Geoscience Frontiers 8 (2017) 299–308

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China University of Geosciences (Beijing)

Geoscience Frontiers

journal homepage: www.elsevier.com/locate/gsf

Research paper

Formation of anorthosite on the Moon through magma ocean fractional crystallization

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ARTICLE INFO

Article history:
Received 5 April 2016
Received in revised form
20 November 2016
Accepted 25 November 2016
Available online 5 December 2016

Keywords:
Moon
Anorthosite
Magma ocean
MELTS
Fractional crystallization

ABSTRACT

Lunar anorthosite is a major rock of the lunar highlands, which formed as a result of plagioclase-floatation in the lunar magma ocean (LMO). Constraints on the sufficient conditions that resulted in the formation of a thick pure anorthosite (mode of plagioclase >95 vol.%) is a key to reveal the early magmatic evolution of the terrestrial planets. To form the pure lunar anorthosite, plagioclase should have separated from the magma ocean with low crystal fraction. Crystal networks of plagioclase and mafic minerals develop when the crystal fraction in the magma (φ) is higher than ca. 40–60 vol.%, which inhibit the formation of pure anorthosite. In contrast, when φ is small, the magma ocean is highly turbulent, and plagioclase is likely to become entrained in the turbulent magma rather than separated from the melt. To determine the necessary conditions in which anorthosite forms from the LMO, this study adopted the energy criterion formulated by Solomatov. The composition of melt, temperature, and pressure when plagioclase crystallizes are constrained by using MELTS/pMELTS to calculate the density and viscosity of the melt. When plagioclase starts to crystallize, the $Mg^{\#}$ of melt becomes 0.59 at 1291 °C. The density of the melt is smaller than that of plagioclase for $P > 2.1$ kbar (ca. 50 km deep), and the critical diameter of plagioclase to separate from the melt becomes larger than the typical crystal diameter of plagioclase (1.8–3 cm). This suggests that plagioclase is likely entrained in the LMO just after the plagioclase starts to crystallize. When the $Mg^{\#}$ of melt becomes 0.54 at 1263 °C, the density of melt becomes larger than that of plagioclase even for 0 kbar. When the $Mg^{\#}$ of melt decreases down to 0.46 at 1218 °C, the critical diameter of plagioclase to separate from the melt becomes 1.5–2.5 cm, which is nearly equal to the typical plagioclase of the lunar anorthosite. This suggests that plagioclase could separate from the melt. One of the differences between the Earth and the Moon is the presence of water. If the terrestrial magma ocean was saturated with H_2O, plagioclase could not crystallize, and anorthosite could not form.

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1. Introduction

In contrast to the Earth where most of the Hadean geological record has been eliminated by tectonics, the lunar surface preserves rocks formed during the early magmatic evolution of the Moon. Understanding how the lunar rocks formed is a key to reveal the early evolution of the terrestrial planets.

The lunar highland crust is mainly composed of anorthosite or anorthositic rocks formed by the floatation of plagioclase in the lunar magma ocean (LMO) (e.g., Wood et al., 1970; Warren, 1990; Ohtake et al., 2009). It has long been debated whether anorthosite is a major rock of the lunar crust. Lunar anorthosite was first discovered from rock fragments in surface regolith sampled during the Apollo 11 mission (Wood et al., 1970), and anorthosite rocks were first found during the Apollo 15 mission (James, 1972). These and other findings suggest that anorthosite is a primary component of the crustal materials of the lunar highlands. The Multiband Imager on the Selenological and Engineering Explorer (SELENE) has enabled investigation of the composition of the lunar crust at global scale (Kato et al., 2008). Ohtake et al. (2009) reported extensive pure anorthosite rock with over 95 vol.% of plagioclase exposed by large impact events; the crustal thickness of anorthosite on the Moon has been estimated to be 50 km based on the calculation of the excavation depth of the anorthosite (Yamamoto et al., 2012).

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Peer-review under responsibility of China University of Geosciences (Beijing).

http://dx.doi.org/10.1016/j.gsf.2016.11.007
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Pure anorthosite has also been reported from lunar rock samples (Warren and Wasson, 1979a). The findings by SELENE suggest that the pure anorthosite is a primary component of the lunar highlands.

Anorthosite can form by floatation of low-density plagioclase in the magma due to the density difference between plagioclase and melt (Wood et al., 1970). The extensive occurrence of anorthosite on the Moon requires a global magma body, referred to as the lunar magma ocean (LMO) (Yamamoto et al., 2012). The "giant impact" scenario has sufficient energy to produce Moon-forming melt (Hartmann and Davis, 1975; Ida et al., 1997; Pritchard and Stevenson, 2000).

However, the actual mechanism that resulted in the formation of the lunar anorthosite is still controversial. To form the pure anorthosite, plagioclase should separate from the LMO with low crystal fraction. If the crystal fraction in magma becomes higher than 40–60 vol.%, crystal networks of plagioclase and mafic minerals develop, making it difficult to separate only plagioclase from the crystal networks (Philpotts et al., 1998, 1999). On the other hand, if the magma ocean was highly turbulent with low crystal fraction, the plagioclase would have likely become entrained in the magma rather than being separated from the melt (Martin and Nokes, 1989; Abe, 1993; Tonks and Melosh, 1993).

In order to constrain the necessary conditions that allow the plagioclase to be separated from the LMO, we have adopted the energy criterion for the onset of fractional crystallization formulated by Solomatov (2015). Sakai et al. (2014) used the same formula to constrain the composition of the initial LMO, with their results indicating that the lunar bulk composition is enriched in FeO compared with the bulk silicate Earth (BSE), which makes a large density difference between plagioclase and melt enough to separate plagioclase from the turbulent LMO. However, Sakai et al. (2014) only discussed the conditions when plagioclase first starts to crystallize. As the LMO solidifies, the melt becomes enriched in FeO, and the density of the melt possibly increases enough to allow plagioclase to float at the upper part of the LMO. In this study, we evaluated the necessary conditions that allow the plagioclase to separate from the melt by taking into account changes in the composition, the density, and the viscosity of melt.

2. Calculation method

2.1. Energy criterion for the onset of fractional crystallization by Solomatov (2015)

In order to constrain the necessary conditions that result in the formation of lunar anorthosite, a theoretical criterion is required, addressing the critical condition when minerals start to separate from the magma.

Several previous studies focused on the comparison between the downward flux of settling velocity of minerals with the upward flux of convective velocity of magma convection (Huppert and Sparks, 1981; Abe, 1993). However, the laboratory experiments on convective suspensions showed that the particles eventually settle down even when the settling velocity is much smaller than the convective velocity (Martin and Nokes, 1988, 1989). This is because the convective velocity vanishes at the boundary layer, and the particles cannot be re-entrained. This experimental observation suggests that the convective speed in the magma ocean is not the parameter scaling the efficiency of the crystal settling.

This study alternatively uses the energy criterion for the onset of fractional crystallization formulated by Solomatov (2015). A condition when fractional crystallization starts was addressed based on the balance between the total amount of energy released per unit time due to crystal settling and the mechanical work most of which is spent to overcome viscous friction associated with convection (Solomatov and Stevenson, 1993; Solomatov, 2015). The critical crystal size above which fractional crystallization starts is shown as:

$$d_f = \left(\frac{18\alpha\eta_l F}{f_\varphi g c_p \Delta\rho^2 \varphi} \right)^{1/2} \quad (1)$$

where f_φ is a hindered settling function, g is the gravity, c_p is the isobaric specific heat, $\Delta\rho$ is the difference of density between crystal and melt, φ is the volume fraction of crystals in the magma, α is the coefficient of thermal expansion of melt, η_l is the viscosity of melt, and F is the surface heat flux. The calculation results of Solomatov and Stevenson (1993) are consistent with the experimental results of Martin and Nokes (1989), though the crystal radius and the density difference between the particles and the fluid were not fully constrained in their experiment.

2.2. Hindered settling function

The hindered settling function represents the phenomenon that settling is hindered by the presence of solid particles and the return flow of the interstitial fluid (Davis and Acrivos, 1985; Huppert et al., 1991). For our calculations, we adopted the most commonly used formula by Garside and Al-Dibouni (1977) and Richardson and Zaki (1954):

$$f(\varphi) = (1-\varphi)^{5.1} \quad (2)$$

It should be noted that it is difficult to simply apply the formula to the magma because the correction also depends on the nature of interparticle forces (Davis and Acrivos, 1985).

2.3. Viscosity and density calculation

In order to constrain the viscosity and the density of the melt while the plagioclase crystallizes from the LMO, its melt composition, temperature, and pressure must be known. This study used thermodynamic equilibrium software, pMELTS for $P > 1$ GPa (Ghiorso et al., 2002) and rhyolite-MELTS for $P < 1$ GPa (Gualda et al., 2012). To calculate compositional change of the melt through the solidification of the magma ocean by using MELTS program, the crystallization model needs to be assumed.

2.3.1. Solidification model of the LMO for MELTS/pMELTS

2.3.1.1. Bulk composition of the initial LMO. To calculate the compositional change of the melt, we assumed a bulk composition of the initial LMO to be the bulk silicate Earth (BSE; McDonough and Sun, 1995).

There has been no agreement about the lunar bulk composition (Taylor, 1982; Jones and Delano, 1989; O'Neill, 1991; Snyder et al., 1992; Lognonné et al., 2003; Warren, 2005; Longhi, 2006). In order to constrain the necessary conditions that result in the separation of the plagioclase from the LMO, the initial bulk composition must be the one with the lowest estimated density. Although most previous works concluded that the composition of the initial LMO is enriched in FeO rather than BSE (Taylor, 1982; Jones and Delano, 1989; O'Neill, 1991; Snyder et al., 1992; Lognonné et al., 2003), $Mg^{\#} = 0.93$ of olivine in the Mg-suite requires $Mg^{\#} = 0.89$ for the lunar bulk composition to generate such a high $Mg^{\#}$ of olivine (Warren, 2005; Longhi, 2006). This suggests that BSE ($Mg^{\#} = 0.89$) can be considered as one of the possible lowest-density compositions of the initial LMO. This is also supported by recent giant-impact modeling on the terrestrial magma ocean, indicating that more materials from the terrestrial magma ocean with high $Mg^{\#}$ contribute to the Moon compared to the impactor (Karato, 2014).

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