LUNAR VOLCANISM: FACTORS CONTROLLING INTRUSION GEOMETRIES AND ERUPTION CONDITIONS. L. Wilson[1,2] and J.W. Head[2], [1]Environmental Science Div., Lancaster Univ., Lancaster LA1 4YQ U.K. (l.wilson@lancaster.ac.uk), [2]Dept. of Geological Sciences, Brown Univ., Providence, RI 02912 U.S.A.

Introduction: Partial melting occurs in planetary bodies due to (1) temperature rise as heat is produced from radioisotope decay; or decompression as parts of the mantle rise due to (2) thermal convection or (3) overturn driven by compositional instability. Case (1) dominated early-forming differentiated asteroids heated by short half-life isotopes [1, 2] but is not relevant to the Moon. There is still debate as to the relative importance of the latter two processes as a function of depth in the Moon and geologic time [3], but both cause progressive accumulation of melt with time.

Melt migration occurs by slow percolation along grain boundaries, but magma geochemistry on Earth suggests [4] that melt migrates more quickly in fracture networks [2] in the upper mantle [5] when the strain rate applied to the host rocks overlying a zone of partial melting drives the rocks from a plastic to an elastic rheological response [6]. Once a fracture is initiated, melt drains into it to form a dike. If the melt viscosity is small enough, melt flow from the source region into the dike may be fast enough to cause the fast-moving dike tip to continue to fracture the host rocks in a brittle fashion. We used this concept to model giant dike swarms inferred to underly graben radiating from Tharsis, Mars [6] and now apply the same principles to the Moon.

Intruded dikes: Figure 1 shows a cross-section of the geometry of a partial melt zone in the mantle feeding a dike propagating along a neutral buoyancy level (NBL) at depth D defined by the density change at the base of the crust. The dike driving pressure P_d is provided by the buoyancy [$g\ H\ \Delta\rho$] of the column of melt extending to the base of the source region at depth H below the NBL (g is the acceleration due to gravity and $\Delta\rho$ is the difference between mantle and magma densities). The dike grows both up into the crust and down into the upper mantle. The NBL does not have to be the base of the crust: Wieczorek et al. [7] have shown that some lunar basalts are less dense than the lower crust and so may be neutrally buoyant within the crust. With the appropriate choice of D the same principles apply.

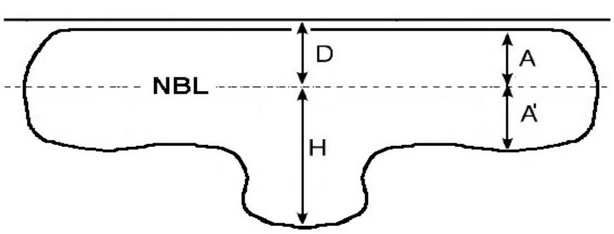

Figure 1. Geometry of dike intrusion.

A key issue is the fact that the magma spreading along the NBL is subject to the excess pressure P_d. Previous models of the spread of NBL dikes in the Moon [3] used the analytical model of [8] that assumes a power law dependence of source volume flux on time and does not include a driving pressure. The unjustified neglect of the driving pressure is remedied here. We do not model the dynamic emplacement of a dike with the geometry of Fig. 1 because once emplacement is complete an elaboration of a static model [9] developed for shallow rift-zone dikes on Earth can be employed.

The upper tip of a dike may fail to reach the surface of a planet for either of two reasons. First, the total buoyancy of the magma relative to its host rocks (positive at depth, negative near the surface [10]) may be insufficient to support a column of magma from the source region to the surface. Second, the stress intensity at the upper tip of the dike, obtained from the distribution of stress across the dike walls [11], may not be great enough to fracture the host rocks. We keep track of both criteria in our dike geometry computer model.

Figs. 2 and 3 show the mean dike widths and the depths to the dike tops expected for a range of magma densities as a function of the depth of the base of the partial melt zone in the mantle. The results assume no regional extension or compression. By comparing the geometries of lunar graben with and without associated volcanic features, Petrycki et al. [12] found mean intruded dike widths of ~50-100 m and a mean depth to dike top of ~400 m, entirely consistent with Figs. 2 and 3 for magma densities near 3000 kg m^{-3} and melt zones extending down to ~70 km. The equivalents of Figs. 2 and 3 can be constructed for any chosen regional lithospheric compressive or tensile stress; tensile stresses lead to greater dike widths and top depths, compressive stresses to smaller values.

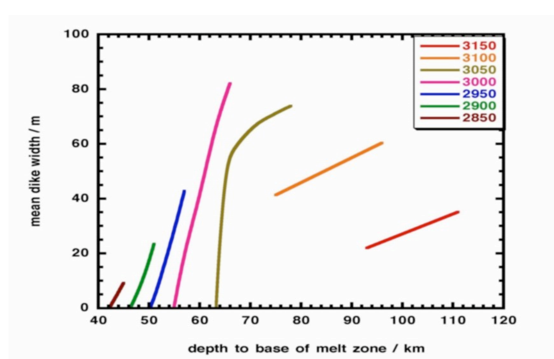

Figure 2. Predicted intruded dike widths.

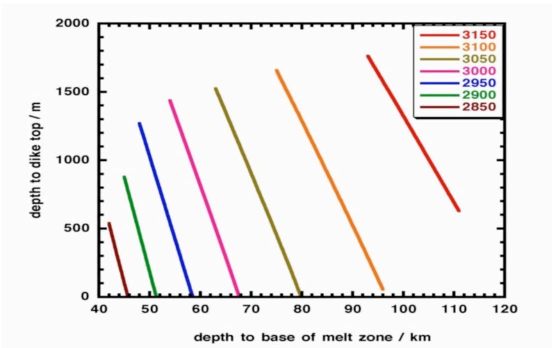

Figure 3. Predicted depths to dike tops.

Dikes feeding eruptions: There is a fundamental difference between the spatially and temporally varying conditions in the magma in a dike while the dike is in the process of establishing its pathway and the tip is ascending from depth [13, 14] and the spatially varying but nearly time independent conditions in an established dike feeding a surface eruption [15]. Current models of magma ascent through planetary lithospheres assume the magma is uniformly buoyant at all depths [8, 16]; no models exist for the geologically common situation of positively buoyant magma ascending through the mantle and invading, or penetrating, the lithosphere where it is negatively buoyant. Our early attempts to model lunar dike propagation [10] were criticized [3] for being based on models of static dikes, but proposed alternatives based on [8] are also incomplete because they neglect negative magma buoyancy and excess magma pressure at shallow depths.

We used the most recent data on lunar basalt volatile contents (250-750 ppm CO, up to 350 ppm S and ~50 ppm H_2O [17, 18]) to compute the pressure at which lunar magmas fragment into pyroclasts and then found the pressure in the vent in the resulting choked flow, P_{ch}. This pressure holds open the surface fissure to a horizontal length L with a maximum width W which can be obtained as a function of the effective fracture toughness K_s ~100 MPa m$^{1/2}$, of the near-surface rocks using relationships summarized in [19]. Using the pressure gradient available to drive magma flow when a dike tip breaks the surface, ~100 Pa/m [20] based on the most recent estimates of lunar crustal structure [3, 7, 21], the magma rise speed U and the total magma volume flux from the fissure V are found.

Table 1 shows the results as a function of the total magma volatile mass fraction n. For magma volatile contents in the range 750-1500 ppm, consistent with current estimates [17, 18], magma volume fluxes feeding mare lava flows are predicted to be in the range 10^5-10^6 m^3 s^{-1}, consistent with our previous estimates [20] based on emplacement dynamics of the observed flows.

Summary: We have developed improved treatments for the Moon of (a) the intrusion of dikes that do not erupt to the surface and (b) the flow of magma through dikes that do erupt. We find predicted geometries of dikes intruded to shallow depths to be consistent with the corresponding properties inferred from the graben that are their induced surface expressions. Similarly the magma volume fluxes implied for erupting dikes match those inferred from lava flow lengths when the stress conditions holding the dikes open are consistent with the near-surface pressures expected for plausible magma volatile contents. Collectively, these results enable us to understand the wide range of volcanic features that we have catalogued on the Moon [22].

Table 1. Fissure eruption conditions. Values of magma pressure in the vent P_{ch}, horizontal fissure length L, maximum fissure width W, magma rise speed at depth U and total volume flux V for various total magma volatile mass fractions n.

n /ppm	P_{ch} /MPa	L /km	W /m	U /(m/s)	V /(m^3/s)
500	0.116	473	10.0	2.9	6.8×10^6
750	0.174	210	6.7	2.4	1.7×10^6
1000	0.232	118	5.0	2.0	6.0×10^5
1250	0.290	76	4.0	1.8	2.8×10^5
1500	0.348	53	3.3	11.7	1.5×10^5

References: [1] McCoy, T.J., Mittlefehldt, D.W., Wilson, L. (2006) pp. 733-745 *in* Meteorites & the Early Solar System II. [2] Wilson, L., Goodrich, C.A., Van Orman, J.A. (2008) GCA **72** (24), 6154. [3] Ch. 4, Rev. Mineral. Geochem. 60, 2006. [4] Kelemen, P.B. et al. (1997) Phil. Trans. R. Soc. A 355, 1. [5] Sleep N.H. (1988) JGR. 93, 10255. [6] Wilson, L., Head, J.W. (2002) JGR 107, (E8), #5057. [7] Wieczorek M.A., Zuber M.T., Phillips, R.J. (2001) EPSL 185, 71. [8] Lister, J.R., Kerr, R.C. (1991) JGR 96 (B6), 10,049. [9] Rubin, A.M., Pollard, D.D. (1987) USGS Prof Pap. 1350, 1449. [10] Head, J.W., Wilson, L. (1992) GCA 56, 2155. [11] Broek, D. (1974) Elem. Eng. Fracture Mech. Noordhoof. [12] Petrycki, J.A., Wilson, L., Head, J.W. (2004) LPS XXXV, #1123. [13] Wilson, L., Head, J.W. (2003) GRL 30, (12), #1605. [14] Wilson, L., Head, J.W. (2007) Nature 447 (7140), 53. [15] Mitchell, K.L. (2005) JVGR 143, 187. [16] Rubin, A.M. (1993) EPSL 119, 641. [17] Nicholis, M.G., Rutherford, M.J. (2006) LPSC XXXVII, #2061. [18] Saal, A.E. et al. (2008) Nature 454 (7201), 192. [19] Schultz, R.A. et al. (2008) JVGR 178, 683. [20] Wilson, L., Head, J.W. (2008) LPS XXXIX, #1104. [21] Hikida, H., Wieczorek, M.A. (2007) Icarus 192, 150. [22] Wilson, L., Head, J.W. (2009) This meeting, #1159.