FORMATION OF MAGMA RESERVOIRS SUPPLYING VOLCANIC COMPLEXES ON THE MOON.

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Introduction: Volcanic calderas indicative of very shallow magma bodies are absent on the Moon [1], but the presence of fields of small cones and domes in areas like the Marius Hills [2] (Fig. 1) suggests the presence of substantial localized reservoirs of magma at much shallower depths [3] than the deep mantle diapirs implicated as the sources of dikes feeding the large-scale mare lava flow fields [4]. Here we explore links between deep mantle magma sources and shallower magma reservoirs.

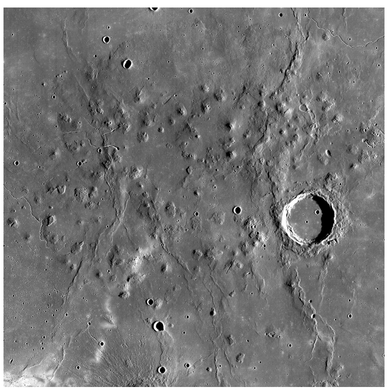

Figure 1. Cones and domes of the Marius Hills volcanic complex. Large crater Marius is 41 km diameter

Analysis: The emplacement of the largest volume mare lava flows can be understood in terms of the growth of giant dikes at the tops of diapiric bodies undergoing pressure-release melting at depths of several hundred kilometers in the mantle [4]. After growing to a critical size these dikes disconnected from their sources and migrated rapidly upward (Fig. 2a) driven by the positive buoyancy of the magma in the dikes relative to that of the host mantle. Vigorous lava eruptions occurred when these dikes were able to penetrate the crust-mantle interface and reach the surface, but to counteract the negative buoyancy of the dike magma in the low-density anorthositic lunar crust these dikes had to have vertical extents at least equal to approximately twice the ~30 km thickness of the nearside lunar crust, implying a minimum magma volume of ~100 km^3. The maximum likely volume of dikes formed in this way was ~1000 km^3 [4]. Dikes with too small a vertical extent ceased rising when the negative buoyancy of the upper part of the dike in the crust was balanced by the positive buoyancy of the lower part of the dike still in the mantle, thus forming a dike-like intrusion centered on the crust-mantle boundary (Fig. 2b).

Dikes formed in this way had another option when they reached the crust-mantle boundary, or indeed any boundary where there was a significant change in the density or elastic properties of the rocks [5], such as the base of the breccia lens beneath an impact crater or basin [6]. As such a dike overshot the boundary, it decelerated as the negative buoyancy of the magma in the crustal part of the dike increasingly cancelled the positive buoyancy of the magma in the mantle part of the dike. This continued until the difference, at the level of the crust-mantle boundary, between the the pressure in the magma in the dike and the horizontal compressive stress in the host rocks became large enough to initiate a sill-like or laccolithic intrusion (Fig. 2c). Magma from the rising lower part of the dike was then diverted into the intrusion. The upper part of the dike rapidly came to rest and may have drained downward to add to the intrusion volume coming from the lower part of the dike. The volume intruded was therefore a large fraction of the total volume of the dike. This process was able to create incipient magma reservoirs at the base of the crust with volumes in the range 100-1000 km^3.

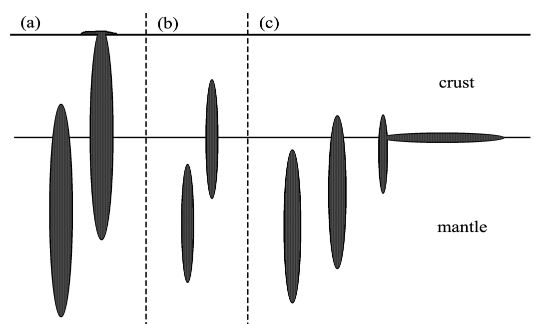

Figure 2. Scenarios for fate of giant dikes derived from deep diapiric sources in the mantle. (a) large dike penetrates completely through the crust and erupts; (b) smaller dike stalls as dike-like intrusion at the crust-mantle boundary; (c) dike interacts with crust-mantle boundary and intrudes a sill-like body at the interface.

The shape of a sill or laccolith depends on how the host rocks deform. If the response is completely elastic the maximum vertical thickness will be ~10^{-3} of the horizontal extent [7]. Table 1 shows the horizontal extents, *D*, average thicknesses, *T*, and total volumes, *V*, of elastic intrusions having total volumes in the range suggested by [4]. Also given are the times, *S*, needed for these magma reservoirs to solidify by conductive cooling, based on the solution to the Stefan Problem [8]. Values are the order of a few centuries.

If the host rocks deform inelastically, the resulting laccoliths will have a greater vertical thickness at the expense of a smaller horizontal diameter. They will take correspondingly longer times to solidify, cooling time being proportional to the square of the thickness. The subsequent thermal history of a laccolith is determined by how much time elapses before another mantle-derived dike arrives sufficiently close in space to the original one that it intercepts the laccolith and enlarges it before it has completely cooled to the local ambient temperature.

The volumes of the 360 cones and 22 domes in the Marius Hills complex identified by [9] range up to 10 km^3 with mean values of 1.55 and 1.12 km^3, respectively. In a model of the eruptive behavior of magma reservoirs, [10] showed that a basaltic magma reservoir whose host rocks behave elastically and have tensile strengths in the range 10-20 MPa and bulk moduli of ~14 GPa will require the addition of a critical volume of magma that is close to 0.1 % of the reservoir volume in order to pressurize the reservoir and trigger failure of its walls, initiating one or more dikes that allow the added magma to escape. Thus, at first sight, the ~1 km^3 mean and ~10 km^3 maximum volumes of the edifices in the Marius Hills are consistent with the triggering of eruptions from the largest of the elastic reservoir volumes in Table 1 by the arrival from the mantle of a series of dikes with volumes in the range 1-10 km^3. However, Table 1 shows that the minimum volumes of deep mantle dikes are likely to be ~100 km^3. It follows that each new addition of magma to the reservoir from the mantle in a dike of this volume would likely initiate numerous simultaneous eruptions at the surface as the roof of the reservoir failed in multiple locations. Thus, while there is no problem with explaining the presence of surface cones and domes with volumes up to ~10 km^3, there is a consequence: the total volume of the 360 cones and 22 domes in the Marius Hills complex estimated from the data in [9] is ~840 km^3. Only one additional dike with a volume close to 1000 km^3 or ~10 additions dikes with volumes of ~100 km^3, or any other permutation between these two examples yielding a total of ~1000 km^3, would be required to explain the formation of the entire complex.

Finally, the model of [10] assumes that the magma involved is less dense than the host rocks so that magma is added to the base of the reservoir and erupts upward from the roof. While lunar basalts (density ~2900-3000 kg m^{-3}) are positively buoyant in the ~3260 kg m^{-3} mantle, a reservoir at the crust-mantle boundary on the Moon will need to erupt the basalt through the low-density (~2550 kg m^{-3}) anorthositic lunar crust to reach the surface, requiring the reservoir to have an excess pressure of at least [2900 - 2550) kg m^{-3} × 1.62 m s^{-2} × crustal thickness]. For areas of nearside crust ~30 km thick this pressure is 17 MPa. A pressure of just this magnitude is needed to cause the reservoir roof to fail, so the negative buoyancy of the lunar basalts in the crust is not a bar to the eruptions taking place. Thicker crust on the lunar farside would present more of a barrier.

Conclusion: Both large volume mare lava flow fields and volcanic provinces containing hundreds of volume edifices that are 2 orders of magnitude smaller than the mare flows can be understood in terms of the same magma production mechanism deep in the lunar mantle. Concentrations of small-volume eruptions, as in the small shield volcanoes in Marius Hills and Mare Tranquillitatis [11], are predicted to be favored by locations of relatively thin crust and lithosphere that allow sill and laccolith-like intrusions at the base of the crust to be triggered by dikes from the deeper mantle into supplying multiple small volumes of magma to the surface.

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Table 1. Parameters of magma intrusions into elastic host rocks: *D*: diameter; *T*: thickness; *V*: volume of intruded magma; *S*: time to solidify intrusion.

D/km	*T*/m	*V*/km^3	*S*/years
50	50	97	107
60	60	181	155
70	70	346	211
80	80	592	275
90	90	920	349
100	100	1328	431