MAGMA OCEAN CRYSTALLIZATION AND ITS CONSEQUENCE ON THE THERMOCHEMICAL EVOLUTION OF THE MOON.

I. Bernt[1], D. Breuer[1], S. Schwinger[1] and A-C. Plesa[1]
[1]Deutsches Zentrum für Luft- und Raumfahrt (DLR), Rutherfordstraße 2, 12489 Berlin; Irene.Bernt@dlr.de

Introduction: The early Moon was covered by a global magma ocean that is assumed to have undergone fractional solidification during cooling - as evidenced by the anorthositic crust. With complete fractionation of the lunar magma ocean (LMO), the solid mantle consists of layers with different compositions, densities, and melting temperatures. After solidification, the layers mix due to solid-state convection, causing some of the mantle material to melt and rise to the surface. The melted material is the origin of the secondary crust, which consists of Mg-suite rocks (ages at 4348 ± 25 Ma [1]) and the mare basalts (ages from 4.0 to 1.2 Ga [2]) that are still found on the lunar surface. Estimates indicate that the Mg-suite represents 6-30% of the total crustal volume [3], while mare basalts constitute only 1-2% [4]. The amount and timing of secondary melt production can therefore be used to constrain the modeling of melting processes in the lunar mantle and improve our understanding of the magmatic evolution of the Moon.

The temporal evolution of crust formation has been studied in previous work by e.g [5] and [6]. The former neglect the formation of a fractionated mantle and assume a simple homogeneous mantle composition - those models typically overestimate the amount of secondary crust. [6], in turn, used the layered mantle structure of [7]. Here, the layers differ in density and concentration of heat-producing elements. This latter study [6] only considers the melting of ilmenite bearing cumulates (IBC), which crystallize below the crust during the last stage of magma ocean solidification. A more recent study by [1] investigates the origin of the Mg-suite but neglects the formation of the mare basalts. In their geodynamical model, they use a layered mantle composition and set the initial temperature profile at the solidus temperature of peridotite.

Model: In the present work we investigate the formation of the secondary crust in more detail. A petrological model provides a layered initial lunar mantle and the initial mantle temperature for the geodynamic model. The latter is consistent with the crystallization temperature from the petrological model. In the geodynamical model we consider individual melting temperatures for all cumulate layers and the latent heat consumption during melting. We also consider the change in solidus and density of the remaining mantle material as melt is extracted to form the crust. Furthermore, we track the timing and amount of melting that occurs throughout lunar evolution and compare it to constraints based on our knowledge of Mg-suite rocks and mare basalts.

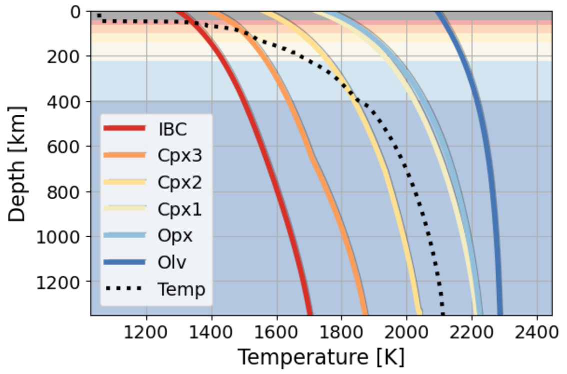

Figure 1: Solidi (solid lines) and initial temperature (dotted line) for the layered mantle structure. The initial layer positions are shown in the background. The colors of the solidi correspond to the colors of the individual cumulate layers.

Petrological Model. In our petrological model [8], we calculate a layered mantle structure assuming fractional solidification of a global magma ocean of 1350 km. As initial composition of the lunar magma ocean we consider the composition of [9] with 10 wt% FeO. Furthermore, we assume that no material is mixed during solidification. The complex stratification can be approximated by 5 main layers: the first layer is an almost pure olivine layer situated above the core-mantle boundary, the second layer above is dominated by orthopyroxene, the third layer is a cumulate mixture containing mainly clinopyroxene, the fourth layer is formed by ilmenite-bearing cumulates and KREEP material, and is situated below the fifth layer which is the anorthositic crust. Within the clinopyroxene layer the density and the melting temperatures change substantially. Therefore, we divided that layer in 3 sub-layers. For each of the compositional layers the density, the thickness, and for mantle layers also the crystallization temperature, the solidus, and the liquidus are calculated (Fig. 1). As many previous studies employed a homogeneous mantle composition, we also investigate such a mantle structure with the composition of [9] with 10%FeO as an input into our geodynamical model and later compare the results to the results obtained with a layered mantle structure. For these models we set the initial mantle temperature in the magma ocean layer to the solidus temperature corresponding to the bulk mantle composition and underneath the magma ocean layer the initial temperature follows an adiabatic profile. We investigated a warm initial temperature profile corresponding to a magma ocean depth of 350 km and a cold profile corresponding to a magma ocean depth of 75 km. These profiles have potential temperatures of 1633K and 1430 K, respectively.

Geodynamical Model. For our geodynamical model we use the mantle convection code GAIA [10] to calculate the thermochemical evolution of the lunar interior. We use an Arrhenius law to calculate the depth- and temperature-dependent viscosity, and we account for core cooling and radioactive decay. If a layered interior is considered, the geodynamic code includes the different layers as derived from the petrological model of lunar magma ocean solidification. The mixing of the layers due to solid-state convection, and thus the change in mantle composition, is tracked using a particle-in-cell method [11], where the tracer particles contain information such as composition, density, mantle depletion, and melting temperatures. We consider the influence of latent heat consumption during melting and the changes in residual materials due to secondary melting. The amount of melt is calculated using a linear interpolation between the solidus and the liquidus.

Results: The results show that both the rapid formation of the Mg-suite rocks and the mare basalts, with ages as young as 1-2 Ga ago, are difficult to achieve with the current models. Neither a homogeneous mantle composition (too much crust is produced or too late in the evolution) nor a layered mantle due to fractional crystallization (olivine cumulates as a reservoir for the Mg-suite cannot be molten and crust formation ends too rapidly) can explain the observations (Fig. 2). In the models with a layered mantle, the main reason for insufficient degrees of melting of the olivine cumulates is the difference between the crystallization temperature and the melting temperatures as a consequence of the fractional crystallization (Fig. 1). Thus, our results do not support the findings of [1] who suggest that the formation of the Mg-suite is caused by the lunar global overturn of the fractionated LMO. In their models, however, [1] assume that the crystallization temperature is equal to the solidus temperature – which is not consistent with the petrological model.

Discussion: To explain the observed volcanic history, we conclude that pure fractional crystallization of the LMO seems unlikely. We discuss possible ways to produce the Mg-suite and prolong the formation of mare basalts, including batch crystallization of a deep magma ocean, the onset of mantle convection already during solidification of the LMO, the presence of a primitive mantle layer beneath the crystallized magma ocean, but also the influence of water and melt in retention.

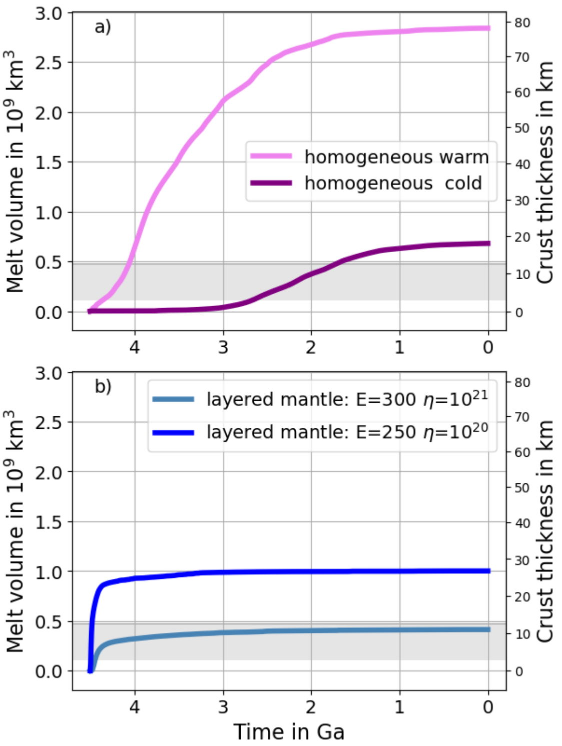

Figure 2 Cumulative amount of secondary melt over time. The left axis gives the amount in km^3 and the right axis in thickness of the secondary crust, if the volume would be distributed over the lunar surface. Panel a shows the results for a homogeneous mantle composition with different initial mantle temperatures, depending on the assumed depth of the magma ocean, which is linked to the initial temperature. Panel b for a layered initial mantle composition with different mantle viscosities (the initial mantle temperature is determined by the crystallization temperature). The legend in panel b contains information on the activation energy E given in kJ/mol and the reference viscosity η in Pa s.

References: [1] Prissel T. C. et al. (2023) *Nat Commun*, *14*, 5002. [2] Hiesinger H. et al. (2011) *Geol. Soc. Am. Spec. Pap.*, *477*, 1-51. [3] Warren P. H. and Kallemeyn G. W. (1993) *JGR Planets*, *98*, 5445-5455. [4] Broquet A. et al. (2024) *Icarus*, *411*, 115954. [5] Laneuville M. et al. (2013) *JGR Planets*, *118*, 1435-1452. [6] Zhao Y. et al. (2019) *EPSL*, *511*, 1-11. [7] Snyder G. A. et al. (1992) *Geochim. Cosmochim. Acta*, *56*, 3809-3823. [8] Schwinger S. and Breuer D. (2022) *PEPI*, *322*, 106831. [9] O'Neill H. St. C. (1991) *Geochim. Cosmochim. Acta*, *55* 1135-1157. [10] Hüttig C. et al. (2013) *PEPI*, *220*, 11-18. [11] Plesa A. C. et al. (2013) *IGI Global*.