolution of Mercury’s silicate shell by virtue of its good insulation performance. The way Megaregolith exerts its influence by prolonging the process of partial melting and reducing the heat loss, resulting in a thicker crust and thinner stagnant lid. As for the deep parts of the silicate shell, it is suggest that more energy is taken away from the mantle due to the longer partial melting, leading to lower temperatures below the crust compared with the case in the absence of Megaregolith, which further helps to advance the formation time of the inner core and promote its final size. In addition, we also carry out a simplified two-dimensional mantle convection simulation as a supplement to the one-dimensional model. The two-dimensional simulation depicts a typical mantle plume fractional melting scenario. Our calculations indicate that Megaregolith may be key to the long-term volcanic activities on Mercury. As far as the Megaregolith itself is concerned, the thermal structure of this particular layer is more sensitive to the thermal conductivity, suggesting that for such a highly fragmented structure, the thermal conductivity coefficient plays a key role in its evolution. Our work emphasizes the importance of Megaregolith to the evolution of Mercury.

Key words: Mercury, thermal evolution, Megaregolith

1 INTRODUCTION

Establishing thermodynamic models is one of the most important methods to study planetary evolutionary history. The most commonly used one is the inversion model based on known detection findings, for example, studies of geometries of shortening tectonic features distributed across Mercury provide constraints on the thermal state of Mercury as well as possible internal compositions (e.g., Watters et al. 2009; Grott et al. 2011; Achille et al. 2012; Michel et al. 2013;
For a long time, since few evidences of younger geological activities have been found, Mercury has been suggested as an inactive planet. However, with in-depth dig of the data returned by NASA’s MErcury Surface, Space ENvironment, GEochemistry, and Ranging (MESSENGER) mission (e.g., Johnson & Hauck 2016), recent studies on the remnants of volcanic activities of Mercury suggested that Mercury may be more active than thought, and its volcanic activities may have continued until less than one billion years ago (Thomas et al. 2014; Thomas & Rothery 2019). These works concluded that there should be some kind of insulation mechanism to keep the planetary inside warm. Geomorphological studies on the impact craters across the plants’ surface indicated that the inner solar system planets experienced intense impact bombardment incidents in its early stages, which profoundly changed and effected these planets’ surface characteristics and evolutionary processes (e.g., Bottke & Norman 2017). Samples from the Moon’s surface and the data obtained from Apollo passive seismic network confirmed the existence of a so-called Megaregolith layer that covers the lunar surface, defined as a thin, highly fragmented structure, which was caused by long term impact events (Petro & Pieters 2006; Lucey et al. 2006; Zhang et al. 2013; Han et al. 2014; Wiggins et al. 2019). Similar to the Moon, it has been suggested that Megaregolith is exist on Mercury’s surface as well (Grott et al. 2011; Bottke & Norman 2017). However, impactors that hit Mercury can have a higher impact velocity, making Mercury’s surface more fragmented and resulting in a higher porosity of Megaregolith than Moon’s (Minton & Malhotra 2010). On the other hand, compared to the underlying crust, Megaregolith shows some different characteristics, such as a much smaller thermal conductivity and a lower density, which makes it easy for Megaregolith to exhibit good thermal insulation properties (e.g., Ziethe et al. 2009). Overall, Megaregolith might be one of the key mechanisms to keep Mercury’s interior warm and influence the evolutionary processes of the planet, especially on its silicate shell (i.e., from the surface to the core-mantle boundary).

In the past few decades, there have been many studies devoted to studying the thermodynamic state of Mercury. Among them, some works focused on the generation of Mercury’s magnetic field (Cao et al. 2014; Breuer et al. 2015; Knibbe & van Westrenen 2018), and some studies focused on the evolution of the overall or local of the planet under the background of planetary contraction (e.g., Tosi et al. 2013; Egea-González & Ruiz 2014). Methodologically, one of the most popular research methods is numerical simulations including one-dimensional (1-D) and two-dimensional simulation (2-D). As an efficient method, the 1-D simulation usually divides the study area into different compositional layers, including the rust, mantle and so on. In recent years, Megaregolith has been increasingly integrated into the 1-D model as an intrinsic structure (e.g., Tosi et al. 2013; Egea-González & Ruiz 2014; Knibbe & van Westrenen 2018). However, few studies have discussed the influence of Megaregolith on the evolution of Mercury’s silicate shell. In addition, with the in-depth analysis of the data returned by MESSENGER and the ground-based observations, our understanding of Mercury is constantly being updated. For example, a Fe-Si-C core instead of the traditional Fe-S core is is gradually being accepted (see below) (e.g., Nittler et al. 2011; Knibbe & van Westrenen 2018). In short, it is necessary to rebuild such thermodynamic model of Mercury with the help of some renewed knowledge, to study the influence of Megaregolith on the thermal evolution of Mercury’s silicate shell.

In this paper, we establish a one-dimensional parametric global thermal evolution model, taking the Megaregolith as the single variable to study the role of Megaregolith. Moreover, we further carry out a simplified two-dimensional convection simulation as a supplement to the one-dimensional model with the help of the fully open finite element convection code – Advanced Solver for Problems in Earth’s ConvectTion, ASPECT (Kronbichler et al. 2012), details for ASPECT, referring to: https://aspect.geodynamics.org. The structure of this paper is as follows, we first introduce the models and methods for the one-dimensional global model and the two-dimensional simulation in section 2. Next, the re-
Fig. 1: Schematic figure. The planet is divided into Megaregolith, crust, stagnant lid, upper thermal boundary (UTB), convecting mantle, lower thermal boundary (LTB), outer core and inner core. a) 1-D radial ideal model of Mercury. b) The non-proportional Radius-Temperature diagram.

Results are presented in section 3. Finally, the summary is discussed in section 4. Other details are contained in the Appendix.

2 MODELS AND METHODS

2.1 1-D model

The one-dimensional model radially divides the planet into several layers, and the schematic diagram is given in Fig 1. The simulation strategy is to describe these structures with their energy-related equations. Finally, these equations are iteratively solved under the constraints of the given initial and boundary conditions until a self-consistent and stable solution is obtained.

2.1.1 Thermal evolution model for the silicate shell

As mentioned in Introduction, Megaregolith is one of the products of impact events. In addition to some differences in material properties (e.g., density, porosity), one can regard Megaregolith as a part of the crust. Therefore, the heat transfer in the crust or Megaregolith is controlled by the steady heat-conduction equation, which is given by:

\[
\frac{\partial}{\partial r} \left[ k_i \frac{\partial T(r,t)}{\partial r} \right] = -Q_i e^{-x} \tag{1}
\]

Where \( k \) is the thermal conductivity, \( Q \) is the radiogenic heating producing rate (RHPR). \( x \) is the thermal depth, which represents an effective thickness of a layer that the accumulated radioactive heat equals to the whole radioactive heat generated in a particular layer. This coefficient indicates the uneven distribution of the radiogenic heating producing elements (RHPE) (Srivastava & Singh 1999). \( i \) is the index representing different layers (i.e., the crust and Megaregolith). If \( i = \text{mega} \), it means the current calculation object is Megaregolith.

Combining with corresponding boundary conditions, this equation can be solved. For Megaregolith, they are:

\[
k_{\text{mega}} \frac{\partial T}{\partial r} \bigg|_{r=R_{\text{mega}}} = q^* \tag{2}
\]
Where $R_{\text{mega}}$ is the radius at the bottom of Megaregolith, $q^*$ is the heat flow from the crust into Megaregolith, $R_p$ is the radius of the planet, and $T_s$ is the surface temperature.

Given the boundary conditions, the temperature field can be solved as follows:

$$T(r) = T_s + \frac{q^*}{k_{\text{mega}}} + \frac{Q_{\text{mega}}D_{\text{mega}}^2}{k_{\text{mega}}} \left(1 - e^{\frac{-r}{D_{\text{mega}}}} - \frac{r}{D_{\text{mega}}} e^{\frac{-r}{D_{\text{mega}}}}\right)$$

Where $D_{\text{mega}} = R_p - R_{\text{mega}}$ is the thickness of Megaregolith.

The above equations allow us to model Megaregolith separately and obtain information such as temperature profile or heat flow profile. Moreover, for a planet with one-plate like Mercury (Byrne et al. 2014), a structure with a high viscosity contrast (stagnant lid) in the mantle is assumed (Patočka et al. 2017). For the stagnant lid, its thickness variation depends on the energy equation at the base of lithosphere (Morschhauser et al. 2010), which is:

$$\rho_m c_m (T_m - T) \frac{dD_l}{dt} + k_m \frac{T_l - T_m}{\delta_u} + k_m \frac{T_c - T_b}{\delta_c} = (\rho_cr_c L_c + \rho_cr_c c_c(T_m - T_l)) \frac{dD_{cr}}{dt}$$

Where $\rho_m$ and $c_m$ are the average density and specific heat capacity of mantle, respectively. $T_m$ is the temperature at the upper convecting mantle, $T_l$ is the temperature at the core-mantle boundary (CMB) and $T_c$ is the temperature at the core-mantle boundary (CMB) and $T_b$ is the temperature at the lower mantle. $\delta_u$ and $\delta_c$ represent the thickness of the upper thermal boundary (UTB) and the lower thermal boundary (LTB), respectively. $\rho_c$ is the average density of the crust, $L_c$ is the latent heat of fusion, and $c_c$ is the specific heat capacity of the crust. The last two terms on the left side of this equation represent the heat flowing into the stagnant lid and through the CMB.

The energy equation of the convecting mantle is:

$$\epsilon_m V_{cm} \rho_m c_m (1 + st) \frac{dT_m}{dt} + (\rho_cr_c L_c + \rho_cr_c c_c(T_m - T_l)) \frac{dD_{cr}}{dt}) A_{cm} = k_m \frac{T_c - T_b}{\delta_c} + V_{cm} Q_m$$

Where $\epsilon_m$ is the ratio of the convecting mantle’s temperature to the average temperature of the convecting mantle, $V_{cm}$ and $A_{cm}$ are the volume and the surface area of this layer, respectively, and $Q_m$ is the RHPR in this layer.

For the 1-D model, the premise of the Eqs. (4) or (5) is that the thickness of Megaregolith or crust remains unchanged. For the sake of simplicity, we set the Megaregolith’s thickness as a constant value. For a typical terrestrial planet, the crust is composed of the primitive crust and the secondary crust. The latter is largely related to partial melting (details see Appendix A), which occurs in the mantle wherever the ambient temperature exceeds the solidus of one or more substances that make up the mantle. Following previous works, peridotite is regarded as the first melting partial material for planets like Mars and Mercury (Morschhauser et al. 2010; Knibbe & van Westrenen 2018). In summary, considering the process of thickening of the crust, we apply a typical heat-condition equation instead of Eqs. (4) or (5) to describe the crustal heat equation, i.e.,

$$\frac{\partial}{\partial t} \left( k_{cr} \frac{\partial T}{\partial r} \right) = -Q_{cr},$$

where $k_{cr}$ is the thermal conductivity of the crust, and $Q_{cr}$ is the RHPR in the crust.
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2.1.2 Thermal evolution for the core

The geodetic studies of Mercury suggested that the Mercury owns a molten outer core and solid inner core (Margot et al. 2007; Knibbe & van Westrenen 2015; Steinbrügge et al. 2018; Genova et al. 2019). In addition, geochemical studies on major chemical elements distributed on the surface of Mercury indicated that the Mercury was formed and differentiated in a highly-reduced environment (Nittler et al. 2011; Weider et al. 2012). Under this background, it is likely that the element silicon (Si) is the dominant light constituent of Mercury’s core, to form a Fe-Si binary or a Fe-Si-X ternary/quaternary core, where X is other small contents of light element, such as sulphur (S) or carbon(C) (e.g., Nittler et al. 2011; Steenstra & van Westrenen 2020; Knibbe et al. 2020). However, due to the lack of understanding of a ternary system under high-temperature and high-pressure conditions, we chose a Fe-Si core instead. For the energy changes involved in the core’s evolutionary processes, only the latent heat and the gravitational energy contribution are taken into account. Lastly, an adiabatic core is assumed, which allows the change of the inner core radius to be associated with the temperature at CMB. We follow the core’s energy equation from Grott et al. 2011.

\[
\begin{align*}
\rho c V c \frac{dT}{dt} &= -A c k m \frac{T - T_b}{R_i} + (E_g + L_c) \rho c \frac{dT}{dt} \\
\rho c V c \frac{dR_i}{dT} &= -A c k m \frac{T - T_b}{R_i}
\end{align*}
\]

Where \( \rho_c, V_c \) and \( c_c \) are the core’s average density, volume and specific heat capacity, \( A_c \) is the surface area of the core, \( E_g \) and \( L_c \) are the gravitational contribution and latent heat generated during the formation of the inner core, \( R_i \) is the inner core’s radius.

In Eq. (8), \( T \) represents the ambient temperature in the core, it can be obtained by an adiabatic relation, which is:

\[
\frac{dT}{dP} = \frac{\alpha_0 K_0}{\rho^2 c_r} \frac{d\rho}{dP}
\]

Where \( \alpha_0 \) and \( K_0 \) are the reference thermal expansion and the isothermal bulk modulus, \( P \) is the ambient pressure, and \( \rho \) is the local density. We can solve \( P \) and \( \rho \) by (Labrosse et al. 2001; Zhang & O’Neill 2016):

\[
\rho = \rho(0) \exp\left(\frac{-2\pi G r^2 \rho(0) \rho_0}{3 K_0 \left( \log \left( \frac{\rho(0)}{\rho_0} \right) + 1 \right)} \right)
\]

\[
g(r) = \frac{4G\pi}{r^2} \int_0^r \rho(x)x^2 dx
\]

Where \( \rho(0) \) is the density at the core’s center, and \( \rho_0 \) is the core’s density at zero pressure. Accordingly, the pressure can be calculated through:

\[
P(r) = \int_r^R g(x) \rho(x) dx
\]

For solutions to these equations of state, see Appendix B.

Last is the melting temperature of the Fe-Si core \( T_{\text{melting}} \), which is given by (Knibbe & van Westrenen 2018):

\[
T_{\text{melting}}(P, X_{si}) = (X_1 - 1000X_{si})(1 + \frac{P}{\beta_1})^\frac{1}{\beta_2}
\]

Where \( X_1 \) varies between 1578K and 1678K, \( \beta_1 \) and \( \beta_2 \) are experimental determined parameters, and \( X_{si} \) is the weight fraction (wt%) of the element silicon in the core, which is constant during our simulation, \( P \) is the local pressure.
As long as the ambient temperature equals to or less than the melting temperature, the solidification begins. Therefore, the inner core radius $R_i$ can be solved by the pressure at the intersection point, which is given by (Breuer et al. 2007):

$$R_i = \sqrt{\frac{2R_c(P(0) - P_i(r))}{\bar{\rho}_c g_{\text{cmb}}}}$$  \hspace{1cm} (14)

Where $P(0)$ is the pressure at the center of the core, $P_i(r)$ is the pressure at the inner-core boundary (ICB), and $g_{\text{cmb}}$ is the gravitational acceleration at CMB.

2.2 2-D model

Regarding the two-dimensional model, as part of the qualitative study, we simplify the configurations of the 2-D model, of which there are two most significant simplifications. The first one is to fix the temperature at CMB ($T_c$) as a constant, rather than evolving with time as in the one-dimensional model. Because the simulation time of 2-D model is not as long as the 1-D simulation, the temperature change at the bottom boundary is not significant. The second is that the model is only divided into three layers (conventionally named compositional fields in ASPECT code), including Megaregolith, crust and mantle. The only criterion to distinguish different layers is their respective thermal conductivities. The geometric model we use is a quarter spherical shell, with inner ($R_c$) and outer ($R_p$) radius of 2020km and 2440km, respectively. This model is heated from the bottom and cooled from the top, with additional heat sources in the form of shear heating etc, and all sides are prescribed as free-slip boundaries. The temperature boundary conditions are prescribed according to the initial values at each side, and a temperature anomaly of 150K is added near the bottom of the shell to make the geometry asymmetric to start the model running.

The 2-D numerical simulation is carried out through the finite element convection code, ASPECT. The framework for calculating the generation and migration of molten materials is provided by Dannberg & Heister 2016. Based on their work, we further make some appropriate adjustments, mainly modifying the code to make the framework to support multiple fields (e.g., crust, Megaregolith) calculation and replacing the melting parameterization by what we apply in the 1-D simulation. For melting parameterization, refer to Appendix C. Additional technical details can be accessed through the link provided by supporting information.

2.3 Parameters

The input values of parameters play a key role in the computational results of a thermal model. There are basically two kinds of parameters. The first category is the data obtained through observations, including ground-based observations and spacecraft in situ detections, namely fixed parameters, such as the average radius of the planet and the mean surface temperature (e.g., Solomon et al. 2007; Peplowski et al. 2011; Perry et al. 2011; Williams et al. 2011). The second category is empirical/experimental parameters, which are characterized by a high degree of uncertainty, called uncertain parameters, such as the thickness of the silicate shell, the concentrations of the RHPE in the mantle (e.g., Michel et al. 2013; Peplowski et al. 2011; Knibbe & van Westrenen 2018; Knibbe et al. 2020).

Among these uncertain parameters, the temperature at CMB ($T_c$) and the weight fractions of the main light constituent of Mercury’s core ($X_{\text{wt.\%}}$) have attracted much attention. An important aspect is that these two parameters are closely related to the formation time and size of the inner core, and the latter can have significant influences on the evolution of the entire planet (Breuer et al. 2015; Dumbrerry & Rivoldini 2014; Genova et al. 2019). According to our arguments in section 2.1.2, the formation of inner core requires the ambient temperature wherever in the core to be lower than (or equivalent to) the melting temperature, which are controlled
by $T_c$ and $X_{\text{wt.}_%}$. In general, we can start from two examples. Previous research proposed a possible solid FeS layer at the top of the core to account for the large average bulk density inferred for Mercury’s silicate shell given by MESSENGER’s XRS measurements (Smith et al. 2012). However, this assumption requires the core is sulfur-rich, so that FeS would precipitate within the core at moderate pressure (Hauck et al. 2013; Breuer et al. 2015). If this hypothesis is true, then the current CMB temperature could be in the range 1600–1700 K, which can provide constraints on the input parameters (e.g., Michel et al. 2013). As we discussed in section 2.1.2, Mercury is more likely to own a Fe-Si-X than a Fe-S dominant core (Nittler et al. 2011; Steenstra & van Westrenen 2020; Knibbe et al. 2020), which deduces that the FeS layer may not exist (Cartier et al. 2017). Therefore, whether there is a FeS layer is still debatable. On the other hand, since Mercury’s magnetic field was found (Anderson et al. 2008, 2011, 2012), multiple implemented dynamo simulations that are consistent with the field intensity of the magnetic field require the inner core is smaller than 1200 km (e.g., Cao et al. 2014; Tian et al. 2015; Takahashi et al. 2019). Moreover, as for the core’s composition, Knibbe et al. 2020 suggested a series of possible combinations of weight fractions of liquid Fe-Si-C metal alloys that meet the constraints by geodetic measurements and dynamo simulations. Although only a few binary composition (Fe-Si) combinations satisfy the constraints, we still can get a rough range of silicon’s concentration around 5 wt. % and $T_c$ about 2000 K. For the rest of fixed parameters, we choose those parameters that are widely used in thermodynamic studies of terrestrial planets (e.g., Grott et al. 2011; Hauck et al. 2013; Tosi et al. 2013; Padovan et al. 2017; Knibbe & van Westrenen 2018). Part of them can be found in Table 1. Since we introduce Megaregolith into our model, it is necessary to explain the key parameters related to this layer. It is obvious from Eq. (1), the RHPR (In fact, the bulk density of Megaregolith, the contents of RHPE), thermal conductivity and the thermal depth control the process of heat-conduction for Megaregolith. Traditionally, high-resolution gravity field and topography data can be used to recast crustal density for terrestrial bodies (Gong et al. 2016; Goossens et al. 2017), but the resolution of Mercury’s field is too low to calculate a precise crustal density (Mazarico et al. 2014). In addition, the X-Ray spectrometer and Gamma-Ray spectrometer (GRS) carried by the MESSENGER spacecraft measured the contents and distribution of the main elements on the surface of Mercury and the main terrane types as well (Nittler et al. 2011; Evans et al. 2012; Kaaden et al. 2017). A new method was proposed to calculate the grain density of the crust by using normative mineralogy, and the bulk density can be obtained through the relation with the grain density: $\rho_b = (1 - \phi) \rho_g$, where $\rho_b$ is the bulk density, $\phi$ is the porosity, and $\rho_g$ is the grain density (Sori 2018). Comparing to the Moon, the impactor that hits the Mercury generally have a higher velocity, resulting in a larger porosity in Mercury’s Megaregolith (e.g., Minton & Malhotra 2010). For this reason, we assume a porosity that is between 18% and 20%, which is consistent with the value used in other studies (e.g., Grott et al. 2011). Generally, for thermal evolution studies, the RHPE is assumed to be uniformly distributed in the crust and mantle (Grott et al. 2011; Tosi et al. 2013; Michel et al. 2013; Egea-Gonzáles & Ruiz 2014; Knibbe & van Westrenen 2018). Taking into account that the materials of Megaregolith originate from the crust, we also assume that the RHPE in Megaregolith is evenly distributed, that is to say, set the thermal depth: $x = D_{\text{mega}}$. At the same time, the contents of RHPE are re-distributed according to the mass ratio of Megaregolith to the crust, i.e., $m_{\text{RHPE}}^{\text{mega}} = \gamma m_{\text{RHPE}}^c$, $0 < \gamma < 1$, where $m_{\text{RHPE}}^{\text{mega}}$ and $m_{\text{RHPE}}^c$ are the mass of RHPE in Megaregolith and the initial mass of RHPE in the crust, respectively. Noting that the thickness used to calculate $m_{\text{mega}}$ (the mass the Megaregolith) varies from 1 km to 5 km (Egea-Gonzáles et al. 2012). The measurement of MESSENGER GRS reveals that the average of the main heating producing elements on the surface of Mercury are 1150 ± 220 ppm K (potassium), 90 ± 20 ppb U (uranium) and 220 ± 60 ppb Th (thorium) at present day (Peplowski et al. 2011). We assume that the present-day surface concentrations are representative of the average crust, resulting in the initial crustal radioactive heating rates on the order of $10^{-10}$ W/kg, which is consistent with other studies’ assumptions (e.g., Peplowski et al. 2011;
Table 1: Part of the parameters used in simulation

<table>
<thead>
<tr>
<th>Symbols</th>
<th>Physical Meaning</th>
<th>Values</th>
<th>Units</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R_p$</td>
<td>Planetary radius</td>
<td>2440</td>
<td>km</td>
<td>Knibbe &amp; van Westrenen 2018</td>
</tr>
<tr>
<td>$g_s$</td>
<td>Surface gravity</td>
<td>3.7</td>
<td>m/s²</td>
<td>Knibbe &amp; van Westrenen 2018</td>
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<tr>
<td>$T_s$</td>
<td>Surface temperature</td>
<td>440</td>
<td>K</td>
<td>Knibbe &amp; van Westrenen 2018</td>
</tr>
<tr>
<td>$R_c$</td>
<td>Outer core radius</td>
<td>2020</td>
<td>km</td>
<td>Charlier &amp; Namur 2019</td>
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<tr>
<td>$\rho_m$</td>
<td>Average mantle density</td>
<td>3500</td>
<td>kg/m³</td>
<td>Tosi et al. 2013</td>
</tr>
<tr>
<td>$\rho_c$</td>
<td>Average crust density</td>
<td>2800</td>
<td>kg/m³</td>
<td>Tosi et al. 2013</td>
</tr>
<tr>
<td>$\bar{\rho}_c$</td>
<td>Average core density</td>
<td>7200</td>
<td>kg/m³</td>
<td>Tosi et al. 2013</td>
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<tr>
<td>$k_m$</td>
<td>Mantle thermal conductivity</td>
<td>4</td>
<td>W/m⁻¹K⁻¹</td>
<td>Grott et al. 2011</td>
</tr>
<tr>
<td>$k_{cr}$</td>
<td>Crust thermal conductivity</td>
<td>1.5-4</td>
<td>W/m⁻¹K⁻¹</td>
<td>Grott et al. 2011</td>
</tr>
<tr>
<td>$k_{megag}$</td>
<td>Megaregolith thermal conductivity</td>
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<td>W/m⁻¹K⁻¹</td>
<td>Breuer et al. 2007; Grott et al. 2011</td>
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<tr>
<td>$c_c$</td>
<td>Core heat capacity</td>
<td>835</td>
<td>J/kg⁻¹K⁻¹</td>
<td>Knibbe &amp; van Westrenen 2018</td>
</tr>
<tr>
<td>$c_m$</td>
<td>Mantle heat capacity</td>
<td>1212</td>
<td>J/kg⁻¹K⁻¹</td>
<td>Knibbe &amp; van Westrenen 2018</td>
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<tr>
<td>$c_{cr}$</td>
<td>Crust heat capacity</td>
<td>1000</td>
<td>J/kg⁻¹K⁻¹</td>
<td>Knibbe &amp; van Westrenen 2018</td>
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<tr>
<td>$L_{cr}$</td>
<td>Latent heat of fusion</td>
<td>$6 \times 10^5$</td>
<td>J/kg⁻¹K⁻¹</td>
<td>Grott et al. 2011</td>
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<tr>
<td>$E_g + L_c$</td>
<td>Latent heat and gravitational heat</td>
<td>$6.5 \times 10^5$</td>
<td>J/kg⁻¹K⁻¹</td>
<td>Bland et al. 2008</td>
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<td>$\alpha_0$</td>
<td>Reference thermal expansion</td>
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<td>K</td>
<td>Knibbe &amp; van Westrenen 2018</td>
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<td>$\kappa_0$</td>
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<td>Gpa</td>
<td>Knibbe &amp; van Westrenen 2015</td>
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<td>$\rho_0$</td>
<td>Core's density at zero pressure</td>
<td>7100</td>
<td>kg/m³</td>
<td>Zhang &amp; O'Neill 2016</td>
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<tr>
<td>$\rho(0)$</td>
<td>The density at core’s center</td>
<td>7300</td>
<td>kg/m³</td>
<td>*</td>
</tr>
</tbody>
</table>

Notes: * Calculated based on Equation (10).

Knibbe & van Westrenen 2018). Meanwhile, the initial concentrations of RHPE in the mantle is assumed to be about 37% of the surface measured value (e.g., Tosi et al. 2013; Padovan et al. 2017), which implies the concentrations in the primordial mantle of 425.5±81.4 ppm K, 33.3±7.4 ppb U and 81.4±22.2 ppb Th. Details of the fractionation of heat producing elements can be found in Morschhauser et al. 2010.

For the thermal conductivity, this coefficient is closely linked to porosity, temperature or pressure and other factors. In the superficial environment of terrestrial planet, giving priority to porosity (e.g., Schumacher & Breuer 2006), the thermal conductivity of Megaregolith and the crust can differ by one or two orders of magnitude, while the former is usually considered in the range of 0.075 to 0.2 W/m⁻¹K⁻¹ (Breuer et al. 2007; Grott et al. 2011). For the remaining parameters, we list them in Table 1. For the 2-D simulation, all the necessary files (including the parameters file) are stored in a repository that can be accessed through the link provided by the Supporting Information. The descriptions of the initial temperature profile can be found below.

3 RESULTS

3.1 1-D model

3.1.1 Representative results

In order to compare the results, we specify that all models have the same initial conditions parameters, which are listed in Table 2. The models are iterated until a self-consistent result is obtained. Hereafter, we show a set of representative results. By default, these displayed results are calculated in the presence of Megaregolith. If there is a suffix $-n$ after the name of the result, it means that the result is computed in the absence of Megaregolith.

As we can see in Fig 2.a, it depicts the temperature versus time of several layers within the silicate shell. A notable feature is that all temperatures decrease with time except for the temperature at the bottom of the crust ($T_{cr}$), which experiences a dramatic rise in the early stages. On the other hand, for the temperature at the CMB ($T_c$), the temperature in the upper con-
Influence of Megaregolith on the thermal evolution of Mercury’s silicate shell

Table 2: Representative Initial Condition Parameters

<table>
<thead>
<tr>
<th>Symbols</th>
<th>Physical Meaning</th>
<th>Values</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>(T_{\odot})</td>
<td>Initial CMB temperature</td>
<td>1950</td>
<td>K</td>
</tr>
<tr>
<td>(T_{m0})</td>
<td>Initial temperature at the upper mantle</td>
<td>1740</td>
<td>K</td>
</tr>
<tr>
<td>(T_{cr0})</td>
<td>Initial temperature at the bottom of the crust</td>
<td>1050</td>
<td>K</td>
</tr>
<tr>
<td>(T_{meg0})</td>
<td>Initial temperature at the bottom of Megaregolith</td>
<td>840</td>
<td>K</td>
</tr>
<tr>
<td>(D_{t0})</td>
<td>Initial thickness of the lithosphere</td>
<td>50</td>
<td>km</td>
</tr>
<tr>
<td>(D_{cr0})</td>
<td>Initial thickness of the crust</td>
<td>5</td>
<td>km</td>
</tr>
<tr>
<td>(D_{meg})</td>
<td>Thickness of Megaregolith</td>
<td>2.5</td>
<td>km</td>
</tr>
<tr>
<td>(D_{hu0})</td>
<td>Initial thickness of the UTB</td>
<td>1</td>
<td>km</td>
</tr>
<tr>
<td>(D_{ltb0})</td>
<td>Initial thickness of the LT (B)</td>
<td>1</td>
<td>km</td>
</tr>
<tr>
<td>(SC_{Th})</td>
<td>Surface detected concentration of Th</td>
<td>0.155</td>
<td>ppm</td>
</tr>
<tr>
<td>(SC_{U})</td>
<td>Surface detected concentration of U</td>
<td>0.090</td>
<td>ppm</td>
</tr>
<tr>
<td>(SC_{K})</td>
<td>Surface detected concentration of K</td>
<td>1288</td>
<td>ppm</td>
</tr>
<tr>
<td>(K_{meg0})</td>
<td>Thermal conductivity of Megaregolith</td>
<td>0.2</td>
<td>W m(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>(X_{si})</td>
<td>The weight fraction of silicon</td>
<td>5</td>
<td>wt.%</td>
</tr>
<tr>
<td>(\gamma)</td>
<td>The mass ratio of Megaregolith to the crust</td>
<td>0.3</td>
<td></td>
</tr>
</tbody>
</table>

Notes: Data of the surface detected concentration of Radioactive Heating Elements comes from Pepłowski et al. 2011

vecting mantle \((T_m)\) and the temperature at the bottom of the stagnant lid \((T_l)\), the decrease of these temperatures begins to slow down due to the energy released during the formation of the inner core at the time of about 2.3Gyr. At present, the value of \(T_c\), \(T_m\) and \(T_l\) are about 1578K and 1462K and 1296K, respectively. At the shallow depth, the current value of \(T_{cr}\) is close to 741K, compared to the temperature at the bottom of the crust without Megaregolith \((T_{cr-n})\), the temperature difference is around 190K. Previous studies have suggested that \(T_{cr-n}\) ranges around 700K to 900K to satisfy the temperature calculated by the ductile strength model in a homogenous crust when faults initiate as early as 3.8Gyr years ago (e.g., Nimmo & Watters 2004; Egea-González et al. 2012), the computed \(T_{cr-n}\) at that time is about 787K, which locates within the temperature range above. For these computed temperatures, we compare them to some parametric works (e.g., Fig 4.a in Tosi et al. 2013; Fig 3.a in Knibbe & van Westrenen 2018), and we find that our results are generally smaller. We think that the possible reasons for this result are as follows. First, we follow Morschhauser et al. 2010 to model the fractionation of RHPE in the crust and mantle, part of RHPE belonging to the mantle enters the crust along with partial melting, thereby reducing the internal heating rate in the mantle (see below). Second, the choice of the core’s composition leads to the late formation and a smaller size of the inner core, which in turn reduces the energy released by the core (see below). The third point is that the difference in the initial conditions will affect the final results. Nevertheless, since this paper focuses on the analysis of the influences brought by Megaregolith, these differences in model settings would not affect the validity of our conclusions.

In terms of Megaregolith, the present value of the temperature at the bottom of Megaregolith \((T_{meg0})\) is close to 566K, and the difference between the initial temperature and the current temperature is not as large as other presented temperatures. We further compute the temperature and heat flux profile with a depth interval of 250m in Megaregolith at present day, which are shown in Fig 2.b and Fig 2.c, respectively. Both results are presented in the form of dataset, where the first data represents depth and the second data represents the temperature or heat flux. It is easy to point out that the precise current value of \(T_{meg}\) and the surface heat flux \((q_s)\) are 566.7K and 10.1 \(mW \cdot m^{-2}\), respectively. In addition, the present derivative of the temperature with respect to depth \((\frac{\partial T}{\partial z})\) is of an order of \(5.0 \times 10^{-2} K \cdot m^{-1}\), and the present derivative of the heat flux with respect to depth \((\frac{\partial q}{\partial z})\) is of an order of \(-2 \times 10^{-7} mW \cdot m^{-3}\) (see Supplementary Fig S1). These results suggest that Megaregolith has a good insulation performance. In order to analyze the key factors that control the thermal structure of Megaregolith, we implement end-member calculations. According to Eq. (4), we mainly discuss the thermal...
Temperature versus time of several layers within Mercury’s silicate shell. $T_c$ is the temperature at CMB, $T_m$ is the temperature at the upper convecting mantle, and $T_l$ is the temperature at the bottom of the stagnant lid, $T_{cr}$ is the temperature at the bottom of the crust, $T_{mega}$ is the temperature at the bottom of Megaregolith. b) The current temperature profile in the Megaregolith. c) The current heat flux profile in the Megaregolith. d) The thickness of several layers within Mercury’s silicate shell and the size of the inner core versus time. $D_{cr}$ is the thickness of the crust, $D_s$ is the thickness of the stagnant lid, and $D_{con}$ is the thickness of the convecting mantle. $R_i$ is the inner core radius. The suffix $-n$ represents the case when Megaregolith does not exist.

conductivity ($k$) and the amount of internal radioactive heating source (represented by $\gamma$, see section 2.3). Our calculation shows that compared to $\gamma$, a slight change in $k$ would bring dozens of temperature changes and a larger fluctuation of heat flux. This calculation concludes that $k$ is the key factor controlling the thermal profile of Megaregolith. Meanwhile, it is clear to figure out that both temperature and heat flux oscillate violently in the early stages of the simulation, and then gradually stabilized until close to zero. The oscillation time is roughly the same as the partial melting time, indicating the relationships between Megaregolith and the process of partial melting (see below). This part of results are shown in Supplementary Fig S2 and Fig S3.

Fig 2.d illustrates the thickness of several layers and the size of the inner core against time. After the model running done, we have a crust with a thickness ($D_{cr}$) of around 32km and a 20km thick crust ($D_{cr-n}$) in the absence of Megaregolith. Both results are located within the range of the mean crustal thickness calculated by geoid-topography ratios (Padovan et al. 2015; Sori 2018). At the same time, a stagnant lid with 276km thick ($D_s$) in the presence
Influence of Megaregolith on the thermal evolution of Mercury’s silicate shell

Fig. 3: a) The partial melting scenario in the presence of Megaregolith. $T_{\text{solidus}}$ is the solidus at the upper convecting mantle. $T_m$ is the temperature at the upper convecting mantle. The amount of the radioactive heating in the crust ($H_{cr}$) and mantle ($H_m$). The suffix $-n$ represents the case when Megaregolith does not exist.

of Megaregolith is also obtained. We further calculate the thickness of the convecting mantle ($D_{\text{con}}$), it is defined as the region between $R_m$ (the radius of the upper convecting mantle) and $R_b$ (the radius at the bottom of convecting mantle) (e.g., Knibbe & van Westrenen 2018). If the thickness is equivalent to zero, the mantle convection stops. Our calculations conclude that the mantle convection ceases near 4Gyr in the presence of Megaregolith, which is around 600Myr longer (i.e., 3.4Gyr) than the case when Megaregolith does not exist. Both of the results are in line with the views of some studies that Mercury’s mantle convection has stopped at least 500Myr ago (e.g., Tosi et al. 2013; Knibbe & van Westrenen 2018; Thiriet et al. 2019).

As mentioned earlier, the geodetic studies suggested that the Mercury owns a solid inner core. We calculate that the inner core’s radius ($R_i$) is about 803.5km, which accounts for 40% of the radius of the entire core ($R_c = 2020$km). Meanwhile, it can be seen from this figure that the inner core is formed at around 2.29Gyr. We compare our results with other parametric studies, and find that our calculations are smaller (e.g., about 1200km in Knibbe & van Westrenen 2018; admissible value of around 1000km in Cao et al. 2014). The possible reasons we suggest are as following. First, we chose a binary Fe-Si core, which excludes other light elements, like sulfur. However, the presence of other light elements (e.g., S and C) can not only change the concentrations of silicon but also effectively reduce the melting temperature within the core (Morard et al. 2007; Buono & Walker 2011; Knibbe et al. 2020). Second, we assume the temperature in the core follows an adiabatic relationship (section 2.1.2). Previous studies implied that Mercury owns a thermally stratified outer core, which results in a sub-adiabatic heat flow at the CMB (e.g., Christensen & Wicht 2008; Dumberry & Rivoldini 2014; Edgington et al. 2019). And such a stratified core would favor an early formation of inner core and a larger size (e.g., Knibbe & van Westrenen 2018). Nevertheless, a precise size of the inner core is still unknown, although several studies suggested that the size is smaller than 1200km (Steinbrügge et al. 2018; Charlier & Namur 2019; Genova et al. 2019; Takahashi et al. 2019). However, the formation of inner core has been suggested to be associated with the generation of the magnetic field (Laneuville et al. 2014; Rückriemen et al. 2015). There were studies found that the Mercury’s magnetic field already existed as early as 3.7Gyr ago (Johnson et al. 2015), the conflict in time suggests that a thermally driven dynamo is needed to support the magnetic field before the inner core’s growth (e.g., Grott et al. 2011; Tosi et al. 2013).
3.1.2 Comparison of the results

We first analyze the temperature differences of several layers within Mercury’s silicate shell that caused by Megaregolith. Noting that we preclude the temperature difference at the bottom of the crust \(TD_{cr}\) (i.e., \(T_{cr} - T_{cr-n}\)), because the discussion on \(TD_{cr}\) can be found in previous section, and its value is too large to cause display problem. The results can be found in supplementary Fig S3.d.

We can find out that, firstly, all these three shown temperature differences are negative, which means that the temperature is lower in the presence of Megaregolith. Secondly, all have a similar variation curve against time, showing a ‘V’ shape. In other words, before the end of partial melting, the temperature differences show a rough tendency to increase. In addition to the temperature difference at the CMB \(TD_c\), both the temperature difference in the upper convecting mantle \(TD_m\) and the temperature at the bottom of the stagnant lid \(TD_l\) have violent oscillations in the early stage of evolution, while \(TD_c\) is more significantly affected by the formation of inner core.

Before we analyze the reasons for this result, our calculations shows that \(TD_m > TD_l > TD_c\), implying that the temperature in the upper convecting mantle \(T_m\) is the most affected. As we know from section 2.1.2, both the temperature at CMB \(T_c\) and the temperature at the bottom of stagnant lid \(T_l\) are related to \(T_m\). Hence, we only analyze \(TD_m\). According to Eq. (7), the factors affecting \(T_m\) can be roughly divided into two categories. The first one is concerned with partial melting, which includes the thickening of the crust, the energy exchange associated with the phase change of mantle material and the migration of RHPE (Morschhauser et al. 2010). The other one is about the heat flow.

We determine whether partial melting occurs according to the relative value between the ambient temperature and the solidus of the main constituent of the lithosphere-mantle (details refer to Appendix A). The partial melting scenarios are shown in Fig 3.a. The region where partial melting occurs is marked in red, while the pink region indicates that the melting lasted longer when Megaregolith was present. The gray dash lines mark the end time of partial melting and the time lag is near to 100Myr. Fig 3.b shows the amount of radioactive heating over time in the convecting mantle \(Q_m\) and the crust \(Q_{cr}\). It is evident that \(Q_m\) increases rapidly with the thickening of the crust (refer to Fig 2.d), and a lower radioactive heating rates in the mantle \(Q_m\) is also obtained (Supplementary Fig S4.a). All of these suggest that more radioactive heats and the energy stored in material itself are extracted from the mantle in the presence of Megaregolith. As for the heat flux, our calculations suggest that the heat flow has limited impacts on the overall temperature compared with the radioactive heating. The results above explain why the temperature differences are negative. In short, we conclude that the reduction of radioactive heating rates in the mantle is the main reason that causes \(TD_m\) to be negative.

Next, we show the differences in thickness and size of several layers due to existence of Megaregolith, and the results are given in supplementary Fig S4.b. In addition to \(D_{cr}\), none of the rest values are less than zero. In fact, a thicker stagnant lid \(D_c < 0\) and a thicker convecting mantle region \(D_{con} > 0\) are the results of less internal heat loss (e.g., Thiriet et al. 2019), and a thicker crust means that more molten materials are created to form a secondary crust (e.g., Tosi & Padovan 2021; Beuthe et al. 2020). We can also find that when Megaregolith exists, the radius of the inner core is about 8.5km larger at present, and the time when the inner core formed is advanced by 20Myr (Supplementary Fig S4.b). The reason is that, the formation of the inner core is the result when the temperature at CMB \(T_c\) is less than or equal to the core’s melting temperature \(T_{melting}\), and the latter is determined by the pressure at the ICB \(P_i\) \((r_{ib})\). One can easily find that the crystallization temperature is fixed at a certain point radius \(r\) when the content of silicon is constant (see section 2.1.2). Therefore, \(T_c\) controls the time of the formation of the inner core and its final size, and it reduces the temperature inside the core to the temperature satisfying the beginning of crystallization through an adiabatic relation. As we discussed earlier, we already know that \(TD_c < 0\), hence, the time of the formation
Influence of Megaregolith on the thermal evolution of Mercury’s silicate shell

Overall, from our 1-D simulation, owing to the good thermal insulation properties of Megaregolith, it has great influences on the evolution of Mercury, especially in the early stages of the evolutionary process. The main way Megaregolith exerts its impacts is to significantly prolong the process of partial melting that occurs in the early stage of evolution and reduce the internal heat loss, so as to obtain like a thicker crust, a larger inner core.

3.2 2-D model

As mentioned in Introduction, the long-term volcanic activities on Mercury suggest some kinds of mechanisms to help produce or preserve molten materials (Thomas et al. 2014; Thomas & Rothery 2019). Although the results from our 1-D indeed suggest that a longer partial melting process can be a result of the existence of Megaregolith, it does not seem to explain whether Megaregolith can help preserve the molten materials. Therefore, we further implement a simplified 2-D simulation aiming at reveal the possible link between them. As required by the 2-D model, a user-defined initial temperature profile is needed. For this purpose, we assume that the thermal profile of the research domain can be determined by the conductive heat transfer equation. The boundary conditions that used to solve the conductive equation (e.g., \( q_c \), \( T_{cr} \)) are taken from our 1-D model (the version without Megaregolith) at 1Myr. The computed initial temperature profile is plotted in Fig 4.a, where the surface temperature is 440K and the bottom temperature is 1945K.

As part of the qualitative study, our 2-D model only simulates 300Myr. Although the simulation time is much less than the 4500Myr of the 1-D model, the 2-D simulation has captured the key features. In principle, this 2-D simulation depicts a typical mantle plume fractional melting scenario (Fig 4.b). Specifically, the plume moves upward from the core-mantle boundary due to its high buoyancy (temperature). The high temperature materials inside the plume reach a shallower place and causes the ambient temperature exceeds the solidus, and melting occurs with the rise of the mantle plume, moving further up and spreading laterally along the base of lithosphere (e.g., Loper 1991; Nikishin et al. 2002; Ziegler & Cloetingh 2004). As the melting goes on, the molten material upwards and accumulates in the top layer while the depleted source rocks remain in a layer below. Once the molten materials freeze, enriched materials will be created (Dannberg & Heister 2016).

As we know, buoyancy is the main driving force for the plume’s ascent (e.g., Brown & Lesher 2014). According to the conclusions obtained from our 1-D model, the existence of Megaregolith drops the cooling rate of the planet, resulting in a higher internal average temperature, which can further reduces the density of the materials in the plume, thereby leading to a greater buoyancy or a greater ascend speed. This allows the melting to start earlier in the presence of Megaregolith, which can be observed in Fig 5. Next, we calculate the value of depletion, it measures the fraction of the source rock that has been molten. The results are shown in Fig 5.a. It is clear to find out that when Megaregolith presents, the value of depletion is larger, meaning that more molten materials are produced from the source rocks. On the other hand, in Fig 5.b, we give the maximal melt fraction, which is used to indicate how much enriched materials are created. Therefore, what the Fig 5.b tells us is that when Megaregolith presents, there is less molten material to freeze into enriched material.

Overall, the existence of Megaregolith helps produce more molten material near to the top of mantle, but due to its good thermal insulation property, it leads to less enriched materials. We speculate that the presence of Megaregolith ensures the long-term existence of molten materials in the mantle. Given the fact that recent remnants of volcanic activities were found on Mercury (Thomas et al. 2014; Thomas & Rothery 2019), Megaregolith may be key to the long-term volcanic activities on Mercury.
4 SUMMARY

In this work, we carry out a one-dimensional global parametric thermal evolution model incorporating Megaregolith to study the influence exerted by Megaregolith on the silicate shell of the planet Mercury. Our results suggest that Megaregolith can effectively affect the thermal evolution of Mercury’s silicate shell by virtue of its good insulation performance. The main way Megaregolith exerts its impacts is to significantly prolong the process of partial melting that occurs in the early stage of the evolution and reduce the internal heat loss in the subsequent simulation. Specifically, the crust, as the closest layer to Megaregolith, is the most directly affected. Owing to the longer partial melting process, more molten materials and RHPE are brought in, making the crust thicker and the temperature higher. In terms of the rest deep parts of the silicate shell, on the one hand, the insulation effect of Megaregolith makes the stagnant lid thinner and delays the termination of mantle convection. On the other hand, a lower temperature of the mantle successfully allows the inner core to form at an earlier time, and eventually results in a larger inner core. We further implement a simplified two-dimensional mantle convection simulation with the help of the fully open finite element code ASPECT. This simulation depicts a typical mantle plume fractional melting scenario. Although we adopt some simplified parameterizations and a much smaller simulation time, this simulation still captures the key features brought by Megaregolith. We conclude that Megaregolith can effectively help to produce more molten materials and to ensure the long-term existence of molten material in
the mantle, which may be key to the long-term volcanic activities found on Mercury. As far as Megaregolith is concerned, the key factors that dominate the thermal structure of Megaregolith are discussed. We mainly analyze the thermal conductivity and the amount of the internal radioactive heating source. The results show that for such a highly fragmented structure, the thermal conductivity plays the key role. Specifically, both temperature and heat flux are more sensitive to the thermal conductivity, and a slight change in thermal conductivity can bring dozens of temperature changes and a larger fluctuation of heat flux. In addition, as its own internal heat production is small (e.g., at present time), the heat flux is more dependent on the flux that flows into Megaregolith.

Our thermodynamic simulations are highly dependent on the input parameters, and the choice of the input parameters usually leads to difference in calculations and even different conclusions. Although with the deepening of planetary studies, we can obtain some solid constraints on the thermodynamic models, we still need to rely on a large number of assumptions to design and implement simulations. In addition, a series of controversies including whether the FeS layer exists continues to plague the study of Mercury (e.g., Hauck et al. 2013; Breuer et al. 2015; Cartier et al. 2017). Nevertheless, as a qualitative research, through the method of controlling
variables, we think our conclusions are reliable.

Finally, our work emphasizes the importance of Megaregolith to the evolution of Mercury. Intuitively, the presence of Megaregolith has undisputed insulation effects on the interior of Mercury, but quantitative comparisons are lacking and meaningful. In the past, we paid more attention to its changes to the environment of the shallow depth in the outermost shell, but the influences on the formation of the inner core reminds us that this effect can be global and profound, and it is worthy of further study. The interaction between Megaregolith and the dynamic phenomenons of the deep inside of the planet will be more comprehensively understood along with the progress in areas like the composition of planetary core. For Megaregolith itself, as one of the results of impact events, its thickness, composition, coverage area and the exchange of energy and matters during the impact may effect its role in the evolution processes. We hope that the new Mercury detection mission BepiColombo can reveal the mysteries for us in the near future (Benkhoff et al. 2010; Milillo et al. 2020).

5 SUPPORTING INFORMATION

Additional Supporting Information can be found through the following link: https://github.com/XieJChris/Megaregolith-Mercury.git

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Appendix A: CRUSTAL THICKENING & MANTLE CONFIGURATION

As we mentioned above, the key to crustal evolution is the partial melting in the mantle. Following by previous studies (Grott et al. 2011; Morschhauser et al. 2010), a peridotite dominant mantle are assumed, therefore, the solidus and liquidus can be expressed as:

\[
T_{\text{sol}} = 1409 + 134.2P - 6.581P^2 + 0.1054P^3
\]  
\[
T_{\text{liq}} = 2035 + 57.46P - 3.487P^2 + 0.0769P^3
\]

Where \(T_{\text{sol}}\) and \(T_{\text{liq}}\) are the solidus and liquidus of peridotite, respectively, \(P\) is the pressure in GPa.

Taking into account the change of the solidus of the mantle caused by the depletion of peridotite, the solidus of the mantle is expressed as follows (Knibbe & van Westrenen 2018):

\[
T_{\text{sol,m}} = T_{\text{sol}} + \frac{D_{\text{cr}}}{D_{\text{ref}}} \Delta T
\]

Where \(T_{\text{sol,m}}\) is the mantle’s solidus, \(D_{\text{cr}}\) is the crustal thickness, \(\Delta T\) is the temperature difference that represents the temperature change of the solidus due to the depletion of peridotite. \(D_{\text{ref}}\) is the reference thickness, defined as (Morschhauser et al. 2010):

\[
D_{\text{ref}} = \frac{4\pi (R_1^3 - R_2^3)}{3} \Gamma
\]

Where \(\Gamma\) is the crustal formation rate (usually assumed to be 0.2).

Once the ambient temperature exceeds the mantle’s solidus, the partial melting occurs, the volume of melting zone is denoted by \(V_{\text{pm}}\), with \(V_{\text{pm}} = \frac{4\pi}{3} (R_1^3 - R_2^3)\), where \(R_1\) (\(R_2\)) is the radius of the lower (upper) boundary where melting is allowed in the mantle. And the volumetrically averaged degree of melting \(\eta_{\text{am}}\) is obtained by solving:
\[ m_{am} = \frac{1}{V_{pm}} \int_{V_{pm}} T(r) - T_{sol,m} \frac{T_{iq} - T_{sol,m}}{V} dV \]  
(A.5)

Finally, the thickness of the crust versus time is calculated through:

\[ \frac{dD_{cr}}{dt} = u_0 \left( \frac{Ra}{Ra_{crit}} \right)^{2/3} \frac{m_{am} V_{pm}}{4\pi R_P^4} \]  
(A.6)

Where \( u_0 \) is the mantle convective velocity scale, \( \beta \) equals to 1/3, and \( Ra (Ra_{crit}) \) is the (critical) Rayleigh number of the mantle.

Following Thiriet et al. 2019, we apply the Arrhenius viscosity law to the mantle, with:

\[ \eta_m = \eta_0 \exp \left( \frac{A(T_{ref} - T)}{RT_{ref}T} \right) \]  
(A.7)

Where \( \eta_0 \) is the reference viscosity, \( A \) is the activation energy, \( R \) is the universal gas constant, and \( T_{ref} \) is the reference temperature, so that \( Ra \) can be represented as:

\[ Ra = \frac{\alpha \rho g s \Delta T \Delta R^3}{\kappa_m \eta_m} \]  
(A.8)

With \( \Delta T = T_m + T_c - T_l - T_b \), and \( \Delta R = R_l - R_c \), and \( \kappa_m \) is the mantle thermal diffusivity.

As for \( T_l \) and \( T_b \), they are:

\[ T_l = T_m - \Theta \frac{RT_m^2}{A} \]  
(A.9)

\[ T_b = T_m + \frac{\alpha g T_m}{\kappa_m} D_{conv} \]  
(A.10)

Where \( \Theta \) is an empirically determined parameter, \( D_{conv} \) is the thickness of the convecting mantle, which is between \( R_m \) and \( R_b \).

Appendix B: CORE CONFIGURATION

Here, we refer to Labrosse et al. 2001 and Zhang & O’Neill 2016 to solve the equation of state of the core. A length scale for the compression and an adiabatic height are defined first:

\[ L = \left( \frac{3K_0 \left( \log \left( \frac{\rho(0)}{\rho_0} \right) + 1 \right)}{2\pi G \rho_0 \rho(0)} \right)^{0.5} \]  
(B.1)

\[ D = \left( \frac{3\rho_c}{2\pi \alpha_0 \rho_c G} \right)^{0.5} \]  
(B.2)

Then, we have the profile of the density and the gravitational acceleration in the core, as:

\[ \rho(r) = \rho_c \exp\left( -\frac{r^2}{L^2} \right) \]  
(B.3)

\[ g(r) = \frac{4\pi}{3} G \rho_c r \left( 1 - \frac{3r^2}{5L^2} \right) \]  
(B.4)

Further, the pressure profile in the core is obtained, it follows:

\[ P(r) = P_{cmb} - \frac{4\pi G \rho_c}{3} \left[ \left( \frac{3r^2}{10} - \frac{L^2}{5} \right) \exp\left( -\frac{r^2}{L^2} \right) - \left( \frac{3R_c^2}{10} - \frac{L^2}{5} \right) \exp\left( -\frac{R_c^2}{L^2} \right) \right] \]  
(B.5)

These symbols involved are explained in previous sections.
Appendix C: MELTING PARAMETERIZATION IN 2-D MODEL

In the 2-D mantle convection model, we generally follow the same melting parameterization used in our 1-D model (Eq. A.1-A.3), except the Eq. A.3 is replaced by:

\[ T_{\text{sol,m}} = T_{\text{sol}} + \Delta T_{\text{pm}} C \]  

(C.1)

Where \( C \) is the depletion or the partial melting rate, and \( \Delta T_{\text{pm}} \) is the solidus change when partial melting reaches 100%, which has a negative value.

The melting rate is computed as the difference between the equilibrium melt fraction and the melt present in the model, with the equilibrium melt fraction:

\[ \frac{T(r) - T_{\text{sol,m}}}{T_{\text{liq}} - T_{\text{sol,m}}} \]  

(C.2)

And refer to Dannberg & Heister 2016 for more details.

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