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The relative timing of Lunar Magma Ocean solidification and the Late Heavy Bombardment inferred from highly degraded impact basin structures

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ABSTRACT

The solidification of the Lunar Magma Ocean (LMO) and formation of impact basins are important events that took place on the early Moon. The relative timing of these events, however, is poorly constrained. The aim of this study is to constrain the formation ages of old impact basins based on inferences of their thermal state. Most proposed basins formed before Pre-Nectarian (PN) 5 stage do not exhibit clear concentric features in either topography or gravity, suggesting substantial viscous lateral flow in the crust. Recent geodetic measurements reveal that the lunar crust is thinner than previously estimated, indicating that an extremely high crustal temperature is required for lateral flow to occur. In this study, we calculate lunar thermal evolution and viscoelastic deformation of basins and investigate the thermal state at the time of basin formation using recent crustal thickness models. We find that a Moho temperature >1300–1400 K at the time of basin formation is required for substantial viscous relaxation of topography to occur; the implied elastic thickness at the time of loading is <30 km. Such a high temperature can be maintained only for a short time (i.e., <50 Myr for most conditions) after solidification of the LMO or after mantle overturn if it took place; relaxed impact basins forming $\geqslant$150 Myr later than LMO solidification are unlikely. This result is in conflict with an intensive Late Heavy Bombardment (LHB) model, which assumes that most impact basins were formed at ~3.9 Ga, since it requires LMO solidification time much later than previous theoretical estimates. Either the LHB was moderate, or the majority of proposed early PN basins were not in fact formed by impacts.

1. Introduction

The very early stage of the evolution of the Moon is thought to be characterized by solidification of the Lunar Magma Ocean (LMO) and formation of large impact basins (e.g., Warren, 1985; Wilhelms, 1987; Shearer et al., 2006). The timings of these events remain poorly constrained despite of many attempts. For example, radiometric ages of lunar pristine sample rocks are varied and can be influenced by later impact heat and shock (e.g., Nemchin et al., 2009). Numerical modeling of the thermal evolution of the LMO also predicts a wide range of solidification ages; solidification of the LMO takes ~200 Myr if a surface conductive lid develops while

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it takes only several tens of Myr if such a lid does not develop (e.g., Solomon and Longhi, 1977; Elkins-Tanton et al., 2011). Tidal heating on the early Moon may contribute to prolong the duration of the LMO for ∼200–300 Myr (e.g., Meyer et al., 2010), though the tidal heating rate depends on the orbital evolution assumed.

The ages of impact basins also have large uncertainties. Based on radiometric ages of impact melts of lunar samples, the concept of a short, intense period of impacts on the Moon at ∼3.9 Ga was proposed (e.g., Tera et al., 1974; Cohen et al., 2000). This heavy bombardment on the Moon is often called the lunar cataclysm or the Late Heavy Bombardment (LHB) and has been debated for decades since it is related to the bombardment on the early Earth and the dynamical evolution of the Solar System (e.g., Stöffler et al., 2006; Gomes et al., 2005). One end-member is an intensive LHB model which assumes that most impact basins, including degraded ones, were formed during this short period (e.g., Ryder, 2002). A LHB model with a less intensive mass flux and a broader peak in time has alternatively been proposed based on a dynamical evolution model (e.g., Bottke et al., 2012; Morbidelli et al., 2012). Another hypothesis, which can explain a peak in ages of impact melts without an increase in impact flux, has also been proposed (e.g., Hartmann, 2003); the high rate of early impacts leads to pulverization of early impact melts, and only late impact melts survive. If this is the case, the formation ages of impact basins may span a long time.

The current structure of lunar impact basins reflects the thermal history since their formation. It has long been known that nearly half the impact basins identified on the Moon exhibit clear positive free-air gravity anomalies (e.g., Müller and Sjogren, 1968). Such basins are often called "mascon" basins and are thought to have large mantle uplifts (e.g., Neumann et al., 1996). Since viscosities of silicates strongly depend on temperature (e.g., Karato, 2008), mantle uplifts can relax depending on the thermal state. Recent numerical calculations (e.g., Balcerski et al., 2010; Melosh et al., 2013; Dombard et al., 2013; Freed et al., 2014) suggest that the formation of a large mantle uplift during the impact is the standard for lunar impact basins, and the main control on relaxation of mantle uplift underneath basins is not impact heating focused on the impact site but is the regional thermal state which controls subsequent long-term viscous relaxation. Thus, highly degraded basins, which do not exhibit clear concentric features in either topography or gravity, suggest that the lunar interior was very hot when such basins formed.

The long-term thermal evolution of the Moon has also been investigated by many authors to explain various observational results, such as prolonged localized mare volcanism and a possible early core dynamo, using conductive and convective models under a wide variety of parameter conditions (e.g., Toksöz and Solomon, 1973; Stevenson et al., 1983; Konrad and Spohn, 1997; Stegman et al., 2003; Grimm, 2013; Laneuville et al., 2013). While the thermal evolution of the deep Moon depends on many factors (e.g., Evans et al., 2014), most model calculations suggest that the upper part of the Moon cooled rapidly (i.e., within several 100 Myr) unless the radiogenic heating rate was anomalously high at the base of the crust (i.e., a 10-km thick layer with >10 ppm thorium concentration) (e.g., Wieczorek and Phillips, 2000).

The goal of this study is to constrain the age of old impact basins following solidification of the LMO. If an impact basin is formed immediately after LMO solidification, the lunar interior is still very hot at the time of basin formation, and thus the initial impact structure would be highly degraded because of viscous relaxation. In contrast, if an impact basin is formed long after LMO solidification, the lunar interior is already cold at the time of basin formation, and thus the impact structure would be clearly preserved.

Viscous relaxation of impact basin topography on the Moon has also been studied for decades (e.g., Solomon et al., 1982; Mohit and Phillips, 2006; Kamata et al., 2013). Previous studies show that the observed degraded topography of an old impact basin can be reproduced well by viscous relaxation of the topography of a young impact basin, suggesting that viscous relaxation, or crustal lateral flow, is a major degradation process for lunar impact basins. The thermal structure at the time of formation of highly degraded basins, however, has not been quantitatively constrained mainly because the spatial resolution of crustal thickness models had been very low.

Recent gravity field data by Kaguya (SELENE) (e.g., Namiki et al., 2009) and by Gravity Recovery and Interior Laboratory (GRAIL) (e.g., Zuber et al., 2013) enable us to estimate crustal thickness variations with a high spatial resolution (e.g., Ishihara et al., 2009; Wieczorek et al., 2013). Crustal thickness models based on GRAIL data and seismology further suggest that the lunar crust is thinner than previous estimates (Wieczorek et al., 2013). Since the timescale for viscous relaxation strongly depends on crustal thickness (e.g., Solomon et al., 1982; Nimmo and Stevenson, 2001), further studies of viscous relaxation on the Moon using new lunar crustal thickness models are very important for constraining the thermal state of the early Moon.

In this study, we investigate viscoelastic deformation of impact basins assuming different thermal evolution scenarios and different basin formation ages using new crustal thickness models. Section 2 shows that older impact basins have higher degrees of degradation using a recent crustal thickness model. Section 3 describes numerical calculation models employed, and Section 4 presents results obtained. Section 5 discusses implications for the LHB and some model dependencies of our work.

2. Highly degraded impact basins

The relative age of impact basins can be determined by a crater counting method; an older basin has a larger number of superposed impact craters. The relative age of basins proposed by Wilhelms (1987) has been widely used and has been broadly confirmed using recent high-resolution topography data for most impact basins (Fassett et al., 2012). This classification divides lunar geologic time into five major periods, and the first three periods cover the formation ages of the lunar impact basins: Pre-Nectarian, Nectarian, Imbrian. Pre-Nectarian (PN) is defined as a period before the formation of Nectaris, and subdivided into nine stages; PN 1 is the oldest, and PN 9 is the youngest. Nectarian is defined as a period between the formations of Nectaris and Imbrium, and subdivided into two stages. Finally, Imbrian is defined as a period after the formation of Imbrium.

Table 1 lists properties of proposed impact basins >450 km in diameter. We adopt center locations and diameters of basins determined based on Lunar Reconnaissance Orbiter (LRO) data if they are available (Head et al., 2010). Otherwise, values are taken from Wilhelms (1987). The relative age reported by Wilhelms (1987) is also summarized in Table 1. From our analysis, Grissom–White and Ingenii are excluded because the effects of crustal thinning due to South Pole-Aitken (an extremely large impact basin that covers a large portion of the southern hemisphere of the lunar farside) cannot be removed from radial profiles of these basins; crustal thickness increases with horizontal distance from the center of the South Pole-Aitken basin.

Using a GRAIL crustal thickness model (34 km on average) (Wieczorek et al., 2013), azimuthally-averaged crustal thickness profiles around these basins are created. To quantify the degree of viscous relaxation, the crustal thickness ratio, which is the minimum crustal thickness (D_{min}) inside the basin divided by the surrounding crustal thickness (D_{cr}), is measured for each basin. Here D_{cr} is defined by the thickness of the crust at 2–3 times basin radius