A MODEL FOR THE ASCENT AND ERUPTION OF LUNAR PICRITIC MAGMA BASED ON EXPERIMENTS AND LUNAR SAMPLE DATA.

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Introduction: The purpose of this paper is to integrate new data on C-O-H-S volatile speciation and solubility in picritic magmas into a revised eruption and dike emplacement model. Models for the eruption of lunar picritic magmas appear to require a column of melt extending deep into the lunar mantle, and a compliment of dissolved volatiles to form the observed glass bead deposits [1,2,3,4]. However, recent experimental studies [5, 6] indicate that gas formation is more complex than indicated in previous studies [4]. Additional data come from analyses of olivine-hosted melt inclusion (M.I.) glasses in the A17 orange glass melt inclusions [7] plotted in Fig. 1.

The Data: The melt inclusions trapped in olivine microphenocrysts of the A17 orange glass (74220) contain dissolved C and H ranging from 4 ppm C and 1400 ppm H_2O [6,7] down to 0.4 ppm C and 800 ppm H_2O (Fig. 1). In the inclusions, S decreases from 650 to 550 ppm as C and H decrease. Experiments on C solubility indicate that the olivine crystals trapped these inclusion melts as the orange glass magma ascended from 50 km depth to ~0.5 km [6]. Very little Cl (0.3 ppm), H (10 ppm), F (10 ppm) and S (100 ppm) was lost during this ascent. However, at ~0.5 km depth, significant losses of H (80 ppm), S (200 ppm), and F (25 ppm) began while the C content went from ~0.5 to 0.4 ppm. The orange glass beads deposited on the lunar surface that contain an average of 0.45 ppm C, 10-30 ppm H_2O, and 250 ppm S, represent the end stage of this two-stage degassing process. The gas lost in stage 1 of the ascent is calculated to be ~13 vol percent of the orange glass magma at 0.5 km; the additional loss of 800 ppm H_2O and 200 ppm S in the final degassing (<0.5 km to quench) produced a magma with ~93 vol% gas bubbles at the vent. Although this gas volume is a mimum estimate, it is more than sufficient to cause fragmentation (bead formation) prior to the final decompression at the surface vent. Gas saturation may have occurred initially at depths greater than the 50 km
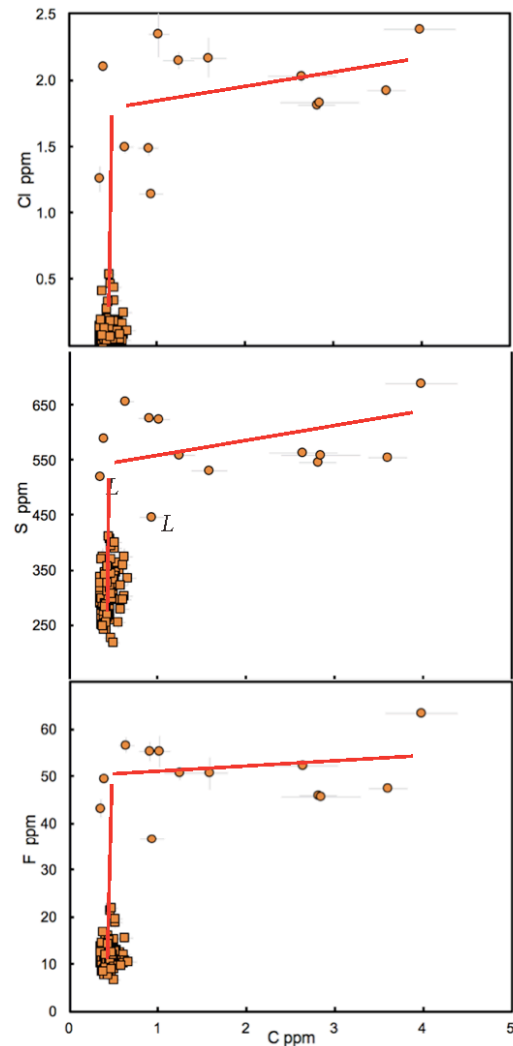

Fig. 1. S, Cl and F vs. C data for the natural picritic lunar glass beads and melt inclusions [7] in 74220 orange glass. Squares are glass beads; the circles are M.I. data. The decrease in dissolved C in the M.I.'s (Horizontal line) marks a decrease in trapping pressure from 50 km to 0.5 km based on S solubility [6]. The two M.I.'s with lowest (*L*) Cl, S and F show evidence of late loss of these elements just prior to M.I. closure. Vertical red line shows volatile loss in last satge of eruption. Error bars are 2 sigma analytic uncertainties.

depth of olivine phenocryst growth and M.I en-trappment.

Ascent and Eruption Model: Previous models for eruption of picritic lunar glass bead deposits [2, 4], used a conduit flow model with a steady state flow in an established conduit. The above gas volume would be sufficient to create an explosive eruption in these models. However, since the orange glass magma separated from its upper mantle source at 550 km depth in the Moon [9] and ascended to the surface with no evidence of reaction or lava flow eruption phase, a rapid dike emplacement also seems required.

Following [8], a partially-melted mantle diapir rising to a depth of 550 km [9] would produce a sufficiently large upward-directed buoyancy force and strain rate on overlying mantle to initiate a dike that would rapidly propagate upward through the mantle. If the melt-connected zone in the diapir extended 25 km vertically [8], the resulting 112 m wide dike would rapidly extend to 75 km above the diapir (450 km depth) where it would probably separate (Fig. 2), but the upward rise would continue. The rapid rise of the dike (~30 m/s) is driven by repeated pressure build-up and cracking at the dike tip, a process that eventually also contributes to gas formation at the dike tip. The M.I. data indicate that the initial CO-rich gas phase would have formed at 50 km depth, but it could have formed even deeper considering the transient low pressures at the dike tip [8], and the possibility of gas phase separation.

The combination of the rapid-dike emplacement model and the associated build up of the gas volume in the upper part of the dike (to ~ 93 vol% assuming closed-system degassing) helps to explain the unique characteristics of the A17 orange glass eruption. Specifically, it explains the explosive nature of the eruption, the wide area of bead distribution, and the apparent lack of similar magma as lava flows. A small amount [10] of olivine phenocryst formation (<1 vol%) with melt entrapment occurred during magma transit through the lunar crust, an ascent accompanied by steady generation of an excess CO-rich gas phase. Fe-metal grains trapped in the olivine phenocrysts require that a reduction reaction occurred in the magma during this phase of the ascent. S partitioning into this gas appears to be the best explanation for this reduction [11]. Once the magma moved above ~0.5 km depth, the rapid loss of the

remaining 800 ppm H_2O, produced an oxidation of the metal beads not included in olivine phencrysts, and a conversion of ferrous to ferric iron in the orange glass [12}.
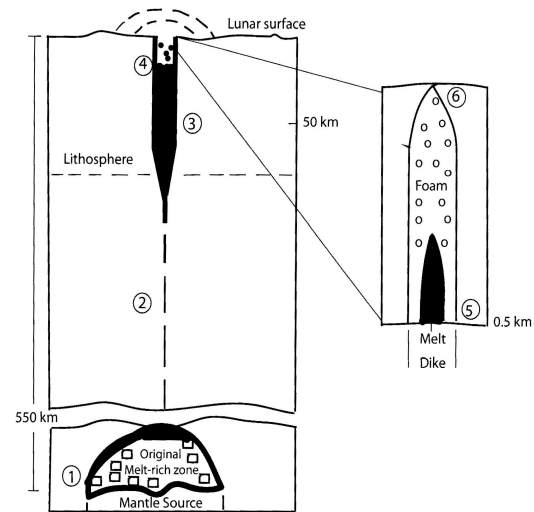

Fig. 2. Orange glass dike emplacement and eruption model. (1) Partial melt last equilibrates with residual mantle diapir at ~550 km, and dike propagation begins at contact of melt-rich zone with overlying mantle, (2) dike reaches ~78 km vertical for a 25 km (vertical) melt zone, then separates from the source and begins to move more rapidly upward, (3) Minimum depth of CO gas saturation ~50 km. Continuous degassing of CO-rich gas with ascent; gas and foam collect at tip, (4) ~13 vol% bubbles minimum in magma at 0.5 km, (5) Stage 3 degassing (H_2O and SO_2, COS and H_2S) begins at 0.5 km depth and initiates fragmentation (bead formation). (6) Dike tip breaches surface releasing melt beads in hot suspension of >93 vol% gas.

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