ORIGIN OF THE ENIGMATIC INA MOUNDS: THREE-STAGE LUNAR SHIELD VOLCANO ERUPTION SEQUENCE AND PRODUCTION AND EXTRUSION OF MAGMATIC FOAM.
L. Wilson[1,2] and J. W. Head,[1] [1]Brown University, Providence RI02912, USA, [2]Lancaster University, Lancaster LA1 4YQ, UK

Introduction: The absence of atmosphere on the Moon means that all magmas approaching the surface attempt to release all volatile species that they contain in solution or generate by chemical reactions at low pressures. Graphite reacts with various metal oxides to produce CO gas as magmas ascend from depths of ~500-600 km [1-7]. Some CO was already present in solution at greater depths [8-10], as were significant dissolved S and H_2O [10]. This gas production ensures that essentially all lunar eruptions begin with an explosive phase. The initial stage of the eruption is fed by a wide dike that is likely to extend completely through the lunar crust into the upper mantle [11] ensuring a high magma discharge rate. The discharge rate, i.e. the volume flux, is the product of the cross sectional area of the dike and the magma rise speed within it. As the initial excess pressure in the dike is lost and the dike begins to close due to the elastic response of the crust, the discharge rate decreases, becoming very small before cooling ends the eruption. As the dike closes, wall friction becomes ever more important and the magma rise speed decreases. Initially, the magma rise speed is so great that gas bubbles nucleating in the magma will have buoyancy-driven rise speeds through the magmatic liquid that are many orders of magnitude less than the rise speed of the magma itself through the dike. Hence a uniform distribution of gas bubbles exists in the magma as it reaches the surface, and the expansion of these bubbles into the lunar vacuum causes the magma to fragment into sub-mm-sized droplets that emerge in a nearly steady Hawaiian-style eruption [11]. However, as the magma speed decreases, the difference between the magma speed and the bubble speeds becomes less, and the greater transit time allows large bubbles to overtake small ones and coalesce with them. At a low enough magma rise speed the process reaches a run-away state, with Strombolian explosions as occasional very large gas bodies emerge though a pond in the vent and burst as they expand into the vacuum. As the magma rise speed at depth approaches zero, no new gas is released, and existing gas bubbles drift upward. In the shallowest several hundred meters, H_2O present in amounts close to 1000 ppm becomes the dominant volatile [12, 13] with gas bubble sizes small enough that surface tension allows them to remain stable against the internal gas pressures and forming a foam with a vesicularity up to ~95%. This is the last material to be extruded. We now discuss each eruptive phase in turn.

Hawaiian explosive phase: Steady Hawaiian fire fountains under lunar conditions had the same morphology as the umbrella-shaped plumes on Io [14] but with a much greater opacity [11]. Total released gas contents in lunar eruptions were up to a few thousand ppm producing maximum pyroclast ranges of 1-10 km, whereas those on Io involve up to 30 mass % volatiles [14] producing ranges up to 300 km. The much greater crowding of sub-mm sized pyroclasts in small lunar fire fountains makes them optically dense; heat cannot escape from the interior and clasts land at magmatic temperatures to coalesce into a lava pond feeding lava flows. An outer shell of cooled pyroclasts can form a cinder/spatter cone around the central lava pond. Some small shield volcanoes on the Moon may be built largely from pyroclastics in this way if the magma volume eruption rate is relatively small and the volatile content relatively large. But if the volume eruption rate is large and the volatile content is small, essentially all pyroclasts degas efficiently as they fall into the pond; pond overflow feeds vesicle-free lava flows that reach cooling-limited lengths of at most a few tens of km, the length being determined by the volume flux [15].

Strombolian explosive phase: As the volume flux declines towards the end of an eruption, the magma rise speed decreases and it becomes ever more likely that gas bubbles will have time to coalesce as they ascend through the rising magma, leading to Strombolian activity. Large bubbles updome the lava pond surface; as they burst, bubble skin fragments are ejected as pyroclasts. Loss of magma from the vent to form these pyroclasts is compensated by the slow rise of magma through the dike and near-surface conduit, and gas bubbles continue to nucleate in the rising magma as it nears the surface. Generalization of the model developed by [16] for terrestrial Strombolian activity to lunar conditions suggests that bubble coalescence will be efficient if magma rise speeds are less than ~5 m s^{-1}, easily satisfied in the late stages of lunar eruptions.

Foam extrusion stage: The final stage of an eruption occurs when the magma rise speed becomes vanishingly small. No further gas is released at depth and a final large bubble emerges at the surface in a final Strombolian explosion. The dike relaxes elastically as residual magma is squeezed slowly towards the surface. Although some overflow of lava from the lava pond in the vent may continue for a while, the surface of the pond becomes stable and a solid crust thickens with time. Bubbles of vapor, mainly H_2O, that have nucleated in the magma beneath the crust expand as a result of decompression as they rise slowly. They also grow by continued diffusion of molecules of H_2O and sulfur species through the magmatic liquid, and larger bubbles grow at the expense of smaller ones with which they are in contact by Ostwald ripening. Cooling of the crust generates ~1% horizontal contraction resulting in fractures up to ~1 meter wide through which magmatic foam can be extruded onto the surface where cooling stabilizes it into lava with an extremely vesicular texture similar to that of reticulite.

Bubbles at the magma surface may explode in a mini-strombolian fashion because they have gas at a finite pressure on one side of the liquid film forming the bubble wall but a hard vacuum on the other side. Disruption is resisted by the surface tension of the magma, σ, ~0.37 J m^{-2}. Bubbles with radius R avoid rupture if the gas pressure inside them is less than a critical value [17], $P_c = [(4\ \sigma) / R]$. The internal pressure and bubble size can be found by assuming the vesicularity of the magma. In experiments exposing melted samples of terrestrial basalts to progressively lower pressures, [18] produced vesicularities of up to 94%, comparable to the 96% vesicularities in reticulite clasts from basaltic eruptions [17]. Bubble sizes in the experimental foams ranged from 100-1000 microns, with some bubbles bursting as the pressure decreased below ~3 kPa. These findings agree well with the above equation and suggest using a vesicularity of 95%. The solubility of water as a function of pressure in mafic magma given by [19] can be used to find the mass fraction exsolved, n_{H2Oe}, at any given pressure given an assumed total amount present pre-eruption, and an assumed magma temperature allows the density of this vapor to be found. The vesicularity is then calculated from the mass fractions and densities of the H_2O vapor and the magmatic liquid, and the pressure, P_f, is varied to reach the required 96% vesicularity. The water vapor bubble radii are found by assuming that they nucleate at the saturation pressure relevant to the assumed total water, n_{H2Ot}, with a radius of 10 microns [20] and expand isothermally to a radius R at the pressure P_f. The value of P_c corresponding to this radius is then found. As long as P_c is greater than P_f the foam is stable. If the reverse is true, bubbles at the top of the foam burst and the bursting process propagates down into the foam producing a loosely-packed layer of bubble-wall magma fragments with a density ρ about half that of the magma, say 1500 kg m^{-3}. The underlying foam becomes stable when the depth of the fragmental layer reaches a value $D = (P_f - P_c)/ (\rho\ g)$ where g is the acceleration due to gravity. The table below shows examples of the relevant parameters.

Conclusion: Late-stage foam development should be characteristic of dikes and conduits beneath summit pit craters on small shield volcanoes [21, 22]. Extrusion and modification of magmatic foams may provide an explanation for some of the unusual textures and surface features observed in these environments. The young ages of volcanic mounds at Ina and elsewhere inferred from impact crater retention [23] are probably due to the anomalous response of the magmatic foam to the impact process, and a re-appraisal leads to ages comparable to those of the main period of lunar mare volcanism [24].

References: [1] Sato, M. (1976) *Proc. Lunar Sci. Conf. 7*, 1323–1344. [2] Sato, M. (1979) *Proc. Lunar Sci. Conf. 10*, 311–325. [3] Fogel, R. and Rutherford, M. (1995) *GCA*, *59*, 201–215. [4] Nicholis, M.G. and Rutherford, M.J. (2006) *LPS XXXVII*, Abstract #2061. [5] Nicholis, M. G. and Rutherford, M. J. (2009) *GCA*, *73*, 5905–5917. [6] Rutherford, M.J. and Papale, P. (2009) *Geology*, *37*, 219–222. [7] Wetzel, D.T. et al. (2015) *Nature Geosci*, *8*, 755–758. [8] Wetzel, D.T. et al. (2013) *Proc. Nat. Acad. Sci.*, *110*, 8010-8013. [9] Yoshioka, T. et al. (2015) *Amer. Min*., *100*, 1641-1644. [10] Hauri, E.H. et al. (2011) *Science*, *333*, 213-215. [11] Wilson, L. and Head, J. (2016) *Icarus*, *283*,146-175. [12] Saal, A. E. et al. (2008) *Nature*, *454*, 192–195. [13] Hauri, E. H. et al. (2015) *Earth Planet. Sci. Lett.*, *409*, 252–264. [14] Kieffer, S. W. (1982) pp. 647–723 in *Satellites of Jupiter*. [15] Pinkerton, H. and Wilson, L. (1994) *Bull. Volcanol*., *56*, 108-120. [16] Parfitt, E. A. and Wilson, L. (1995) *Geophys. J. Internat*., *121*, 226-232. [17] Mangan, M. T. and Cashman, K. V. (1996) *J. Volcanol. Geotherm. Res*., *73*, 1–18. [18] Fielder, G. et al. (1967) *Planet. Space Sci*., *15*, 1653-1666. [19] Dixon, J. E. (1997) *Amer. Min*., *82*, 368–378. [20] Sparks, R. S. J. (1978) *J. Volcanol. Geotherm. Res.*, *3*, 1–37. [21] Head, J. and Gifford, A. (1980) *Moon and Planets*, *22*, 235-258. [22] Head, J. and Wilson, L. (2016) *Icarus*, *283*, 176-223. [23] Braden, S. et al. (2014) *Nature Geosc*i., 7, 787-791. [24] Qiao, L. et al. (2017) *Geology*, in press.

Table: Water vapor pressure, P_f, in bubbles near surface of 95% vesicular magmatic foam as a function of total pre-eruption magma H_2O mass fraction, n_{H2Ot}. Mass fraction of water exsolved into bubbles is n_{H2Oe}, typical bubble radius is R, and capillary pressure preventing foam rupture is P_c. If $P_f > P_c$, e.g. if n_{H2Ot} = 1000 ppm, a thickness D of accumulated pyroclastic debris is needed to stabilize the foam.

n_{H2Ot} /ppm	P_f /kPa	n_{H2Oe} /ppm	R /m	P_c /kPa	D /m
1000	85.4	807	21.9	67.5	7.36
500	40.8	385	20.1	73.4	no debris needed
250	19.2	182	18.6	79.6	no debris needed