EXPEDITED COOLING OF THE LUNAR MAGMA OCEAN DUE TO IMPACTS. V. Perera[1], A.P. Jackson[2], T.S.J. Gabriel[1], L.T. Elkins-Tanton[1], and E. Asphaug[1], [1]School of Earth and Space Exploration, Arizona State University (PO Box 876004, Tempe, AZ 85287-6004. Email: viranga@asu.edu), [2]Centre for Planetary Sciences, University of Toronto, Toronto, ON, Canada.

Introduction: The Moon likely coalesced from the debris released by a giant impact between the proto-Earth and another planetary body [1, 2]. Over years of refinement, the favored scenario had settled on a low velocity, glancing impact by a Mars-sized body [e.g. 3]. However, in this scenario the majority of lunar material would have been derived from the impacting body, which appears at odds with more recent isotopic data [e.g. 4]. A number of alternative scenarios and equilibration mechanisms [e.g. 5, 6, 7] have been proposed to reconcile the giant impact hypothesis with isotopic data, but the general mechanism of a collision between proto-Earth and a large impactor remains strongly favored [8].

Anorthosite fragments that were discovered in early geochemical analyses of Apollo 11 samples gave credence to the giant impact hypothesis. It was proposed that the anorthositic lunar crust formed by fractional crystallization of a lunar magma ocean (LMO) [9]. A LMO is consistent with the impact formation scenario, which provided a thermal state that would generate a liquid lunar interior [10]. If the magma ocean solidified through fractional crystallization, olivine would have crystalized first and started sinking to the base, while plagioclase would have started to solidify after 70–80% by volume of the LMO had fractionally crystallized. Plagioclase would have been less dense than the magma and thus would have floated to the surface forming the primordial crust of the Moon.

The time required for the magma ocean to crystallize is important since it regulates the age of the earliest lunar crust. That age may be recorded in the oldest lunar samples [e.g. 11]. So long as the surface was molten, the Moon would have solidified 80% by volume in about 1 kyr [12]. Final crystallization of the LMO would have been much slower since the floating plagioclase would have formed a thermally conductive, global lid on the Moon. Previous work found that it would have taken 10 Myrs for the last of the LMO (i.e. a 100 km thick layer of magma beneath nascent crust) to crystalize [12]. However, that time period is insufficient to explain the 200 Myr range of lunar sample ages. To explain this discrepancy, mechanisms such as tidal heating have been suggested [13, 14]; nonetheless, explaining the 200 Myr crustal age range remains an open problem.

Another factor that could have affected the cooling of the LMO is the debris produced after the Moon forming impact. Even the relatively gentle Moon-forming collision provides a substantial output of both bound and unbound debris; however, little focus is turned towards the latter in existing literature due to the computational hurdles in resolving the distribution of released debris products. Despite this, recent work that accounts for the eccentric debris disk generated by the Earth-Moon collision has shown that a substantial amount of the debris evolves along heliocentric orbits and accretes at much later times [15].

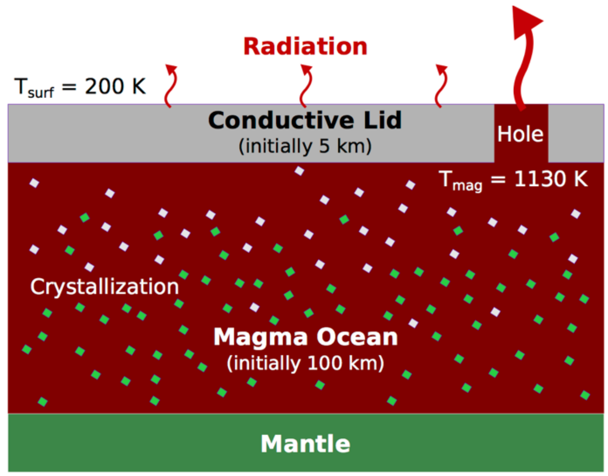

Figure 1: Schematic of the upper layers of the Moon after it had solidified 80% by volume (not to scale). Craters puncturing the thermally conductive lid would have greatly accelerated the cooling of the lunar magma ocean. Assuming a surface temperature of 200 K and a magma temperature below the lid of 1130 K [12], the luminosity of a single Tycho-sized hole is approximately 20 times that of the rest of the Moon (limited by conduction). Therefore, if re-impacting debris is capable of keeping even one such hole open, the cooling of the magma ocean will be much faster than what existing models predict.

According to simulations of the canonical model [3] around 10^{23} kg (1.3 lunar masses) of material achieves sufficient velocities to escape the Earth-Moon system and go into heliocentric orbits. This can be considered a lower limit, as other Moon-forming scenarios produce even more heliocentric debris [8]. This material has low relative velocities and high impact probabilities with the Earth-Moon system. Even at 1 Myr after the impact the Moon is accreting $>10^{13}$ kg/yr of material, for reasonable assumptions about collisional grinding within the debris distribution.

This impacting material likely affected the cooling rate of the LMO and in turn the formation of the primordial lunar crust. Early impacts could have added thermal energy (thus prolonging the cooling time) and could have mechanically stirred the magma (causing advection and affecting the chemistry of the crystals that formed [16]). Impacts with sufficient energy to puncture the lid would have sped up the cooling time by exposing the magma (see **Figure 1**). Radiative cooling is $\propto T^4$ so that a ten-fold difference in temperature is a 10,000-fold difference in luminosity, meaning that a few holes can transport considerable heat from the LMO.

Methods: Similar to earlier work [12], we start with a 100 km thick LMO and a 5 km thick crust. We calculate the energy released by cooling and crystallizing 1/1000 of the LMO mass until the LMO is solidified. We adopt a simplified scheme neglecting latent and radiogenic heating and assume that a constant fraction of the crystallizing magma goes into building crust, while the remainder builds solid mantle. For the infinitesimal energy released by crystallizing a fraction of the LMO, we iteratively calculate the total time required for that energy to be released through conduction and radiation.

For conductive cooling, the crustal thickness is updated at each iteration. For radiative cooling, we set a time span (in Myrs) that we expect impacts to puncture holes. For that time span, we assume that a constant portion of the Moon's surface has holes. N-body code results show that bombardment of the Moon by re-impacting debris should be frequent. We make a very conservative assumption that the total surface area of open holes sustained by this final sweep-up is a few multiples of Tycho crater (an 85 km diameter crater that covers about 0.02% of the lunar surface). By fixing a constant covering fraction of holes during the hole-puncture period, we assume that the rate of hole closure and the rate of creation of new holes are approximately balanced. If the time required to release the infinitesimal energy is longer than the defined time span for holes, then only conductive cooling is utilized since there are no holes on the surface for thermal radiation. At each iteration, the magma ocean temperature is updated by subtracting the energy released due to cooling and crystallizing of magma from the total thermal energy of the magma ocean. The calculations are terminated once the crustal thickness has developed to at least 40 km (about the average crustal thickness of the Moon).

Results: Since we adopted a simplified scheme neglecting latent and radiogenic heating, we obtained slightly longer cooling times for the LMO without impacts than previous work [12]. However, we found that 5 Tycho-sized holes lasting for 2 Myrs is sufficient to reduce the LMO cooling time by a factor of two and that 20 Tycho-sized holes lasting for 1 Myrs is sufficient to reduce the LMO cooling time by a factor of six. It is interesting that only a small fraction of the lunar surface with holes dramatically reduces the cooling time of the LMO. This likely has implications for the thermal evolution of the Moon.

Future work: We are in the process of including impacts into a more sophisticated magma ocean code [i.e. 12]. We will input the debris rate from N-body codes and incorporate the results of impact modeling using the iSALE code (see *LPSC 2017* abstract by Jackson et al.).

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