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Letter

Insights into mare basalt thicknesses on the Moon from intrusive magmatism

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ABSTRACT

Magmatic intrusions preferentially spread along interfaces marked by rigidity and density contrasts. Thus the contact between a lunar mare and its substratum provides a preferential location for subsequent magmatic intrusions. Shallow intrusions that bend the overlying layer develop characteristic shapes that depend on their radius and on the overlying layer flexural wavelength and hence on their emplacement depth. We characterize the topography of seven, previously identified, candidate intrusive domes located within different lunar maria, using data from the Lunar Orbiter Laser Altimeter. Their topographic profiles compare very well with theoretical shapes from a model of magma flow below an elastic layer, supporting their interpretation as intrusive features. This comparison allows us to constrain their intrusion depths and hence the minimum mare thickness at these sites. These new estimates are in the range 400–1900 m and are generally comparable to or thicker than previous estimates, when available. The largest thickness ($\geqslant$1700 m) is obtained next to the Hortensius and Kepler areas that are proposed to be the relicts of ancient volcanic shields.

1. Introduction

Determining the thicknesses of the lunar maria is crucial to constrain the amount of melt produced in the lunar interior and the thermal evolution of the Moon. Unfortunately, they are still poorly constrained. Direct methods use radar sounding to detect interfaces within the subsurface. The lunar-orbiting Apollo 17 radar sounding experiment first detected two interfaces at apparent depths of 0.9 and 1.6 km in Mare Serenitatis and one at ~1.4 km in Mare Crisium (Peeples et al., 1978). More recently, the Lunar Radar Sounder (LRS) of the SELENE mission has detected horizontal subsurface interfaces at only a few hundred meters depth within all major lunar maria (Ono et al., 2009; Pommerol et al., 2010). The radar onboard of Chang'E 1 also detected several subsurface interfaces in the northern Mare Imbrium, the deepest one at ~360 m (Xiao et al., 2015). However, this method is not able to discriminate between the base of the mare and interfaces in between different flow units (Peeples et al., 1978; Ono et al., 2009; Xiao et al., 2015).

Regional constraints on mare basalt thicknesses have been placed by evaluating the difference between the observed topography of flooded basins and craters and their initial one, estimated by comparison with basins and craters the same size (Head, 1982; De Hon, 1979). However, the crater depth to radius relationship is associated to relatively large uncertainties, especially for large basins whose morphologies largely depend on the impactor and target properties (Milković et al., 2013). Regional constraints on mare thicknesses can also be given by studying the gravitational signal recorded above the maria. Lunar maria are, however, often coincident with lunar mascons, and a major difficulty is to extract the signal due to the mare basaltic flows themselves from the signal coming from the processes of impact basin excavation, relaxation and cooling (Melosh et al., 2013; Gong et al., 2016).

Local constraints on mare thicknesses can be given by studying the geological processes that deform them. For instance, impacts lead to craters that open – or not – a window through the mare on its substratum, providing in any case a constraint on mare thickness at the impact site (Budney and Lucey, 1998; Thomson et al., 2009). Alternatively, magmatic intrusions, intruding below a lunar mare, can create characteristic surface deformations that depend on the local mare thickness.

A dozen of low-slope lunar domes have been proposed as candidate intrusive domes, i.e. as being formed by magma spreading below an elastic layer as laccolith intrusions (Wöhler et al., 2009; Michaut, 2011). Most of them are situated within, though on the side of, different maria. The absence of spectral contrast between these domes and their surroundings and their elongated outlines

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compared to typical lunar extrusive domes were the first arguments in favor of their intrusive origin (Wöhler et al., 2009). Their morphologies (shape, slope, radius) have previously been determined by morphometric measurements using telescopic images acquired at oblique illumination (Wöhler et al., 2009). These measurements reveal low flank slopes between 0.2 and 0.6° and large diameters between 10 and 30 km compared to terrestrial laccoliths that are typically twice as small (Wöhler et al., 2009; Michaut, 2011). Using a model of magma spreading below an elastic layer, Michaut (2011) has characterized the surface deformations induced by shallow magmatic intrusions and has shown that the size discrepancy between terrestrial laccoliths and candidate intrusive domes on the Moon could be well explained by differences in gravity and magma viscosity in between lunar and terrestrial settings, supporting the intrusive interpretation.

This episode of magmatism occurred after extrusion of the maria which have been raised by the intruding magma. The magma could have been emplaced at the contact between the mare and it substratum or in between different mare units, but we suggest that the former is the most plausible scenario. By studying the deformation affecting the mare basalt layer overlying the intrusion, we place constraints on the local mare thickness. In particular, the theory of magma flow below an elastic plate predicts that the intrusion shape depends on its radius and on the overlying layer flexural wavelength, that itself mainly depends on its elastic thickness (Michaut, 2011; Michaut et al., 2013; Lister et al., 2013; Thorey and Michaut, 2014; Hewitt et al., 2015). Here, we study the topography of these low-slope lunar domes, located in different maria, using the high-resolution Lunar Orbiter Laser Altimeter data from the Lunar Reconnaissance Orbiter. The observed shapes well correspond to the theoretical shapes predicted by the model and we use these shapes to constrain the local mare thickness.

2. Materials and methods

2.1. Model summary

We consider the axisymmetric spreading of magma below an overlying layer of constant elastic thickness d and above a rigid layer. We summarize below the model, which is described in detail in several papers (Michaut, 2011; Michaut et al., 2013; Lister et al., 2013; Thorey and Michaut, 2014). The flow driving pressure P is the sum of the bending pressure due to the elastic deformation of the overlying layer and the pressure due to magma weight:

$$P(r,z,t) = \frac{Ed^3}{12(1-\nu^2)} \nabla_r^4 h(r,t) + \rho_m g (h(r,t) - z) \tag{1}$$

where r and z are radial and vertical coordinates, t is time, ρ_m magma density, $h(r,t)$ intrusion thickness, g gravity, E Young's modulus, ν Poisson's ratio, and $\nabla_r^4 h = \frac{1}{r}\frac{\partial}{\partial r}\left(r\frac{\partial}{\partial r}\left(\frac{1}{r}\frac{\partial}{\partial r}\left(r\frac{\partial h}{\partial r}\right)\right)\right)$. By applying flow momentum and mass conservation, the following equation for the evolution of the flow thickness with time and radial coordinate has been obtained (Michaut, 2011; Michaut et al., 2013; Lister et al., 2013; Thorey and Michaut, 2014):

$$\frac{\partial h}{\partial t} = \frac{Ed^3}{12(1-\nu^2)12\eta r}\frac{\partial}{\partial r}\left(rh^3\frac{\partial}{\partial r}(\nabla_r^4 h(r,t))\right) + \frac{\rho_m g}{12\eta r}\frac{\partial}{\partial r}\left(rh^3\frac{\partial h}{\partial r}\right) + w(r) \tag{2}$$

where η is magma viscosity and $w(r)$ injection velocity, i.e. the velocity given by a constant flux Poiseuille flow through a cylindrical central feeder conduit of diameter a.

Eq. (2), made dimensionless using a characteristic horizontal scale Λ, vertical scale H and time scale τ, becomes

$$\frac{\partial h}{\partial t} = \frac{1}{r}\frac{\partial}{\partial r}\left(rh^3\frac{\partial}{\partial r}(\nabla_r^4 h(r,t))\right) + \frac{1}{r}\frac{\partial}{\partial r}\left(rh^3\frac{\partial h}{\partial r}\right) + \mathcal{H}\left(\frac{\gamma}{2}-r\right)\frac{32}{\gamma^2}\left(\frac{1}{4}-\frac{r^2}{\gamma^2}\right) \tag{3}$$

where $\mathcal{H}$ is the Heaviside function, $\gamma = a/\Lambda$ and

$$\Lambda = \left(\frac{Ed^3}{12\rho_m g(1-\nu^2)}\right)^{1/4} \quad H = \left(\frac{12\eta Q_0}{\pi\rho_m g}\right)^{1/4} \quad \tau = \frac{\pi\Lambda^2 H}{Q_0} \tag{4}$$

where Q_0 is the constant volume flux through the feeder conduit and Λ the flexural wavelength of the overlying elastic layer.

This equation was resolved using a pre-wetting film of negligible thickness h_f compared to flow thickness to avoid divergent viscous stresses at the contact line between the rigid support and the elastic overlying layer (Lister et al., 2013; Hewitt et al., 2015). Numerical results show that the flow evolves following different regimes (Fig. 1) (Michaut, 2011). In the first regime, the flow is controlled by the elastic deformation of the overlying layer and develops a bell shape described by the following function:

$$h(r,t) = h_0(t)\left(1-\frac{r^2}{R(t)^2}\right)^2 \tag{5}$$

where $h_0(t)$ is the flow height at the center $r = 0$ and $R(t)$ is the flow radius. In this bending regime, gravity is negligible, the pressure is constant over the flow interior and pressure gradients are localized at the front. The maximum height h_0 evolves with the radius R following a power-law relationship with an exponent equal to 8/7 (Lister et al., 2013), i.e. very close to 1 (Fig. 1). When the flow reaches a critical size of $\sim 4\Lambda$, the current enters an intermediate regime where bending and gravity contributes almost equally to the flow. In this regime, the height decreases with time, the current flattens and the dimensional flow shape is described by (Lister et al., 2013):

$$h(r,t) = h_0(t)\left[1 - exp\left(-\frac{R-r}{\sqrt{2}\Lambda}\right)\left(\cos\left(\frac{R-r}{\sqrt{2}\Lambda}\right) + \sin\left(\frac{R-r}{\sqrt{2}\Lambda}\right)\right)\right] \tag{6}$$

Finally, when magma weight becomes dominant, the flow evolves as a gravity current; the height evolves toward a constant while the radius evolves with time following $R(t) \propto t^{1/2}$ (Huppert, 1982; Michaut, 2011; Lister et al., 2013).

2.2. Dome topography

We draw the topographic profiles of candidate intrusive domes proposed by Wöhler et al. (2009) using the 128 ppd ($\sim$236 m/pixel) LOLA gridded topography, obtained from the Planetary Data System Geosciences Node and verified our results using the 256 ppd data set. We compare their shapes to theoretical shapes derived from the model, assuming the dome shape reflects the intrusion shape. A bell-shape dome would characterize a dome that has solidified in the bending regime and its radius R would be such that $R \leq 4\Lambda$. On the contrary, a flat-top dome would have a final radius $R \gg 4\Lambda$. Finally, the shape of an intrusion in the intermediate regime would depend on Λ and hence on d following (6).

Over a total of 12 domes that have been observed by Wöhler et al. (2009, 2010) using morphometric data, we have been able to isolate and characterize the topography of 7 (Fig. 2). The other domes were mixed with other types of relief and particularly difficult to isolate. Cross-sections going through the highest point of each dome were realized in all directions. Topographic profiles