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# Thermal evolution of the lunar magma ocean

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## Abstract

The energy of the giant impact was large enough to generate an initially fully molten Moon. During the solidification of this lunar magma ocean (LMO), an anorthosite crust formed by flotation of light anorthite crystals. Lunar anorthosites show crystallization ages as young as 4.360 Gyr, suggesting a long-lived LMO or a rather young Moon. Existing models for LMO solidification are for a specific phase diagram based on one compositional model. However, the LMO solidification timescale depends on the lunar bulk composition and on the appearance of anorthite in the crystallization sequence.

Here, we propose a physically robust 1D model for LMO evolution based on a simple anorthite-olivine eutectic phase diagram. Cumulates first settle at the ocean base for about a thousand years. This first stage results in an unstable thermal profile for the cumulates that can lead to their overturn. In the second stage, simultaneous crystallization of anorthite and cumulates leads to the formation of a buoyant lid that considerably slows down LMO cooling.

We explore the impact of an initially hydrated composition, which reduces the stability of plagioclase, of the eutectic position and the crust thermal conductivity. We show that cumulates overturn may reduce or extend the LMO solidification time depending on its duration. The total LMO solidification timescale ranges between 45 and 250 Myr. Given the most reliable age of 4.360 Gyr for FAN sample 60025, which derives from more than 99% of crystallization, we estimate an age of 4400 to 4560 Myr for the Moon.

*Keywords:* Moon, magma ocean, crystallization time, anorthosite crust

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## 1. Introduction

The energy from the giant impact between the proto-Earth and a Mars-sized body was sufficient to give rise to an initially hot Moon with a global magma ocean (Pritchard and Stevenson, 2000; Canup, 2012; Āuk and Stewart, 2012; Nakajima and Stevenson, 2014). A global magma ocean appears particularly inevitable if the formation of the Moon occurred in a Synestia, a planetary structure resulting from a high-energy, high-angular momentum giant impact, as recently suggested (Lock and Stewart, 2017; Lock et al., 2018). The lunar Highlands, forming the pale and bright regions on the Moon’s surface, are remnants of the initial crust formed by the cooling and crystallization of this lunar magma ocean (LMO). Samples collected by astronauts during the Apollo missions, meteorite samples, and remote-sensing studies indicate that the lunar Highlands are predominantly made of anorthosite, containing more than 90% of Ca-rich anorthite (Warren and Wasson, 1979; Ohtake et al., 2009). In the LMO solidification scenario, this primary crust is formed by aggregation of buoyant anorthite minerals floating on the surface and insulating the magma ocean beneath while, cumulates form by sedimentation of dense olivine and pyroxene crystals (Wood et al., 1970) (fig. 1c). Fractional crystallization of the LMO would have resulted in the formation of a residual liquid, highly enriched in incompatible elements (K: Potassium, REE: Rare Earth Elements and P: Phosphorus), which could explain the singular composition of the Oceanus Procellarum KREEP terrane. LMO fractional crystallization would also lead to an unstable density profile in the cumulates, resulting both from the decrease in the LMO liquidus temperature and its progressive enrichment in iron as crystallization proceeds (Hess and Parmentier, 1995; Parmentier et al., 2002; Boukaré et al., 2018; Morison et al., 2019).

Dating of ferroan anorthosite, believed to be the oldest crustal rock type on the Moon, yield ages ranging from 4.29 to 4.55 Gyr (Alibert et al., 1994; Borg et al., 2020; Marks et al., 2019; Nyquist et al., 2006), hence spanning over  $\sim 250$  Myr. It has been argued that these ages may not reflect the true crystallization ages of anorthosites as different chronometers do not always give overlapping results (Borg et al., 2015). The intense early bombardment of the lunar crust could have perturbed the crystallization and metamorphic history of these rocks (Carlson, 2019). Nonetheless, the sample FAN 60025 shows one of the youngest ages of 4.360 Gyr, consistent for several radiometric systems (Borg et al., 2011). Borg et al. (2011) proposed that this age could reflect a relatively late formation of the Moon, more than 100 Myr after Solar System formation. Numerous studies summarized in Borg and Carlson (2023) have also shown that many lunar rocks show ages that cluster around

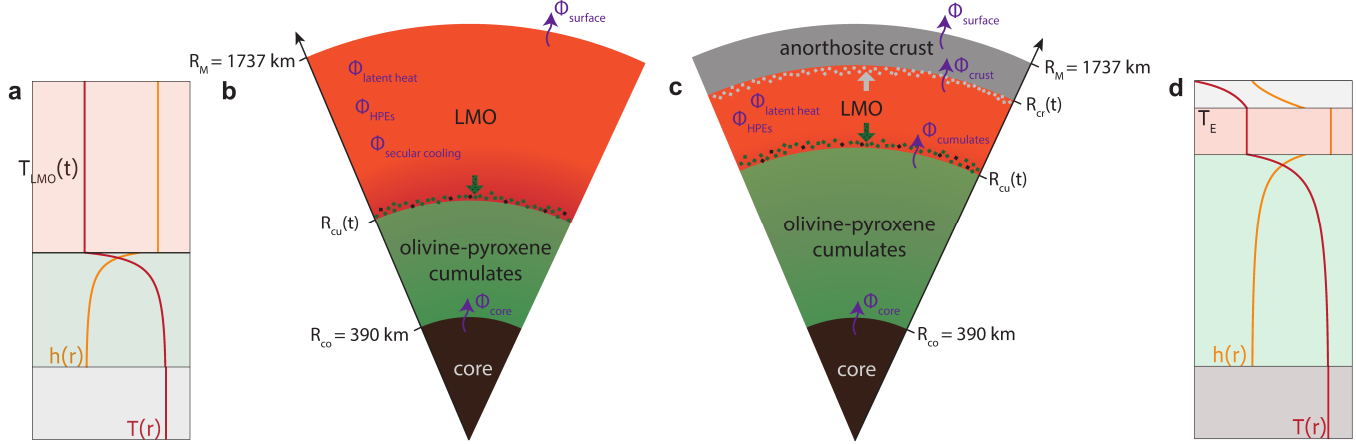
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7 4.33-4.360 Gyr suggesting rapid cooling of a magma ocean. These observations were  
8 used to argue that the Moon formation took place shortly before 4.360 Gyr. The  
9 debate about the age of the Moon has been ongoing since then and is closely tied to  
10 the timescale of solidification of the LMO, approximately 200 Myr after the formation  
11 of the Solar System. Indeed, this young age for a ferroan anorthosite can also imply  
12 that the LMO took longer to solidify than previously thought (Maurice et al., 2020)  
13 or that the lunar crust did not form by flotation but by a process closer to serial  
14 magmatism (Longhi, 2003; Borg et al., 2011; Gross et al., 2014), for instance by  
15 extraction of melts from a slushy magma ocean which cools down over a longer  
16 time-scale than in the classical LMO solidification scenario (Michaut and Neufeld,  
17 2022).

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19  
20 Existing models provide variable estimates for the solidification time in its clas-  
21 sical scenario, from 10 Myr (Elkins-Tanton et al., 2011) to 100-200 Myr (Maurice  
22 et al., 2020; Zhang et al., 2021b). These models are based on only one specific phase  
23 diagram associated to the compositional model of Elkins-Tanton et al. (2011). How-  
24 ever, the shortest estimate does not account for the production of radioactive heat.  
25 In this model, anorthite appears rather late in the crystallization sequence, when  
26 80% of the lunar magma ocean has solidified. Other fractional crystallization exper-  
27 iments have been carried out, based on slightly different bulk compositions for the  
28 Moon, or different LMO depths (considering a whole magma ocean rather than a  
29 shallower one as in Elkins-Tanton et al., 2011).

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32 They notably differ in the overall final proportion of anorthite and the specific  
33 point in the crystallization sequence when anorthite emerges (Snyder et al., 1992;  
34 Elkins-Tanton et al., 2011; Charlier et al., 2018). Adding water to the bulk composi-  
35 tion of the Moon reduces the stability of plagioclase and delays its appearance in the  
36 crystallization sequence (Lin et al., 2017). Such differences in composition not only  
37 result in different final possible thicknesses for the anorthosite crust, they also affect  
38 the time at which the flotation crust starts to insulate the magma ocean beneath,  
39 and hence the total amount of latent and radioactive heat that must be conducted  
40 away through this lid. In some models, anorthite starts to form when only  $\sim 66-70\%$   
41 of the LMO has crystallized (Lin et al., 2017; Rapp and Draper, 2018; Johnson et al.,  
42 2021; Schmidt and Kraettli, 2022), i.e. notably earlier in the crystallization sequence  
43 than in the model of Elkins-Tanton et al. (2011), which could prolong the lifetime of  
44 the LMO.

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47 To evaluate the consequences of different compositional models as well as the con-  
48 sequences of an overturn of the cumulate layer on the LMO solidification timescale,  
49 we propose a physically robust 1D model for the evolution of the Moon in its magma  
50 ocean stage. Our model is based on a simple anorthite/olivine-pyroxene eutectic  
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75 phase diagram, which naturally results in a two-stage cooling process for the LMO.  
76 Before anorthite becomes a stable phase, the LMO cools down rapidly by radiative  
77 loss of heat at its surface (Solomatov, 1999). In the second stage, the simultaneous  
78 crystallization of anorthite and olivine-pyroxene cumulates leads to the formation of  
79 a buoyant anorthositic lid that considerably slows down the cooling of the magma  
80 ocean (fig. 1c). We solve for heat conservation in the different layers of the lunar  
81 body to follow the evolution of the crust, LMO and cumulates thicknesses and tem-  
82 peratures and evaluate the LMO final solidification time.



34 Figure 1: Schematic representation of the two stages of crystallization of the lunar magma ocean.  
35 Left two panels (a, b): first, “radiative”, stage. Right two panels (c, d): second, “conductive”, stage.  
36 a) and d): Radial profiles of the temperature and heat-producing element concentration during the first and second stage respectively. b) and c): Schematic of the structure of the Moon, the heat  
37 sources and fluxes in the first and second stage, respectively. The core is shown in dark gray, the  
38 olivine-pyroxene cumulates in green, the LMO in red, and the conductive crust in light gray.

## 83 2. Compositional Model

### 84 2.1. Phase diagram and mass conservation

85 We consider a simple binary eutectic phase diagram with two components: anor-  
86 thite on the one hand and olivine-pyroxene on the other. The liquidus temperature  
87  $T_{\text{Liq}}(t)$  is assumed to be linearly related to the mass fraction in anorthite component  
88  $C(t)$  of the liquid (fig. 2a):

$$T_{\text{Liq}}(t) = T_{\text{OL}} - mC(t), \quad (1)$$

89 where  $T_{OL}$  is the liquidus temperature for the olivine-pyroxene component and  $m =$   
 90  $\frac{T_{OL}-T_E}{C_E}$  characterises the liquidus slope.  $T_E$  and  $C_E$  correspond respectively to the  
 91 eutectic temperature and composition. As the LMO convects vigorously, it is well-  
 92 mixed and we assume that its temperature is uniform and equal to the liquidus  
 93 temperature  $T_{LMO}(t) = T_{Liq}(t)$ . Considering the onset of crystallization as the initial  
 94 time for our calculation, we note  $C_0$  the initial LMO content in anorthite component  
 95 and  $T_{LMO}^0$  the corresponding liquidus temperature. The pressure range inside the  
 96 LMO is small because of the low lunar gravity, in particular during the second stage  
 97 where the thickness of the LMO at the beginning of the second stage varies between  
 98 250 and 50 km (see section 4.2); we thus neglect the effect of pressure on the phase  
 99 diagram, which would have a second order influence on our results.

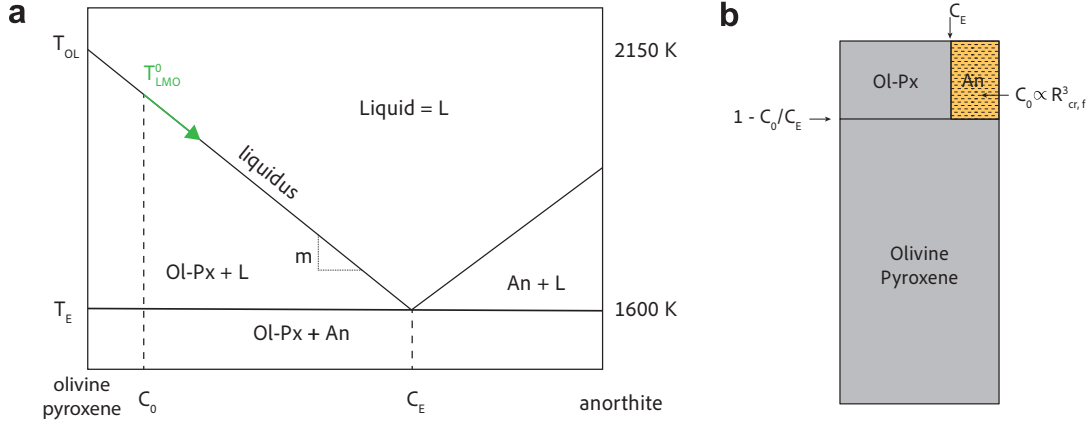


Figure 2: a) Simplified olivine-pyroxene/anorthite phase diagram with liquidus  $T_{liq}(C) = T_0 -$   
 $mC(t)$  as a function of anorthite content of the liquid  $C(t)$ , where  $T_{OL} = 2150$  K is the liquidus  
 temperature of the olivine-pyroxene component,  $T_E = 1600$  K is the eutectic temperature,  $C_E \in$   
 $[0.2, 0.7]$  the eutectic composition and  $m$  represents the slope of liquidus. The initial conditions  
 are  $C_0 \in [0.05, 0.1]$  and the corresponding initial temperature of the LMO  $T_{LMO}^0$ , adapted from  
 O'Driscoll et al. (2010). b) Simplified solidifying mantle mineral assemblage adapted from Elkins-  
 Tanton et al. (2011).

100 Two stages of crystallization naturally appear from this simplified phase diagram  
 101 (fig. 2a). In the first stage, as the magma ocean cools down along the liquidus, dense  
 102 crystals of olivine and pyroxene settle at the bottom leading to a crystallization of  
 103 the LMO from the bottom-up. The LMO becomes richer in anorthite component as  
 104 it crystallizes until it reaches the eutectic composition  $C_E$  and temperature  $T_E$ . Con-  
 105 servation of the anorthite component provides a relation between the upper radius

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106 of the cumulate layer  $R_{\text{cu}}(t)$  and the LMO content in anorthite component  $C(t)$ :

$$\left(R_{\text{M}}^3 - R_{\text{co}}^3\right)C_0 = \left(R_{\text{M}}^3 - R_{\text{cu}}^3(t)\right)C(t), \quad (2)$$

107 where  $R_{\text{M}}$  is the radius of the Moon and  $R_{\text{co}}$  that of the core, and where, for simplicity,  
108 we consider a uniform average density for the different layers of the Moon. At the  
109 end of the first stage  $C(t) = C_{\text{E}}$  and the ratio  $(C_{\text{E}} - C_0)/C_{\text{E}}$  thus corresponds to the  
110 total percentage of LMO crystallized at that time. Equations (1) and (2) can be used  
111 to obtain an analytical expression for the temperature variation in the cumulates as  
112 a function of the radial coordinate, assuming that diffusion is negligible because this  
113 first stage is short. During stage 1, the temperature profile in the cumulates is thus  
114 considered frozen and depends on the radius as the temperature at the cumulates-  
115 LMO boundary evolves along the liquidus (fig. 2):

$$T_{\text{cu}}(r) = T_{\text{OL}} - mC_0 \frac{R_{\text{M}}^3 - R_{\text{co}}^3}{R_{\text{M}}^3 - r^3} \quad \text{for} \quad R_{\text{co}} \leq r \leq R_{\text{cu}}. \quad (3)$$

116 As the LMO crystallizes, the cumulate temperature profile becomes super-isentropic  
117 and thus prone to instability and overturn.

118 When the eutectic composition is reached in the LMO, anorthite starts to crys-  
119 tallize and olivine-pyroxene crystals continue to form in eutectic proportions. This  
120 constitutes the second stage of the thermal evolution. Light anorthite crystals then  
121 form a floating crust, while the LMO remains at the eutectic temperature  $T_{\text{E}}$ . Con-  
122 servation of the anorthite component gives the final radius of the crust  $R_{\text{cr,f}}$ :

$$\left(R_{\text{M}}^3 - R_{\text{cr,f}}^3\right) = \left(R_{\text{M}}^3 - R_{\text{co}}^3\right)C_0, \quad (4)$$

123 where we assume that all available anorthite floats to form a pure anorthositic crust.  
124 This buoyant and conductive lid then insulates the LMO, which significantly slows  
125 down its rate of crystallization.

## 126 *2.2. Parameters of the phase diagram*

127 The parameter  $C_0$  represents the available amount of anorthite. As we assume  
128 all available anorthite reaches the crust,  $C_0$  controls the final thickness of the crust.  
129 This parameter depends on the exact composition of the bulk silicate Moon, its  
130 water content in particular, as water tends to delay and diminish the appearance  
131 of anorthite. To estimate a range for  $C_0$ , we refer to the various phase diagrams  
132 published in the literature (Snyder et al., 1992; Elkins-Tanton et al., 2011; Lin et al.,  
133 2017; Charlier et al., 2018; Rapp and Draper, 2018; Johnson et al., 2021; Schmidt

and Kraettli, 2022); these studies suggest the overall range  $C_0 \in [0.05, 0.1]$ . The larger  $C_0$ , the thicker the anorthositic crust, the slower the heat conduction through the lid and the longer it takes to solidify the LMO. Assuming a global magma ocean, a water concentration of 1600 ppm in the LMO, corresponding to a value of  $C_0 = 0.075$ , may be necessary to explain the reduced crustal thickness of the Moon, as constrained from gravity and topography data (Wieczorek et al., 2013; Lin et al., 2017). This appears significantly larger than available estimates based on the water content measured directly in lunar rocks (Saal et al., 2008; Hauri et al., 2011; Hui et al., 2013). Alternatively, a fraction of anorthite crystals may have not reached the lunar crust and remained trapped in the cumulates ; the parameter  $C_0$  then represents the effective concentration of anorthite forming the lunar primary crust. The second key parameter of the phase diagram is  $C_E$  which represents the percentage of anorthite component crystallizing during the second stage. As  $C_E$  decreases, more olivine and pyroxene crystallize during the second stage and more latent heat must be evacuated through the floating crust. While  $C_E$  varies from  $\sim 0.3$  to  $0.5$  among the different proposed compositional models (table 1), we explore the wider range  $[0.2 - 0.7]$  based on the phase diagram of O’Driscoll et al. (2010).

study	$C_0$	$C_E$	% crystallization	initial LMO depth [km]	crustal thickness [km]
Snyder et al. (1992)	0.088	0.4	78	400	28
Elkins-Tanton et al. (2011)	0.089	0.43	80	1000	49
Lin et al. (2017) (no water)	0.1	0.31	68	700	47
Lin et al. (2017) ( $\sim 3000$ ppm)	0.06	0.2	78	700	27
Charlier et al. (2018)	0.1	0.5	78	600	43
Rapp and Draper (2018)	0.1	0.37	74	1347	60
Johnson et al. (2021)	0.084-0.92	0.27-0.36	70-75	1347	50-55
Schmidt and Kraettli (2022)	0.14	0.3-0.5	72	600-1150	60-82

Table 1: Values of  $C_0$ ,  $C_E$ , percentage of LMO crystallized at the end of the first stage, initial LMO depth and final crustal thickness for different studies.

### 2.3. Heat Producing Elements (HPEs)

The decay of radioactive Heat Producing Elements (HPEs) contributes to a non-negligible source of heat during the magma ocean evolution and must be considered. Uranium ( $^{235}\text{U}$  and  $^{238}\text{U}$ ), Thorium ( $^{232}\text{Th}$ ), and Potassium ( $^{40}\text{K}$ ) are the main HPEs in the LMO. These highly incompatible elements tend to remain in the liquid phase as their solid/liquid partition coefficients are low (Sun et al., 2017). For simplicity, we consider all of these elements as a single one with a given radiogenic decay constant  $\lambda$  and distribution coefficient  $D$  between the liquid phase  $l$  and the crystallized layer  $s$ :

$$D = \frac{[HPE]_s}{[HPE]_l}. \quad (5)$$



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160 HPEs are assumed to be homogeneously distributed in the LMO, but not in the  
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161 solid layers, where their distribution varies radially and is calculated based on eq. (5)  
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162 and on the rate of crystallization (see section 3.2). We track the radial distribution  
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163 of the heat production from HPEs  $h_i(t)$  during magma ocean crystallization, in the  
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164 two solid layers  $i$ , where index  $i$  stands for crust (cr) or solid cumulates (cu).

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165 During the first stage, as the cumulates grow with time, the total content of the  
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166 LMO in HPEs decreases. Morison et al. (2019) and Boukaré et al. (2018) give an  
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167 analytical solution for the radial distribution of the heat production from HPEs as a  
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168 function of the radius of the cumulates, which can be converted in a function of the  
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169 radial position in the cumulate  $r$ :

$$h_{\text{cu}}(r) = h_0 \left( \frac{R_{\text{M}}^3 - R_{\text{co}}^3}{r^3 - R_{\text{co}}^3} \right)^{1-D} \quad \text{for } R_{\text{co}} \leq r \leq R_{\text{cu}}, \quad (6)$$

170 with  $h_0$  the heat production rate per unit volume of the initial LMO. This expression  
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172 neglects radioactive decay during the first, very rapid, stage. The actual heat pro-  
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174 duction as a function of time in the second stage is obtained by taking into account  
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176 radioactive decay as  $he^{-\lambda t}$ , with  $\lambda$  an effective mean radioactive decay constant. We  
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178 call  $h_0$  and  $h_{\text{cu}}$  heat production coefficients for the initial magma ocean and the cumu-  
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180 lates, which are the products of the concentration in HPEs by their heat production  
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182 rates.

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184 During the second stage, the concentration in HPEs increases in the LMO as  
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186 the two solid layers grow. Conservation of HPEs gives the variation of the heat  
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188 production coefficient in the LMO induced by LMO solidification:

$$\frac{dh_{\text{LMO}}}{h_{\text{LMO}}} = (1 - D) \frac{dV_{\text{cryst}}}{V_{\text{LMO}}}, \quad (7)$$

189 where  $V_{\text{LMO}}$  is the LMO volume and  $-dV_{\text{cryst}}$  its volume variation. We follow the  
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191 evolution in radiogenic heat production in the LMO from eq. (7), and compute that  
192  
193 in the crust and cumulates from the partition coefficient (see Appendix A).

1	Symbol	Description	Value	References
2	<b>Parameters</b>			
3				
4	$R_M$	Radius of the Moon	1737 km	
5	$R_{co}$	Radius of the core	390 km	
6	$T_{OL}$	Liquidus temperature for the ol-px component	2150 K	O'Driscoll et al. (2010)
7	$T_E$	Eutectic temperature	1600 K	O'Driscoll et al. (2010)
8			$(T_{OL} - T_E)$	
9	$m$	Liquidus slope	$m = \frac{C_E}{(T_{OL} - T_E)}$	
10	$h_0$	Initial heat production	25 pW kg <sup>-1</sup>	Taylor (1982)
11		Initial uranium abundance	33 ppb	Taylor (1982)
12		Initial thorium abundance	125 ppb	Taylor (1982)
13		Initial potassium abundance	400 ppb	Taylor (1982)
14	$\lambda$	Effective radiogenic decay constant	$5.8 \times 10^{-10} \text{ Myr}^{-1}$	
15	$D$	Solid/liquid partition coefficient of HPEs	10 <sup>-3</sup>	Sun et al. (2017)
16	$g$	Gravity acceleration	1.62 m s <sup>-2</sup>	
17	$\rho$	Density	$3.3 \times 10^3 \text{ kg m}^{-3}$	
18	$c_p$	Heat capacity	$1 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$	
19	$L$	Latent heat	$5 \times 10^5 \text{ J kg}^{-1}$	Weill et al. (1980)
20	$\sigma$	Stefan-Boltzman constant	$5.67 \times 10^{-8} \text{ J K}^{-1}$	
21	$\varepsilon$	Emissivity	1	
22	$A$	Albedo	0.12	
23	$T_{\odot}$	Sun surface temperature	5780 K	
24	$R_{\odot}$	Sun radius	$700 \times 10^3 \text{ km}$	
25	$D_{\odot}$	Moon-Sun distance	$1495 \times 10^5 \text{ km}$	
26	$T_{\infty}$	Black body equilibrium temperature (eq. (11))	270 K	
27	$\alpha$	Thermal expansion coefficient	10 <sup>-5</sup> K <sup>-1</sup>	Maurice et al. (2020)
28	$\mu_{LMO}$	Viscosity	1 kg m <sup>-1</sup> s <sup>-1</sup>	
29	$k$	Thermal conductivity of the LMO	4 W m <sup>-1</sup> K <sup>-1</sup>	
30	$k_{cu}$	Thermal conductivity of cumulates	4 W m <sup>-1</sup> K <sup>-1</sup>	Maurice et al. (2020)
31	$\beta$	Thermal boundary layer coefficient	1/3	Gastine et al. (2016)
32	$\gamma$	Scaling prefactor	0.07	Gastine et al. (2016)
33	$c_{p, co}$	Heat capacity of the core	840 J kg <sup>-1</sup> K <sup>-1</sup>	Maurice et al. (2020)
34	$\rho_{co}$	Density of the core	$7.2 \times 10^3 \text{ kg m}^{-3}$	
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36				
37	<b>Explored parameters</b>			
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39	$C_0$	Initial content in anorthite component	$\in [0.05, 0.1]$	
40	$C_E$	Eutectic composition	$\in [0.2, 0.7]$	
41	$\kappa_{cr}$	Thermal diffusivity of the crust	$\in [5.5, 8.2] 10^{-7} \text{ m}^2 \text{ s}^{-1}$	Roy et al. (2021)
42	$k_{cr}$	Thermal conductivity of the crust	$\in [1.8, 2.7] \text{ W m}^{-1} \text{ K}^{-1}$	Branlund and Hofmeister (2012)
43				Roy et al. (2021)
44	$Q$	Energy released at the beginning of the overturn	$\in [3 \times 10^{14}, 2 \times 10^{12}] \text{ W}$	
45	$\tau_{ov}$	Overturn time	$\in [1.8, 180] \text{ Myr}$	
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47	<b>Variables stage 1</b>			
48	$T_{LMO}(t)$	LMO temperature	eq. (1)	
49	$R_{cu}(t)$	Radius at the top of the cumulates	from eq. (13)	
50	$T_s(t)$	Surface temperature	eq. (10)	
51	$h_{cu}(t)$	HPEs concentration in the cumulates	eq. (15)	
52				
53	<b>Variables stage 2</b>			
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55	$R_{cu}(t)$	Radius at the top of the cumulates	eq. (24)	
56	$R_{cr}(t)$	Radius at the base of the crust	eq. (23)	
57	$T_{cr}(t, r)$	Temperature in the crust	eq. (15)	
58	$T_{cu}(t, r)$	Temperature in the cumulates	eq. (15)	
59	$h_{cr}(r)$	HPEs production in the crust	eq. (7) and Appendix A	
60	$h_{cu}(r)$	HPEs production in the cumulates	eq. (7) and Appendix A	
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Table 2: List of the parameters and variables of the model.

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### 183 3. Thermal model

#### 184 3.1. Stage 1

##### 185 3.1.1. Surface temperature

186 The LMO forms a spherical shell of outer radius  $R_M$  and inner radius  $R_{cu}(t)$ ,  
187 which increases with time as the LMO cools down and solidifies. During the first  
188 stage, the heat flux lost by radiation at the surface is balanced by the flux supplied  
189 by convection in the LMO. We parameterize the convective heat flux following the  
190 boundary layer theory which assumes that convection near one boundary is indepen-  
191 dent of processes near the other boundary and which provides a simple relationship  
192 between the Nusselt number  $Nu$  and the Rayleigh number  $Ra$ :

$$Nu = \gamma Ra^\beta, \quad (8)$$

193 where  $Nu$  is the ratio between the convective heat flux and the characteristic conduc-  
194 tive flux in the LMO,  $\gamma$  the scaling prefactor and,  $\beta = 1/3$  is the exponent given by  
195 the thermal boundary layer theory at the limit of weakly rotating convection (Gas-  
196 tine et al., 2016). The Rayleigh number  $Ra$  characterises the strength of convection  
197 and is defined as:

$$Ra = \frac{\alpha \rho g (T_{LMO} - T_s) d^3}{\kappa \mu} \quad (9)$$

198 where  $\alpha$  is the thermal expansion coefficient,  $\rho$  the LMO density,  $T_s$  the surface  
199 temperature,  $d = R_M - R_{cu}$  the LMO thickness,  $\kappa$  the thermal diffusivity, and  $\mu$   
200 the LMO viscosity,  $\alpha, \rho, \kappa, \mu$  being assumed constant. As  $T_{LMO}$  and  $d$  decrease with  
201 time, the Rayleigh number is time-dependent. From eqs. (8) and (9), the surface  
202 equilibrium between the radiative and convective heat fluxes is expressed as:

$$\epsilon \sigma (T_s^4 - T_\infty^4) = k \frac{T_{LMO} - T_s}{d} \gamma Ra^\beta, \quad (10)$$

203 where  $\epsilon$  is the emissivity,  $\sigma$  the Stephan-Boltzman constant,  $k$  the thermal conduc-  
204 tivity.  $T_\infty$  is the equilibrium temperature, derived from the equality between the  
205 solar flux absorbed by the Moon and the radiative heat flux it emits:

$$T_\infty = (1 - A)^{1/4} T_{\odot} \sqrt{\frac{R_{\odot}}{2D_{\odot}}}, \quad (11)$$

206 where  $T_{\odot}$  is the surface temperature of the Sun,  $R_{\odot}$  the radius of the Sun,  $D_{\odot}$  the  
207 Moon-Sun distance, and  $A$  the surface albedo of the Moon.

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208 *3.1.2. Heat budget in the magma ocean*

209 During the first stage, only olivine and pyroxene cumulates crystallize (see sec-  
210 tion 2). We neglect pressure effects and assume a homogeneous temperature for the  
211 LMO. As the first stage is very rapid ( $\sim 300$  yr, which is verified a posteriori), we  
212 neglect heat conduction in the cumulates.

213 In the LMO, the heat lost by radiation into space at the surface is balanced by the  
214 sum of the latent heat released by crystallization of olivine-pyroxene cumulates, the  
215 heat produced by radiogenic decay and secular cooling of the LMO, that is written  
216 as in [Lister and Buffett \(1995\)](#) ; we neglect the heat flux associated to compositional  
217 changes in the LMO:

$$\begin{aligned} \epsilon\sigma 4\pi R_M^2 (T_s(t)^4 - T_\infty^4) &= \rho L \frac{dV_{\text{cryst}}}{dt} + h_{\text{LMO}}(t)V_{\text{LMO}}(t) \\ &\quad - \rho c_p \left( \int_{R_{\text{cu}}(t)}^{R_M} \frac{\partial T_{\text{LMO}}}{\partial t} 4\pi r^2 dr \right), \end{aligned} \quad (12)$$

218 where  $L$  is the latent heat of crystallization,  $h_{\text{LMO}}(t)$  the LMO radioactive heat  
219 production per unit volume,  $c_p$  the heat capacity. As the LMO temperature is  
220 uniform, the third term representing secular cooling on the right hand side of eq. (12)  
221 can be simplified using the expression of  $T_{\text{LMO}}$  from eq. (3). Given eq. (3) and eq. (12),  
222 the heat budget can be simplified as:

$$\begin{aligned} 4\pi R_M^2 \sigma \epsilon (T_s(t)^4 - T_\infty^4) &= 4\pi R_{\text{cu}}^2 \frac{dR_{\text{cu}}}{dt} \left[ \rho L + (T_{\text{OL}} - T_{\text{LMO}}(t)) \rho c_p \right] \\ &\quad + h_{\text{LMO}}(t) \frac{4}{3} \pi (R_M^3 - R_{\text{cu}}^3(t)). \end{aligned} \quad (13)$$

223 Equations (10) and (13) are solved numerically to calculate the temporal evolution  
224 of the cumulates radius  $R_{\text{cu}}$  and of the surface temperature  $T_s$  during the first stage.  
225 At the end of the first stage, the amount of heat stored in the cumulates that can be  
226 released during an overturn episode is calculated from:

$$\begin{aligned} E_{\text{cu}} &= \rho c_p \int_{R_{\text{co}}}^{R_{\text{cu}}} 4\pi (T_{\text{cu}}(r) - T_{\text{E}}) r^2 dr \\ &= \rho c_p \frac{4}{3} \pi \left[ (T_{\text{OL}} - T_{\text{E}})(R_{\text{cu}}^3 - R_{\text{co}}^3) - mC_0 (R_M^3 - R_{\text{co}}^3) \ln \left( \frac{R_M^3 - R_{\text{co}}^3}{R_M^3 - R_{\text{cu}}^3} \right) \right], \end{aligned} \quad (14)$$

227 where we use eq. (3).

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228 *3.2. Stage 2*

229 *3.2.1. Heat conduction in the crust and cumulates*

230 In the second stage, cooling of the LMO occurs by diffusion of heat through the  
231 floating anorthositic lid. Diffusion of heat in the cumulates is not negligible anymore  
232 and we solve for the 1D time-dependent conduction equation in both the crust and  
233 cumulates to obtain their temperature profiles as a function of time:

$$\frac{\partial T_i(t, r)}{\partial t} = \frac{\kappa_i}{r^2} \frac{\partial}{\partial r} \left( r^2 \frac{\partial T_i(t, r)}{\partial r} \right) + \frac{h_i(t, r)e^{-\lambda t}}{\rho c_p}, \quad (15)$$

234 where index  $i$  stands for crust or solid cumulates,  $\kappa_i = \frac{k_i}{\rho c_p}$  is the thermal diffusiv-  
235 ity and  $h_i(r)$  the heat production coefficient per unit volume, computed from the  
236 partition coefficient  $D$ , eq. (5) (see section 2.3).

237 For the anorthositic crust, the boundary conditions are  $T|_{r=R_{\text{cr}}} = T_{\text{E}}$  and  $T|_{r=R_{\text{M}}} =$   
238  $T_{\text{s}}$  (see section 2), where  $T_{\text{s}}$  is given by the balance between the radiative and con-  
239 ductive fluxes at the surface:

$$\epsilon \sigma (T_{\text{s}}^4 - T_{\infty}^4) = -k_{\text{cr}} \frac{\partial T_{\text{cr}}(t, r)}{\partial r} \Big|_{r=R_{\text{M}}}. \quad (16)$$

240 For the solid cumulates, we use:  $T|_{r=R_{\text{co}}} = T_{\text{co}}$  and  $T|_{r=R_{\text{cu}}} = T_{\text{E}}$  (see section 2). The  
241 diffusion equation is solved using a fully implicit finite volume scheme on a regular  
242 grid with a front-fixing method to deal with the growth of the solid layers with time  
243 (Appendix A).

244 *3.2.2. Core*

245 The core is considered as a sphere of uniform temperature  $T_{\text{co}}$ . Heat conservation  
246 in the core gives the evolution of  $T_{\text{co}}$ :

$$\rho_{\text{co}} c_{\text{p,co}} \frac{4}{3} \pi R_{\text{co}}^3 \frac{dT_{\text{co}}}{dt} = -4\pi R_{\text{co}}^2 q_{\text{co}}, \quad (17)$$

247 where  $\rho_{\text{co}}$  is the core density and  $c_{\text{p,co}}$  the heat capacity and the heat flux out of the  
248 core  $q_{\text{co}}$  is deduced from the temperature gradient at the base of the cumulates:

$$q_{\text{co}} = -k_{\text{cu}} \frac{\partial T_{\text{cu}}}{\partial r} \Big|_{r=R_{\text{co}}}. \quad (18)$$

249 Equation (17) is used to calculate the thermal evolution of the core, which gives the  
250 boundary condition at the base of the cumulates eq. (18).

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251 *3.2.3. Conservation of heat in the LMO in stage 2*

252 In stage 2, the magma ocean is a spherical shell of inner radius  $R_{\text{cu}}(t)$  and outer  
253 radius  $R_{\text{cr}}(t)$ . Heat conservation in the LMO requires that the heat flux evacuated by  
254 conduction through the anorthositic lid balances the sum of the heat fluxes coming  
255 from the diffusion of heat in the cumulates, latent heat released by crystallization in  
256 the LMO and radiogenic decay of HPEs (fig. 1c):

$$S_{\text{cr}}q_{\text{cr}} = S_{\text{cu}}q_{\text{cu}} + \rho L \frac{dV_{\text{cryst}}}{dt} + h_{\text{LMO}}e^{-\lambda t}V_{\text{LMO}}. \quad (19)$$

257  $S_{\text{cr}} = 4\pi R_{\text{cr}}^2(t)$  is the surface of the LMO-crust interface,  $S_{\text{cu}} = 4\pi R_{\text{cu}}^2(t)$  is the  
258 surface between the LMO and the cumulate layer,  $q_{\text{cu}}$  and  $q_{\text{cr}}$  are respectively the  
259 heat flux conducted away by the crust and the heat flux brought by conduction in  
260 the cumulates. These conductive fluxes are given by:

$$q_i = -k_i \left. \frac{\partial T_i}{\partial r} \right|_{r=R_i, i=\text{cr}, \text{cu}}, \quad (20)$$

261 and are computed at each time step from the temperature profiles in the crust and  
262 cumulates (section 3.2.1). We also consider the case of a thermal overturn of the  
263 cumulates and then modify the heat flux from the cumulates  $q_{\text{bot}}$  as described in sec-  
264 tion 3.3.  $V_{\text{LMO}}$  and  $h_{\text{LMO}}$  depend on time as the LMO crystallizes and concentration  
265 of HPEs increases (section 2.3).  $\frac{dV_{\text{cryst}}}{dt}$  is the total rate of crystallization, which is  
266 the sum of the rate of crystallization of anorthite and olivine-pyroxene components.  
267 Since anorthite and olivine-pyroxene are in eutectic proportion, we have:

$$\frac{dV_{\text{An}}}{dt} = C_{\text{E}} \frac{dV_{\text{cryst}}}{dt} = -4\pi R_{\text{cr}}^2 \frac{dR_{\text{cr}}}{dt}, \quad (21)$$

$$\frac{dV_{\text{Ol-Px}}}{dt} = (1 - C_{\text{E}}) \frac{dV_{\text{cryst}}}{dt} = 4\pi R_{\text{cu}}^2 \frac{dR_{\text{cu}}}{dt}. \quad (22)$$

269 Using eq. (21) into eq. (19), we obtain an equation for the evolution of  $R_{\text{cr}}$  as a  
270 function of time:

$$-\frac{dR_{\text{cr}}}{dt} = \frac{C_{\text{E}}}{\rho L} \left( -k_{\text{cr}} \left. \frac{\partial T_{\text{cr}}}{\partial r} \right|_{r=R_{\text{cr}}} - \frac{R_{\text{cr}}^3 - R_{\text{cu}}^3}{3R_{\text{cr}}^2} e^{-\lambda t} h_{\text{LMO}} + k_{\text{cu}} \frac{R_{\text{cu}}^2}{R_{\text{cr}}^2} \left. \frac{\partial T_{\text{cu}}}{\partial r} \right|_{r=R_{\text{cu}}} \right). \quad (23)$$

271 In case of an overturn, the last term on the right-hand-side is modified according  
272 to section 3.3. The evolution of  $R_{\text{cu}}$  is then derived from eqs. (21) and (22):

$$\frac{dR_{\text{cu}}}{dt} = - \left( \frac{1 - C_{\text{E}}}{C_{\text{E}}} \right) \frac{R_{\text{cr}}^2}{R_{\text{cu}}^2} \frac{dR_{\text{cr}}}{dt}. \quad (24)$$

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7 273 *3.3. Case of cumulates overturn during the stage 2*

8 274 At the end of the first stage, the temperature profile in the cumulates is super-  
9 275 isentropic and can induce a thermal overturn. The timescale for the onset of cumu-  
10 276 lates overturn may be short, as short as several thousand years (Elkins-Tanton et al.,  
11 277 2011; Boukaré et al., 2018; Morison et al., 2019), which is however longer than the  
12 278 first stage of LMO solidification (see section 3.1). Consequently, heat released by  
13 279 the overturn can be considered in the second stage only and heat conservation in the  
14 280 LMO eq. (19) then writes:

$$S_{\text{top}}q_{\text{top}} = \Phi_{\text{ov}}(t) + \rho L \frac{dV_{\text{cryst}}}{dt} + h_{\text{LMO}}e^{-\lambda t}V_{\text{LMO}}, \quad (25)$$

21 281 where  $\Phi_{\text{ov}}$  represents the rate of heat released at the surface of cumulates during over-  
22 282 turn and varies with time as the overturn progresses. As the onset time of overturn  
23 283 is likely short compared to the timescale of the second stage of LMO solidification  
24 284 (thousands of years compared to millions of years), and to explore the effect of over-  
25 285 turn decay time, we parameterise this rate of heat released with an exponential decay  
26 286 function:

$$\Phi_{\text{ov}}(t) = Qe^{-\frac{t}{\tau_{\text{ov}}}}, \quad (26)$$

30 287 where  $Q$  is the heat flux from the overturn at the start of stage 2, and  $\tau_{\text{ov}}$  its  
31 288 characteristic decay time. Assuming that the additional heat stored in the cumulates  
32 289 is fully released during the overturn, we have:

$$E_{\text{cu}} = \int_0^{\infty} \Phi_{\text{ov}}(t)dt = Q\tau_{\text{ov}}. \quad (27)$$

35 290 The total overturn energy available,  $E_{\text{cu}}$ , is determined by the choice of phase diagram  
36 291 through eq. (14). The values of  $Q$  and  $\tau_{\text{ov}}$  can then vary with the constraints that  
37 292 their product must equal  $E_{\text{cu}}$ , eq. (27).

44 293 **4. Results**

46 294 *4.1. Reference case*

47 295 We first show the thermal evolution of a reference case, considering an initial  
48 296 global magma ocean, a final crustal thickness of 44 km and 80% of crystallization at  
49 297 the end of the first stage (i.e.:  $C_0 = 0.075$ ,  $C_E = 0.37$ ), no overturn in the cumulates  
50 298 and a low thermal conductivity for the crust,  $k_{\text{cr}} = 2 \text{ W m}^{-1} \text{ K}^{-1}$ , corresponding to a  
51 299 thermal diffusivity of  $\kappa_{\text{cr}} = 6.06 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$  (Branlund and Hofmeister, 2012; Roy  
52 300 et al., 2021).

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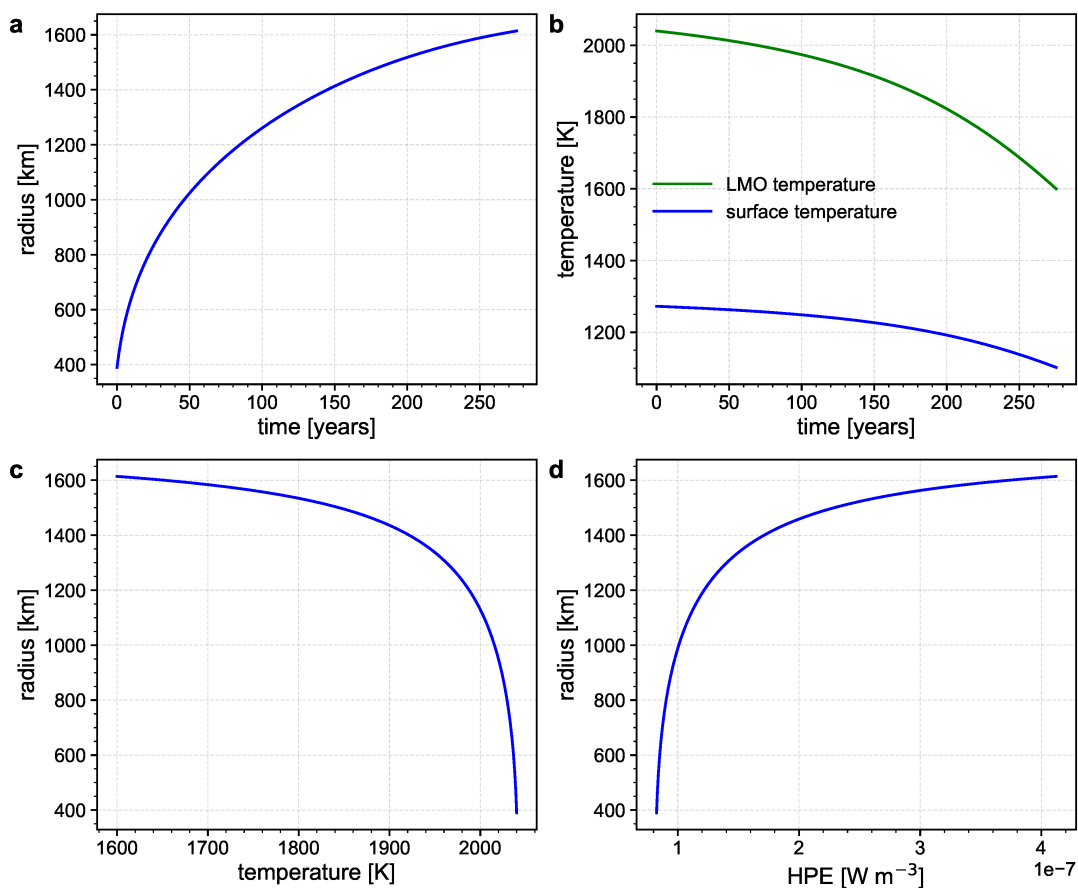


Figure 3: Temporal evolution of a) the radius of the cumulates and b) the surface temperature in blue and the LMO temperature in green. Radial evolution of c) the temperature in the cumulates and d) the heat production in the cumulates.



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301 *4.1.1. Stage 1*

302 The first stage is very fast: in 275 yr, 80% of the LMO is crystallized and the  
303 eutectic temperature is reached (Figure 3a). This timescale is much shorter than  
304 the diffusion timescale in the cumulates, it is also shorter than the onset time for  
305 cumulates overturn (Morison et al., 2019; Hess and Parmentier, 1995; Elkins-Tanton  
306 et al., 2011) ; this justifies that we neglect the heat flux from the cumulates for  
307 this stage. The temperature in the cumulate layer decreases with radius following  
308 the decrease of the LMO temperature with time, which is associated to its gradual  
309 enrichment in anorthite (fig. 3 b and c). As the LMO solidifies, its concentration  
310 in HPEs increases and therefore the concentration in the cumulates increases with  
311 radius (fig. 3d).

312 *4.1.2. Stage 2*

313 The total solidification time at the end of the second stage is 166 Myr in the  
314 reference case (fig. 4a). Owing to the presence of the conductive lid, the heat loss  
315 from the LMO is significantly reduced compared to stage 1 (fig. 4a and b). The  
316 temperature profile within the crust gradually approaches a linear conduction profile  
317 (inset in fig. 4c), typical of steady-state conduction in a thin shell with negligible  
318 heat production. In the cumulates, the larger heat production in the solid at the end  
319 of LMO crystallization tends to curve the temperature profile with time (fig. 4c).  
320 As the crust thickness increases, its basal heat flux decreases and the growth rate of  
321 both the crust and cumulates decreases over time, which implies a decrease of the  
322 latent heat release (purple curve on fig. 4b). The heat flux due to radioactive heat  
323 production in the LMO remains quasi-constant throughout the solidification process  
324 and, although latent heat released by crystallization first dominates the LMO heat  
325 budget, radioactive heat production prevails after 25 Myr (red dashed line fig. 4b).  
326 Thus, radioactive heat production must be taken into account to properly compute  
327 the solidification time of the LMO.

328 *4.2. Exploration of the model parameters*

329 Having presented a typical evolution for the reference case, we now explore the  
330 effects of varying the main control parameters: initial composition  $C_0$ , or equiva-  
331 lently final crust thickness, eutectic composition  $C_E$ , and thermal conductivity  $k_{cr}$   
332 (or equivalently thermal diffusivity  $\kappa_{cr}$ ), within the range given in table 2.

333 As the LMO solidification time is controlled by the timescale for diffusion of  
334 heat through the crust,  $\tau_{diff}^{cr} = (R_M - R_{cr})^2 \rho_{cr} c_p / k_{cr}$ , it strongly depends on the  
335 crustal conductivity and final crustal thickness. As shown by Maurice et al. (2020)  
336 and Zhang et al. (2021b), the lower the conductivity  $k_{cr}$  and the diffusivity  $\kappa_{cr}$ ,

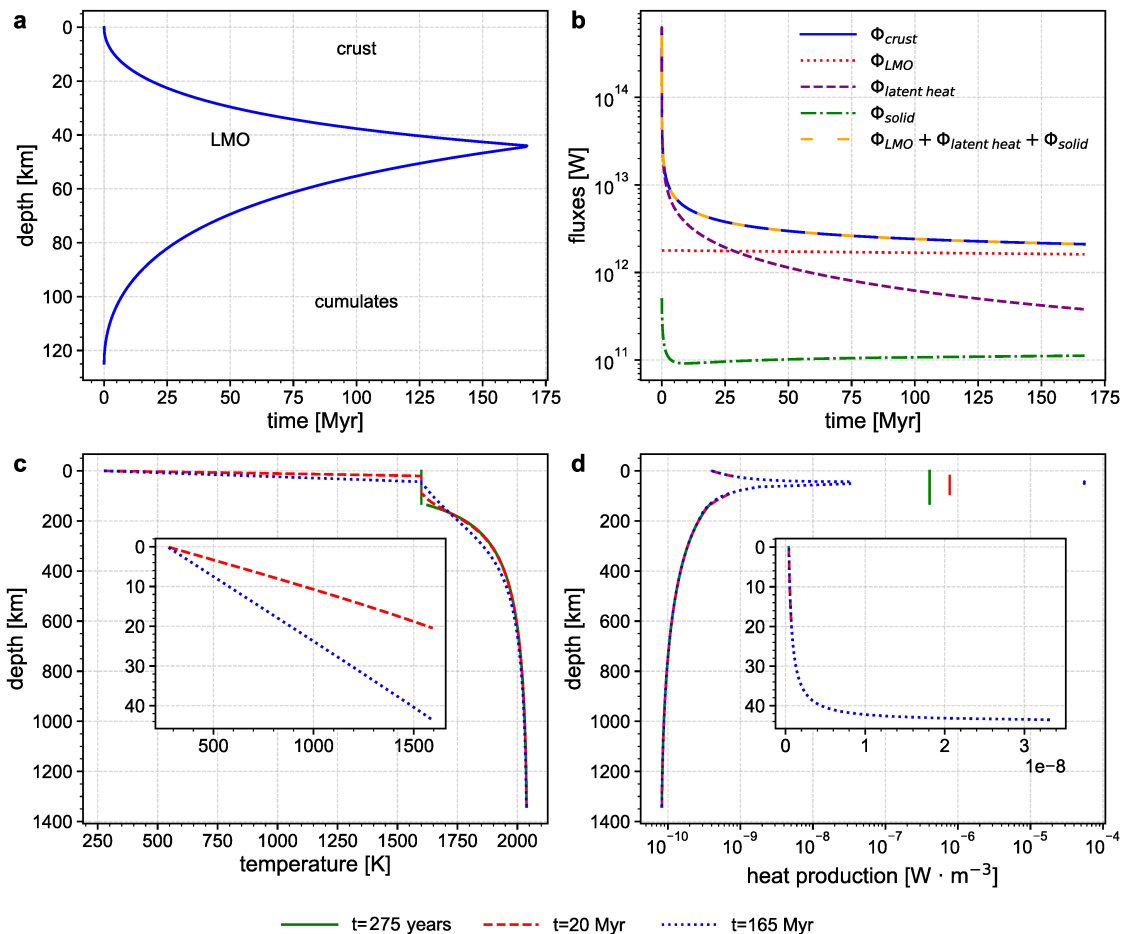


Figure 4: Temporal evolution of a) the crustal radius and cumulates radius, b) the various contributions to the LMO heat budget: conductive flux at the base of the crust  $\Phi_{cr}$  (blue solid line), heat flux from radioactive decay of HPEs in the LMO  $\Phi_{LMO}$  (red dotted line), latent heat flux  $\Phi_{latent\ heat}$  (dashed purple line), conductive flux from the cumulates  $\Phi_{cu}$  (dash-dotted green line), sum of  $\Phi_{LMO}$ ,  $\Phi_{latent\ heat}$  and  $\Phi_{cu}$  (dash-dotted green line). c) Full temperature profile at times  $t = 275$  yr (green line),  $t = 20$  Myr (in red dashed line) and at  $t = 165$  Myr (blue dotted line). Temperature profiles in the crust are shown in the inset. d) Heat production distribution as a function of radius at the same times as in panel c) The inset shows the radial heat production distribution in the crust.

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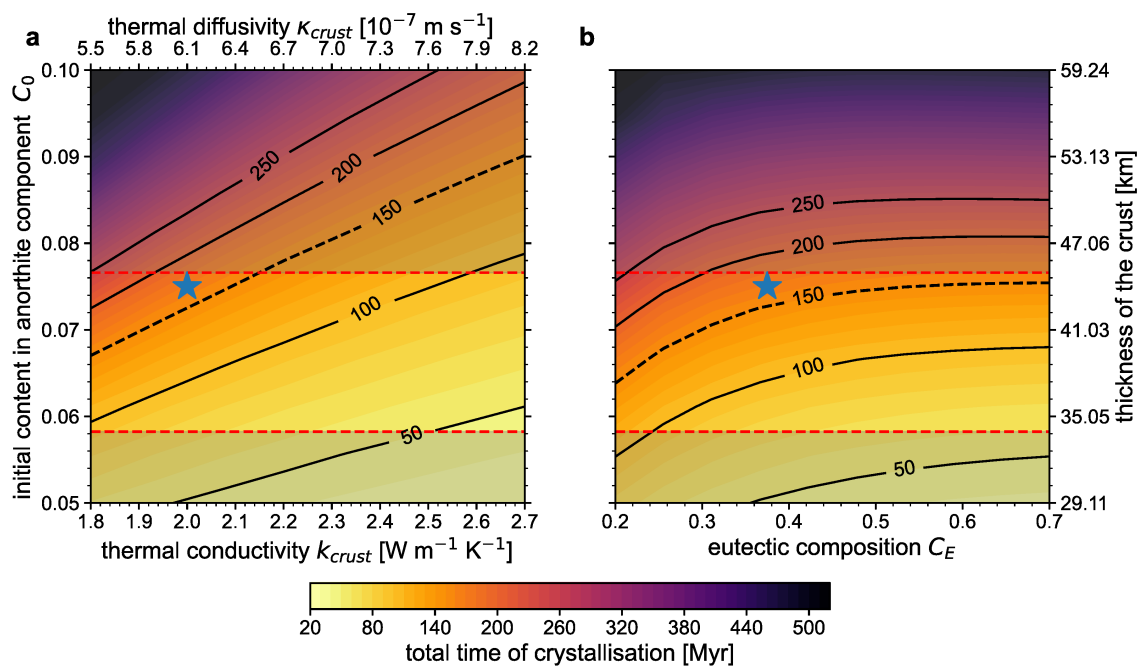


Figure 5: a) LMO solidification time as a function of initial anorthite content  $C_0$  and crust thermal conductivity  $k_{cr}$  for a fixed eutectic composition corresponding to  $C_E = 0.37$  of crystallization at the end of the first stage. b) Total crystallization time as a function of initial anorthite content  $C_0$  and eutectic composition  $C_E$  for a fixed value of  $k_{cr} = 2 \text{ W m}^{-1} \text{ K}^{-1}$ . The bright area between the red dashed lines is the acceptable range of crustal thickness as estimated from the inversion of topography and gravity data by [Wieczorek et al. \(2013\)](#). The dashed lines corresponds to the estimated values of the solidification time according to [Fu et al. \(2023\)](#). The blue star corresponds to our reference case.

337 the more difficult it is to extract heat through the crust and the longer it takes  
 338 for the LMO to crystallize (fig. 5a). The thermal conductivity of anorthosites does  
 339 not appear to depend on temperature (Roy et al., 2021), we thus use a constant  
 340 value. Thermal conductivity measurements on anorthosites vary between 1.5 and  
 341  $2.5 \text{ W m}^{-1} \text{ K}^{-1}$ , while the thermal diffusivity range between  $5 \times 10^{-7}$  and  $8 \times 10^{-7} \text{ m}^2$   
 342  $\text{s}^{-1}$  (Roy et al., 2021; Clauser and Huenges, 1995). We thus vary  $k_{\text{cr}}$  between 1.8 and  
 343  $2.7 \text{ W m}^{-1} \text{ K}^{-1}$ , which gives values for  $\kappa_{\text{cr}}$  between  $5 \times 10^{-7}$  and  $8 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$  given  
 344 our chosen density and heat capacity value (see table 1). Increasing  $k_{\text{cr}}$  from 1.8 to  
 345  $2.7 \text{ W m}^{-1} \text{ K}^{-1}$  the solidification timescale decreases by 160 Myr (going from 250 Myr  
 346 to 90 Myr) for a 45 km thick crust (fig. 5a). The LMO solidification time increases  
 347 with the initial anorthite content  $C_0$  (fig. 5a and b) because the larger  $C_0$ , the thicker  
 348 the anorthositic crust, the slower the heat transfer through the crust is. For a crustal  
 349 thickness increasing from 34 to 45 km, the range of crustal thicknesses compatible  
 350 with the inversion of gravity and topography data (Wieczorek et al., 2013), the  
 351 LMO crystallization time increases from 70 to 170 Myr using  $k_{\text{cr}} = 2 \text{ W m}^{-1} \text{ K}^{-1}$  and  
 352  $C_{\text{E}} = 0.37$ .

353 The LMO solidification time also increases as the anorthite eutectic composition  
 354  $C_{\text{E}}$  decreases: if  $C_{\text{E}}$  is smaller, more olivine and pyroxene crystallize during the  
 355 second stage because the crust starts to form earlier for a given value of  $C_0$  (i.e. a  
 356 given final crustal thickness), and more latent heat must be extracted through the  
 357 anorthositic lid. For instance, decreasing by 0.2 the value of  $C_{\text{E}}$  in our reference case,  
 358 which corresponds to the estimated difference between the compositional models of  
 359 Charlier et al. (2018) and Lin et al. (2017) (see table 1), makes the total solidification  
 360 time increase from 160 Myr to 200 Myr for a final crustal thickness of 45 km (fig. 5b).

361 Overall, the solidification time can range between 44 Myr and 250 Myr.

### 362 4.3. Effect of an overturn

363 Here, we examine the effect of cumulates overturn considering the same parame-  
 364 ters as in our reference case. The heat flux from the cumulates is then not diffusive  
 365 but decreases exponentially with time (section 3.3). Using the phase diagram of our  
 366 reference case, we estimate that the additional heat stored in the cumulates that  
 367 can be released during overturn is of  $1.8 \times 10^{28} \text{ J}$ , eq. (14). This is significantly  
 368 larger than the total amount of latent heat released during the second stage, which  
 369 is  $0.7 \times 10^{18} \text{ J}$ , showing that, depending on its amplitude and decay rate, cumulate  
 370 overturn may induce melting of the crust. Given our parametrisation, the initial  
 371 heat flux  $Q$  increases as the characteristic decay time of the overturn decreases fol-  
 372 lowing eq. (27). We explore decay times between 1.8 and 180 Myr. For a short-lived  
 373 overturn such that  $\tau_{\text{ov}} \lesssim 5 \text{ Myr}$ , the solidification time of the LMO is reduced when

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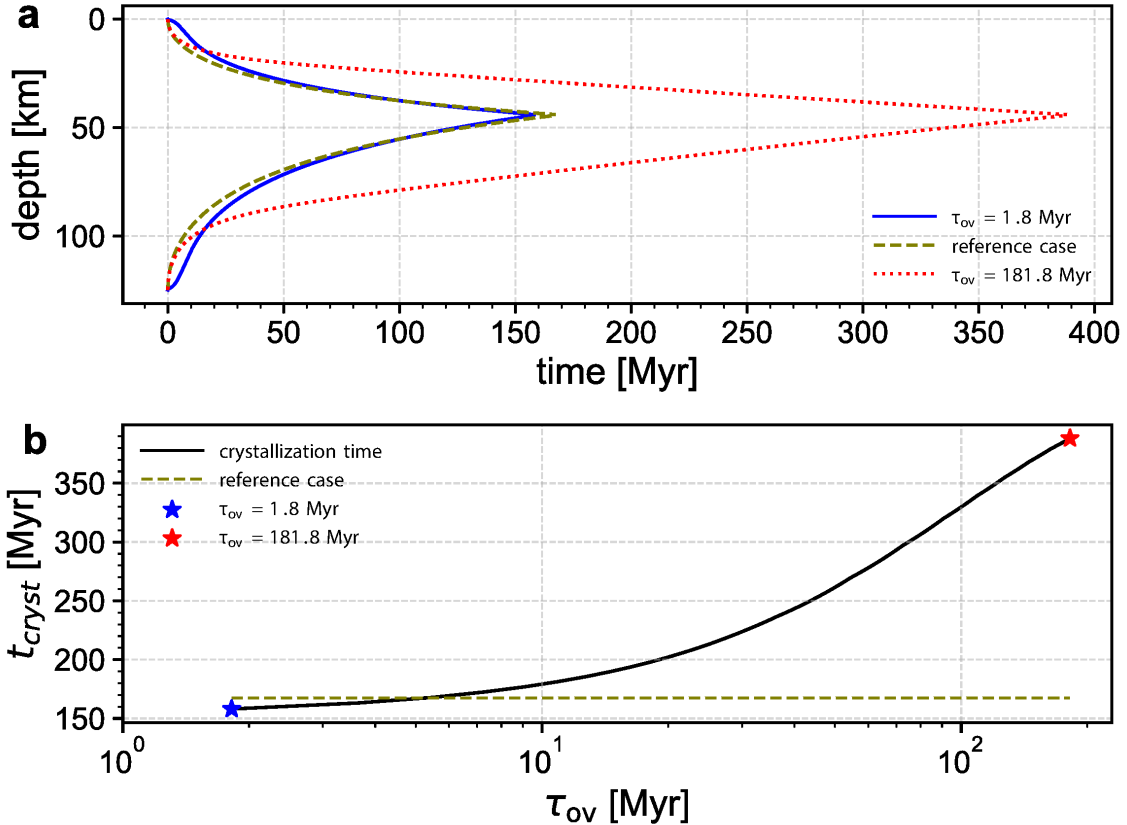


Figure 6: a) Temporal evolution of the crustal radius and cumulates radius, in the reference case assuming a conductive heat flux from the cumulates fig. 4a (olive green dashed line), in the case of a short-lived overturn (blue line), and for a long-lived overturn (dotted red line). The parameters are the same as for the reference case, *i.e.*:  $C_0 = 0.075$ ,  $k_{cr} = 2 \text{ W m}^{-1} \text{ K}^{-1}$  and  $C_E = 0.37$ . b) Crystallization time (black line) as a function of  $\tau_{ov}$  the decay time of the overturn. The olive green dashed line corresponds to the LMO solidification time in the reference case and the blue and red stars correspond to the extremums of a).

374 accounting for an overturn compared to our reference case because the heat released  
 375 by the overturn is quickly evacuated through the initially very thin crust, marked by  
 376 a large temperature gradient and basal heat flux (fig. 6b). The growth rate of the  
 377 crust remains negligible for the first million years and then follows the trend of the  
 378 reference scenario (fig. 6a) once the overturn has died out. When the crust is thicker  
 379 and more insulating, no conductive heat flux remains to be evacuated on the contrary  
 380 to the reference case, which explains the shorter solidification time for a short-lived  
 381 overturn. For decay times smaller than  $\sim 1.8$  Myr, the heat flux  $Q$  becomes too large  
 382 to be evacuated through the initially thin crust and can melt it, possibly entirely  
 383 if the overturn proceeds quickly enough. This would increase the heat flux at the  
 384 LMO surface and accelerate even more its cooling. If, on the contrary, the decay  
 385 time of the overturn exceeds 5 Myr, a significant fraction of the heat stored in the  
 386 cumulates is then released when the crust becomes thick and insulating, and this  
 387 can significantly prolong the LMO solidification time: for  $\tau_{\text{ov}} = 100$  Myr, the LMO  
 388 solidification timescale reaches 388 Myr (fig. 6).

## 389 5. Discussion

### 390 5.1. LMO solidification time

391 The LMO solidification time depends strongly on the crustal conductivity, initial  
 392 anorthite content and eutectic composition. For a crustal thickness of 44 km, a  
 393 percentage of crystallization of 80 % at the end of the first stage, as in [Elkins-Tanton  
 394 et al. \(2011\)](#), and using  $k = 2 \text{ W m}^{-1} \text{ K}^{-1}$ , the LMO solidification time is 166 Myr.  
 395 This is much larger than the timescale of  $\sim 10$  Myr estimated by [Elkins-Tanton et al.  
 396 \(2011\)](#), who neglect heat production by radioactive decay, an important heat source  
 397 in the LMO energy budget (fig. 4b), and assume an initial LMO depth of 1000 km  
 398 and a larger thermal diffusivity of  $10^{-6} \text{ m}^2 \text{ s}^{-1}$  corresponding to a crustal conductivity  
 399 of  $k_{\text{cr}} = 3.3 \text{ W m}^{-1} \text{ K}^{-1}$  in our model. Neglecting the heat flux from the cumulates,  
 400 radioactive heat production and assuming a steady-state conduction in the crust, an  
 401 analytical solution for the total solidification time  $\tau_{\text{sol}}$  can be obtained ([Appendix B](#)):

$$46 \tau_{\text{sol}} = -\frac{\rho L}{R_{\text{M}} k_{\text{cr}} \Delta T} \frac{1}{C_{\text{E}}} \left[ \frac{R_{\text{M}} R_{\text{cr, f}}^2}{2} - \frac{R_{\text{cr, f}}^3}{3} - \frac{R_{\text{M}}^3}{6} \right]. \quad (28)$$

402 Using  $C_{\text{E}} = 0.37$ ,  $R_{\text{crust, final}} = 1693 \text{ km}$  and  $k_{\text{cr}} = 3.3 \text{ W m}^{-1} \text{ K}$ , this solidification  
 403 time is  $\sim 30$  Myr, which is still significantly larger than in [Elkins-Tanton et al.  
 404 \(2011\)](#). Results for our reference case are very similar to that of [Maurice et al.  
 405 \(2020\)](#), who use the more complex, i.e. pressure-dependent, version of the phase

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406 diagram of [Elkins-Tanton et al. \(2011\)](#): for a magma ocean extending down to the  
407 core and accounting only for heat conduction in the cumulates as assumed here, their  
408 solidification timescale is 165 Myr (fig. S6 of [Maurice et al., 2020](#)), which is very close  
409 to our estimate. Small differences arise from the variation of the phase diagram with  
410 pressure, which corresponds to a variation of  $C_E$  during the second stage, as well as  
411 from a different distribution in heat producing elements, whose partition coefficient  
412 is not constant and depends on the crystallizing minerals in [Maurice et al. \(2020\)](#).  
413 The influence of the pressure-dependence of the phase diagram or of a varying HPEs  
414 partition coefficient during LMO solidification are thus clearly of second order given  
415 our current knowledge of the average lunar crust thickness, crust bulk conductivity  
416 and phase diagram (fig. 5). Our simplified, though physically robust, approach, with  
417 a binary eutectic phase diagram, appears thus of appropriate complexity to estimate  
418 the LMO solidification timescale.

419 With a range of LMO solidification time between 45 and 250 Myr for crustal  
420 thicknesses between 34 and 45 km (fig. 5), the lifetime of the LMO appears increased  
421 compared to initial estimates by [Elkins-Tanton et al. \(2011\)](#). Our results show that  
422 the crustal conductivity value as well as the occurrence and timescale of cumulates  
423 overturn can significantly affect estimates of the LMO solidification time. The ther-  
424 mal conductivity of anorthosites is low compared to other crustal rocks ([Clauser and  
425 Huenges, 1995](#)) which increases LMO insulation. Analyses of GRAIL gravity data  
426 additionally shows that the lunar crust is highly porous, with an average porosity  
427 of  $\sim 12\%$  over a few tens of kilometers depth ([Wieczorek et al., 2013](#)). Accounting  
428 for such a large porosity would significantly decrease the crustal conductivity and  
429 prolong the LMO duration ([Zhang et al., 2021b](#)). However, the lunar megaregolith  
430 most likely formed after LMO solidification, concomittantly to large basins through  
431 impacts and ejectas ([Liu et al., 2022](#)). Impacts large enough to deeply fracture the  
432 crust and forge a megaregolith may in fact create holes in a flotation crust, which  
433 would accelerate the cooling of the underlying LMO ([Perera et al., 2018](#)).

434 The LMO solidification time can be significantly extended if the overturn has  
435 a long duration (fig. 6). However, the impact of the overturn is highly sensitive  
436 to its lifetime, it may even reduce the LMO solidification time by  $\sim 10$  Myr if its  
437 characteristic decay timescale is smaller than 5 Myr, because all the heat from the  
438 cumulates is then quickly evacuated through the thin floating crust. Furthermore,  
439 the decay timescale of the thermal overturn is not likely to be several orders of  
440 magnitude longer than its initiation timescale, which is likely to be less than 1 Myr  
441 ([Morison et al., 2019](#); [Boukaré et al., 2018](#)).

442 Studies of LMO solidification often use an initial depth of 1000 km, based on a  
443 potential seismic discontinuity ([Khan et al., 2000](#); [Lognonné, 2005](#)). Using our model

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444 and varying the initial LMO depth still leads to a solidification time larger than  
445 50 Myr (fig. 7), using  $k_{\text{cr}} = 2 \text{ W m}^{-1} \text{ K}^{-1}$ , as it mainly depends on the final crustal  
446 thickness. Our results thus suggest a long-lived LMO and show that this classical  
447 scenario, which explains a wide range of lunar crust characteristics, from the large  
448 feldspar enrichment of the lunar crust to the formation of the KREEP Province, can  
449 also explain the age range of anorthosites, in particular given dating uncertainties  
450 (Borg and Carlson, 2023). By comparison, the model of Michaut and Neufeld (2022),  
451 which propose that anorthosites form by extraction of melts in a stagnant lid from  
452 a slushy magma ocean could lead to even longer crustal formation timescales and  
453 account for the compositional heterogeneities of anorthosites, a characteristic that is  
454 more difficult to explain with the flotation scenario (Gross et al., 2014; Russell et al.,  
455 2014). More data on lunar rocks are thus needed to fully understand the lunar crust  
456 formation scenario.

### 457 5.2. Formation age of FAN 60025 and age of the Moon

458 Based on the compositional profiles of pigeonite and augite, McCallum and  
459 O'Brien (1996) estimated a maximum cooling rate of  $18 \text{ K Myr}^{-1}$  for ferroan anorthosite  
460 sample 60025. Using our model, we calculate the cooling rate in the crust as a func-  
461 tion of depth and time (fig. 8). Cooling rates are initially large, of the order of a few  
462 hundreds of  $\text{K Myr}^{-1}$  and decrease with time as the crust thickens. Calculated cool-  
463 ing rates are consistent with the upper limit given by McCallum and O'Brien (1996)  
464 and show that sample 60025 formed at depths larger than 15 km and at least 20 Myr  
465 after the onset of LMO crystallization for a closure temperature ranging between  
466 1070 K and 1300 K.

467 Fu et al. (2023) argued that the Ca and Mg isotopic composition of anorthosite  
468 sample 60025 is compatible with this sample marking the end of LMO differentia-  
469 tion and indicates a formation after more than 99% of solidification. Such a large  
470 percentage of crystallization is reached after at least 40 Myr, in the configuration of  
471 the shortest solidification timescale (fig. 8b), and up to 200 Myr after Moon forma-  
472 tion (longest crystallization time, fig. 8a). Anorthosites crystallizing after 99% of  
473 LMO solidification would form at more than 30 km depth and cool down at rates  $\sim$   
474  $10 \text{ K Myr}^{-1}$  if considering the shortest solidification timescale, or  $<1 \text{ K Myr}^{-1}$  for the  
475 "longest scenario". These cooling rates are compatible with estimates of McCallum  
476 and O'Brien (1996).

477 Sample 60025 has been dated at 4.51 Gyr using the U-Pb method and the pla-  
478 gioclase fraction by Hanan and Tilton (1987). A later study by Borg et al. (2011)  
479 reported an age of 4.360 Gyr using the U-Pb and Sm-Nd methods. However, as ar-  
480 gued by Borg and Carlson (2023), the age reported by Hanan and Tilton (1987) can



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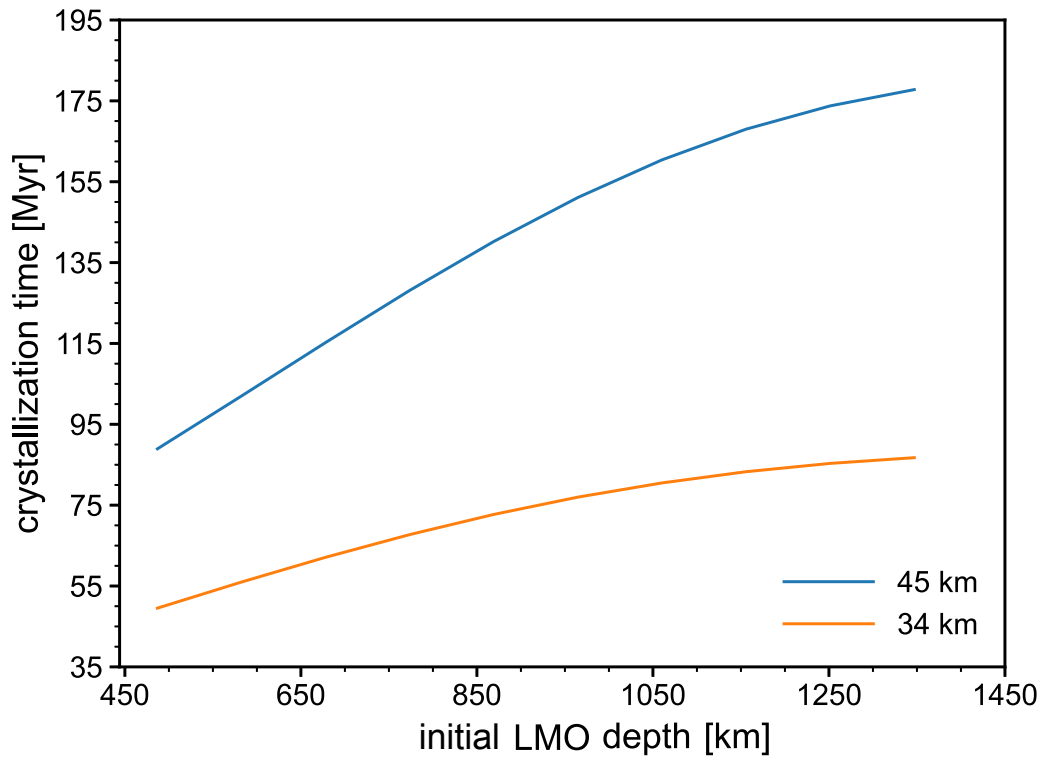


Figure 7: LMO solidification time as a function of initial LMO depth for the two extreme values of average crustal thickness according to [Wieczorek et al. \(2013\)](#): a 45 km thick crust (blue line) and a 34 km thick crust (orange line).  $C_E = 0.37$ ,  $k_{cr} = 2 \text{ W m}^{-1} \text{ K}^{-1}$  and  $C_0$  is calculated using eq. (4).

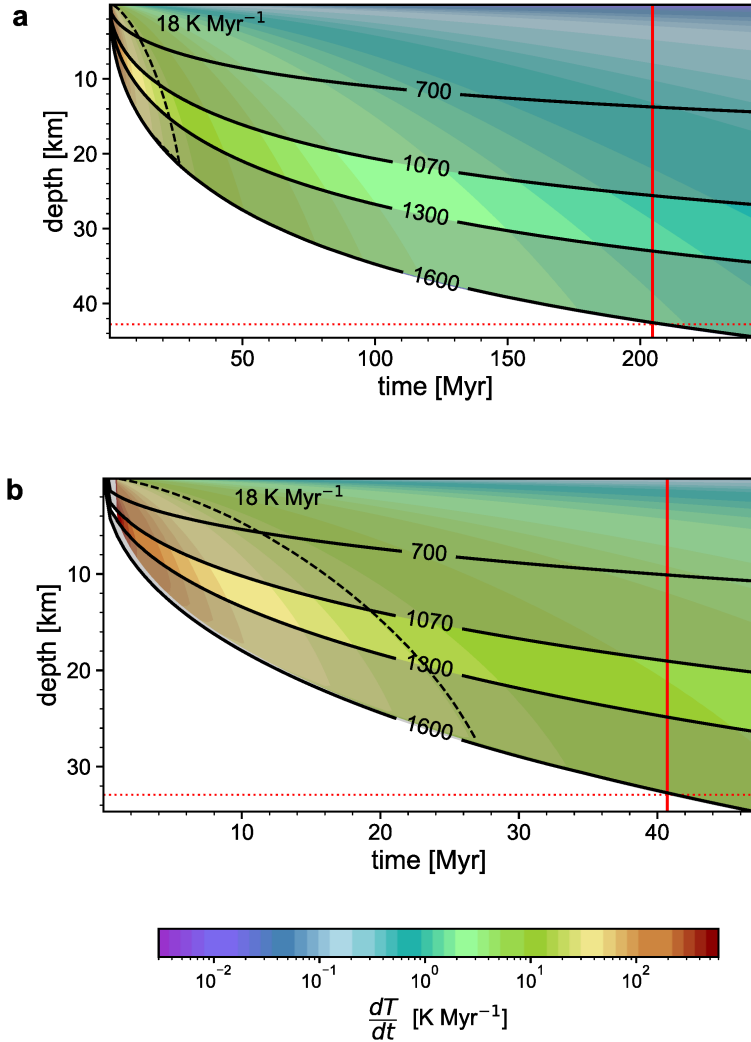


Figure 8: Cooling rate as a function of depth and time a) for the "longest scenario" where the final crustal thickness is 45 km and the total solidification time is 246 Myr, and where we use  $C_0 = 0.0765$ ,  $C_E = 0.37$  and  $k_{cr} = 1.8 \text{ W m}^{-1} \text{ K}^{-1}$  and b) for a thinner crustal thickness of 35 km and a shorter LMO solidification time of 47 Myr, and where we use  $C_0 = 0.06$ ,  $C_E = 0.37$  and  $k_{cr} = 2.7 \text{ W m}^{-1} \text{ K}^{-1}$ . These cases correspond to the shortest and longest total solidification time (fig. 5). The black dashed line corresponds to the maximum cooling rate of  $18 \text{ K Myr}^{-1}$  inferred for FAN sample 60025 by McCallum and O'Brien (1996). The solid black lines represent different isotherms and the range of temperatures corresponding to the cooling rates determined by (McCallum and O'Brien, 1996) are shown in the brighter area. The red solid and dotted lines correspond respectively to the times and crustal thicknesses reached when 99% of the LMO has crystallized.

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481 be questioned as it assumed that the initial Pb isotope composition of the Moon was  
482 identical to that of the Canyon Diablo iron meteorite, commonly used to represent  
483 the starting composition of the Solar System. There are many reasons to think that  
484 the Pb isotope composition of the Moon did not result from a single stage process  
485 and that the Moon precursors must have initially evolved with a U/Pb higher than  
486 that of Canyon Diablo. A multistage evolution would certainly modify the age cal-  
487 culated by [Hanan and Tilton \(1987\)](#). Our modeling results are indeed not consistent  
488 with the age of [Hanan and Tilton \(1987\)](#) and are more consistent with an age of  
489  $4.360 \pm 0.003$  Gyr for the last stages of the LMO crystallization ([Borg et al., 2011](#)).

490 Our model depicted in [fig. 5](#) indicate that 99% of crystallization for a whole  
491 magma ocean (depth extending to 1300 km) is reached after 40 to 200 Myr, depending  
492 on the crustal thickness (34-45 km) and thermal conductivity. This means that the  
493 age of the Moon formation would have to be older than 4400-4560 Myr. This inference  
494 is at odds with the conclusions of [Borg and Carlson \(2023\)](#) who argued that the  
495 clustering of age for lunar rocks between 4.33-4.360 Gyr was a strong indication that  
496 the lunar magma ocean had cooled rapidly and that the Moon forming impact should  
497 have taken place at 4.360 Gyr. Our study suggests that an alternative interpretation  
498 of the chronological record for lunar rocks is necessary. First, the clustering of ages is  
499 indeed around 4.360 Gyr but, based on the data compiled by [Borg and Carlson \(2023\)](#)  
500 and including all uncertainties stemming from different dating methods, we estimate  
501 that the range of possible ages is 4297 to 4402 Myr suggesting that the duration  
502 of LMO crystallization could actually be longer than what is argued in that study.  
503 Second, apart from the age of mare basalts based on the  $^{146}\text{Sm}$ - $^{142}\text{Nd}$  chronometer, all  
504 the ages correspond to rocks formed at a late stage of LMO crystallization (>80%) or  
505 simply post-date LMO crystallization (KREEP, FAN, Mg- and alkali suites). Thus,  
506 the most problematic age is the  $^{142}\text{Nd}$  model age of lunar mare basalts because the  
507 mare basalts are generally thought to originate from the melting of the first cumulates  
508 of olivine and pyroxene, that should have the oldest age of LMO crystallization ([Borg  
509 et al., 2019](#)). Looking more closely at this dataset (as compiled in [Borg et al., 2019](#)),  
510 several questions arise: in their [fig. 6](#), if one considers only the mare basalts, a  
511 dichotomy appears between the low Ti-basalts and the high Ti-basalts, with the high-  
512 Ti basalts characterized by high Sm/Nd ratios that strongly constrain the  $^{142}\text{Nd}$   
513 age. Four of the low-Ti basalts plot in the field of high Ti-basalts. However, by  
514 looking at their actual Ti contents, all of these samples are characterized by Ti  
515 contents greater than 4 wt% and cannot represent melts derived from an olivine +  
516 orthopyroxene cumulates, as is commonly thought for low-Ti basalts. They rather  
517 represent intermediate Ti basalts. As argued by [Shearer et al. \(2006\)](#) the intermediate  
518 Ti basalts with 4 wt%  $\text{TiO}_2$  require a Ti-rich phase in their sources derived from the

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519 late stage Ti-rich cumulates. Thus, one cannot consider that these rocks formed early  
520 during LMO crystallization, similarly to true low-Ti basalts. Consequently, it is not  
521 surprising that their apparent ages are very similar to those of more differentiated  
522 rocks such as anorthosites, or KREEP rocks. If one reexamines the  $^{146}\text{Sm}$ - $^{142}\text{Nd}$   
523 isochron in light of these considerations, it becomes obvious that the only rocks  
524 representative of the early stage of LMO crystallization are the low-Ti mare basalts  
525 with  $\text{TiO}_2 < 2\text{wt}\%$ . This is represented by the rocks that have a nearly chondritic  
526 Sm/Nd source ratio and are as expected to be mildly fractionated (in incompatible  
527 elements) from a bulk Silicate Moon source. From these rocks, it becomes very  
528 difficult to extract an isochron age because there is little spread in their  $^{147}\text{Sm}/^{144}\text{Nd}$   
529 ratios. To the very least, their variable Ti contents could still be used to argue  
530 that the residual  $^{142}\text{Nd}/^{144}\text{Nd}$  versus  $^{147}\text{Sm}/^{144}\text{Nd}$  source trend represents a mixing  
531 line. These considerations strongly relax the argument that the LMO life span was  
532 extremely short and that requires that the Moon formation only shortly precedes the  
533 time of late LMO crystallization.

534 Third, there are new observational evidence for zircons that have crystallized  
535 at a time that precedes the late stage crystallization of the magma ocean (Zhang  
536 et al., 2021a), yielding an age of  $4460 \pm 31$  Myr. This age was confirmed by a closer  
537 examination of the same sample using an atom probe that demonstrated that there  
538 was no clustering of radiogenic Pb (Greer et al., 2023) that would have resulted from  
539 a disturbance of radiogenic lead distribution. Thus, these observations suggest that  
540 the age of the Moon is at least 4460 Ma, which is well before the age of FAN and  
541 Mg-suite samples reported above, and consistent with our estimated age range for  
542 the Moon, based on our model.

543 Last, the studies of Jacobson et al. (2014) and Woo et al. (2024) have shown that  
544 the late accretion to the Earth that post-dates the Moon forming impact, indicated  
545 by highly siderophile element (HSE) enrichment in the bulk Silicate Earth, is not  
546 compatible with a late age for the formation of the Moon, as argued by Borg and  
547 Carlson (2023). Incidentally, these ages are also in agreement with the recent Rb-Sr  
548 age determination of the Moon (Yobregat et al., 2024), showing that the Moon-  
549 forming impact should have taken place no later than 79 Myr after the beginning of  
550 the Solar System.

## 551 6. Conclusion

552 We study the thermal evolution of the lunar magma ocean (LMO) using a phys-  
553 ically robust 1D model. Our model is based on the sequential crystallization of  
554 olivine-pyroxene cumulates and a floating anorthositic crust, while taking into ac-  
555 count conductive heat fluxes through the crust and from the cumulates, latent heat

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556 released by crystallization and radiogenic heat production. The first stage of crys-  
557 tallization is very fast: in 275 yr, 80 % of the LMO is crystallized. The second stage  
558 is slowed down by the formation of the buoyant crust with a low thermal diffusivity.  
559 For our reference case, the total LMO solidification time is 166 Myr. This solidifi-  
560 cation time depends on the crust conductivity and on the parameters of the phase  
561 diagram: the initial content in anorthite component  $C_0$ , which controls the final crust  
562 thickness, and the eutectic composition  $C_E$ , which affects the LMO thickness at the  
563 initiation of anorthosite crust growth. Considering these parameters variability, the  
564 LMO solidification time ranges from 45 Myr to 250 Myr and is long-lived.

565 The occurrence of an overturn during the second stage of LMO evolution may  
566 reduce this solidification time if short-lived (i.e. less than  $\sim 5$  Myr), or extend  
567 it if long-lived. Ferroan anorthosite sample 60025 has been consistently dated at  
568 4360 Myr by different radiometric systems. Considering that this sample derives from  
569 more than 99% of LMO crystallization, as suggested from its Ca and Mg isotopic  
570 composition, we estimate that the Moon formed between 4400 Myr and 4560 Myr.

## 571 Acknowledgments

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576 edges support of the Institut Universitaire de France.

## 577 Appendix A. Numerical solution

578 To obtain the flux at the base of the crust and at the top of cumulates, we need  
579 to solve the diffusion equation in the two layers:

$$580 \frac{\partial T_i(t, r)}{\partial t} = \frac{\kappa_i}{r^2} \frac{\partial}{\partial r} \left( r^2 \frac{\partial T_i(t, r)}{\partial r} \right) + \frac{h_i(r, t)e^{-\lambda t}}{\rho c_p}, \quad (\text{A.1})$$

581 where subscript  $i$  stands for crust or cumulates. To deal with the growing thickness  
582 of the layer, we use a front-fixing method. This method consists in rescaling the  
583 radial coordinate  $r$  into a dimensionless radial coordinate  $y \in [1, 2]$ :

$$584 y = \frac{r - R^-}{R^+ - R^-} + 1, \quad (\text{A.2})$$

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7 where  $R^+$  is the radius at the top and  $R^-$  the radius at the bottom of the layer. For  
8 the crust  $R^+ = R_M$  and  $R^- = R_{\text{cr}}(t)$ , for the cumulates  $R^+ = R_{\text{cu}}(t)$  and  $R^- = R_{\text{co}}$ .  
9 We also define the dimensionless temperature  $\tilde{T}$  using:

$$\tilde{T} = \frac{T - T^+}{T^- - T^+} \in [0, 1], \quad (\text{A.3})$$

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14 where  $T^+$  and  $T^-$  the temperatures at the top and bottom of the spherical shell,  
15 with  $T^- > T^+$ . Introducing eq. (A.2) and eq. (A.3) into eq. (A.1) gives, in the case  
16 of the crust:  
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$$\frac{\partial \tilde{T}}{\partial t} = - \frac{\partial \tilde{T}}{\partial y} \frac{\partial R^-}{\partial t} \left( \frac{y-2}{R^+ - R^-} \right) + \frac{\kappa}{r^2 (R^+ - R^-(t))^2} \frac{\partial}{\partial y} \left( r^2 \frac{\partial \tilde{T}}{\partial y} \right) + \frac{h_i(y, t) e^{-\lambda t}}{\rho c_p}, \quad (\text{A.4})$$

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22 A similar equation is also obtained for the cumulate layer. The change in variable  
23 introduces an advection term related to the growth rate of the solid layer. Similarly,  
24 we follow the evolution of the radial distribution in heat production in the spherical  
25 shell using:  
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$$\frac{\partial h_i(y, t)}{\partial t} = -u(y, t) \frac{\partial h_i(y, t)}{\partial y}, \quad (\text{A.5})$$

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31 where subscript  $i$  stands for crust or cumulates,  $u_{\text{cr}} = \frac{dR^-}{dt} \frac{y-2}{R^+ - R^-(t)}$  and  $u_{\text{cu}} =$   
32  $-\frac{dR^+}{dt} \frac{y-1}{R^+(t) - R^-}$ . These equations are solved using a fully implicit time scheme  
33 and a first-order finite-volume scheme on a regular grid.  
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36 The thermal evolution code developed and used for this study is available in open  
37 source at <https://github.com/LineColin/NEMMO.git>.  
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## 40 41 **Appendix B. Analytical solutions**

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43 An analytical approximate relation between the evolution of the radius at the  
44 base of the crust and the time can be found for the second stage when considering  
45 several simplifications to the model. Considering no radiogenic heating and no heat  
46 flux from the cumulates, eq. (23) in Stage 2 becomes:  
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$$\frac{dR_{\text{cr}}}{dt} = \frac{C_E}{\rho L} k_{\text{cr}} \frac{dT}{dr}. \quad (\text{B.1})$$

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52 Considering a steady-state in the crust, the solution to the diffusion equation is:

$$T(r) = -\frac{A}{r} + B \quad (\text{B.2})$$

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604 where  $A$  and  $B$  are constant. With  $T(r = R_M) = T_s$ , where  $T_s$  is constant, and  
605  $T(r = R_{cr}) = T_E$ , we have  $A = \Delta T \frac{R_M R_{cr}}{R_M - R_{cr}}$  and  $B = \frac{A}{R_M} + T_s$  with  $\Delta T = T_s - T_E$ .  
606 The temperature gradient at the base of the crust can be expressed as:

$$\frac{dT(r)}{dr} = \Delta T \frac{R_M R_{cr}}{R_M - R_{cr}} \frac{1}{r^2} \Big|_{r=R_{cr}}, \quad (\text{B.3})$$

607 Considering the following initial condition, at  $t = 0$ ,  $R_{cr} = R_M$  and introducing  
608 eq. (B.3) into eq. (B.1) and integrating, we obtain:

$$t = -\frac{\rho L}{R_M k_{cr} \Delta T} \frac{1}{C_E} \left[ \frac{R_M R_{cr}^2(t)}{2} - \frac{R_{cr}^3(t)}{3} - \frac{R_M^3}{6} \right], \quad (\text{B.4})$$

$$\Rightarrow \tau_{sol} = -\frac{\rho L}{R_M k_{cr} \Delta T} \frac{1}{C_E} \left[ \frac{R_M R_{cr, f}^2}{2} - \frac{R_{cr, f}^3}{3} - \frac{R_M^3}{6} \right]. \quad (\text{B.5})$$

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**Declaration of interests**

- The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.**
- ~~The author is an Editorial Board Member/Editor in Chief/Associate Editor/Guest Editor for *[Journal name]* and was not involved in the editorial review or the decision to publish this article.~~
- ~~The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:~~