LUNAR MARE BASALT VOLCANISM: NEW CONSTRAINTS ON MAGMA ASCENT AND ERUPTION.

L. Wilson[1] and J. W. Head[2], [1]Lancaster Environment Centre, Lancaster University, Lancaster LA1 4YQ, U.K. (l.wilson@lancaster.ac.uk), [2]Dept. Geological Sciences, Brown University, Providence RI 02912, U.S.A. (James_Head@Brown.edu).

Introduction: Ascent and eruption of lunar mare basalt magmas was significantly influenced by crustal and lithospheric structure. GRAIL has recently measured the broad spatial variation of the bulk density structure of the crust of the Moon [1], finding a mean crustal density of ρ_c = 2550 kg m^{-3}. The liquidus densities of mare basalts and lunar picritic magmas span the ranges 2775-3025 and 2825-3150 kg m^{-3}, respectively [2, 3], making the average magma density ρ_m = ~2950 ± 200 kg m^{-3}. Thus essentially all lunar magmas were negatively buoyant everywhere within the lunar crust.

Implications: Positive excess pressures must have been present in melts at or below the crust-mantle interface to enable them to erupt. The source of such excess pressures is clear: melt in any region experiencing partial melting, or in any region containing accumulated melt, behaves as though an excess pressure is present at the top of the melt column if (a) the melt is positively buoyant relative to the host rocks and (b) the melt forms a continuously interconnected network. The latter means that, in partial melt regions, probably at least a few percent melting must have taken place.

Magma sources: Petrologic evidence suggests that both mare basalts and picritic glasses may have been derived from polybaric melting of source rocks in regions extending vertically for at least a few tens of km [3]. This is not surprising: the vertical extent of a region containing inter-connected partial melt produced by pressure-release melting is inversely proportional to the acceleration due to gravity [4]. Translating the ~50 km vertical extent of melting in a rising mantle diapir on Earth [5] to the Moon then implies that melting could have taken place over a vertical extent of up to 300 km. If convection were absent, melting could have occurred throughout any region in which heat from radioisotope decay was accumulating; in the extreme this could have been most of the mantle.

Melt pressures: The maximum excess pressure that can be reached in a magma body depends on its environment. If melt percolates upward from a partial melt zone and accumulates as a magma reservoir, either at the density trap at the base of the crust or at the rheological trap at the base of the elastic lithosphere, the excess pressure at the top of the magma body will exert an elastic stress on the overlying rocks. This will eventually cause them to fail in tension when the excess pressure has risen to close to twice the tensile strength of the host rocks [6], perhaps up to ~10 MPa, allowing a dike to propagate upward from this point.

However, if partial melting occurs in a large region deep in the mantle, connections between melt pockets and veins may not occur until a finite amount, probably a few %, of melting has occurred. When interconnection does occur, the excess pressure at the top of the partial melt zone will rise abruptly to a high value, again initiating a brittle fracture, i.e. a dike. That excess pressure is the product of the vertical extent of the melt zone, the density difference between host rocks and melt, and the acceleration due to gravity. Thus if the vertical extent of the melt zone is the 300 km noted above, the excess pressure due to a typical 360 kg m^{-3} density difference between magma and host mantle would rise abruptly to (300 km × 360 kg m^{-3} × 1.62 m s^{-2} =) ~174 MPa, vastly greater than the value needed to initiate a dike. We therefore explore excess source pressures over the range of 10 to at least 100 MPa.

Eruption requirements: If eruptions take place through dikes extending upward from the base of a crust of thickness C, the mantle magma pressure at the point where the dike is initiated must exceed the pressure due to the weight of the magma column, $P_m = (\rho_m g C)$, where the acceleration due to gravity is g = 1.62 m s^{-2}. The pressure due to the weight of the crust is $P_c = (\rho_c g C)$ and so the excess pressure must be at least $[(\rho_m - \rho_c) g C]$. Using the average magma density of ρ_m = ~2950 ± 200 kg m^{-3} and ρ_c = 2550 kg m^{-3}, $(\rho_m - \rho_c)$ is 400 ± 200 kg m^{-3}. Thus on the nearside, with C = ~30 km, the excess pressure, P_e, must be at least ~19 ± 9 MPa, and on the farside the corresponding range of minimum excess pressures is ~29 ± 15 MPa. If the top of the magma body feeding an erupting dike is a little way below the base of the crust, slightly smaller excess pressures are needed because the magma is positively buoyant in the part of the dike within the upper mantle.

Even the smallest of these excess pressures is probably greater than the ~10 MPa likely maximum value in a magma reservoir at the base of the crust or elastic lithosphere, but the values are easily met by the excess pressures in extensive partial melt zones deeper within the mantle. Thus magma accumulations at the base of the crust would have been able to intrude dikes part-way through the crust, but not able to feed eruptions to the surface; in order to be erupted, magma must have been extracted from deeper mantle sources, which seems consistent with petrologic evidence.

We saw earlier that deep mantle sources could have provided excess pressures of up to 100 MPa. If only 30 MPa were available at the base of the 30 km nearside crust, where the ambient lithostatic pressure is P_c = 123.9 MPa, then the pressure at the base of the magma column would be 153.9 MPa. The weight of the 30 km column of magma corresponds to a pressure of P_m = 143.4 MPa, and so the difference, 10.5 MPa, would have been available to drive magma flow through the 30 km long dike, producing a pressure gradient of 352 Pa m^{-1}. Table 1 gives pressure gradients, dP/dz, driving magma flow for a range of excess magma pressures.

Dike widths: The formulae of [7] approximate the width of a dike penetrating completely through the crust. The dike extends from the magma neutral buoyancy depth both upward to the surface and downward into the upper mantle in the way described for giant dike swarms by [8]. The model of [7] provides the mean thickness of the dike and the stress intensities at the upper and lower tips. The stress intensity must exceed the fracture toughness of the host rocks for dike growth; values between the growth threshold and zero correspond to dike stability and a negative value leads to dike closure. Table 1 shows values of mean dike thickness, W, for dikes extending from the base of the crust for a range of plausible excess magma pressures; all of these dikes would be stable.

Magma rise speed: Given the pressure gradient driving magma flow and the mean dike width, the magma rise speed can be found. It is necessary to evaluate the speed using the formulae for both laminar and turbulent flow; whichever gives the smaller speed is correct [9]. In all of the present cases magma motion is turbulent for any plausible lunar magma viscosity and the rise speed $U = [(W\, dP/dz)/(f\, \rho_m)]^{1/2}$, where the friction factor f is conservatively taken as 0.03; the speeds shown in Table 1 are of order 10 m s^{-1}.

Magma volume fluxes: The lava volume eruption rates, V, corresponding to the above magma rise speeds when dikes open to the surface involve the horizontal lengths of active fissures. Observable lunar features interpreted as fissure vents (e.g. elongate sinuous rille source depressions) are rarely longer than ~10 km [10, 11]. Longer fissures may have existed and been drowned by fluid lavas, but we use 10 km as a conservative example. The final column of Table 1 shows implied volume fluxes, F_{10}, of order 10^5 to 10^6 m^3 s^{-1}.

Discussion: Methods of deriving volume fluxes in lunar eruptions from lava flow thicknesses and surface slopes or rille lengths and depths were summarized by [12-14]. They found volume fluxes of 10^5 to 10^6 m^3 s^{-1} for volume-limited lava flows and >10^4 to 10^5 m^3 s^{-1} for sinuous rilles, with dikes widths of ~50 m. These lava flow values are entirely consistent with the results in Table 1. However, the lower end of the volume flux range for sinuous rilles corresponds to magma rise speeds approaching the limit set by the fact that excessive cooling would occur during flow up a 30 km long dike kept open by a very low excess pressure. These eruptions were probably fed by partial melt zones deep in the mantle. The pressures holding wide dikes open would then be similar to those implied by Table 1. However, the volume flux of magma would be limited not by friction during flow in the dike, the criterion assumed here, but instead by the rate at which melt could be fed into the base of the dike by percolation through the partial melt zone itself.

Conclusions: (1) Essentially all lunar magmas were negatively buoyant everywhere within the crust. (2) Positive excess pressures of at least 20-30 MPa must have been present in mantle melts at or below the crust-mantle interface to drive magmas to the surface. (3) Such pressures are easily produced in zones of partial melting by pressure-release during mantle convection or simple heat accumulation from radioisotopes. (4) Magma volume fluxes available from melts accumulating at the tops of partial melt zones are consistent with the 10^5 to 10^6 m^3 s^{-1} volume fluxes implied by earlier analyses of surface flows. (5) Eruptions producing thermally-eroded sinuous rille channels involved somewhat smaller volume fluxes of magma where the supply rate was limited by the rate of extraction of melt percolating through partial melt zones.

Table 1: See text for definitions.

P_e /MPa	W /m	dP/dz /(Pa/m)	U /(m/s)	F_{10} /(m^3/s)
20	24	19	2.3	5.46×10^5
25	33	185	8.4	2.82×10^6
30	43	352	13.1	5.63×10^6
40	62	685	21.9	1.35×10^7

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