

1 **The influence of tidal heating on the Earth’s thermal**
2 **evolution along the dynamical history of the**
3 **Earth-Moon system**

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15 **Key Points:**

- 16 • Geological evidences allow to track back a possible dynamical evolution path of
17 the Earth-Moon system and to evaluate tidal heating.
18 • We found that tidal heating could have played a significant role in the Earth’s
19 thermal history.
20 • Our results suggest that the onset of plate tectonics may have occurred between
21 4.5 and 3.5 billion years in the past.

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Abstract

Several geological evidences, such as tidal rhythmites and bivalve shells, allow to track back the evolution paths of both the major semiaxis of the Moon’s orbit and the Earth’s spin rate. However, the data is scarce and with large uncertainty and the orbital evolution of the Moon is still controversial. The aim of this work is to evaluate how significant could have been the effect of bodily tides on the Earth’s mantle thermal evolution. To this end, different thermal models of the Earth’s interior were proposed. We explore plate tectonics and stagnant lid regimes. These models take into account both tidal and radiogenic heat sources. In order to compute tidal dissipation, we made use of three realistic rheological models of Earth mantle and proposed three different dynamical evolution paths for the lunar major semiaxis and terrestrial length of day. It was found that the impact of tidal interaction could have been specially appreciable on the first hundreds million of years of the Earth’s history, provided that the mantle was at a higher temperature. In addition, we found that thermal evolution of Earth’s interior is mainly controlled by the rheological behavior of the mantle, which controls the amount of tidal heat produced, and by the dynamical evolution of the Earth-Moon system.

Plain Language Summary

We aim to try to answer the question of how could the Moon have influenced the onset of plate tectonics in our planet. In our work, we take under consideration the effect of bodily tides, i.e. the deformation of the Earth as a whole due to the gravitational attraction exerted by the Moon. This deformation heats the Earth’s interior due to friction. We also take into account the heat generated by the decay of radioactive isotopes. These sources of heat are counteracted by thermal convection. The interaction between the heating and cooling mechanisms determines the time evolution of the internal temperatures of the Earth which, in turn, controls the lithosphere’s thickness, which give clues when plate tectonics could have started. The evaluation of tidal heating depends on the combination of rotational period of the Earth and the orbital period of the Moon, which changed over time. We used geological data to obtain a possible evolution path of both periods. We found that tidal heating could have had a significant role in the early stages of Earth’s thermal history and that the plate tectonics may have started between 4.5 and 3.5 billion years in the past.

1 Introduction

Since its formation, the dynamical evolution of the Earth-Moon system was characterized by the progressive increase of the mutual distance between both members of the system, the probably rapid capture of the Moon in the spin-orbit resonance 1:1 (Makarov, 2013), and the gradual decrease of the angular spin rate of the Earth.

Several geological evidences offer clues about the possible evolution path of the Earth-Moon system, such as the one found in bivalve fossil records and in rocks called *tidal rhythmites*, which account for the tidal cycles (Williams, 2000; López de Azarevich & Azarevich, 2017). The stratified structure of these rocks are interpreted as variations in the sea level as a consequence of tidal interaction and, hence, the lunar cycle. The bivalve fossils growing cycles are interpreted as an evidence of the length of day (LOD). However, the data is scarce and with large uncertainty. In addition, the orbital evolution of the Moon is still controversial.

Rigorous analysis of the aforementioned evidences has allowed the estimation of variations in the length of the day, the orbital period of the Moon and the orbital period of the Earth around the Sun throughout time. This information serve as a “ground

70 truth” for contrasting the theoretical models with empirical evidence, a fundamental
71 procedure in natural sciences.

72 The main objective of this work is to evaluate how large could have been the
73 impact of tidal heating, arising from bodily tides raised by the Moon, on the thermal
74 evolution of the Earth. To that end, we developed different thermal models that
75 includes an up-to-date expression for the rate of tidal heating or, in other words, the
76 rate at which heat is produced due to internal dissipation within Earth’s mantle.

77 The Darwin-Kaula formalism of Tidal Theory (Efroimsky & Makarov, 2013)
78 offers a complete theoretical framework for the computation of the heat production
79 rate due to internal dissipation excited by tidal loading.

80 Tidal interaction is originated by the deformation of a celestial body due to
81 gravitational forces acting on it, exerted by other bodies. This deformation, in turn,
82 modifies the gravitational field of the Earth, thus disturbing the orbital motion of the
83 Moon and the LOD. Since the deformation depends on the rheology we investigate
84 different rheological models (see the Supplementary information). We assume, on one
85 hand, that the Earth is spherical with a concentric layered interior structure containing
86 homogeneous core and mantle and, eventually, a lithosphere. On the other hand, the
87 Moon is assumed to be a point mass distribution.

88 Thermal evolution of the Earth’s interior is described by using simple parametric
89 one-dimensional models of thermal convection. These models consider the boundary
90 layer theory to describe heat transfer within both the Earth’s core and mantle and are
91 explained in detail in Section 2.

92 Later, the implementation of the aforementioned thermal model is explained in
93 detail in Section 3, which also offers a discussion about the compatibility of obser-
94 vational data with the dynamic characteristics imposed by tidal interaction. Then,
95 the results obtained from the numerical simulations carried out are presented and
96 discussed in Section 4. Lastly, our conclusions are offered in Section 5.

97 **2 Thermal evolution models of Earth’s interior**

98 As we have pointed out in Section 1, we model the Earth’s interior as a two layers
99 structure composed by the core and the mantle as shown in Figure 1. Two different
100 regimes will be considered, namely the plate tectonics (PT) and the stagnant lid (SL).
101 In the former, the Earth is assumed to be formed by the two aforementioned layers
102 (Figure 1a). In the stagnant lid regime an extra layer is included in the thermal model
103 representing a thick elastic lithosphere (Figure 1b).

104 The study of the time evolution of the temperatures inside the Earth makes use
105 of the continuity equation for energy flow and energy balance within both the core
106 and the mantle. On geological time scales, the dominant heat transfer mechanism is
107 convection (Stacey & Davis, 2008; Turcotte & Schubert, 2014). In this sense, we will
108 consider simple parametric one-dimensional models to describe the temporal evolution
109 of the mean temperatures of both the core and the mantle due to convection. These
110 models, for both the plate tectonics and stagnant lid regimes, will be presented in the
111 following subsections.

Let us then consider the continuity equation expressing the principle of energy
conservation:

$$\frac{\partial u}{\partial t} + \nabla \cdot \mathbf{q} = \varrho(\mathbf{r}, t), \quad (1)$$

112 where u is the energy density, in other words the amount of energy per unit volume,
113 and \mathbf{q} is the heat flux vector, whose magnitude is equal to the amount of energy per
114 unit area and unit time that crosses an specified surface, its direction is perpendicular

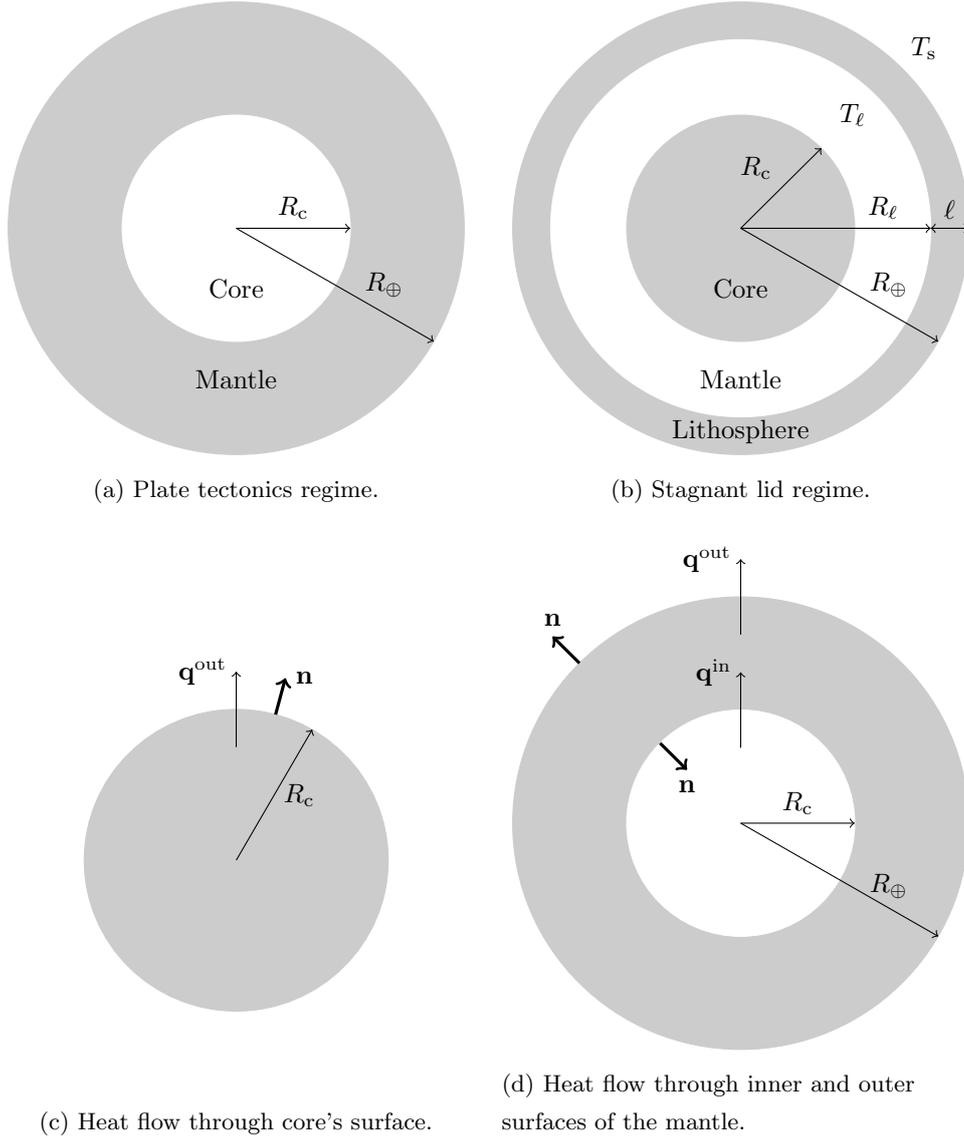


Figure 1: Schematics of the Earth’s interior structure for both the plate tectonics regime (a) and the stagnant lid regime (b). Heat flow through the core surface (c) and the inner and outer surfaces of the mantle (d).

115 to that surface and its sense is outwards. In addition, $\varrho(\mathbf{r}, t)$ is the heat production
 116 rate per unit volume which, in principle, could be a function of the position and time.

117 As the principle of energy conservation implies that energy is neither destroyed
 118 nor created, but only transformed from one type to another, by using the term
 119 “sources” we simply mean *sources of thermal energy*. In this work we will consider
 120 two of them, namely that from the decay of radioactive nuclides and tidal interaction.

A special form of Equation (1) is obtained when there are no sources:

$$\frac{\partial u}{\partial t} + \nabla \cdot \mathbf{q} = 0. \quad (2)$$

121 The particular functional form of $\varrho(\mathbf{r}, t)$ will then depend on the source or sources
 122 considered. For the Earth, we will assume that heat sources are uniformly distributed
 123 in the mantle and we will neglect the presence of sources in the core and in the
 124 lithosphere. In this sense, Equation (2) will allow us to derive the differential equation
 125 giving the time evolution of the core's temperature, while the equation describing the
 126 time evolution of the mantle's temperature will be obtained from Equation (1).

If we assume that changes in the amount of internal energy per unit volume Δu
 is only due to temperature changes (ΔT), in other words we neglect phase transitions,
 then Δu is given by:

$$\Delta u = \rho c \Delta T, \quad (3)$$

where ρ is the density and c is the specific heat. If we now divide both members of
 Equation (3) by the time interval during which the change of the temperature and,
 consequently, the change in the internal energy density takes place, and then we take
 the limit when $\Delta t \rightarrow 0$, we obtain:

$$\frac{\partial u}{\partial t} = \rho c \frac{\partial T}{\partial t}. \quad (4)$$

127 The partial time derivative is considered due to the possibility that u and T depend
 128 on other variables like the point at which they are measured.

Let us consider first the thermal evolution of the Earth's core. In order to obtain
 the corresponding differential equation giving the time derivative of core's temperature,
 we should insert Equation (4) into Equation (2), which leads to:

$$\rho c \frac{\partial T}{\partial t} + \nabla \cdot \mathbf{q} = 0. \quad (5)$$

Then we should integrate over the whole volume of the core. Certainly, core's physical
 parameters, such as its density and specific heat, as well as its temperature, can vary
 both temporally and spatially. However, as we want to track the time evolution of
 core's mean temperature, $\langle T \rangle_c$, we will take the mean values of its density and specific
 heat and will assume that $\langle \rho \rangle_c$ and $\langle c \rangle_c$ are constant in time. Thus, the result of
 the volume integral of the first term on the left hand side of Equation (5) is equal to
 the product of $\rho c \partial T / \partial t$ and the core's volume. In addition, the volume integral of
 the second term on the left hand side of Equation (5) is transformed into a surface
 integral by virtue of the divergence theorem, which should be evaluated on the core's
 surface. If we assume that heat flows outwards only in the radial direction, as shown
 in Figure 1c, then $\mathbf{q}^{\text{out}} \cdot \mathbf{n} = q^{\text{out}}$. Thus, Equation (5) can be rewritten as:

$$\rho_c c_c \frac{d \langle T \rangle_c}{dt} V_c + q^{\text{out}}(t) A_c = 0, \quad (6)$$

where V_c and A_c are the volume and external surface area of the core, respectively.
 Equation (6) can also be expressed as:

$$\frac{d \langle T \rangle_c}{dt} = - \frac{3}{\rho_c c_c} \frac{q^{\text{out}}(t)}{R_c}. \quad (7)$$

129 It should be pointed out that in Equations (6) and (7) we omitted the mean value
 130 symbol for the core's density and specific heat for the sake of simplicity.

131 Concerning the thermal evolution of the mantle, an analogous reasoning is fol-
 132 lowed. We begin with Equation (1) in which we insert Equation (4) and then we
 133 integrate on both members over the volume of the mantle. Then, by assuming that
 134 the mean density, the specific heat, and the temperature of the mantle, as well as the
 135 rate of heat production over its volume, are homogeneous the volume integral of the
 136 first term on the left hand side of Equation (1) is equal to $\rho_m c_m \frac{d \langle T \rangle_m}{dt} V_m$, where ρ_m
 137 and c_m are the mean density and mean specific heat of the mantle, respectively, and

138 V_m is the mantle's volume. In addition, we have assumed that mantle temperature
 139 depends only on time and, consequently, we track the time evolution of the mean
 140 mantle's temperature, $\langle T \rangle_m$.

141 The volume integral of the second term on the left hand side of Equation (1) is
 142 also transformed into a surface integral by virtue of the divergence's theorem. As can be
 143 noted in Figure 1d, the Earth's mantle is enclosed by two concentric spherical surfaces,
 144 the inner of radius R_c and the outer of radius R_\oplus . This implies that the aforementioned
 145 surface integral must be split into two separate surface integrals corresponding to each
 146 limiting surface.

Analogously to the case of the Earth's core, we assume that heat flows outwards
 only in the radial direction and that it depends only on time. Hence, we have that
 $\mathbf{q}^{\text{in}} = q^{\text{in}}(t) \mathbf{n}$ and $\mathbf{q}^{\text{out}} = q^{\text{out}}(t) \mathbf{n}$. Thus, $\mathbf{q}^{\text{in}} \cdot \mathbf{n} = -q^{\text{in}}(t)$ and $\mathbf{q}^{\text{out}} \cdot \mathbf{n} = q^{\text{out}}(t)$. As
 result, Equation (1) is expressed as:

$$\rho_m c_m \frac{d \langle T \rangle_m}{dt} V_m - q^{\text{in}} A^{\text{in}} + q^{\text{out}} A^{\text{out}} = \varrho_m(t) V_m, \quad (8)$$

147 where A^{in} and A^{out} are the areas of the inner and outer spherical surfaces, respectively.

Equation (8) can be reshaped in a form similar to Equation (7). On one hand,
 the mantle's volume is:

$$V_m = \frac{4}{3} \pi R_\oplus^3 (1 - x^3), \quad (9)$$

where

$$x = \frac{R_c}{R_\oplus}. \quad (10)$$

On the other hand, the areas of the inner and outer limiting surfaces are:

$$A^{\text{in}} = 4 \pi R_c^2, \quad (11a)$$

$$A^{\text{out}} = 4 \pi R_\oplus^2. \quad (11b)$$

By inserting Equation (9) and (11) into Equation (8), we obtain:

$$\frac{d \langle T \rangle_m}{dt} = \frac{1}{\rho_m c_m} \left(3 \frac{[x^2 q_m^{\text{in}}(t) - q_m^{\text{out}}(t)]}{R_\oplus (1 - x^3)} + \varrho_m(t) \right). \quad (12)$$

148 Now that we have derived Equations (7) and (12), which give the time evolution of the
 149 core and mantle of the Earth, respectively, the next step is to obtain the corresponding
 150 expressions of the outgoing thermal flux from the core and those entering into and
 151 outgoing from the mantle.

The general expression of the heat flux due to convection is:

$$q^{\text{conv}} = k \frac{\Delta T}{d}, \quad (13)$$

where

$$d = \left(\frac{\text{Ra}_{\text{cr}} \kappa \eta}{2^4 \alpha \rho g \Delta T} \right)^{\frac{1}{3}}. \quad (14)$$

See the Supplementary information for a detailed derivation of Equations (13) and
 (14). Insertion of Equation (14) into Equation (13) leads to:

$$q^{\text{conv}} = k \left(\frac{2^4 \alpha \rho g \Delta T}{\text{Ra}_{\text{cr}} \kappa \eta} \right)^{\frac{1}{3}} \Delta T. \quad (15)$$

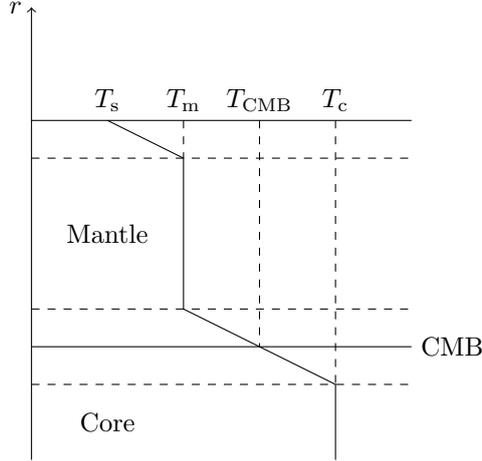


Figure 2: Vertical temperature profile corresponding to the plate tectonics regime due to convection. The earth is assumed to be internally differentiated into two layers, namely the core and the mantle. Both are divided by an imaginary spherical surface called core-mantle boundary (CMB).

The heat flux outgoing from the core will have the same form in both the plate tectonics and stagnant lid regimes. For that reason, we consider its expression here. Taking into account Figure 2 and Equation (15), we find that:

$$q_c^{\text{out}}(t) = k_c \left(\frac{2^4}{\text{Ra}_{\text{cr}}} \frac{\alpha_c g_c}{\kappa_c \nu_c} \right)^{\frac{1}{3}} [\langle T \rangle_c(t) - T_{\text{CMB}}(t)]^{\frac{4}{3}}, \quad (16)$$

where

$$\nu = \frac{\eta}{\rho}$$

152 is the kinematic viscosity, T_{CMB} is the mean temperature at the core-mantle boundary
 153 (CMB) and $\langle T \rangle_c$ is the mean temperature of the core, which is obtained from Equa-
 154 tion (7), and correspond to the isothermal temperature of the thermal convection
 155 model considered in our work.

The heat flux incoming into the mantle has the same functional form for both the considered regimes pointed out above as well. By taking into account Figure 2 and Equation (15) once again, we arrive to:

$$q_m^{\text{in}}(t) = k_m \left(\frac{2^4}{\text{Ra}_{\text{cr}}} \frac{\alpha_m \rho_m g_m}{\kappa_m \eta_m} \right)^{\frac{1}{3}} [T_{\text{CMB}}(t) - \langle T \rangle_m(t)]^{\frac{4}{3}}, \quad (17)$$

where $\langle T \rangle_m$ is the mean (isothermal) temperature of the mantle. The mean temperature at the CMB is obtained by equating the right hand side of Equations (16) and (17), leading to:

$$T_{\text{CMB}} = \frac{\langle T \rangle_m + \varphi \langle T \rangle_c}{1 + \varphi} \quad (18)$$

where

$$\varphi = \left(\frac{k_n}{k_m} \right)^{\frac{3}{4}} \left(\frac{\alpha_n g_n \kappa_m \eta_m}{\alpha_m \rho_m g_m \kappa_n \nu_n} \right)^{\frac{1}{4}}. \quad (19)$$

156 It worth to point out that mantle's dynamical viscosity depends on the T_m according
 157 to Equation (43).

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2.1 Thermal evolution in the plate tectonics regime

The expression of the outgoing thermal flux from the mantle is (see Supplementary material):

$$q_m^{\text{out}}(t) = k_m \left(\frac{2^4}{\text{Ra}_{\text{cr}}} \frac{\alpha_m \rho_m g_m}{\kappa_m \eta_m} \right)^{\frac{1}{3}} [\langle T \rangle_m(t) - T_s]^{\frac{4}{3}}, \quad (20)$$

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where T_s is the temperature at the Earth surface, which is assumed to be constant.

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Equations (16), (17), (18), (19), (20) and (43), together with Equations (7) and (12), allow to perform the numerical simulations of the thermal evolution of Earth's interior once the particular form of ϱ_m has been considered. The latter issue will be approached in Section 3.

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2.2 Thermal evolution in the stagnant lid regime

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The thermal evolution model of the stagnant lid regime includes Equations (16), (17), (18), (19) and (7) without modifications. However, Equations (12) and (20) have to be refined. The main difference lies on the consideration of an external elastic lid, the lithosphere, ranging from the external surface of the mantle up to the Earth surface, of thickness ℓ (see Figure 1b). The lithosphere is assumed to be an external stagnant lid of the Earth in the sense that it has no lateral motions, but its thickness changes on time.

The equation giving the time evolution of the lithosphere's thickness is (Schubert et al., 1979; Spohn, 1991; Grott & Breuer, 2008; Stamenković et al., 2012):

$$\rho_m c_m (\langle T \rangle_m - T_\ell) \frac{d\ell}{dt} = q_\ell - q_m^{\text{out}}, \quad (21)$$

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where T_ℓ and q_ℓ are the temperature and thermal flux at the base of the lithosphere, respectively. Thus, provided that $\langle T \rangle_m$ will always be greater than T_ℓ , the lithosphere's thickness will grow or decrease according to if q_ℓ is greater or lower than q_m^{out} , respectively.

The expression of the outgoing thermal flux of the mantle is slightly different from that of Equation (20). As the lithosphere is no longer part of the convecting mantle, as in the plate tectonics regime, the upper bound temperature is not T_s , but T_ℓ . In consequence, we have:

$$q_m^{\text{out}}(t) = k_m \left(\frac{2^4}{\text{Ra}_{\text{cr}}} \frac{\alpha_m \rho_m g_m}{\kappa_m \eta_m} \right)^{\frac{1}{3}} [\langle T \rangle_m(t) - T_\ell(t)]^{\frac{4}{3}}. \quad (22)$$

The time dependence of the temperature at the base of the lithosphere is due to its dependence on the mean mantle's temperature, provided that it will be computed following the work by Stamenković et al. (2012):

$$T_\ell(t) = \langle T \rangle_m(t) - \frac{\ln 10 R_{\text{gas}} \langle T \rangle_m^2(t)}{E^*}. \quad (23)$$

The expression of the thermal flux entering the lithosphere (q_ℓ) is given by:

$$q_\ell = k_m \frac{(T_\ell - T_s)}{\ell}, \quad (24)$$

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where we assume that the thermal conductivity of the lithosphere is equal to that of the mantle. We may point out that both Equation (13) and (24) were derived assuming steady state regime of heat transport, which is described by $\nabla \cdot \mathbf{q} = 0$. It arises the question about the validity of this approach since we are considering time evolution of the internal temperatures of the Earth. However, as is pointed out by

181 Stamenković et al. (2012), the aforementioned approximation can be justified by the
 182 fact that relaxation time of the lithosphere is typically on the order of 100 Ma. In
 183 contrast, the relaxation time of the mantle is of the order of few Ga, in other words,
 184 the latter is much greater than the former.

185 Another important difference between the thermal model in the plate tectonics
 186 regime and that of the stagnant lid regime, lies on the fact that as the lithosphere
 187 thickness changes on time, also change both the mantle's and lithosphere's volumes.
 188 The change in the mantle's volume affects the expression of the time evolution of the
 189 $\langle T \rangle_m$.

The volume of the mantle in the stagnant lid regime is given by:

$$\begin{aligned} V_m &= \frac{4}{3}\pi \left[(R_\oplus - \ell)^3 - R_n^3 \right], \\ &= \frac{4}{3}\pi R_\oplus^3 (y^3 - x^3), \end{aligned} \quad (25)$$

where we have defined y as:

$$y = 1 - \frac{\ell}{R_\oplus}. \quad (26)$$

Analogously, the surface area of the mantle is computed as:

$$A_m = 4\pi R_\oplus^2 y^2. \quad (27)$$

Bearing in mind Equations (25) and (26), by a reasoning similar to that leading to Equation (12), we arrive at:

$$\frac{\partial \langle T \rangle_m}{\partial t} = \frac{1}{\rho_m c_m} \left(3 \frac{[x^2 q_m^{\text{in}}(t) - y^2 q_m^{\text{out}}(t)]}{R_\oplus (y^3 - x^3)} + \varrho_m(t) \right). \quad (28)$$

190 Thus, the equations governing the thermal evolution in the stagnant lid regime are
 191 Equations (7) and (28) together with Equations (10), (16), (17), (18), (19), (21), (22),
 192 (23), (26) and (43).

193 It worth to note that Equation (12) can be obtained from Equation (28) by
 194 setting $y = 1$ or, equivalently, $\ell = 0$. This means that either there is no lithosphere
 195 at all or we are assuming that the lithosphere is part of the conductive lid of the
 196 convecting mantle as in Subsection 2.1.

197 2.3 Heat sources

198 The functional form of the heat production rate per unit volume and per unit
 199 time depends on the heat sources assumed to be present in the Earth's mantle. It
 200 worth to recall that we assume that heat sources are uniformly distributed in the
 201 mantle. In our work, we will take into account two sources of heat, namely the heat
 202 produced by the decay of radioactive isotopes and tidal heating.

203 On Earth, the most thermally relevant isotopes Uranium (235 and 238), Thorium
 204 and Potassium. On the other hand tidal heating is originated by internal friction within
 205 the tidally loaded Earth's mantle.

If heat is produced only by the decay of radioactive isotopes, then we have that:

$$\varrho_m(t) = \rho_m H(t). \quad (29)$$

206 where $H(t)$ is given by Turcotte and Schubert (2014) and it is reproduced in the
 207 Supplementary material.

However, if we assume that heat is produced by both the decay of radioactive elements and tidal heating, then $\varrho_m(t)$ is given by:

$$\varrho_m(t) = \rho_m H(t) + \frac{\langle P \rangle_m^{\text{tide}}}{V_m}, \quad (30)$$

where $\langle P \rangle_m^{\text{tide}}$ is the mean tidal heating rate in the Earth's mantle. The formula that we will use to compute the aforementioned quantity was developed by Efroimsky and Makarov (2014). However, that formula gives the mean tidal heating rate over the whole volume of the considered body, the Earth in our case. Still, we can use the expression by Efroimsky and Makarov (2014) taking into account that both tidal heating rates are related by (Renaud & Henning, 2018):

$$\frac{\langle P \rangle_m^{\text{tide}}}{V_m} = \frac{\langle P \rangle_{\oplus}^{\text{tide}}}{V_{\oplus}}, \quad (31)$$

where $\langle P \rangle_{\oplus}^{\text{tide}}$ is given by:

$$\begin{aligned} \langle P \rangle_{\oplus}^{\text{tide}} = & \frac{G M_{\zeta}^2}{a} \sum_{l=2}^{\infty} \left(\frac{R_{\oplus}}{a} \right)^{2l+1} \sum_{m=0}^l \frac{(l-m)!}{(l+m)!} (2 - \delta_{0m}) \\ & \times \sum_{p=0}^l F_{lmp}^2(i) \sum_{q=-\infty}^{\infty} G_{lpq}^2(e) \omega_{lmpq} K_{\text{I}}(l, \omega_{lmpq}), \end{aligned} \quad (32)$$

208 where G is the gravitation constant, M_{ζ} is the mass of the Moon, R_{\oplus} is the mean
 209 Earth radius, a is the major semiaxis, δ_{0m} is the Kronecker delta distribution, $F_{lmp}(i)$
 210 and $G_{lpq}(e)$ are the inclination (Gooding & Wagner, 2008) and eccentricity (Giacaglia,
 211 1976) functions, respectively.

The factor $K_{\text{I}}(l, \omega_{lmpq})$ describes the rheological response of Earth, i.e. the modification of its gravitational field due to the deformation caused by the gravitational attraction forces exerted by the Moon, and is given by:

$$K_{\text{I}}(l, \omega_{lmpq}) = -\frac{3}{2} \frac{1}{l-1} \frac{B_l \Im [\bar{J}(\chi)] \operatorname{sgn}(\omega_{lmpq})}{(\Re [\bar{J}(\chi)] + B_l)^2 + (\Im [\bar{J}(\chi)])^2}, \quad (33)$$

where $\Re(\bar{z})$ and $\Im(\bar{z})$ are the real and imaginary parts of the complex number \bar{z} , respectively, and χ_{lmpq} are the physical tidal frequencies which are defined through:

$$\chi_{lmpq} = |\omega_{lmpq}|, \quad (34)$$

where the tidal modes ω_{lmpq} are given by:

$$\omega_{lmpq} = (l-2p)\dot{\omega} + (l-2p+q)\dot{\mathcal{M}} + m(\dot{\delta} - \dot{\theta}), \quad (35)$$

where $\dot{\mathcal{M}}$, $\dot{\omega}$ and $\dot{\delta}$ are the time rates of the mean anomaly \mathcal{M} , the argument of perigee ω and the longitude of ascending node δ (Efroimsky, 2012; Efroimsky, 2015). The latter two angles, as well as the inclination of the Moon's orbital plane i , are defined with respect to the Earth's equatorial plane. In many applications $\dot{\omega} \approx 0$ and $\dot{\delta} \approx 0$ while $\dot{\mathcal{M}} \approx n$. Consequently the approximation

$$\omega_{lmpq} \approx (l-2p+q)n - m\dot{\theta}, \quad (36)$$

is generally valid and frequently used in the specialized literature. In Equation (36), n is the mean orbital frequency which for the unperturbed two-body problem is defined by the mathematical expression of Kepler's third law:

$$G(M_{\oplus} + M_{\zeta}) = n^2 a^3, \quad (37)$$

where M_{\oplus} is the mass of the Earth. In addition, B_l is given by:

$$B_l = \frac{R_{\oplus} (2l^2 + 4l + 3)}{l G M_{\oplus} \rho_{\oplus}}, \quad (38)$$

where ρ_{\oplus} is the mean density of the Earth. Lastly, $\bar{J}(\chi)$ is the complex creep-response or complex compliance function (Efroimsky, 2012; Efroimsky, 2015), which is defined through:

$$\bar{J}(\chi) = \int_0^{\infty} \dot{J}(t-t') \exp[-i\chi(t-t')] dt', \quad (39)$$

where the over-dot means differentiation with respect to t' and $i = \sqrt{-1}$ is the imaginary unit. The particular form of the kernel $J(t-t')$ depends on the particular rheological model considered. However, it is generally given by:

$$J(t-t') = J(0) \Theta(t-t') + \text{viscous and hereditary terms}, \quad (40)$$

212 where $J(0)$ is the instantaneous value of the compliance which, in its turn, is the
 213 reciprocal value of the instantaneous rigidity $\mu(0)$, and $\Theta(t-t')$ is the Heaviside step
 214 function (Efroimsky, 2012).

2.4 Mantle viscosity

The dynamic viscosity, η , of the mantle depends, in general terms, on the temperature (T) and pressure (P) (Valencia et al., 2006; Henning et al., 2009; Stamenković et al., 2012). The functional form of this dependence is usually expressed in the form of Arrhenius' law:

$$\eta(T, P) = b \exp\left(\frac{E^* + P V_{\text{eff}}^*(P)}{R_{\text{gas}} T}\right), \quad (41)$$

216 where b is a constant, E^* and V_{eff}^* are the energy and effective volume of activation,
 217 respectively, and R_{gas} is the Universal constant of ideal gases.

As can be seen in Equation (41), the activation energy describes the coupling between viscosity and temperature, while the effective activation volume describes the coupling between pressure and viscosity. For reasons of simplicity, only the dependence of viscosity on temperature will be considered. In this sense, we set $V_{\text{eff}}^* = 0$ and, consequently, Equation (41) is written as:

$$\eta(T) = b \exp\left(\frac{E^*}{R_{\text{gas}} T}\right). \quad (42)$$

Following Stamenković et al. (2012), the latter equation can be rewritten in a more convenient way by considering a viscosity reference value corresponding to a temperature reference value, i.e. $\eta_{\text{ref}} = \eta(T_{\text{ref}})$. In this way the constant b can be eliminated and Eq (42) becomes:

$$\eta(T) = \eta_{\text{ref}} \exp\left[\frac{E^*}{R_{\text{gas}}}\left(\frac{1}{T} - \frac{1}{T_{\text{ref}}}\right)\right]. \quad (43)$$

3 Model set up and implementation

219 This section is dedicated to discuss relevant details on the geological evidence,
 220 which allows the estimation of the tidal heating rate. In addition, we also present
 221 details concerning the numerical implementation of the thermal models we consider in
 222 our work, including the initial conditions and the assumed values of the parameters.

3.1 Geological evidences of the dynamical evolution of the Earth-Moon system

223 Detailed studies of a special type of rocks, called tidal rhythmites, and of bivalve
 224 shells have allowed the estimation of the time evolution of the rotational period of the
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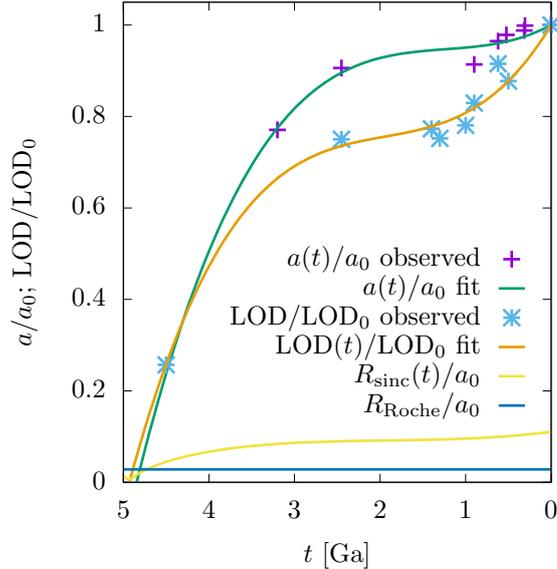


Figure 3: Normalized values of the major semiaxis of the lunar orbit and of the terrestrial length of day, estimated by studying the record of tidal cycles present in tidal rhythmites rocks, (dots) along the Earth’s history. The corresponding polynomial fits of degree three are also shown, as well as the time evolution of the radius of the synchronous orbit, R_{sinc}/a_0 , and the Roche limit R_{Roche}/a_0 .

227 Earth and the orbital period of the Moon. In Figure 3, the estimated values of the
 228 major semiaxis (a) of the Moon’s orbit with respect to the Earth and the length of
 229 day (LOD) of the latter are shown as a function of time during the Earth’s history.
 230 Both data sets are normalized with respect to their current values, a_0 and LOD_0 ,
 231 respectively. The values of a were taken from the work by López de Azarevich and
 232 Azarevich (2017), while those of the LOD were gathered from the works by Williams
 233 (2000) and Spalding and Fischer (2019).

234 In addition, in Figure 3 the plots of the normalized radius of the synchronous
 235 orbit, R_{sinc}/a_0 , and the Roche limit, R_{Roche}/a_0 , are also included in order to show
 236 that the dynamical evolution of the Earth-Moon system, obtained from the geological
 237 evidences mentioned above, is consistent with the dynamical constraints imposed by
 238 tidal theory (Murray & Dermott, 1999).

239 In order to compute tidal dissipation rate as a function of time, along the dy-
 240 namical history of the Earth-Moon system, we fit both data set with third degree
 241 polynomials. The main reason behind this choice is that a third degree polynomial is
 242 the simplest model that better fits the available data. Particularly, such polynomials
 243 represent the apparent change of concavity in the plot of the data. Interestingly, if we
 244 take the time derivative of the polynomial that fits the values of the major semiaxis
 245 along time, and evaluate it at the present, we find that it lies within the uncertainty
 246 of the current accepted value of the time rate of lunar recession measured by Lunar
 247 Laser Ranging (LLR) experiments (Bills & Ray, 1999). Thus, we obtain the major
 248 semiaxis of the lunar orbit and the Earth’s spin rate as smooth functions of time and,
 249 consequently, we can evaluate tidal modes by using Equation (36) where the mean
 250 motion n is computed through Equation (37).

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3.2 Thermal model set up

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The tables with the values of all the parameters necessary to perform the required numerical computations are available at the Supplementary information. These includes the rheological parameters, such as the mean rigidity of the Earth’s mantle μ , the reference shear viscosity η_{ref} (at the reference temperature T_{ref}), the activation energy E^* and the Andrade parameter α . In addition, the parameters and its respective assumed values for the evaluation of the time derivatives of the Earth’s core and mantle temperatures, given in Equations (7), (12) and (28), are also available.

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In order to evaluate the time rates of tidal heating, which are expressed infinite series over the l and q indexes, that is the right hand side of Equation (32), we need to truncate those series.

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As can be noted in Equation (32), the leading terms of the corresponding expansions are of order $(R_{\oplus}/a)^5$ while the next to leading order term is of order $(R_{\oplus}/a)^7$. Currently, $R_{\oplus}/a_0 \approx 0.016$. However, near the origin of the Earth-Moon system the major semiaxis was about $0.3a_0$. Thus, the initial value of $R_{\oplus}/a \approx 0.055$. In consequence, $(R_{\oplus}/a)^5 \approx 5 \times 10^{-7}$ while $(R_{\oplus}/a)^7 \approx 1.55 \times 10^{-9}$. In other words, terms with $l = 3$ are at least two orders of magnitude lesser than those of the leading terms. We performed several numerical experiments comparing results using terms up to $l = 2$ and $l = 3$ in order to assess if appreciable differences appear between both sets of results. Even though differences are naturally expected, we however found that these differences are small enough to be considered negligible. Consequently, we can safely keep terms up to $l = 2$, obtaining simpler expressions of Equation (32).

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Truncation of series in q depends strongly on the eccentricity (see e.g. Noyelles et al., 2014; Veras et al., 2019; Luna et al., 2020; Renaud et al., 2021). In our work, we will assume a fixed eccentricity, equal to its current value. Owing the low value of the lunar orbit eccentricity, it will be enough to take term up to $q_{\text{max}} = 10$, where q_{max} is the maximum value of $|q|$ in the sums over q (Luna et al., 2020). In other words, the sums over q extend from $q = -10$ up to $q = 10$. In general, $|q| \leq q_{\text{max}}$. In a future work, we will explore thermal and dynamical evolution considering different values of the eccentricity.

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Another important issue concerning the values of the lunar orbital eccentricity (e) and inclination (i) with respect to the equatorial plane of the Earth must be addressed. For exploratory purposes, we have considered the current values for the two orbital parameters and assumed them to be fixed. Such an assumption could seem very simplistic, because all the parameters of the Moon’s orbit evolve in time due to perturbations ranging from those of tidal origin to the gravitational attraction from other bodies, principally that of the Sun. In addition, grater values of both e and i can enhance tidal heating by activating modes corresponding to higher-than-synchronous spin-orbit resonances (Renaud et al., 2021). However, tidal heating is much more sensitive to variation of the eccentricity than of the inclination (Efroimsky & Makarov, 2014; Makarov & Efroimsky, 2014; Renaud et al., 2021). In this sense, we can expect that time evolution of the Moon’s orbital inclination with respect to the Earth’s equator has a negligible effect on the thermal evolution of our planet.

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In what respects to the time evolution of the eccentricity, some recent works on the dynamical evolution of the Earth-Moon system that takes into account tidal evolution and solar perturbations (Ćuk & Stewart, 2012; Cuk et al., 2016; Wisdom & Tian, 2015; Tian et al., 2017, and references therein) obtain as result that lunar eccentricity varies between 0 and 0.1 or 0.2. In addition, the work by Rufu and Canup (2020) obtain higher eccentricities. However, it has to be pointed out that the aforementioned works make use of very simplistic rheological models, namely the Constant Phase Lag (CPL) (Wisdom & Tian, 2015; Tian et al., 2017) and the Constant Time

302 Lag (CTL) models (Rufu & Canup, 2020). These models do not describe appropriately
 303 the rheological response of a rocky planet or satellite (Efroimsky & Williams, 2009;
 304 Efroimsky, 2012; Efroimsky & Makarov, 2013). We would like to emphasize that the
 305 Darwin-Kaula expansion of the tidal disturbing potential from which Equation (32) for
 306 the rate of tidal heating (Efroimsky & Makarov, 2014) is derived, has the advantage of
 307 allowing the inclusion of any linear rheology as the ones considered in our work, such
 308 as the Maxwell-Andrade model which is a more realistic rheological model describing
 309 the response of a rocky material under stress (Efroimsky, 2012; Efroimsky, 2015; Re-
 310 naud & Henning, 2018; Renaud et al., 2021). In addition, Maxwell-Andrade model has
 311 proven to be more dissipative than CPL and CTL ones (Renaud & Henning, 2018).
 312 Therefore, we can expect that numerical simulations using realistic rheologies could
 313 lead to a more enhanced eccentricity damping. Under the light of the aforementioned
 314 considerations, setting a constant value for the lunar eccentricity equal to its current
 315 value is a conservative assumption and the resulting rate of tidal heating and, con-
 316 sequently, the thermal evolution of the Earth’s interior can be considered as a lower
 317 bound estimation.

318 In order to identify clearly the results obtained considering each regime, we will
 319 assign them a label and a number. The results obtained under the plate tectonics
 320 regime will be labeled with PT and the number 1 will be added when it is assumed
 321 that heat is only produced by the decay of radioactive isotopes, while the number 2
 322 is included when tidal heating is allowed for, together with the considered rheological
 323 model. Analogously, in the stagnant lid regime, the corresponding results will be
 324 labeled with SL and a number with the same agreement as for the plate tectonics
 325 regime, i.e. when the label is followed by the number 1, only the radiogenic heat
 326 source is being considered. When the SL label is followed by number 2, radiogenic
 327 and tidal heat sources are taken into account, also indicating the particular rheological
 328 model. Each rheological model gives the functional form of the complex compliance, as
 329 is explained in the Supplementary information, and thus defines particular expressions
 330 of the complex Love numbers $K_I(l, \omega_{lmpq})$ which in turn enter the expression for the
 331 tidal heating rate, i.e. Equation (32).

332 Thus, as an example, the results of the thermal evolution considered under the
 333 plate tectonics regime and assuming both tidal and radiogenic heat sources will be
 334 labeled with “PT2”. Similarly, the results considering only the radiogenic heat source
 335 in the stagnant lid regime will be labeled with “SL1”.

336 Regarding the initial conditions, we set the initial mean temperatures to be
 337 $\langle T \rangle_c(0) = 5000$ K and $\langle T \rangle_m(0) = 2000$ K for both the plate tectonics and the stagnant
 338 lid regimes. In addition, for the latter regime we set the initial lithosphere thickness
 339 $\ell(0) = 50$ km.

340 It is important to point out that the chosen values for the reference viscosity η_{ref}
 341 and temperature T_{ref} resulted in an excessive mantle’s temperature increasing that
 342 equates that of the core’s leading to a numerical collapse in our preliminary numerical
 343 experiments. In order to avoid this inconveniences we decided to change the value of
 344 the reference viscosity to $\eta_{\text{ref}} = 4.5 \times 10^{24}$ Pa.s. This practice is used to mimic the
 345 dependence of the viscosity on the pressure (see e.g. Walterová & Běhouňková, 2020).

346 4 Results and discussion

347 We performed numerical simulations considering only the radiogenic heat source
 348 and then adding tidal heating for each of the three considered rheological models and
 349 for both the plate tectonics and stagnant lid regimes.

350 Figure 4 shows the evolution of the mean temperature over time for the Core
 351 and Mantle. As can be noted, Earth’s core and mantle behave differently. Earth’s

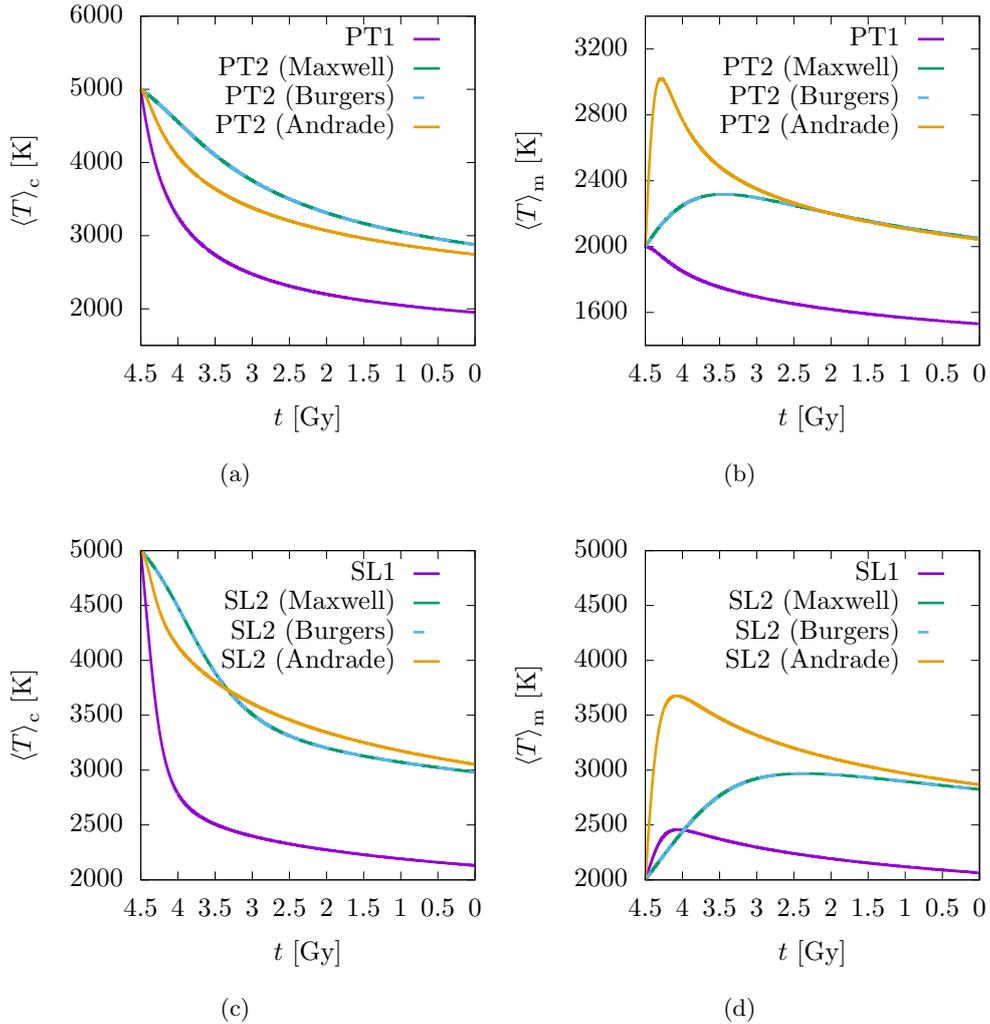


Figure 4: Thermal evolution of the Earth in the plate tectonics regime for the core (a) and mantle (b). Thermal evolution of the Earth in the stagnant lid regime for the core (c) and (d) mantle.

352 core temperature decreases in an exponentially-like fashion. However, the core cools
 353 down slower when tidal heating is taken into account (Figures 4a and 4c). Thermal
 354 evolution of the mantle is greatly affected by tidal heating. The shape of the curve in
 355 the evolution is appreciably changed. The impact of the tidal heating is stronger in
 356 the stagnant lid regime than in the plate tectonics one (see Figures 4b and 4d). When
 357 tidal heating is considered the mantle starts to cool down later than the case where
 358 tidal heating is neglected.

359 Considering that the major contribution of tidal heating is observed in the mantle
 360 (Henning & Hurford, 2014), we will focus our discussion on this layer. If we ignore
 361 the contribution of tidal heating, in the plate tectonics regime, mantle's temperature
 362 decreases uniformly from its initial value, as can be seen in Figure 4b, the corresponding
 363 line is the one labeled with PT1. However, if tidal heating is taken into account,
 364 then the thermal evolution of the mantle differs significantly. Using the Maxwell and
 365 Burgers rheological models, $\langle T \rangle_m$ increases over the first billion of years of evolution,

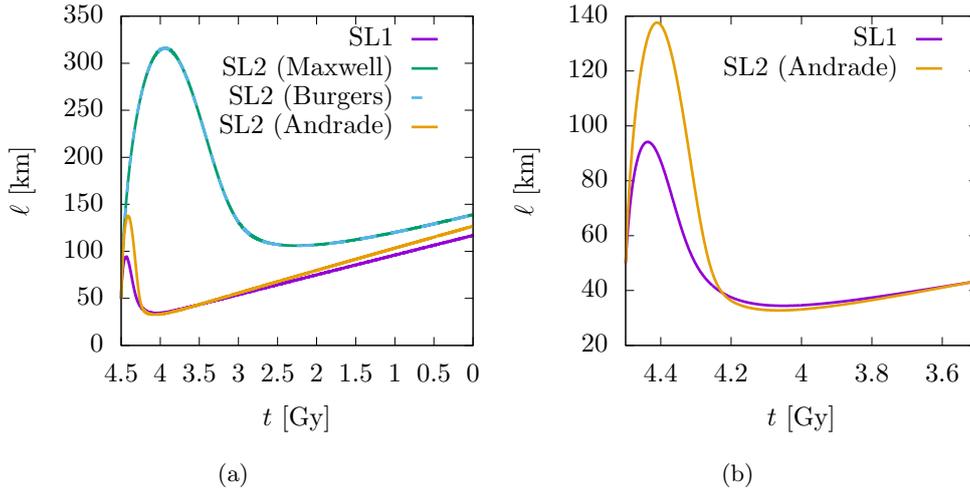


Figure 5: (a) Time evolution of the lithosphere thickness over the Earth’s history. (b) Comparison between numerical experiments SL1 and SL2, using the Maxwell-Andrade rheology, over the time interval from 4.5 to 3.5 Gy in the past.

366 then the temperature starts decreasing. It has to be pointed out that the corresponding
 367 lines of the aforementioned models are overlapped. A more realistic rheology, such as
 368 the Maxwell-Andrade one, results in a temperature overshoot in the first 0.5 Gy before
 369 the temperature begin to decrease.

370 Contrary to plate tectonics regime, in the stagnant lid regime the thermal evolu-
 371 tion shows an increasing of temperature in the first billion of years of evolution either
 372 with or without taking into account tidal dissipation. Nevertheless the increment in
 373 temperature without tidal dissipation is lower reaching only 2500 K, as is shown in
 374 Figure 4d by the curve labeled with SL1.

375 Both the Maxwell and the Burgers rheological models give the same thermal
 376 evolution for the mantle and for each thermal regime. In the plate tectonics regime,
 377 mantle’s temperature begins to decrease after 1 Gy for the Maxwell and Burgers mod-
 378 els, and after approximately 0.25 Gy for the Maxwell-Andrade model. In the stagnant
 379 lid regime, mantle begins to cool down somewhat later. For the Maxwell and Burgers
 380 model mantle’s temperature starts decreasing after 2 Gy, and for the Maxwell-Andrade
 381 models it begins cooling after 0.5 Gy. Again, we observe a temperature overshooting of
 382 about 1600 K in the first 0.5 Gy in the mantle temperature when the Maxwell-Andrade
 383 model is considered in the SL regime.

384 Such temperature increasing could result in partial melting of at least some of
 385 the materials composing the primordial Earth’s mantle. As we have pointed out when
 386 presenting the thermal models (Section 2), these do not take into account phase tran-
 387 sitions. However, for this exploratory work, we are considering only the time evolution
 388 of the average mantle temperature. This means that there could be partial melting
 389 locally in some regions of the mantle, but in no way such temperature increasing or
 390 overshooting means that the mantle is totally melt.

391 Concerning the time evolution of the lithosphere thickness, we observe in Fig-
 392 ure 5a that for all the cases ℓ increases rapidly in the very first 0.5 Gy and then decreases
 393 up to a minimum and then starts increasing again but at a lower rate. It is interesting

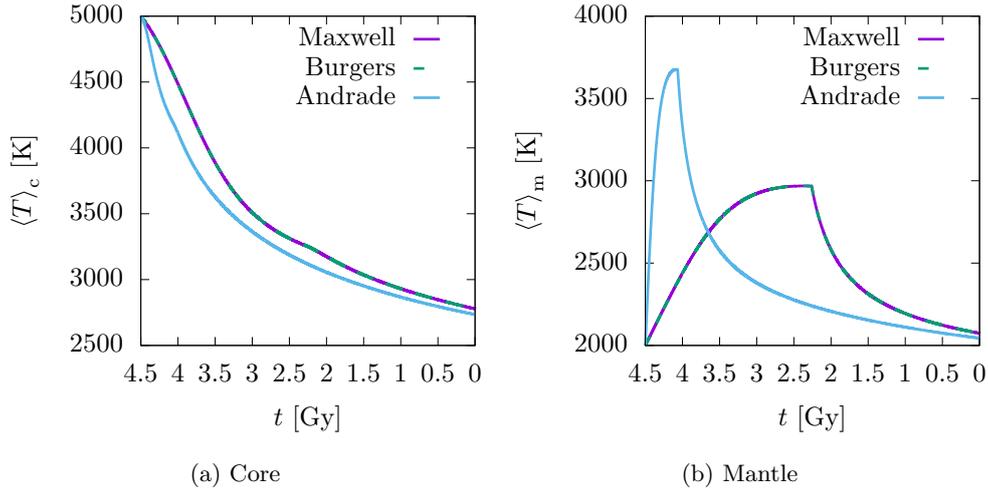


Figure 6: (a) Thermal evolution of the Earth’s core and (b) mantle as result from the combined thermal regime model, exploring the Maxwell, Burgers and Maxwell-Andrade rheological models.

394 to note that when tidal heating is ignored (SL1 curve) and when it is considered, for
 395 the Maxwell-Andrade model, at least for the set of parameters explored, nearly the
 396 same minimum is obtained after the first 500 My of evolution, where the lithosphere
 397 thickness is less than 50 km. However, a somewhat thinner lithosphere is obtained
 398 considering tidal heating with the Maxwell-Andrade model, as can be observed in
 399 Figure 5b.

400 In our model of plate tectonics we cannot evaluate the evolution of lithosphere
 401 due to the inclusion of this layer in the global convection of the mantle. Yet, we
 402 can study the impact of the onset of plate tectonics on the thermal evolution of the
 403 Earth’s interior. To this end, we propose a “combined” thermal model to describe the
 404 transition from the stagnant lid regime to the plate tectonics one. We assume that the
 405 aforementioned regime transition occurs when the curve describing the time evolution
 406 of the lithosphere thickness reaches a local minimum (Figure 5). In this sense, the
 407 change from one regime to the other is forced in the considered model. Although this
 408 is an arbitrary criterion, it corresponds to an “educated guess” because we expect that
 409 a thin lithosphere yields to a more favorable scenario for the plate subduction and,
 410 consequently, the onset of plate tectonics.

411 We performed three numerical simulations, taking in account the tidal heating
 412 and using the same initial conditions for the core’s and mantle’s temperatures and
 413 lithosphere thickness, exploring the three rheological models proposed. The results
 414 are shown in Figure 6a and 6b. The former depicts the time evolution of the mean
 415 core temperature, while the latter shows the time evolution of the mean mantle
 416 temperature.

417 In total correspondence with Figure 5, it can be seen in Figure 6b that the
 418 Maxwell-Andrade model leads to a regime transition about 2 Gy earlier than Maxwell
 419 and Burgers models. The regime transition occurs after the first 0.5 Gy of evolution
 420 when the Maxwell-Andrade model is considered. The Maxwell and Burgers models
 421 predict a regime transition between 2 and 2.5 Gy in the past.

422 Maxwell-Andrade model also produces a temperature overshoot, as in the pre-
 423 vious results, which is rapidly damped, as well as the temperature increase produced
 424 by the Maxwell and Burgers models, showing the high efficiency of convection in the
 425 plate tectonics regime to remove heat from the mantle. In Figure 6a we can observe
 426 that thermal evolution of the core reflects that of the mantle. Figure 6b shows that
 427 using the Burgers or Maxwell rheologies the Mantle cooling delays until 2 Ma while
 428 the Andrade rheology predicts a minimum lithosphere and mantle cooling as early as
 429 4 Gy ago.

430 In Section 3.1 we considered a dynamical evolution model (DEM) which describes
 431 the time evolution of the major semiaxis of the Moon’s orbit and the length of Earth’s
 432 day. That model was obtained by fitting the geological data to a cubic polynomial.
 433 However, we may now wonder how would be affected the thermal history of the Earth
 434 if the Moon recess and the Earth spins down in different ways. In order to answer
 435 that question, we proposed two different linear models for the dynamical evolution.
 436 We identify the three models by the label DEM and an ID number. Thus, DEM1
 437 correspond to the three cubics fitting the geological data. DEM2 is obtained
 438 by assuming both that the Moon is spiraling outwards at a constant rate equal to
 439 the current value of the time derivative of the major semiaxis with respect to time,
 440 obtained by LLR, i.e., $(da/dt)_0 = 3.82 \times 10^{-2} \text{ m yr}^{-1}$ (Bills & Ray, 1999) and that
 441 the LOD increases linearly from its estimated initial value of 6.15 hours (4.5 Gy ago)
 442 to its current value of 23.93 hours. It worth to point out that the latter DEM implies
 443 an initial value for the major semiaxis of about $2.125 \times 10^8 \text{ m}$. Finally, DEM3 also
 444 correspond to constant recession of the Moon from an arbitrary value of $20 R_{\oplus}$ to
 445 its present value (Table S2), while LOD increases linearly from only 2.53 hours to
 446 its current value (Tian et al., 2017). In Figures S3 and S4 of the Supplementary
 447 information we plot the assumed time evolution of the major semiaxis of the lunar
 448 orbit and the Earth’s LOD, respectively.

449 We evaluated these three evolution models of Moon’s major semiaxis and Earth’s
 450 rotation period and how they impacted in Earth interior temperature applying the
 451 combined model which considers the stagnant lid regime until a minimum value of ℓ
 452 is reached and then continues the integration considering the plate tectonics regime.
 453 The results are shown in Figure 7. As can be noticed in Panels (a), (b), (d) and
 454 (e) of Figure 7, we find that there is no difference between the thermal histories of
 455 the Earth’s core and mantle when the Maxwell and Burgers rheological models are
 456 considered. However, it is very interesting how relevant can be the influence of the
 457 dynamical history of the Earth-Moon system on the thermal evolution of the Earth’s
 458 interior when the Maxwell-Andrade model is used to model the rheology of the mantle.
 459 As it is shown in Figure 7c and 7f three different thermal evolution are obtained, each
 460 corresponding to a particular DEM.

461 The curve corresponding to DEM1 in Figure 7f, describes the same thermal
 462 history of the mantle as that of Figure 6b. Both figures show a mean temperature
 463 overshoot of about 1700 K over its initial value and then start decreasing rapidly due
 464 to the change from the stagnant lid regime to the plate tectonics one. This reflects the
 465 fact that DEM1 is characterized by a swift increase of the mean distance between Earth
 466 and Moon. If our natural satellite moves away more slowly, as prescribed by DEM2,
 467 we obtain a somewhat higher temperature overshoot that becomes damped later than
 468 the latter case. As expected, if the Moon was farther away at the beginning of the
 469 dynamical history of the system it forms with Earth, and receding at a constant rate
 470 equal to the current value, then the mean mantle temperature increases up to 3000 K
 471 and the regime transition takes place about 2 Gy later than the aforementioned cases.

472 It worth to point out that geological evidences support the hypothesis of an
 473 initially fast rotating Earth. In addition, several theories were capable to reproduce
 474 the current dynamical state of the Earth-Moon system assuming the formation of the

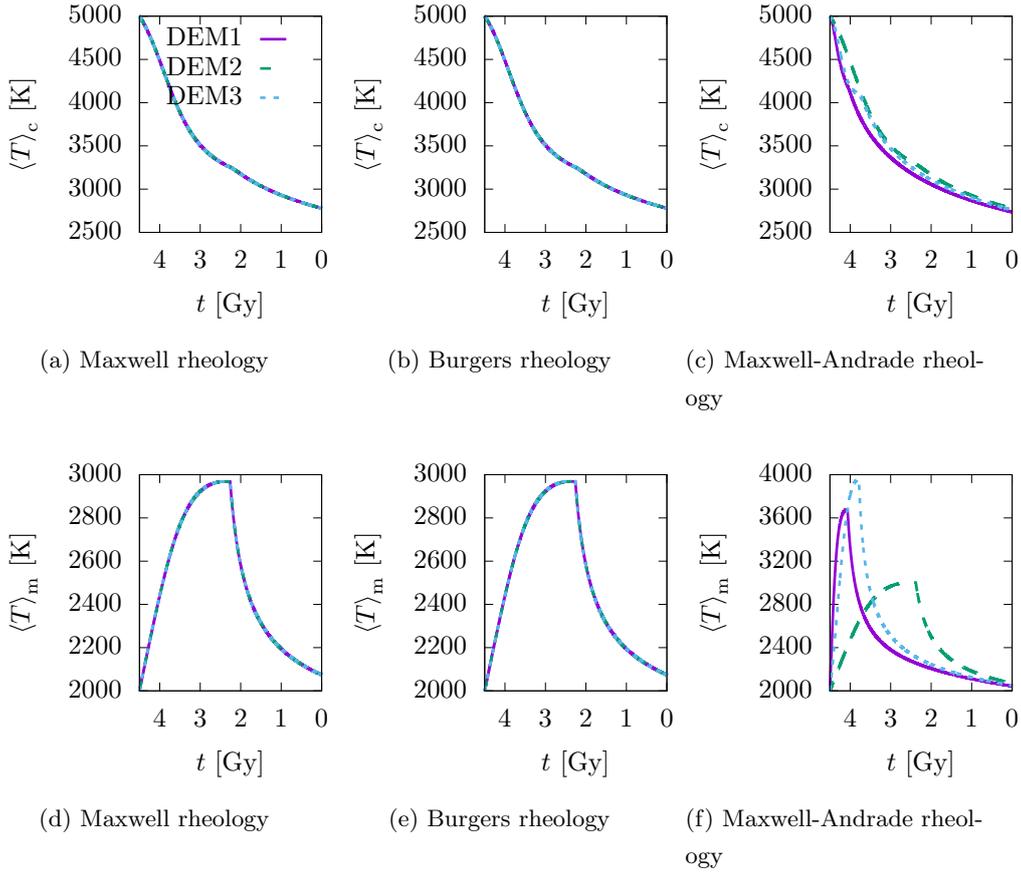


Figure 7: Thermal evolution of the Earth’s core –(a), (b) and (c)– and mantle –(d), (e) and (f)– as result of the numerical experiments performed considering three different dynamical evolution models (DEM1, DEM2 and DEM3), for the three rheological models considered in our work.

475 Moon from a giant impact (Hartmann & Davis, 1975; Cameron & Ward, 1976; Canup,
476 2004, 2008).

477 However, the so-called “canonical” theory has two main difficulties. On one
478 hand, it can not explain satisfactorily the isotopic similarities between Earth and
479 Moon (Lugmair & Shukolyukov, 1998; Wiechert et al., 2001; Touboul et al., 2007;
480 Meier, 2012; Zhang et al., 2012; Young et al., 2016). On the other hand, that theory,
481 which is based on numerical simulations of the Moon-forming giant impact, assumes
482 the conservation of the total angular momentum of the Earth-Moon system along its
483 whole history.

484 In order to reconcile the giant impact theory with isotopic similarities between
485 the Earth and Moon, alternative models have been proposed (Pahlevan & Stevenson,
486 2007; Čuk & Stewart, 2012; Canup, 2012; Reufer et al., 2012). In particular, the works
487 by Canup (2012) and Čuk and Stewart (2012) propose that the aforementioned diffi-
488 culties may be overcome by loosening the angular momentum conservation constraint
489 assuming a rapidly rotating Earth after the impact. The results of the numerical
490 simulations performed under the latter assumption not only agrees with the geologi-

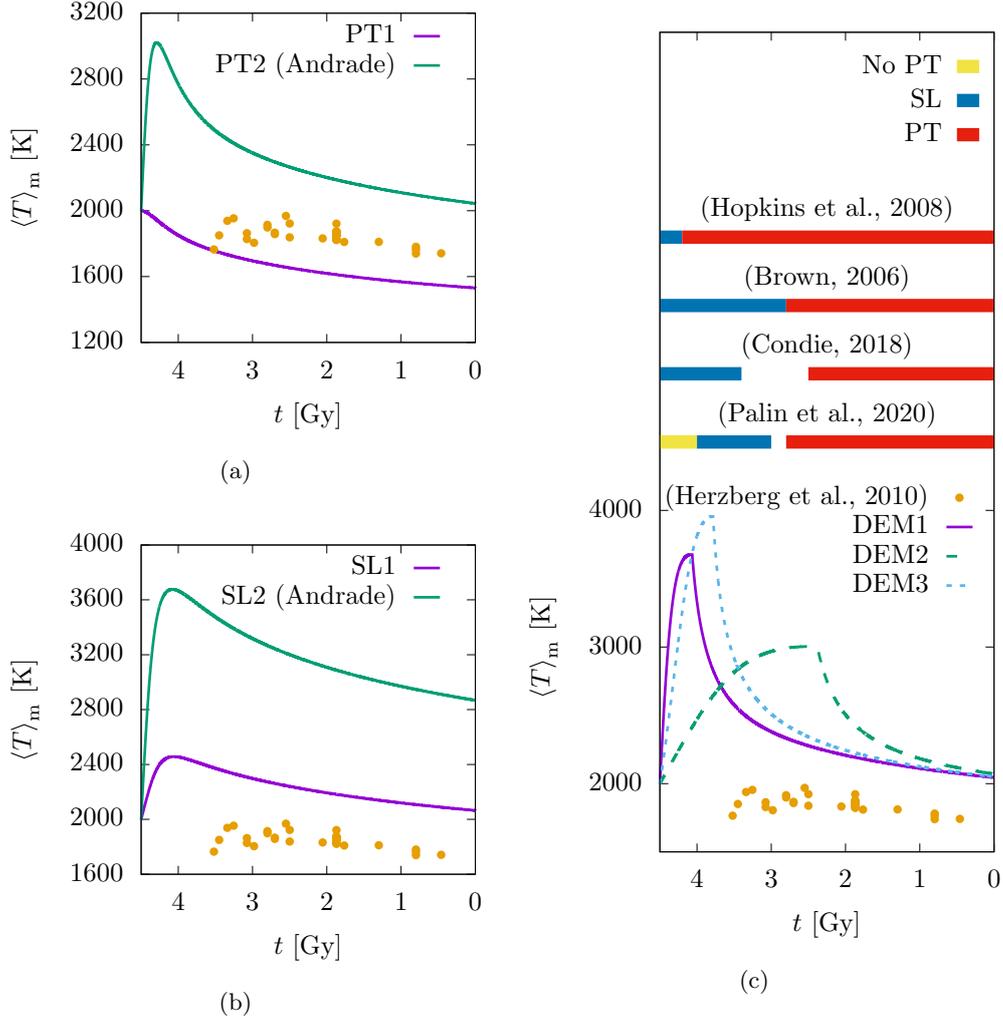


Figure 8: Comparative plots between the results obtained from our numerical simulations, under the plate tectonics (a), stagnant lid (b) and the combined model (c), and the results from the work by Herzberg et al. (2010).

491 cal data, but also agree with the empirical fact that the Moon is formed mainly by
 492 mantle-derived materials.

493 5 Conclusions

494 In this work, we explored the influence of tidal heating on the thermal evolu-
 495 tion of Earth’s interior. To this end, we investigate two end members of thermal
 496 model regimes, namely plate tectonics and stagnant lid regimes. Moreover, for each
 497 of those scenarios we evaluate different rheological models for the Earth’s mantle. For
 498 these models we have considered the decay of the most thermally relevant radioactive
 499 isotopes and tidal heat sources.

500 As result of the numerical simulations carried out, we obtained that tidal heating
 501 could have played a relevant role in the thermal history of the Earth in the first billions

502 of years. In addition, the study of the time evolution of the lithosphere thickness
 503 opens the possibility of studying the consequence of tidal heating on the onset of plate
 504 tectonics.

505 The importance of the rheological model applied for modeling should be empha-
 506 sized. In this sense, the influence of tidal heating is not trivial and its relevance for
 507 the Earth’s thermal evolution depends on the model used. Moreover, our results show
 508 that it becomes crucial to know the evolution of the Moon’s orbit and Earth’s rotation
 509 properly, to be able to model the terrestrial thermal evolution.

510 Taking into account that the closer the Moon to the Earth the more tidal heat
 511 it brings to the latter, we conclude that the orbital evolution controls not only the
 512 magnitude of tidal heating but also the rate at which mantle cools down. Throughout
 513 our numerical experiments we have demonstrated that the steeper the slope of the
 514 recession rate the greater the temperature loss.

515 After our analysis, it is interesting to compare the results with geological evi-
 516 dences on the onset of plate tectonics and terrestrial geodynamics. In this sense, we
 517 compared our results with the data from the work of Herzberg et al. (2010) for mantle
 518 temperature at different periods of Earth history.

519 As shown in Figure 8a the mantle temperature data are higher than our plate
 520 tectonic model without tidal heating, and at the same time lower but close to the plate
 521 tectonic model taking into account tidal heating. This is interesting since the data from
 522 the work by Herzberg et al. (2010) represent the upper mantle (shallower than 300 km
 523 depth), while our model represents the average mantle temperature. Thus, the fact
 524 that the data are above the TP1 curve suggest that an extra heat source is needed
 525 to reproduce those data since the mean mantle temperature must always be higher
 526 than that of its outermost regions. The TP2 curve laying above the data suggests
 527 that tidal heating input could have been necessary in the Earth’s history. In Figure 8b
 528 it is observed that, on one hand, thermal models considering the stagnant regime
 529 describe a higher mantle temperature since the SL1 and SL2 model curves are above
 530 the data. However, a stagnant lid regime over the entire Earth’s history is infeasible.
 531 On the other hand, tidal heat would not be needed in the stagnant lid regime. In
 532 addition, thermal model considering the stagnant lid regime including tidal heating
 533 would have generated a temperature rise in the early Gy and a slow decline (O’Neill,
 534 2020) obtaining a similar pattern reminiscent of the data.

535 In the previous section, we analyzed different dynamical evolution models (DEMs)
 536 of the Earth-Moon system, concluding that according to the model used, the recession
 537 rate of the Moon has a significant influence on the thermal evolution of the mantle
 538 (Figure 7f). Comparing these results with the different estimations for the origin of
 539 plate tectonics, (Figure 8c) we observe that our models coincide with the time window
 540 that different authors propose for the onset of plate tectonics (Brown, 2006; Hopkins et
 541 al., 2008; Condie, 2018; Palin et al., 2020). In addition, we also conclude that since the
 542 thermal evolution of the mantle strongly depends on the dynamical evolution model,
 543 it may be interesting to use geological evidence and thermal models to constrain the
 544 dynamical evolution of the Earth-Moon system.

545 According to our models, the minimum of lithospheric thickness is reached within
 546 the proposed time window for the onset of plate tectonics. As Korenaga (2013) points
 547 out, in the past the buoyancy of the lithosphere would have been greater due to greater
 548 thickness, preventing plate tectonics scenario. Before this time our models also predict
 549 a greater thickness which would lead to greater buoyancy and inhibition of subduction,
 550 as was also proposed by some authors (Korenaga, 2013, and reference therein).

551 Taking into account different models of dynamical evolution to evaluate the tem-
 552 perature of the mantle, we conclude that the thermal trend of the Earth become spiky,

553 with larger amplitudes when the tidal heating is considered (Figures 4 and 5). This ex-
 554 tra heat could have had some impact in the large geologic and geochemical changes that
 555 would have occurred between 2 and 3 Gy ago (Condie, 2018; Palin et al., 2020). This
 556 leads us to propose for a future work that using geological and petrological evidence
 557 of the temperature of the Earth’s mantle could give clues to lunar orbital evolution
 558 and Earth’s rotational evolution by using this type of models backwards.

559 Our work shows that tidal heating could have played a predominant role in the
 560 early stages of terrestrial evolution and should be considered in thermal modeling.
 561 However, the magnitude by which it would have been affected is highly dependent
 562 on the rheological model used. We conclude that our results can contribute to the
 563 understanding of the Earth’s dynamics in the hadean/archean and should be taken
 564 into account when studying the origin of plate tectonics on Earth.

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