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boundary in the initial steps of the modeling.

Once a lithospheric layer is formed, the thermal properties of the upper layer need to be adjusted accordingly. Above the mantle, we included an outer layer in our model representing the atmosphere. The temperature beyond our atmospheric layer was equal to that of outer space.

2.1.1. INITIAL TEMPERATURES

In our model, the initial temperatures are constant in each of the layers, including the atmosphere (Figure 1). The initial temperatures of our model are based on the results of a smoothed particle hydrodynamics (SPH) simulation (Canup, 2008). Figure 1a illustrates the thermal results of a collision with the canonical impact model (Canup, 2008) and the temperature profiles that we used as initial conditions in our models. The temperature profiles of the silicates after impact are shown in blue and those of iron components in red in Figure 1a. The orange and green lines are the temperature profiles we use to constrain initial maximum and minimum temperatures for modeling. Although these lines are not the exact envelopes of the thermal results of the collision, they are representative of the maximum and minimum temperatures of most of the fragments that were accreted to form the Earth. Those profiles are shown as the red and blue lines, respectively, in Figure 1b.

The atmosphere temperatures of 2000 and 4000 K were chosen because they are reasonably expected values after a large collision, in addition to being the most used in previous works (*e.g.*, Nakazawa *et al.*, 1985; Abe, 1993; Canup and Esposito, 1996; Sleep *et al.*, 2001; Zahnle, 2006; Zahnle *et al.*, 2007). All the values used during the modeling are shown in Table 1.

2.1.2. HEAT TRANSFER

The balance between the heat generated in the formation of the Earth and the heat released into space allows us to understand its thermal evolution (Hofmeister, 2020). Heat transfer can occur through three different mechanisms: conduction, radiation and convection. Radiation is the emission of energy in the form of electromagnetic waves from surfaces of finite temperature, which can travel through a vacuum.

Conduction is the transfer of energy within a medium through the movement and interaction of particles. As constantly happens when nearby molecules collide, a transfer of energy in the direction of decreasing temperature must occur. The random molecular motion that causes the net transfer of energy can be seen as energy diffusion.

Finally, convection is the transfer of heat through two different mechanisms, diffusion and advection. Advection is the movement of the fluid and diffusion allows the transfer of energy by molecular movement (Incropera *et al.*, 2007). It is necessary to clarify that the movement of the material does not imply the transfer of heat, but the way in which a greater quantity of particles of abundant energy make contact with particles of lower energy. In the presence of a temperature gradient, the movement of the material contributes to heat transfer. That is, convection can be considered as conduction favored by movement (Çengel, 2006).

In general, heat flow is proportional to the temperature differences between two locations, flowing from areas with higher heat to those with lower heat. This always happens regardless of the heat transport mechanism. Thus, heat transport by the different mechanisms can be described using the following expressions (Incropera *et al.*, 2007):

$$q_{rad} = h_{radiative}\,(T_2 - T_1) \tag{1a}$$

$$q_{cond} = -k\frac{dT}{dx} = -\frac{k}{L}\,(T_2 - T_1) \tag{1b}$$

$$q_{conv} = h_{convective}\,(T_2 - T_1) \tag{1c}$$

Equations (1) show on the left side the heat flow for each mechanism. On the right side, a proportionality constant (depending on the heat transport mechanism) multiplies the temperature difference

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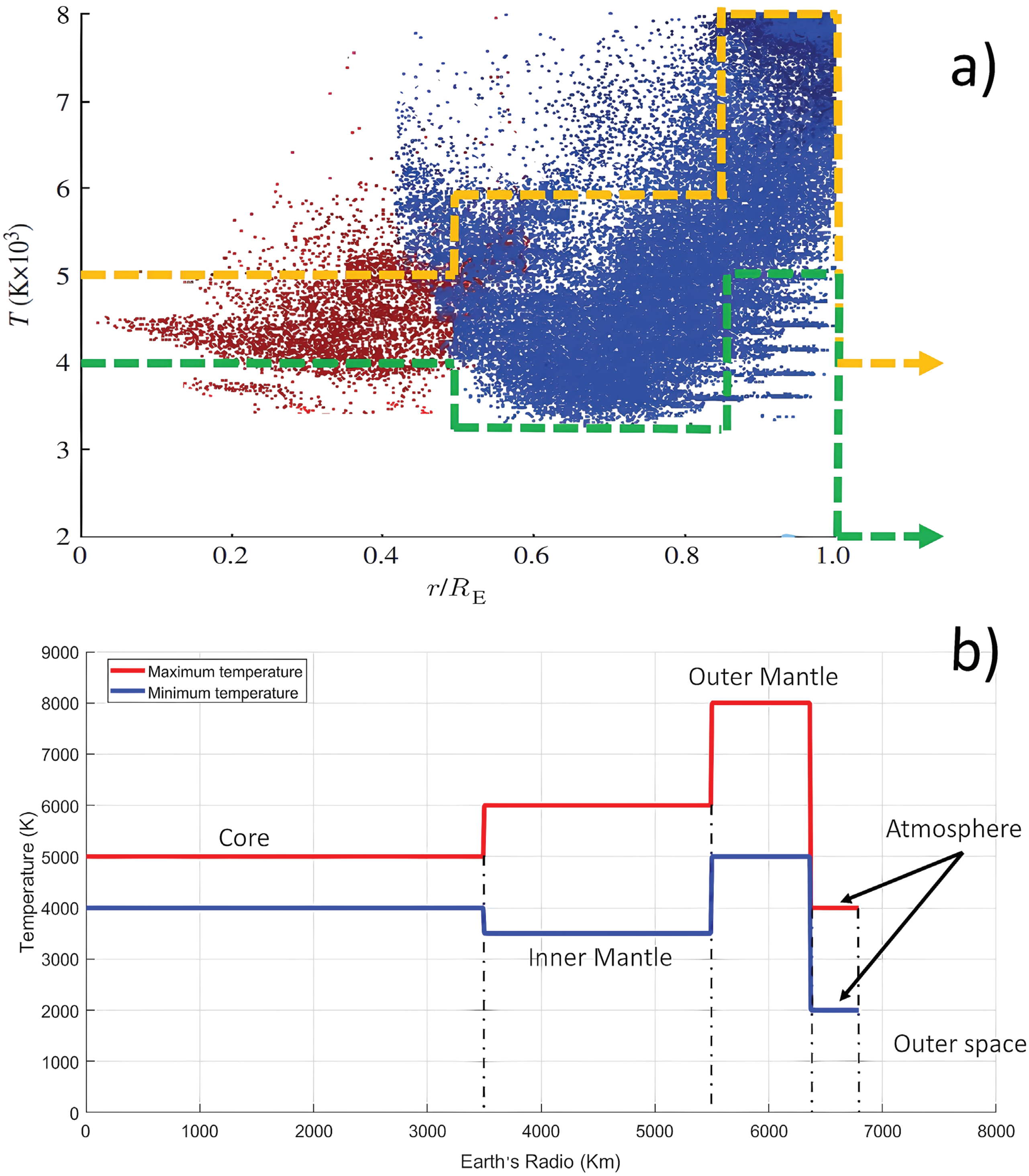

Figure 1 a) temperature in the proto-Earth as a function of depth (where r is the distance from the planet's center and RE 6378 km) from Canup (2008). Red and blue points show the original proto-Earth material at the simulation's final time step (31 h), while dark blue points (upper right) are silicate particles originating from the impactor that are accreted by the proto-Earth. The green and orange lines are the temperature profiles we use to constrain initial maximum and minimum temperatures for our model. Those profiles are shown in b) without the points from the SPH simulation.

between two places. Heat flow will occur in any medium or between media where there is a temperature gradient. The proportionality constant depends on the particular mechanism; in equation 1a, $h_{radiative}$ is associated with the Stefan Boltzmann constant and the surface emissivity. In equation 1b, thermal conductivity is the material property that determines the speed of heat transfer. In equation 1c, $h_{convective}$ depends on the characteristics of the material, the type of convection or the phase changes it may experience.

Equations (1) have the same general form because "heat transfer (or heat) is thermal energy in transit due to a spatial temperature difference" (Incropera *et al.*, 2007). Thus, regardless of the details of the mechanism of transport, heat transfer can be modeled with the appropriate choice of the proportionality constant. This proportionality constant is known in engineering as effective thermal conductivity. Its numerical value can be expressed as a multiple of the thermal conductivity of the material being modeled (see below). Although this approximation cannot be used to appreciate the mechanical aspects of convection (*i.e.* the displacement of the fluid involved), it provides a solid argument in relation to its thermal effects.

2.1.3. HEAT EQUATION

Our model uses the heat diffusion equation to calculate the temperature as a function of radial distance from the center of the planet. In general, the heat diffusion equation can be written as:

$$\frac{\partial T}{\partial t} = \frac{1}{r^2}\alpha\frac{\partial}{\partial r}\left(r^2\frac{\partial T}{\partial r}\right) + R \quad (2)$$

where a is thermal diffusivity, T is temperature, t is time, R includes the external heat sources and r is the radius of a sphere. During the magma ocean phase modeled in this work, the radiogenic heat production rate (R) is negligible, and therefore it could be excluded from the calculations without affecting the main conclusions (see appendix A). Nevertheless, for the sake of completeness, the influence of this term is considered below.

Thermal diffusivity is defined in terms of the thermal conductivity (k) and the product of the specific heat of the body (Cp) and its density (ρ): $a = k/(Cp*\rho)$. Diffusivity measures the ability of a material to conduct thermal energy relative to its ability to store thermal energy (Incropera *et al.*, 2007). When the diffusivity is small the material has a slow response to energy transfer; a large diffusivity has a fast response to heat transfer (Hofmeister and Criss, 2019; Hofmeister, 2020). Thus, thermal diffusivity governs how quickly thermal fields change, while thermal conductivity describes the amount of thermal energy that moves down a thermal gradient (Whittington, 2019; Hofmeister, 2020).

The radiogenic heat production rate is calculated with equation 3, using the parameters from Chapter 2 in Hofmeister (2020).

$$R = 3.51 * 10^{-3}\xi_{K,ppm}e^{0.5543t} + 0.0263\,\xi_{Th,ppb}e^{0.0495t} + \xi_{U,ppb}[0.0943\,e^{0.1551t} + 0.00421e^{0.9849t}] \quad (3)$$

$\xi_{K,ppm}$, $\xi_{Th,ppb}$ and $\xi_{U,ppb}$ are the present-day concentrations of the bulk elements in the rock in the indicated units. t is the age of interest. In the present work, the production of heat by radiogenic elements is homogeneously distributed throughout the mantle and R is reported as picowatts per kg of rock.

2.1.4. EFFECTIVE THERMAL CONDUCTIVITY

During the magma ocean phase on Earth, after the impact, the planet was most likely completely molten due to the high temperatures that were reached (Tonks and Melosh, 1993; Canup, 2008). Estimating a specific thermal diffusivity for this early period on the planet is difficult. Nevertheless, as a first approximation, we can use an effective thermal conductivity capable of representing the effects of the convective movement of the mantle. Effective thermal conductivity has been used before to model convection during magma ocean solidification with novel results (Monteux *et al.*, 2016; Monteux *et al.*, 2020). In our case,

solidify. In panels a) and b) of Figure 2 the slow heat transfer from the mantle causes surface solidification regardless of how effective the atmosphere is in transmitting heat to space.

3.2. MINIMUM TEMPERATURE ENVELOPE

This section shows cases for the minimum temperatures as initial conditions. To facilitate comparison, the k_{conv} values and color pattern in Figures 2 and 3 are the same. In panel a) the mantle has weak convection and an atmosphere with slow heat transfer. Because the heat flow from the mantle is very low the surface takes a long time to solidify. Nevertheless, we observe that the lithosphere can form while the planet remains in a liquid state. In panel b) the heat transfer in the atmosphere is faster, so the solidification of the surface is more easily achieved while also keeping the planet mostly liquid. When convection is vigorous as in panels c and d, interior temperatures drop rapidly. In these scenarios, the temperatures in the atmosphere are lower than in Figure 2, so the cooling is faster, causing the solidification of the surface, preserving liquid material inside. This happens regardless of whether the atmosphere exhibits fast or slow heat transfer.

The formation of the lithosphere favors the preservation of a greater part of the mantle in a liquid state for a certain time. This can be easily seen if we compare the blue curves in panels c and d of Figure 3. In c) the planet's surface remains liquid for almost 300000 years while in d) it remains liquid for only 100 years. In either case, the solid lithosphere serves as a relatively insulating layer that keeps heat inside the planet. Although this difference in the amount of liquid material appears to fade over time (red curves in panels c and d of Figure 3), it may be important in the early stages of the planet's solidification.

3.3. TIME OF LITHOSPHERE SOLIDIFICATION

In the previous section, only the extreme scenarios of our 98 simulations were shown. To visualize the results of all our simulations, we concentrate on the estimates of the solidification time yield by each simulation. The surface solidification time is

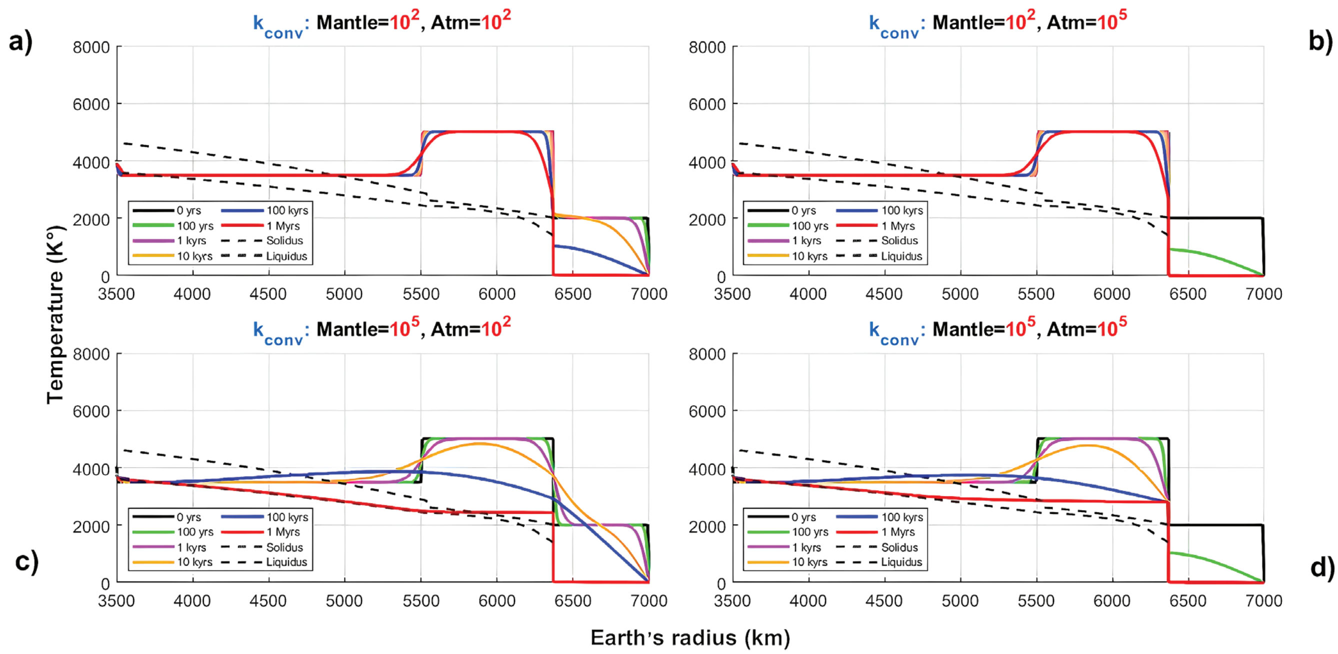

Figure 3 Thermal profiles of the post-impact thermal evolution using the minimum temperature envelope as initial condition. All symbols and panels as in Figure 2.

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obtained as the time in which the geotherm drops below the solidus temperature at the surface (1400 K). We also calculated the solidification time of the mantle as the time it takes to reach >90% crystalline material throughout its entire length (*i.e.*, there is no point in the entire mantle with more than 10% liquid material). Although this time of reference does not mark a total solidification in the strictest sense of the word, it suffices to ensure that there are no longer areas of abundant liquid material in the mantle. This also allowed us to reduce the computational calculation time to < 200 Myrs after the impact (in most cases).

In Figure 4 the colors represent the post-impact time of solidification for the lithosphere (upper two diagrams) and for the entire mantle (lower two diagrams). On the horizontal axes are the k_{conv} values for the atmosphere and on the vertical axis are the mantle values for each simulation.

The time of lithosphere formation is shown numerically in years (Figures 4a and 4b; Tables 2 and 3) while millions of years are used for the mantle (Figures 4c and 4d; Tables 2 and 3). The results based on the maximum and minimum temperature envelopes as initial conditions are shown in the left side and right side, respectively. As observed in those diagrams, the effect of the atmosphere is crucial in the first stages of solidification of the planet. It is observed that the formation of the lithosphere occurs rapidly when the k_{conv} value of the atmosphere is > 10^3 regardless of the value of the k_{conv} of the mantle. Importantly, even when the time of solidification of the lithosphere may be large, all the simulations indicated that the interior of the magma ocean remained liquid for much longer. Also, the importance of the atmosphere in controlling the speed of cooling of the interior of the planet decreases once the lithosphere has formed.

To establish a clear relationship between the time it takes for the solidification of the mantle and the surface, we calculated their ratio. The

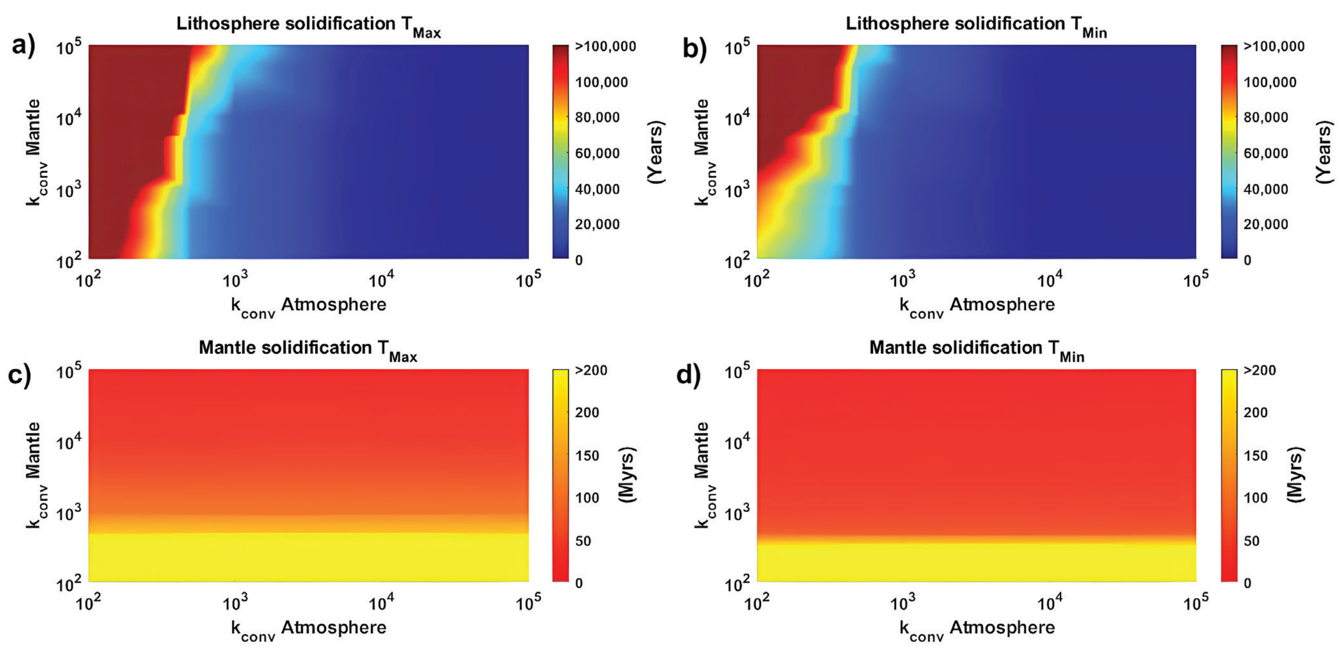

Figure 4 Time of lithosphere solidification as a function of different values of the convective thermal conductivity k_{conv} for the atmosphere and the mantle. Left maximum temperature envelope, right minimum temperature envelope. Above is the solidification time of the lithosphere (a and b), down the solidification time of the mantle (c and d). Data from this picture is reported in Tables 2 and 3.

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Despite this, the difference in surface solidification time does not have as much impact on the solidification of the interior as the formation of the lithosphere. That is, the formation of the lithosphere occurs before the complete solidification of the mantle regardless of the speed at which the atmosphere transmits heat to space. Therefore, the solidification of the planet's interior can be modified by the appearance of an insulating layer such as the lithosphere.

4.1. ATMOSPHERE AFTER THE GIANT IMPACT

Normally, it is considered that the degassing of the magma ocean produced the greenhouse gases that kept the magma ocean on the planet's surface. Nevertheless, the formation of a proto lithosphere can limit the outward flow of heat, which would lead to increasingly cold climatic conditions. Thus, if the outgassing did not take place quickly, the rapid cooling of the atmosphere could have caused the formation of the lithosphere before the mantle outgassed (Scenarios b and d of Figures 2 and 3). The chemical evolution of the atmosphere therefore, would not have included the release of greenhouse gases, and further cooling from the top of the magma ocean would have been promoted. On the other hand, if the magma ocean can get rid of its volatiles quickly, the greenhouse gas content in the atmosphere would increase. As our results show (Scenario c in Figures 2 and 3), this would become a positive feedback process in which the magma ocean causes the atmosphere to maintain its insulating condition and lengthens the lifetime of the magma ocean. The evidence of liquid water on the planet's surface from zircons and the existence of life suggest that this path was not followed by Earth (Wilde *et al.*, 2001). However, this process could have been followed by Venus. The difference between Earth and Venus, could have been an order of magnitude in the heat transfer from the atmosphere to outer space in the beginning of the magma ocean cooling. Potentially the difference in the solar radiation received by the two planets could have contributed to their different evolutions of those two planets (Hamano *et al.*, 2013).

It is remarked that the formation of the Earth's atmosphere is a difficult problem still to be solved. The atmosphere could have formed from planetary accretion before the dissipation of the solar nebula, so it could have acquired a nebular composition. It may also have been formed by outgassing from accretional impacts during the creation of proto-Earth (Ahrens *et al.*,1989; Zahnle *et al.*, 1988). A third possibility is that it could be a product of the degassing of the magma ocean after the impact that gave rise to the Moon (Elkins-Tanton, 2012). In addition, it could have originated from the outgassing of cometary impacts in its last phase of accretion after the giant impact (Wetherill, 1980). Finally, it could have been the result of a random combination of all the processes that have been mentioned. Although there is isotopic evidence supporting each of the different possibilities, it is difficult to encompass all of them in a single model (Pollack and Yung, 1980).

One of the reasons why the argument has been made for a strong greenhouse atmosphere is because of the faint young Sun problem (Zahnle *et al.*, 2010). If the solar radiation in the early years of the Earth was weaker than it is today, a greenhouse effect of the atmosphere would have been necessary to allow the formation of liquid water. This forces us to explain the capture of all the CO_2 that potentially formed that powerful atmosphere to initiate habitable climatic conditions. Furthermore, it is necessary to consider that liquid water on the surface can only form in certain proportions of H_2O and CO_2 (Miyazaki and Korenaga, 2022b). Some zircons from 4.4 Gyrs ago point to the existence of continental lithosphere and liquid water on the planet's surface (Wilde *et al.*, 2001). The appropriate climatic conditions had to be reached quickly. Therefore, the capture of all the CO_2 must have been equally fast. Charnay *et al.* (2020) mention that the explanation may be related to the effect of the atmosphere moderately rich in CO_2, potentially helped by additional warming processes. Although he does not express

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it directly, in the presence of higher heat flow to the surface a smaller amount of greenhouse gases will be necessary for the formation of liquid water on the surface. This scenario would fall on the cases shown in Figures 3c and 4c.

Also, there are several reasons to consider scenarios of an evolving Earth without a post-impact atmosphere (*e.g.*, Wang *et al.*, 2022 and references therein). The depletion of rare gases in Earth's atmosphere compared to cosmic abundances indicates that any primary atmosphere must have been lost (Kasting, 1993). The Earth was able to obtain a mass capable of retaining an atmosphere perhaps until after the dissipation of the solar cloud, suggesting that accretion probably occurred mostly in a gas-free environment (Canuto *et al.*, 1983). On the other hand, the "boil-off" theory has been proposed in which the sudden disappearance of the solar nebula could expand the primordial atmosphere and then undergo an efficient escape (Owen and Wu, 2016; Wang *et al.*, 2022). Although part of the primary atmosphere (nebular atmosphere) had survived, it must have been subjected to a lot of accretional impacts that could have eroded it (Chen and Ahrens, 1997). For some simulations the estimated atmosphere loss fraction for the giant impact is only 20% (Genda and Abe, 2003). In the canonical Moon-forming impact, only around 10% of the atmosphere would have been lost from the immediate effects of the collision (Kegerreis *et al.*, 2020). Nevertheless, there are certain conditions that can increase the loss of the atmosphere. For example, the existence of an ocean of water before the impact could amplify the loss of atmosphere and volatiles during the collision (Genda and Abe, 2003). In addition, the thermal consequences of the impact can reduce the atmospheric mass between 50 and 100% for impactors with 10% of the mass of the planet (Biersteker and Schlichting, 2019). Although this is applicable for H/He dominated atmospheres that expand thermally on impact, this type of atmosphere is what one might expect to exist before the end of Earth's accretion. Something that must be considered is that an impact capable of causing the loss of the atmosphere would inevitably also cause the loss of crust or part of the mantle (Denman *et al.*, 2020). In any case, there is enough evidence justifying the assessment of an evolutionary trend that does not include a highly insulating post-impact atmosphere. As shown by panels b and d of Figures 2 and 3, this scenario inevitably triggers the formation of a lithosphere before the complete solidification of the mantle.

It should be noted that the solidification of the lithosphere before the complete solidification of the mantle does not prevent the formation of the atmosphere. Volcanic degassing (Tajika, 1998; Gaillard *et al.*, 2021), meteoric impact and the process of intrusive magmatism has a strong influence on the total degassing into the atmosphere of the early Earth (Vulpius and Noack, 2022). A more gradual development of the atmosphere would also gradually change the rate at which heat is transferred on this layer. Although this possibility should be kept in mind, given the wide range of conditions upon which the top of the magma ocean cooled before its interior parts, such gradual evolution of the atmosphere would not modify the main trends reported in Figures 2 and 3.

If the atmosphere does not form immediately after the impact, or if it is not as insulating, it would favor the formation of the lithosphere. If there is an insulating atmosphere, but not enough to transport heat more slowly than the magma ocean does, the conditions to form the lithosphere would also have been reached. If turbulent convection of magma ocean favored lateral movements instead of vertical ones as shown by some experimental evidence (Davaille and Limare, 2007), the lithosphere would also form before the total solidification of the planet. Water vapor degassing may not be as abundant as previously speculated (Miyazaki and Korenaga, 2022a). This is an important greenhouse gas, so the atmosphere was possibly less insulating. Even if we consider the large-scale circulation of the mantle, total degassing could have taken longer times than those considered by some previous models (Salvador and Samuel, 2023).

4.2. MAGMA OCEAN CONVECTION

The magma ocean just after the impact is expected to have had a very turbulent, poorly organized, and not large-scale convection. Therefore, the heat transfer could be carried out in a chaotic manner and not as organized as has been proposed until now (Elkins-Tanton, 2012; Solomatov, 2015; Lebrun *et al.*, 2013). This implies that heat transfer can be very different even for nearby points in the planet's interior during this chaotic period. The flow of heat that would come out of the planet's surface has been speculated to reach extremely high amounts (Elkins-Tanton, 2012; Solomatov, 2015). This assumes vigorous convection capable of creating a large-scale circulation that could move large amounts of material from the planet's interior very quickly. But a turbulent convection could decrease the total heat flux reaching the surface of the magma ocean (Davaille and Limare, 2007). Within that context, it is important to mention that our results show a clear relationship between the rate of mantle heat transfer and surface solidification.

The main argument for assuming a magma ocean with a large-scale circulation is based on the Rayleigh number (Solomatov, 2015). Although this dimensionless number can be useful for different modeling, it is important to note that it is not directly applicable to systems with co-existing solids and fluids (Hofmeister, 2020). If we consider this, there is a possibility that large-scale circulation was not as rapid or as vigorous as has been thought. As shown by our results, the lithosphere could form even if large-scale convection in the mantle exists.

4.3. COMPARISON WITH PREVIOUS MODELS

A methodology recurrently used to model the solidification of the magma ocean is described by Solomatov (2015). In his work he considered that the magma ocean behaves as an adiabatic system from the beginning. This is equivalent to a magma ocean that presents a large-scale circulation throughout the entire portion of molten material. Almost simultaneously, due to the imposed convection, the mantle degasses until it forms an atmosphere made up mainly of CO_2 and H_2O. This atmosphere must have had an incredibly strong greenhouse effect to keep the planet's surface completely liquid. Such initial conditions are certainly contrasting with the results of SPH simulations for large impacts (Canup, 2008; Carter *et al.*, 2020). One of the main differences between the behavior of the adiabatic curves and the results of the SPH simulations is found in the surface of the planet, where the temperature can be higher due to secondary impacts (*e.g.*, Carter *et al.*, 2020). This makes it difficult to consider an adiabatic curve as an initial condition without explaining how it arrived at.

In addition to the assumed shape of adiabatic curves, there are other implications of the assumed vigorous convective mantle. There are numerous simulations where the planet solidifies in the same way regardless of whether an atmosphere exists or not (*e.g.*, Lebrun *et al.*, 2013). This implies that the model is being forced to have a vigorous convention even without having a greenhouse atmosphere that causes thermal insulation. Adiabatic behavior depends on the insulating boundaries that allow vigorous convection (Incropera *et al.*, 2007; Lienhard and Lienhard, 2008; Çengel, 2006). In contrast, in the different simulations shown above the atmosphere is not insulating enough to prevent the formation of the lithosphere. However, the lithosphere can cause the insulating effect necessary to modify the solidification of the interior. Such dependence on the conditions of the atmosphere seems to be more realistic, as no specific conditions are imposed on the behavior of the mantle in our model.

Furthermore, the physical effect of having a contrast between a gaseous material that distributes and displaces heat better than liquid magma have not been taken into account in previous models. This contrast proves to be extremely important in deciding the rate of cooling of the surface of the magma ocean. Even if we consider that the atmosphere could be preserved as an insulating

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system, heat transfer towards outer space would have continued to take place through a much more conductive layer than those representing the mantle-lithosphere. Consequently, no matter how low the viscosity could have been in the post-impact liquid mantle, liquid convection is likely to distribute heat more slowly than the gaseous materials of the atmosphere. Consequently, the conditions for solidification from the top would have been met more easily than those favoring cooling only from the bottom. Therefore, the scenario with two solidification fronts is a very possible scenario for the Earth (Intermediate greenhouse effect).

Finally, it is noted that the lithosphere could have served as an insulating layer, favoring liquid convection in the interior, and storing a greater amount of primordial heat. If the primordial heat did not leave the planet as quickly as thought, this directly impacts geochemical observations (Boyet and Carlson, 2006; Boehnke *et al.*, 2015; Holland *et al.*, 2009), calculations of radiogenic elements in the mantle (KamLAND Collaboration 2011), and our understanding of mantle evolution (Davies, 2007). It could also shed new light on how the Earth's electromagnetic field could have been kept functioning for at least 3.5 Gyrs (Landeau *et al.*, 2022).

5. Conclusions

The most important points that we can highlight from this work are:

- The formation of the lithosphere may be possible before the complete solidification of the magma ocean. Hence, from a purely thermal point of view Earth could have had a magma ocean beneath a solid lithosphere early in its history.
- The formation of a lithosphere could delay the complete cooling of the planet.
- The best conditions for the formation of a lithosphere are reached when the atmosphere transports heat faster than the magma ocean, conditions that are easily met.

Although our model includes many simplifications, it serves as a first approach to the problem posed by the cooling of the planet based on fewer assumptions than many current models. Consequently, our results open the door to possibilities that have not been considered before, but that nonetheless are physically possible and therefore should be explored with attention.

In particular, our approximation shows how the great contrast that exists between the physical properties of the mantle and the atmosphere may have had a great influence on the formation of a lithosphere on the surface of the planet.

Supplementary data

Carapia, M., 2023, Atmospheric_influence_on_lithosphere_formation (Version 1). (Software), figshare. https://doi.org/10.6084/m9.figshare.24558262

Contributions of authors

Both authors contributed to the conceptualization, methodology, research, writing of the manuscript, preparation of figures, revision of the manuscript and interpretation of results.

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Conflicts of interest

The authors declare that there are no competing interests, financial or otherwise, that could have influenced the reported research.

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Appendix A

A1.1 RADIOGENIC ELEMENTS DURING LITHOSPHERE SOLIDIFICATION

This section shows the impact of the presence of radiogenic elements inside the magma ocean during lithosphere formation. The results of our simulations suggest that lithosphere solidification occurred in less than a million years. Therefore, the temperature difference for some simulations during this period with and without radioactive heat generation has been compared. This has been done by subtracting the temperature of simulations with heat generation by radiogenic elements from the results of the same simulation without them. An average of this difference has been obtained for the entire mantle, since it is here where radiogenic elements are homogeneously distributed.

Figure A1 shows a series of curves representing temperature differences averaged throughout the mantle for simulations of maximum (*solid*) and minimum (*dotted*) temperatures. For simplicity, only some of the simulations performed in this work have been exemplified. The red curves are those with the weakest heat transfer (k_{conv}=10^2), the magenta curves have the fastest heat transfer (k_{conv}=10^5) and finally, the green (k_{conv}=10^3) and blue (k_{conv}=10^4) curves are intermediate cases. All curves show a clear upward trend, indicating that the impact of radiogenic elements increases with time. However, the difference between all these simulations is less than 1.5°K for the first million years. In other words, the temperature difference between considering the influence of radiogenic heat production and neglecting it, is less than 0.1% of the solidification temperature considered in this work (1400 °K).

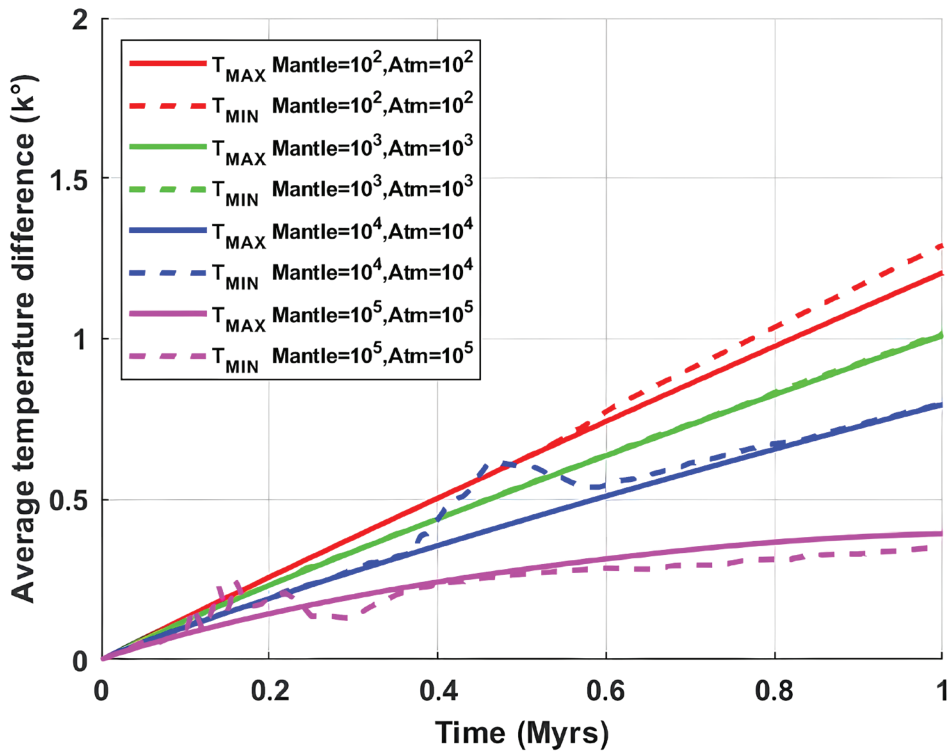

Figure A1 A series of curves representing the average temperature difference for the mantle. The temperature difference was obtained between simulations with identical initial parameters, only distinguished by the presence of radiogenic elements. These differences have been averaged for the entire mantle over time. The solid curves represent the maximum initial temperatures and the dashed curves the minimum initial temperatures.