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8 **Generation, Ascent and Eruption of Magma on the Moon:**
9 **New Insights Into Source Depths, Magma Supply, Intrusions**
10 **and Effusive/Explosive Eruptions (Part 1: Theory)**
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Abstract:

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49 We model the ascent and eruption of lunar mare basalt magmas with new data on crustal
50 thickness and density (GRAIL), magma properties, and surface topography, morphology and
51 structure (Lunar Reconnaissance Orbiter). GRAIL recently measured the broad spatial variation
52 of the bulk density structure of the crust of the Moon. Comparing this with the densities of lunar
53 basaltic and picritic magmas shows that essentially all lunar magmas were negatively buoyant
54 everywhere within the lunar crust. Thus positive excess pressures must have been present in
55 melts at or below the crust-mantle interface to enable them to erupt. The source of such excess
56 pressures is clear: melt in any region experiencing partial melting or containing accumulated
57 melt, behaves as though an excess pressure is present at the top of the melt column if the melt is
58 positively buoyant relative to the host rocks and forms a continuously interconnected network.
59 The latter means that, in partial melt regions, probably at least a few percent melting must have
60 taken place. Petrologic evidence suggests that both mare basalts and picritic glasses may have
61 been derived from polybaric melting of source rocks in regions extending vertically for at least a
62 few tens of km. This is not surprising: the vertical extent of a region containing inter-connected
63 partial melt produced by pressure-release melting is approximately inversely proportional to the
64 acceleration due to gravity. Translating the ~25 km vertical extent of melting in a rising mantle
65 diapir on Earth to the Moon then implies that melting could have taken place over a vertical
66 extent of up to 150 km. If convection were absent, melting could have occurred throughout any
67 region in which heat from radioisotope decay was accumulating; in the extreme this could have
68 been most of the mantle.

69 The maximum excess pressure that can be reached in a magma body depends on its
70 environment. If melt percolates upward from a partial melt zone and accumulates as a magma
71 reservoir, either at the density trap at the base of the crust or at the rheological trap at the base of
72 the elastic lithosphere, the excess pressure at the top of the magma body will exert an elastic
73 stress on the overlying rocks. This will eventually cause them to fail in tension when the excess
74 pressure has risen to close to twice the tensile strength of the host rocks, perhaps up to ~10 MPa,
75 allowing a dike to propagate upward from this point. If partial melting occurs in a large region
76 deep in the mantle, however, connections between melt pockets and veins may not occur until a
77 finite amount, probably a few percent, of melting has occurred. When interconnection does
78 occur, the excess pressure at the top of the partial melt zone will rise abruptly to a high value,
79 again initiating a brittle fracture, i.e. a dike. That sudden excess pressure is proportional to the
80 vertical extent of the melt zone, the difference in density between the host rocks and the melt,
81 and the acceleration due to gravity, and could readily be ~100 MPa, vastly greater than the value
82 needed to initiate a dike. We therefore explored excess pressures in the range ~10 to ~100 MPa.

83 If eruptions take place through dikes extending upward from the base of the crust, the mantle
84 magma pressure at the point where the dike is initiated must exceed the pressure due to the
85 weight of the magmatic liquid column. This means that on the nearside the excess pressure must
86 be at least $\sim 19 \pm 9$ MPa and on the farside must be $\sim 29 \pm 15$ MPa. If the top of the magma body
87 feeding an erupting dike is a little way below the base of the crust, slightly smaller excess
88 pressures are needed because the magma is positively buoyant in the part of the dike within the
89 upper mantle. Even the smallest of these excess pressures is greater than the ~10 MPa likely
90 maximum value in a magma reservoir at the base of the crust or elastic lithosphere, but the
91 values are easily met by the excess pressures in extensive partial melt zones deeper within the
92 mantle. Thus magma accumulations at the base of the crust would have been able to intrude
93 dikes part-way through the crust, but not able to feed eruptions to the surface; in order to be

94 erupted, magma must have been extracted from deeper mantle sources, consistent with petrologic
 95 evidence.

96 Buoyant dikes growing upward from deep mantle sources of partial melt can disconnect from
 97 their source regions and travel through the mantle as isolated bodies of melt that encounter and
 98 penetrate the crust-mantle density boundary. They adjust their lengths and internal pressure
 99 excesses so that the stress intensity at the lower tip is zero. The potential total vertical extent of
 100 the resulting melt body depends on the vertical extent of the source region from which it grew.
 101 For small source extents, the upper tip of the resulting dike crossing the crust-mantle boundary
 102 cannot reach the surface anywhere on the Moon and therefore can only form a dike intrusion; for
 103 larger source extents, the dike can reach the surface and erupt on the nearside but still cannot
 104 reach the surface on the farside; for even larger source extents, eruptions could occur on both the
 105 nearside and the farside. The paucity of farside eruptions therefore implies a restricted range of
 106 vertical extents of partial melt source region sizes, between ~16 to ~36 km. When eruptions can
 107 occur, the available pressure in excess of what is needed to support a static magma column to the
 108 surface gives the pressure gradient driving magma flow. The resulting typical turbulent magma
 109 rise speeds are ~10 to a few tens of m s^{-1} , dike widths are of order 100 m, and eruption rates from
 110 1-10 km long fissure vents are of order 10^5 to $10^6 \text{ m}^3 \text{ s}^{-1}$.

111 Volume fluxes in lunar eruptions derived from lava flow thicknesses and surface slopes or
 112 rille lengths and depths are found to be of order 10^5 to $10^6 \text{ m}^3 \text{ s}^{-1}$ for volume-limited lava flows
 113 and $>10^4$ to $10^5 \text{ m}^3 \text{ s}^{-1}$ for sinuous rilles, with dike widths of ~50 m. The lower end of the
 114 volume flux range for sinuous rilles corresponds to magma rise speeds approaching the limit set
 115 by the fact that excessive cooling would occur during flow up a 30 km long dike kept open by a
 116 very low excess pressure. These eruptions were thus probably fed by partial melt zones deep in
 117 the mantle. Longer eruption durations, rather than any subtle topographic slope effects, appear to
 118 be the key to the ability of these flows to erode sinuous rille channels.

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

129

130 **1. Introduction**

131 The role of the generation, ascent and eruption of magma in shaping the surface of the
 132 Earth has long been studied, and the major environments of emplacement and extrusion
 133 (lithospheric plate boundaries, intraplate volcanic centers, and Large Igneous Provinces) are well
 134 known. Prior to the advent of the Space Age in 1957, the Moon was the first laboratory beyond
 135 Earth in which fundamental questions about the generation, ascent and eruption of magma could
 136 be considered in an independent and different planetary environment (e.g., unknown origin,
 137 uncertain interior structure, smaller size, different gravity, lack of an atmosphere, etc.). Was the
 138 Moon accreted hot or cold? How did the lunar nearside and farside compare? Were interior
 139 heating and extrusive volcanism important? What was the origin of the tens of thousands of

140 craters: volcanic (interior) or impact (exterior)? How did the resurfacing history of the Moon
141 compare with that of the Earth? The advent of the Space Age (with the first lunar mission, Luna
142 1, in 1959) immediately led to an era of exploration missions that rapidly addressed these
143 questions using orbital remote sensing, human and robotic surface exploration, deployment of
144 geophysical instruments and analysis of returned samples (see background and reviews in
145 Fielder, 1961; Baldwin, 1963; Toksoz et al., 1974 and Wilhelms, 1987).

146 In the first twenty five years of the Space Age, a wide variety of morphologic features and
147 deposits representing a range of volcanic eruption styles was identified and documented on the
148 Moon, mostly in and associated with the lunar maria (e.g., Schultz, 1976a; Guest and Murray,
149 1976; Head, 1976). By 1981, analysis of these data and returned samples led to an initial
150 understanding of the basic principles of the generation, ascent and eruption of magma on the
151 Moon, and how they compared with those of the Earth (Wilson and Head, 1981). Indeed, a
152 synthesis was compiled comparing the processes of basaltic volcanism on the Earth, Moon and
153 the terrestrial planets (BVSP, 1981), and the emerging picture of lunar evolution was used as a
154 planetary frame of reference (e.g., Taylor, 1982).

155 On the basis of the diversity and abundance of data about volcanism on the Moon, and the
156 clear distinction between magmatic styles on the Earth and one-plate planets like the Moon
157 (Solomon, 1978), the Moon has become a reference body for the understanding of crustal
158 formation and evolution, magmatism (plutonism and volcanism), and the thermal evolution of
159 one-plate planetary bodies. We now know that lunar mare basalt deposits cover ~17% of the
160 lunar surface, occur preferentially on the nearside and in topographic lows, and have a total
161 volume estimated at $1 \times 10^7 \text{ km}^3$ (Head and Wilson, 1992a). Returned samples and remote
162 sensing studies show that mare volcanism began prior to the end of heavy impact cratering (the
163 period of cryptomare formation; Whitten et al., 2015a,b), in pre-Nectarian times (Wilhelms,
164 1987), and continued possibly into the Copernican Period (Hiesinger et al., 2011), a total
165 duration approaching 3.5–4 Ga. Stratigraphic analyses (e.g., Hiesinger et al., 2011) show that the
166 volcanic flux was not constant, but peaked in early lunar history, during the Imbrian Period
167 (which spans the period 3.85-3.2 Ga) (Head and Wilson, 1992a). Average lunar volcanic output
168 rate during this peak period, $\sim 10^{-2} \text{ km}^3/\text{a}$, was very low relative to the present global terrestrial
169 volcanic output rate (comparable to the present local output rates for individual volcanoes such
170 as Vesuvius, Italy, and Kilauea, Hawai'i) (Head and Wilson, 1992a). On the other hand, volcanic
171 landforms indicate that peak fluxes were often extremely different from average fluxes. Some
172 eruptions associated with sinuous rilles (e.g., Hurwitz et al., 2012, 2013) were of large volume
173 and are estimated to have lasted on the order of a year and emplaced 10^3 km^3 of lava,
174 representing the equivalent in one year of about 100,000 years of the average flux (Head and
175 Wilson, 1992a). Due primarily to the low frequency of dike intrusions into a single specific area
176 of the crust, shallow magma reservoirs were uncommon (Head and Wilson, 1991); those
177 observed are related to intrusions of sills into low-density breccia zones below impact craters
178 (e.g., Schultz, 1976b; Jozwiak, 2012, 2015).

179 The asymmetry of mare deposits between the nearside and farside appears to be due
180 largely to differences in crustal thickness (Head and Wilson, 1992a; Whitten et al., 2011).
181 Magma ascending from the mantle or from a buoyancy trap at the base of the crust should
182 preferentially extrude to the surface on the nearside, but should generally stall and cool in dike
183 intrusions in the farside crust, extruding only in the deepest basins. Dikes that establish
184 pathways to the surface on the nearside should have very large volumes, comparable to the

185 volumes associated with many observed flows (Schaber, 1973; Moore and Schaber, 1975;
186 Bugiolacchi and Guest, 2008) and sinuous rille eruptions (e.g., Hurwitz et al., 2012, 2013).

187 As the Moon thermally evolves and loses heat dominantly by conduction (Solomon and
188 Head, 1982; Spohn et al., 2001; Ziethe et al., 2009), the interplay between thermal contraction
189 and differentiation leads to net cooling and ultimate contraction of the outer portions of the
190 Moon, resulting in a regional horizontal compressive stress acting on the lunar crust (Solomon
191 and Head, 1982). In addition, overall cooling leads to deepening of sources requiring the
192 production of ever-larger volumes of magma in order to reach the surface. Crustal stresses
193 became large enough with time so that few intruded dikes could open to the surface, causing
194 eruptive activity to be severely diminished in the Eratosthenian, and to cease in the Copernican
195 Period. Lunar mare deposits provide an example of the transition from primary crusts to
196 secondary crusts (Taylor, 1989) relevant to the ascent and eruption of magma and they illustrate
197 the significance of several factors in the evolution of secondary crusts, such as crustal density,
198 variations in crustal thickness (Wieczorek and Phillips, 1998; Wieczorek and Zuber, 2001;
199 Hikida and Wieczorek, 2007; Wieczorek et al., 2013), presence of impact basins, state and
200 magnitude of stress in the lithosphere, and general thermal evolution. These factors are also
201 responsible for the extremely low lunar volcanic flux, compared with Earth, even during periods
202 of peak extrusion (Head and Wilson, 1992a).

203 In parallel with the documentation of surface volcanic features and deposits, numerous
204 analyses have treated the petrology and geochemistry of the generation of mare basalts (e.g.,
205 summary in Shearer et al., 2006) and the physical processes associated with their ascent and
206 eruption (e.g., Wilson and Head, 1981; Head and Wilson, 1992a,b; Head and Wilson, 1994;
207 Wilson and Head, 2003a). In particular, it has been shown that the main path for the ascent and
208 eruption of magma from mantle source regions is through magma-filled cracks, i.e. dikes (Head
209 and Wilson, 1992a). Lacking, however, has been an up-to-date treatment of the generation,
210 ascent and eruption of magma that includes the full assessment of dike initiation in the deep
211 mantle, volatile sources and effects, and the behavior of dikes that penetrate to the shallow
212 subsurface, but do not penetrate fully to the surface to form significant effusive eruptions.

213 In this paper we present such an updated treatment of the generation, ascent and eruption
214 of magma using: 1) new data on lunar crustal thickness and structure from the Gravity Recovery
215 and Interior Laboratory (GRAIL) mission (Zuber et al., 2013), 2) new data on lunar rock and
216 melt density (Kiefer et al., 2012), 3) updated treatments of the generation of magma and the
217 initiation and propagation of magma-filled cracks (dikes) (Weinberg and Regenauer-Lieb, 2010;
218 Bouilhol et al., 2011; Havlin et al., 2013), 4) new data on the production of volatiles during
219 magma ascent and eruption (Wilson and Head, 2003a; Rutherford and Papale, 2009; Saal et al.,
220 2008; Wetzel et al., 2015), 5) the global topography of the Moon from new altimetry (Lunar
221 Orbiter Laser Altimeter, LOLA; Zuber et al., 2010; Smith et al., 2010), and 6) detailed
222 characterization of lunar volcanic features and deposits using new imaging (Lunar
223 Reconnaissance Orbiter Camera, LROC; Robinson et al., 2010), altimetry (Lunar Orbiter Laser
224 Altimeter, LOLA; Zuber et al., 2010; Smith et al., 2010), and spectral reflectance (Moon
225 Mineralogy Mapper, M3; Pieters et al., 2009) data. We begin with an updated assessment of
226 lunar crustal structure, and then provide a new assessment of the modes of dike initiation and
227 propagation from magma sources. Using this framework, we document the theoretical basis for
228 the ascent, intrusion and eruption of magma, including deep and near-surface processes of gas
229 release. Finally, we summarize the main themes, findings and predictions about the generation,
230 ascent and eruption of magma on the Moon and conclude with a discussion of the major factors

231 that are important in explaining the spectrum of lunar volcanic structures and deposits and how
 232 they differ from those on Earth and other planets. In a separate analysis (Head and Wilson,
 233 2015), we compare this theoretical treatment and its predictions with the variety and context of
 234 observed volcanic features, structures and deposits in order to test the predictions and refine the
 235 principles of ascent and eruption and to provide an interpretative framework for the major
 236 characteristics of mare basalt and related volcanism on the Moon. We also treat there the wide
 237 (up to 40 km) linear dike-like features that are interpreted to have been emplaced much earlier in
 238 lunar history than lunar mare basalts (Andrews-Hanna et al., 2013) and the intrusion of early
 239 lunar pre-mare Mg-suite magmas (Shearer et al., 2006).

240

241 **2. The influence of the structure of the Moon on volcanism**

242 The early thermal evolution of the Moon's interior has been modelled (Solomon and Head,
 243 1980; Hess and Parmentier, 1995, 2001; Spohn et al., 2001; Wieczorek et al., 2006; Shearer et
 244 al., 2006; Ziethe et al., 2009) assuming conductive cooling through the crust (the volume of mare
 245 lavas is too small a fraction of the total crustal volume for advective heat transfer to be an
 246 important contributor) and heating from a convecting rather than a conducting mantle (the option
 247 that maximises upward heat flow). All such models imply that the elastic lithosphere must have
 248 been 100-150 km thick during the main period of mare volcanism, mostly in the interval 3.9 to 3
 249 Ga before present, with minor activity as recent as ~1 Ga ago (Hiesinger et al., 2000, 2011),
 250 whereas the thickness of the crust would have already become fixed by solidification of an initial
 251 magma ocean to lie in the present-day range of ~30-50 km (Wieczorek et al., 2013). Thus
 252 plumes in a convecting mantle would have encountered a rheological boundary (Figure 1a) well
 253 below the base of the compositionally-defined lunar crust (Hess, 2000). It is therefore clear on
 254 theoretical rock-mechanical grounds (e.g. Pollard, 1988; Rubin, 1993) that magma transport at
 255 all depths shallower than this rheological boundary, not only through the shallow crust of the
 256 Moon but also through the upper mantle, must have taken place by flow through dikes held open
 257 by elastic stresses in rocks that behaved in a brittle fashion.

258 Other mechanisms of magma transport must have operated in the deeper interior (Hess,
 259 1991). The main alternative to flow through brittle fractures is porous flow along grain
 260 boundaries in regions of partial pressure-release melting accompanied by compaction of the
 261 matrix (Richter and McKenzie, 1984; Bouilhol et al., 2011). Insertion of plausible values for the
 262 parameters involved into the equations governing this process (Shearer et al., 2006) yields melt
 263 transport speeds within an order of magnitude of 1 m per year, in stark contrast to the likely rise
 264 speeds of magmas in brittle dikes which, under lunar conditions, are likely to be of order 3 m s^{-1}
 265 (Wilson and Head, 1981). A hybrid state must exist in the upper part of a region of partial
 266 melting where concentration of melt enlarges some small veins at the expense of others leading
 267 to rapid melt migration in a small number of large veins (Sleep, 1988). Indeed, a complex
 268 hierarchy of veins with a wide range of sizes may develop (Brown, 2004; Maaløe, 2005)
 269 encouraging melt percolation (Schmelting, 2006) until ductile fracture mechanisms allow a brittle
 270 fracture to nucleate (Weinberg and Regenauer-Lieb, 2010). These mechanisms are encouraged
 271 by the creation of local non-hydrostatic pressures as a result of the volume increase that most
 272 silicate minerals undergo on melting, but this process was probably not important on the Moon.
 273 This is because such excess pressures are important only if the melting rate is so fast (as was the
 274 case in early-forming asteroids heated by decay of short-lived ^{26}Al), that plastic flow of the
 275 mantle surrounding the partial melt zone could not occur fast enough to relax the developing

276 stresses (Wilson et al., 2008). An integrated model of the change from melt percolation to dike
 277 initiation in the lunar interior is developed by Havlin et al. (2013).

278 Mantle convection provides an obvious mechanism to cause melting by pressure release.
 279 However, it is not guaranteed that mantle convection was possible at all times in lunar volcanic
 280 history (Stevenson, 2003) and an alternative is melt formation by accumulation of radiogenic
 281 heat in finite regions due to the concentration of radioactives driven by gravitational overturn and
 282 negative diapirism of density-stratified cumulates (Delano, 1990; Wagner and Grove, 1997).
 283 Melt migration and upward concentration by porous flow is possible within such regions, but
 284 models involving this mechanism (Hess, 1991) generally place the melt sources even further
 285 below the elastic lithosphere. The frequency of mare lava eruptions (one large-volume eruption
 286 every 10^6 years – Head and Wilson, 1992a) implies that the time scale for the accumulation of a
 287 sufficient volume of magma to trigger an eruption is commonly much shorter than the Ga time
 288 scales of large-scale crustal deformation driven by global cooling. The boundary within a planet
 289 between elastic and plastic responses to applied stresses is strain-rate- as well as temperature-
 290 dependent, and so the rheological boundary for magma percolation should be deeper than the
 291 base of the elastic lithosphere as usually defined. This emphasizes the requirement for magma
 292 transfer in dike-like conduits in at least the upper part of the mantle. Perhaps the most
 293 compelling evidence of the need for such pathways (Wilson and Head, 2003a) is the petrologic
 294 implication that the picritic melts forming the orange, green and black pyroclasts found at the
 295 Apollo 15 and 17 sites were transported to the surface from sources at depths of 250-600 km on
 296 time scales of hours to days (Spera, 1992) without significant chemical interaction with the rocks
 297 through which they passed. This is a similar argument to that proposed for kimberlite eruptions
 298 on Earth (Wilson and Head, 2007a). The experimental verification (Beck et al., 2006) that
 299 relatively porous dunite channels should develop quickly over a large range of depths in the
 300 lunar mantle during a protracted melt-extraction episode provides an attractive explanation for
 301 the connection between deep inter-grain porous flow and shallower transport in dikes formed by
 302 brittle fracture.

303 Ideas on the relative importance of magma buoyancy and magma source pressure in lunar
 304 eruptions have evolved considerably with improvements in values for the density and thickness
 305 of the lunar crust (Wieczorek et al., 2013) and for the densities of the erupted magmas
 306 (Wieczorek et al., 2001; Shearer et al., 2006; Kiefer et al., 2012). Current estimated values of
 307 key parameters having a bearing on the physics of lunar volcanism can be summarized as
 308 follows. The lunar crust varies in total thickness, probably being on average ~30 km thick on the
 309 Earth-facing hemisphere and ~50 km thick on the far side (Fig. 3 of Wieczorek et al., 2013).
 310 This thickness for the near side crust may be a slight over-estimate due to the neglect of the mare
 311 lava fill in the analysis of the GRAIL mission data. The average crustal density, ρ_c , is ~2550 kg
 312 m^{-3} and the bulk density of the mantle, ρ_m , is inferred to lie between 3150 and 3370 kg m^{-3}
 313 (Wieczorek et al., 2013). Finally, the liquidus densities, ρ_l , of mare basalts and picritic magmas
 314 have been calculated by Wieczorek et al. (2001) and Shearer et al. (2006) to span the ranges
 315 2775 to 3025 and 2825 to 3150 kg m^{-3} , respectively. Kiefer et al. (2012) measured the densities
 316 of some returned lunar basalt samples in the range 3010 to 3270 kg m^{-3} , which would correspond
 317 to ~2980 to 3240 kg m^{-3} at liquidus temperatures. Thus with no exceptions mare basalts were
 318 negatively buoyant in the crust of the Moon. For subsequent modeling we take the density of the
 319 mantle to be the average of the estimated range, 3260 kg m^{-3} , and based on the various estimates
 320 we assume the density of erupting basalts to be either 2900 or 3010 kg m^{-3} .

321 The earliest estimates of the density structure of the Moon led to the suggestion that an
 322 excess pressure was required in the magma source region to enable melts to erupt at the surface,
 323 irrespective of whether those melts traveled directly from source to surface in a single event
 324 (Solomon, 1975; Wilson and Head, 1981) or were temporarily stored in a reservoir at some
 325 intermediate depth (Head and Wilson, 1992a,b). The new data confirm that such excess
 326 pressures are important. Their presence is understandable, because the melt in any region
 327 experiencing partial melting or containing accumulated melt will behave as though an excess
 328 pressure is present at the top of the melt column provided that the melt is positively buoyant
 329 relative to the host rocks and forms a continuously connected network. The value of this
 330 effective excess pressure is the product of the finite vertical extent of the region, the difference in
 331 density between the host rocks and the melt, and the acceleration due to gravity at the relevant
 332 depth. Petrologic evidence suggests that both mare basalts and picritic glasses may have been
 333 derived from polybaric melting of source rocks in regions extending vertically for at least a few
 334 tens of km (Shearer et al., 2006). This is not surprising: the vertical extent of an inter-connected
 335 partial melt region is expected to be approximately inversely proportional to the acceleration due
 336 to gravity (Turcotte and Schubert, 2002) and hence should be ~6 times larger on the Moon than
 337 on the Earth if no other factors intervene. Deciding what vertical extent of partial melting on
 338 Earth to use in such scaling is not trivial. Almost all melt production on Earth is associated with
 339 plate tectonics and the melting region is subject to horizontal shearing. Maaløe (2005) suggested
 340 that the vertical extent of melting in an unsheared rising mantle diapir on Earth could be as small
 341 as ~2 km; in contrast, estimates of the depths over which partial melting takes place under
 342 Hawai'i are ~25-40 km (Farnetani et al., 2010). We adopt the lower end of this range as being
 343 most likely to be relevant, and so melting in lunar diapirs might be expected to extend over a
 344 vertical distance of up to $\sim(6 \times 25 =) 150$ km. The excess pressure due to a typical $\sim 300 \text{ kg m}^{-3}$
 345 density difference between magma and host mantle would then be $1.62 \text{ m s}^{-2} \times 150 \text{ km} \times 300 \text{ kg}$
 346 $\text{m}^{-3} = \sim 73 \text{ MPa}$. We therefore include the possibility of excess magma source pressures up to
 347 $\sim 100 \text{ MPa}$ in our modeling. These values apply whether the magma source is a partial melt zone
 348 deep in the mantle or the melt at the top of a diapiric body stalled at a rheological trap at or
 349 beneath the base of the elastic lithosphere.

350 Excess pressures would also be present in bodies of melt trapped in magma reservoirs at
 351 the compositional discontinuity at the base of the crust or at other neutral buoyancy levels within
 352 the crustal lithosphere if these exist. In this latter case these excess pressures would be limited
 353 because they would only increase by the addition of more magma until the elastic deformation of
 354 the host rocks caused their fracture toughness to be exceeded somewhere, at which point a dike
 355 (or sill) would begin to propagate. As a result, excess pressures in these crustal magma bodies
 356 would probably be no more than twice the tensile strength of the host rocks (Tait et al., 1989),
 357 perhaps up to $\sim 10 \text{ MPa}$, an order of magnitude smaller than those in deep partial melt zones. We
 358 use the above concepts and numerical values in subsequent analyses.

359

360 **3. Modes of dike initiation and development from magma sources**

361 **3.A General considerations**

362 Dikes are initiated when the rocks overlying a melt body fracture in a brittle mode. As
 363 suggested above, in the mantle this may be the result of the excessive strain rate imposed on
 364 mantle rocks above a diapir rising to a rheological boundary where the decreasing temperature of
 365 the host rocks forces them to cease to respond in a plastic manner. In shallower bodies of melt
 366 already accumulated in elastic host rocks, fracture is likely to be the result of tensile failure of the

367 host rocks as the pressure in the already-accumulated magma increases, either as a result of the
368 arrival of more magma from depth or because the volume of the already-accumulated magma
369 increases as chemical or thermal evolution occurs. Cooling of magma and crystallization of
370 dense minerals causes a reduction of volume and a pressure decrease (Carslaw and Jaeger, 1947),
371 but the attendant chemical changes may in principle force volatile exsolution and hence a volume
372 and pressure increase. While significant for many magmas on Earth, this is not likely to be an
373 important process for lunar magmas due to their low contents of dissolved volatile species and
374 the fact that their commonest volatile, CO, was produced in a chemical reaction requiring
375 absolute pressures less than ~40 MPa (Sato, 1976, 1979; Fogel and Rutherford, 1995; Nicholis
376 and Rutherford, 2006, 2009; Wetzel et al., 2015).

377 The geometry of a growing dike that develops from an initial fracture will be dictated by
378 the excess pressure, if any, in the magma source region, the stress regime in the host rocks, and
379 the relative densities of the magma and host rocks (Pollard, 1988). The density of the host rocks
380 will in general be a function of depth below the surface as described in Section 2. The density of
381 the magma (neglecting the small effects of pressure-dependent compressibility) will be a
382 function of the way volatiles (if available) are released from, and accumulate in, the magmatic
383 liquid, and will vary with position and time along the growing dike until the upper dike tip either
384 reaches the surface or ceases to propagate (Wilson and Head, 2003a). Treatments allowing
385 predictions of dike propagation conditions exist for only limited ranges of circumstances and
386 focus on two scenarios: the rise of magma that is everywhere positively buoyant relative to the
387 host rocks and has no excess source pressure (Spera, 1980; Spence et al., 1987; Lister, 1990a,
388 1990b, 1991; Lister and Kerr, 1991; Mériaux and Jaupart, 1998; Dahm, 2000a, 2000b; Menand
389 and Tait, 2002; Rivalta and Dahm, 2006; Chen et al., 2007; Roper and Lister, 2007), and the
390 mainly lateral spreading of magma in dikes centered on a level of neutral buoyancy, again with
391 no excess source pressure (Lister, 1990b; Lister and Kerr, 1991). Roper and Lister (2005)
392 proposed a model of upward dike propagation including source pressure, but again only for
393 positively buoyant magmas. Chen et al. (2011) and Taisne et al. (2011) discuss the arrest of
394 buoyant dikes propagating upward from shallow and deep sources, respectively, when they
395 encounter a neutral buoyancy level. Taisne et al. (2011) also address the limitations on dike
396 propagation distance due to changes in dike shape, and Maccaferri et al. (2011) discuss sill
397 formation linked to host rock density and stress changes.

398 Early attempts to use the static dike models of Weertman (1971) to approximate all types
399 of propagating dikes (Head and Wilson, 1992a; Wilson and Head, 2001) were legitimately
400 criticized (Shearer et al., 2006) on the grounds that they ignored the dynamic aspects of dike
401 propagation. In particular, in the magma in a propagating dike there must be a pressure gradient
402 driving magma motion against wall friction (Lister, 1990a). The pressure at the base of a dike
403 connected to a large-volume magma reservoir is essentially fixed by the pressure at the top of the
404 reservoir, and so the pressure at the propagating upper tip must decrease to a low value to
405 maximize the magma flow rate. Lister and Kerr (1991) and Rubin (1993) inferred that this
406 minimum value should be the pressure at which the most soluble magmatic volatile present
407 (commonly water in magmas on Earth) is just saturated, and that the uppermost part of a dike
408 will consist of an elongate cavity containing pure gas at this saturation pressure. Wilson and
409 Head (2003a) pointed out that in addition there must be a zone of magmatic foam beneath the
410 gas-filled tip cavity. We enlarge on this model for propagating dikes feeding the opening stages
411 of lunar eruptions below.

412 Unless the magma in it suffers excessive cooling during transport, a dike containing
413 magma that is everywhere buoyant in its host rocks would inevitably reach the surface and erupt
414 until the supply of magma ceases. However, if the magma is not positively buoyant at all depths,
415 and the dike is relying to some extent on an excess pressure in the magma source to aid its
416 growth, the upper tip of the dike may cease to propagate for a number of reasons in addition to
417 thermal limitations. The stress intensity at the dike tip may no longer be able to fracture the
418 overlying rocks; or the combination of source pressure and magma-host-rock density contrast
419 may not be able to support the magma column any closer to the surface. In principle, another
420 option that could apply to both positively and negatively buoyant magma is that all of the
421 available magma might be removed from the source region. However, this would require the
422 rocks surrounding the source region to deform on a time scale, and by an amount, consistent with
423 the flow speeds of magmas in dikes (commonly within a factor of 10 of $\sim 3 \text{ m s}^{-1}$ on the Moon
424 and Earth - Wilson and Head, 1981) and the duration of the eruption. The deformation speed of
425 the mantle rocks surrounding a deep mantle source are likely to be similar to those inferred for
426 mantle convection, within 2 orders of magnitude of $\sim 0.1 \text{ m a}^{-1}$ (Crowley and O'Connell, 2012) on
427 Earth and presumably nearly an order of magnitude less on the Moon because of the smaller
428 acceleration due to gravity. This contrast by a factor of order at least 10^8 suggests that only a
429 very small fraction of the total available magma can be extracted quickly from a deep mantle
430 reservoir. Extraction would be aided if there were an excess pressure in the melt as a result of
431 elastic stresses in the host rocks, but even in the case of very shallow magma reservoirs, where
432 the host rocks behave entirely elastically, Blake (1981) showed that only $\sim 0.1\%$ of the reservoir
433 is likely to be erupted before the excess internal pressure is relaxed. Numerous individual mare
434 lavas flows with volumes up to $\sim 200 \text{ km}^3$ are observed (Head, 1976). Mare lava ponds have a
435 wide range of sizes from 15 to 1045 km^3 , with mean pond volumes of 190 km^3 in the Smythii
436 basin, 270 km^3 in the Marginis basin, 240 km^3 in Mare Orientale and 860 km^3 in the South Pole-
437 Aitkin basin (Yingst and Head, 1997, 1998; Whitten et al., 2011). We infer that if they behaved
438 elastically, the source regions feeding these eruptions must have had total volumes (assuming
439 they contained $\sim 0.1\%$ by volume melt) of order 10^5 to 10^6 km^3 . These volumes would, for
440 example, be consistent with spherical bodies of diameters ~ 60 to 125 km or flattened ellipsoids
441 of greater horizontal extent. It is difficult to imagine spherical magma bodies of this size being
442 present at the base of a 30-50 km thick crust without producing surface consequences, but
443 diapiric bodies of this size might well be present deeper in the mantle. An option for a shallow
444 source might be crustal underplating forming an areally extensive sill, but a 10 km thick example
445 of such a sill would need to have a diameter of ~ 350 to 1100 km .

446 The relative values of crust, mantle and magma densities for the Moon quoted in Section 2
447 imply that essentially all lunar magmas are negatively buoyant everywhere in the crust. Dikes
448 containing these magmas that originate from diapirs that have stalled at rheological boundaries in
449 the mantle will rise as far as the level of neutral buoyancy at the crust-mantle interface; as long
450 as the least principle stress is horizontal they will then spread out both vertically and laterally. In
451 some cases the upper tips of such dikes can reach the surface, provided that the positive
452 buoyancy in the part of the dike in the mantle is great enough. However, in other cases the upper
453 tip must remain below the surface. This is a circumstance commonly encountered in the lateral
454 rift zones of shield volcanoes on Earth. A model of the growth of such a system is given by
455 Lister (1990b) and Lister and Kerr (1991) and a model of the final configuration is given by
456 Rubin and Pollard (1987). The model of Lister (1990b) and Lister and Kerr (1991) assumes that
457 the growing dike intrusion is fed from a point source at the level of neutral buoyancy, the source

458 having no excess pressure, whereas in fact such a dike will always bring with it an internal
 459 excess pressure acquired in its deep mantle source zone, ensuring that there is a positive
 460 pressure, in excess of the local lithostatic load, at the neutral buoyancy depth. This excess
 461 pressure is not included in Lister and Kerr's (1991) dynamic model, which led Wiczorek et al.
 462 (2001) to conclude, we infer incorrectly, that lateral intrusions at neutral buoyancy depths can
 463 never reach the surface. The option of including such an excess pressure is part of the Rubin and
 464 Pollard (1987) static model describing the final configuration of such a dike, and we use this
 465 treatment to model the final geometry of intruded dikes that do not erupt at the surface and to
 466 estimate the eruption conditions in such dikes that do breach the surface. Dikes able to intrude
 467 the crust or erupt are likely to have been the norm during the first quarter of lunar history when
 468 interior heating and global expansion induced extensional stresses in the crust (Solomon and
 469 Head, 1980), and, indeed, this period overlaps the main era of mare volcanism (Hiesinger et al.,
 470 2003). In cases where the least principle stress is not horizontal, most likely during the latter
 471 three-quarters of lunar history when global cooling induced compressive stresses in the crust
 472 (Solomon and Head, 1980), mantle dikes encountering the density discontinuity at the crust-
 473 mantle boundary are more likely to have initiated sills underplating the crust. The possibility
 474 exists that some of these magma bodies evolved chemically in ways that subsequently allowed
 475 them to inject rare high-silica dikes into and through the crust (e.g., Wilson and Head, 2003b).

476 A final but critical issue concerns dikes that grow from the tops of diapiric bodies deep in
 477 the mantle. These dikes can in principle grow upward to a great length, albeit slowly because
 478 they are being fed by melt migrating through the unmelted mineral fabric of the diapir.
 479 However, if a dike becomes too vertically extensive under these slow-growth conditions (Figure
 480 1b), the overall stress distribution can cause the dike to pinch off from its source while the upper
 481 tip is still extending (Weertman, 1971; Muller and Muller, 1980; Crawford and Stevenson,
 482 1988). The slow growth rate means that this stage of the dike's development can be adequately
 483 treated by the static stress models of Weertman (1971). After disconnecting from its source, the
 484 dike migrates as a discrete body of fixed volume, with the host rocks fracturing and opening
 485 above and closing behind the dike. This process leads to a limitation on the maximum volume of
 486 melt that can be transferred upward from the deep mantle in a single dike-forming event. This
 487 volume limitation would not necessarily apply to a dike reaching the surface or near-surface
 488 from a shallower diapir to which it was still connected.

489 The above considerations suggest that we need to address the following scenarios for the
 490 Moon (Figure 1): (a) dikes growing from magma sources so deep that the dike pinches off from
 491 the top of the magma source before the upper tip of the dike is arrested by any of the
 492 mechanisms discussed above, and (b) dikes growing from magma sources sufficiently shallow
 493 that a continuous dike pathway can exist between the top of the magma source region and the
 494 upper tip of the dike, irrespective of how far the dike is able to penetrate into the crust and
 495 whether or not it reaches the surface. The first of these scenarios is a guide to the major issues
 496 involved.

497

498 **3.B Stability and sizes of dikes growing from deep mantle sources**

499 Consider a dike that has grown upward to a length L from a diapiric magma body of
 500 vertical extent E_d . The stress intensity at the upper tip of the dike, K_u , is given by

501

$$502 \quad K_u = (\pi L)^{1/2} [P_d + (g \Delta\rho L)/\pi] \quad (1),$$

503

504 where $\Delta\rho$ is the density difference between host mantle and magma, ($\rho_m - \rho_l$), and P_d is the
 505 driving pressure at the dike inlet, given by

$$506 \quad P_d = g \Delta\rho E_d \quad (2).$$

507
 508
 509 Given the earlier arguments, it seems reasonable to assume that the diapiric body is undergoing
 510 partial melting over a vertical distance of at least $E_d = 10$ km; in that case, with $\rho_m = 3260 \text{ kg m}^{-3}$
 511 and $\rho_l = 2900 \text{ kg m}^{-3}$, so that $\Delta\rho = 360 \text{ kg m}^{-3}$, and $g = 1.62 \text{ m s}^{-2}$, P_d will be 5.8 MPa. The
 512 requirement for fracturing to occur at the upper dike tip, allowing it to grow, is that K_u must be
 513 greater than the fracture toughness of the host rocks, K_{crit} . Values of K_{crit} measured in laboratory-
 514 scale samples are $\sim 3 \text{ MPa m}^{1/2}$, and for crustal-scale masses of volcanic rocks values have been
 515 estimated at $\sim 100 \text{ MPa m}^{1/2}$ (Rubin, 1995). With $P_d = 5.8 \text{ MPa}$, K_u would exceed K_{crit} if L were
 516 ~ 11.5 cm for the lower fracture toughness value and ~ 93 m for the higher value. Given that we
 517 are assuming that partial melt occupies a region extending vertically for at least 10 km, having
 518 interconnected melt veins ready to form an embryonic dike with a vertical extent of even the
 519 larger of these values does not seem likely to be a problem.

520
 521 The stress intensity at the lower dike tip is K_l given by

$$522 \quad K_l = (\pi L)^{1/2} [P_d - (g \Delta\rho L)/\pi] \quad (3).$$

523
 524
 525 Initially the vertical dike length L is small and the term $[(g \Delta\rho L)/\pi]$ is much less than
 526 P_d , so that the stress intensities at both ends are similar and equal to $[(\pi L)^{1/2} P_d]$. As the dike
 527 grows, i.e. L increases, K_u constantly increases because both P_d and $[(g \Delta\rho L)/\pi]$ in equation (1)
 528 are positive. In contrast, the negative second term in equation (3) causes K_l to go through a
 529 maximum and eventually decrease, reaching zero when L reaches a critical value L_m such that P_d
 530 $= (g \Delta\rho L)/\pi$, i.e.,

$$531 \quad L_m = (\pi P_d)/(g \Delta\rho) \quad (4a),$$

532
 533
 534 We note here that equations (1) and (3) above differ from the equivalent formulae given by
 535 Muller and Muller (1980) by a factor of 2 in the second term on the right-hand side; we have
 536 made this change as the only way of reconciling expression (4a) with the more detailed analysis
 537 given by Weertman (1971). Figure 2 shows an example of how K_u and K_l vary with L , again for
 538 $P_d = 5.8 \text{ MPa}$: K_u increases continuously as L increases, whereas K_l increases at first to a
 539 maximum of $705 \text{ MPa m}^{1/2}$ when $L = 10.5$ km and then decreases to zero at $L = L_m = 31.4$ km.
 540 The fact that K_l reaches zero means that the stress acting on the lower tip of the dike causes its
 541 width to go to zero. At this point the dike, containing positively buoyant magma throughout its
 542 length as long as its top has not reached the crust-mantle interface, decouples from the diapiric
 543 source region and migrates upward as a discrete entity (Figure 1b). Rocks ahead of it fracture
 544 and dilate to provide a path and close back together behind it. In practice a small amount of
 545 magma is likely to be left on the walls of the closing crack, so that the volume of the dike is
 546 steadily depleted, but this is likely to be a small effect for dikes of the sizes relevant here. If we
 547 substitute for P_d from equation (2) into equation (4a), we find

$$548 \quad L_m = \pi E_d \quad (4b).$$

550

551 Thus the maximum vertical extent of a stable dike is slightly more than three times the vertical
 552 extent of the diapiric body that feeds it. Furthermore, L_m is independent of the density difference
 553 between magma and mantle host rocks, as long as there is a difference. If E_d is 10 km, as in the
 554 above example, L_m is ~ 31 km, and if E_d were as large as 100 km, L_m would be 314 km. But even
 555 in that case, it would still not be possible for a continuous dike pathway to exist between a deep
 556 mantle source region, at a depth of ~ 500 km, and the surface. Note that these results depend
 557 only on the criterion that the stress intensity at the pinch-off point is zero, and are completely
 558 independent of the fracture toughness assumed for the host rocks.

559 Large dikes that have decoupled from their source regions will have complex shapes
 560 because the need to do work against wall friction requires a pressure gradient to drive magma
 561 flow, and we do not model these shapes in detail during the passage of the dikes through the
 562 mantle. However, we can estimate the mean dike widths and initial volumes of magma in dikes
 563 of this kind as they decouple from the source by noting that their slow growth up to this point
 564 suggests that they will have the "penny" shape often cited as likely for static dikes, in which the
 565 horizontal extent is equal to the vertical length, L_m , and the mean thickness, W , can be obtained
 566 by numerically integrating Weertman's (1971) equation (20) and is well-approximated by
 567

$$568 \quad W = (\pi/6) [(1 - \nu)/\mu] L_m P_d \quad (5).$$

569

570 where ν is the Poisson's ratio and μ is the shear modulus of the host rocks. Adopting $\nu = 0.25$
 571 and $\mu = 4$ GPa (Bieniawski, 1984; Rubin, 1990) gives conservative estimates of W . The volume,
 572 V , of magma in the dike is then given by $[(\pi/4) L_m^2 W]$, i.e.

573

$$574 \quad V = (\pi^2/24) [(1 - \nu)/\mu] L_m^3 P_d \quad (6).$$

575

576 Using the above relationships, Table 1 shows the implied dike-base driving pressures, vertical
 577 lengths and volumes of dikes produced from diapiric partial melt zones of a range of vertical
 578 extents. For later reference the table also contains the driving pressure at the dike center, P_0 ,
 579 which is greater than P_d by an amount $(g \Delta\rho 0.5 L_m)$. Using equations (2) and (4b),

580

$$581 \quad P_0 = [1 + (\pi/2)] (g \Delta\rho E_d) = [1 + (\pi/2)] P_d \quad (7).$$

582

583 Two implications can be drawn from Table 1. First, unless source regions are many tens of
 584 km in vertical extent, the volumes of magma available to reach shallow depths in a single
 585 extraction episode are a few thousand km^3 . Only a fraction of this is likely to be erupted at the
 586 surface - we return to this issue later. These magma volumes are large enough that the
 587 limitations on isolated dike propagation due to changing shape discussed by Taisne et al. (2011)
 588 do not apply. Second, the widths of the propagating isolated dikes are so large that magma
 589 motion within them is likely to be turbulent and thus not controlled explicitly by the magma
 590 viscosity. If, as generally assumed, the pressure in the upper tip of a propagating dike decreases
 591 to a low value to induce the pressure gradient, dP/dz , driving the magma motion, the pressure
 592 gradient must be of order (P_d/L_m) , and equation (4a) shows that this will be $(g \Delta\rho)/\pi$, $\sim 185 \text{ Pa m}^{-1}$,
 593 in all cases. The turbulent flow speed U of magma in a dike of width W under a pressure
 594 gradient dP/dz is given by
 595

$$U = [(W \, dP/dz)/(f \, \rho_l)]^{1/2} \quad (8)$$

where f is a friction factor close to 0.02. For the range of values of W in Table 1, U spans the range from $\sim 4 \, \text{m s}^{-1}$ to $\sim 70 \, \text{m s}^{-1}$. At an intermediate speed of $30 \, \text{m s}^{-1}$, a propagating isolated dike would require only 4.6 hours to reach the surface from a depth of 500 km, a speed of $\sim 100 \, \text{km/hour}$.

3.C Isolated dikes encountering the crust-mantle interface density trap

As mentioned earlier, the volumes of all but the very smallest dikes shown in Table 1 are large enough that we do not need to consider the details of their evolving shapes while they are rising through the mantle, a general issue addressed by Taisne et al. (2011). Instead we focus next on what happens when an isolated dike reaches the base of the crust. As long as the least principle stress is horizontal, the dike penetrates the crust and, as long as it does not erupt, stabilizes with its center at or very close to the crust-mantle density discontinuity. This geometry was modeled by Rubin and Pollard (1987). The criteria for stability are that the stress intensity at the tips of the dike should be equal to the fracture toughness of the host rock. In the present case the more important end of the dike is the upper tip because the lower tip is located in rocks that have already fractured to allow passage of the dike. There is an added requirement, that the driving pressure should adjust until the thickness of the dike is such that the volume of magma that it contains is equal to the volume of magma that was in the dike when it left the mantle source zone. The equations specifying the stress intensities at the upper and lower dike tips, K_{top} and K_{base} , respectively, and the new mean dike thickness, W_n , are:

$$K_{\text{top}} = P_n (A_u A_l)^{1/4} - g [(\pi^{-1} + 4^{-1}) (\rho_l - \rho_c) A_u^{3/2} + (\pi^{-1} - 4^{-1}) (\rho_m - \rho_l) A_l^{3/2}] \quad (9),$$

$$K_{\text{base}} = P_n (A_u A_l)^{1/4} - g [(\pi^{-1} - 4^{-1}) (\rho_l - \rho_c) A_u^{3/2} + (\pi^{-1} + 4^{-1}) (\rho_m - \rho_l) A_l^{3/2}] \quad (10),$$

$$W_n = [(1 - \nu)/\mu] \{0.5 \pi P_n (A_u A_l)^{1/2} - 0.33 g [(\rho_l - \rho_c) A_u^2 + (\rho_m - \rho_l) A_l^2]\} \quad (11),$$

where P_n is the new driving pressure at the crust-mantle boundary and A_u and A_l are the extensions of the dike above and below that boundary, respectively. Setting $K_{\text{base}} = 0$,

$$P_n (A_u A_l)^{1/4} = g [(\pi^{-1} - 4^{-1}) (\rho_l - \rho_c) A_u^{3/2} + (\pi^{-1} + 4^{-1}) (\rho_m - \rho_l) A_l^{3/2}] \quad (12).$$

Inserting this expression for $P_n (A_u A_l)^{1/4}$ into (9), and setting $K_{\text{top}} = K_{\text{crit}}$,

$$K_{\text{crit}} = 0.5 g [(\rho_m - \rho_l) A_l^{3/2} - (\rho_l - \rho_c) A_u^{3/2}] \quad (13),$$

which gives a relationship between A_u and A_l :

$$A_l^{3/2} = \{[K_{\text{crit}} (0.5 g)] + (\rho_l - \rho_c) A_u^{3/2}\} / (\rho_m - \rho_l) \quad (14).$$

These equations can be solved with the following steps: (i) an estimate is made of A_u ; (ii) equation (14) is used to find the corresponding value of A_l ; (iii) the values of A_u and A_l are inserted into equation (12) to find P_n ; (iv) the values of A_u , A_l and P_n are inserted into equation (11) to find W_n ; (v) the magma volume implied by these values, V_n , approximated by $(\pi A_u A_l$

642 W_n), is calculated and compared with the original volume leaving the mantle source, V , given by
 643 equation (6); (vi) the estimate of A_u in step (i) is varied until the two volumes are equal. This
 644 process is readily implemented in a spreadsheet.

645 Table 2 shows the results using $K_{crit} = 100 \text{ MPa m}^{1/2}$. Part (a) of the table assumes a
 646 magma density, ρ_1 , of 2900 kg m^{-3} and part (b) assumes $\rho_1 = 3010 \text{ kg m}^{-3}$. The values of E_d , L_m ,
 647 W , V and P_0 are repeated from Table 1 for comparison with the values of the dike lengths above
 648 and below the crust-mantle interface, A_u and A_l , the mean dike width, W_n , and the driving
 649 pressure at the dike center, P_n , after it has intruded the crust. For $\rho_1 = 2900 \text{ kg m}^{-3}$, the trend is
 650 for both the mean width of the dike and its driving pressure to decrease, by $\sim 30\%$ and slightly
 651 more than a factor of 2, respectively. The total vertical length of the dike, $A_d = (A_u + A_l)$,
 652 increases by $\sim 15\%$. For $\rho_1 = 3010 \text{ kg m}^{-3}$, both the mean width of the dike and its driving
 653 pressure decrease, by $\sim 17\%$ and $\sim 52\%$, respectively, and the total vertical length of the dike, (A_u
 654 $+ A_l$), increases by $\sim 11\%$. These results are not strongly dependent on the value assumed for
 655 K_{crit} . Reducing the value by an order of magnitude to $10 \text{ MPa m}^{1/2}$ to be more consistent with
 656 values determined in laboratory experiments only change the values of dike length, width and
 657 driving pressure in the part of the table that we shall show to be of most importance by at most
 658 5% .

659 Table 2 is of critical importance in understanding why eruptions are concentrated on the
 660 nearside of the Moon. The upper group of values in italics in both parts of Table 2 are solutions
 661 for which A_u is less than 30 km, meaning that the upper tips of dikes do not reach the surface on
 662 either the near- or farsides of the Moon. These conditions do not lead to eruptions but instead
 663 represent intrusions. The tops of these intrusions are generally at shallower depths on the
 664 nearside, with depths ranging up to ~ 13 km, than on the farside, where depths range up to ~ 43
 665 km. The lower group of values in italics in both parts of the table represent solutions for which
 666 A_u is greater than 50 km, meaning that upper dike tips could reach the surface on the farside, so
 667 that eruptions would have readily occurred on both the near- and farsides of the Moon. These
 668 solutions are not in agreement with the observation that farside eruptions are very rare. The non-
 669 italic values in the center of each table are solutions where eruptions can occur on the nearside
 670 but not on the farside, as observed. We infer that these represent the actual conditions during the
 671 main period of mare volcanism. They imply that the vertical extents of the mantle diapiric
 672 source regions that produced the erupted magmas lay in the restricted range of 17 to 27 km for a
 673 basalt density of 2900 kg m^{-3} and 22 to 36 km for a basalt density of 3010 kg m^{-3} . If we include
 674 dikes that intruded the crust but did not erupt, the implication is that to allow intrusions
 675 anywhere on the Moon, but to allow eruptions on the nearside while at the same time forbidding
 676 eruptions on the farside of the Moon, mantle diapiric sources could have had any vertical extents
 677 up to a limiting value of $\sim 32 \pm \sim 5$ km.

678

679 ***3.D Erupted volumes and eruption rates from isolated dikes breaching the surface***

680 The values of parameters in Table 2 can be used to make estimates of the expected
 681 volumes of magma erupted when the tops of dikes stalled at the crust-mantle boundary reached
 682 the surface and caused a fissure vent to become active. Once the surface was breached, the
 683 conditions in the dike changed in several ways. The excess internal pressure was relaxed as
 684 magma spilled onto the surface, and the eruption continued until an equilibrium was reached
 685 between the horizontal stress state in the crust and the negative buoyancy, relative to the crust, of
 686 the magma in the dike. Recall that dikes capable of reaching the surface were intruded during
 687 the period of lunar history when internal heating had led to expansion of the interior and the

688 production of a net tensional deviator, relative to hydrostatic stresses, in the crust. Solomon and
 689 Head (1980, their Fig. 20) estimated that this tensional deviator, T , was present in the interval
 690 between 3.8 and 3.0 Ga ago and reached a maximum value of ~ 20 MPa near the middle of this
 691 time interval. The tensional deviator replaced the internal excess pressure as the stress holding
 692 the intruded dike open, the upper part of the dike occupied the full thickness of the crust, C , and
 693 the extent of the part of the dike in the mantle shrank to a new final value, A_{if} , such that the stress
 694 intensity at the lower dike tip was exactly zero. By analogy with equation (10) this requires

$$695 \quad K_{\text{base}} = 0 = T (C A_{if})^{1/4} - g [(\pi^{-1} - 4^{-1}) (\rho_l - \rho_c) C^{3/2} + (\pi^{-1} + 4^{-1}) (\rho_m - \rho_l) A_{if}^{3/2}] \quad (15),$$

696 which allows A_{if} to be found for any given values of T and C . The mean width of the final
 697 intrusion, by analogy with equation (11) is W_f where

$$700 \quad W_f = [(1 - \nu)/\mu] \{0.5 \pi T (C A_{if})^{1/2} - 0.33 g [(\rho_l - \rho_c) C^2 + (\rho_m - \rho_l) A_{if}^2]\} \quad (16),$$

701 Setting $C = 30$ km for the nearside crust, Figure 3 shows how A_{if} varies with T and Figure 4
 702 shows how W_f varies with T , in each case for magma densities 2900 and 3010 kg m⁻³. Figure 4
 703 implies that if T is less than some limiting value, ~ 6.2 MPa for $\rho_l = 2900$ kg m⁻³ and ~ 7.1 MPa
 704 for $\rho_l = 3010$ kg m⁻³, no stable residual dike can exist, and so all of the magma in the potential
 705 intrusion is forced out onto the surface as lava. To establish the actual volumes involved we
 706 need to know the horizontal extent of the residual dike. In keeping with the concept used earlier
 707 of the rising dike having had a penny-like shape, the horizontal length is assumed to be still
 708 approximated by the total vertical height before the surface was reached, $A_d = (A_u + A_l)$. The
 709 volume of the residual dike is therefore $V_f = [(C + A_{if}) W_f A_d]$, and the volume of magma erupted,
 710 V_e , is the difference between the original volume of the dike in the mantle, V given by equation
 711 (6), and the residue in the intrusion, V_f . Figures 5(a) and (b) show how V_e varies with T for
 712 magma densities of 2900 and 3010 kg m⁻³, respectively. In each case the graphs are labeled with
 713 the range of values of E_d , the mantle source zone extent, that allows eruptions on the nearside but
 714 not the farside. As expected, for large values of T much of the magma remains in the residual
 715 intrusion and does not erupt, but for values of T less than the limiting values implied by Figure 4,
 716 essentially all of the magma erupts. The range of erupted volumes extends from the order of tens
 717 of km³ up to more than 750 km³ for a magma density of 2900 kg m⁻³ and to more than 1600 km³
 718 for a magma density of 3010 kg m⁻³. These values are entirely consistent with the range of lava
 719 flow and lava pond volumes reported in Section 3.A.

720 Estimating the erupted volume fluxes during the eruptions produced by these dikes is also
 721 not easy without a much more elaborate model than that presented here. However, an estimate
 722 of the eruption rate soon after the surface is first breached can be found by evaluating the
 723 pressure at the base of the magma column in the dike, subtracting the static weight of the magma
 724 in the dike, and noting that the difference, if positive, is the pressure differential available to
 725 drive magma upward against fluid friction with the dike walls. Dividing this by the vertical
 726 extent of the dike gives the pressure gradient that can be used to give the magma rise speed in the
 727 dike. The pressure at the base of the magma column in the dike is P_b where

$$730 \quad P_b = \rho_c g C + \rho_m g A_l \quad (17).$$

731 The pressure due to the weight of the magma is P_w where

$$P_w = \rho_l g (C + A_l) \quad (18).$$

The pressure difference driving the magma upward is $dP = (P_b - P_w)$, and the length of the magma column is $(C + A_l)$, so the pressure gradient driving magma motion, dP/dz , is

$$dP/dz = g \{ [A_l (\rho_m - \rho_l) - C (\rho_l - \rho_c)] / (C + A_l) \} \quad (19).$$

We cannot decide *a priori* if the motion of the magma in the dike is laminar or turbulent and so adopt the procedure shown to work by Wilson and Head (1981): we calculate the speed using both assumptions, i.e.,

$$U_{\text{lam}} = (W_n^2 dP/dz) / (6 \eta) \quad (20),$$

$$U_{\text{turb}} = [(W_n dP/dz) / (f \rho_l)]^{1/2} \quad (21),$$

and adopt whichever is the smaller speed. In these equations η is the magma viscosity, taken as 1 Pa s, and f is a dimensionless wall friction factor close to 0.02. Finally, the volume flux of magma being erupted from a fissure vent of horizontal extent L_e is F_e given by

$$F_e = \text{MIN}(U_{\text{lam}}, U_{\text{turb}}) W_n L_e \quad (22).$$

Theoretically, there is no reason to think that the length, L_e , of the surface fissure from which magma erupts will be the same as the entire subsurface horizontal extent of the dike, A_d . Penny-shaped dikes are by definition convex-upward where they approach the surface. If the shape were preserved, half of the magma in the dike would have to be erupted before L_e became as large as A_d . Most of the pressure gradient driving the eruption would have been relaxed by this time and a much shorter fissure would be able to accommodate the magma volume flux. There are very few well-preserved examples of volcanic vents on the Moon, many vents being drowned in the late stages of eruptions (*Head, 1976; Head and Wilson, 1992a; Head and Wilson, 2015*). Perhaps the most useful evidence comes from the sizes of the source depressions feeding sinuous rilles. These depressions are interpreted to be the results of thermal erosion at the bases of lava ponds fed by explosive eruptions (*Wilson and Head, 1980; Head and Wilson, 1980*). A survey of the asymmetries of 15 elongate sinuous rille source depressions (*Head and Wilson, 1981* and unpublished data) suggests that the fissure vents that fed them had lengths that ranged from 200 to 7000 m, with a median value of 1600 m. Adopting this value for L_e , Figure 7 shows the inferred magma rise speeds, U_e , all of which are found to be turbulent, and the corresponding erupted volume fluxes, F_e . For a magma density of 2900 kg m^{-3} , erupted volume fluxes lie in the range 10^5 to $10^6 \text{ m}^3 \text{ s}^{-1}$; for a magma density of 2900 kg m^{-3} , the volume fluxes are all of order $10^6 \text{ m}^3 \text{ s}^{-1}$. We show later that volume fluxes of order $10^5 \text{ m}^3 \text{ s}^{-1}$ are implied in the formation of sinuous rilles, and that fluxes of order $10^6 \text{ m}^3 \text{ s}^{-1}$ are required to form the longest mare lava flows.

Note the smaller range of values of U_e and F_e for the higher density magma in both parts of Figure 7. This dikes in which this magma rises are capable of reaching the surface, but it is so dense that its positive buoyancy in the mantle cannot compensate completely for its negative buoyancy in the crust. In practice, such a situation would lead to volatile exsolution in the

780 magma in the upper part of the dike and the formation of a gas pocket overlying a column of
 781 foam, effectively reducing the magma density and making it possible for an eruption to begin;
 782 we discuss these dynamic effects in more detail shortly. The rate at which magma was expelled
 783 from the bulk of the dike would be a function of the rate at which the crustal host rocks relaxed
 784 in response to the changing stress conditions. It seems likely, based on the trend of the values of
 785 U_e and F_e in the rest of the figure, that eruptions from these dikes would have taken place at
 786 volume fluxes in the range 10^4 to 10^5 $\text{m}^3 \text{s}^{-1}$.

787

788 ***3.E Eruptions and intrusions when dikes are connected to their melt source zones.***

789 Our previous treatments of dikes erupting at the surface (Wilson and Head, 1981; Head and
 790 Wilson, 1992a) assumed that they were still connected to their magma source zones, a condition
 791 which we have seen above is not possible for magma sources deep in the mantle. However, for
 792 shallower mantle sources such a scenario is still possible, but imposes restrictions on the possible
 793 depths and sizes of magma sources. The model of the Moon's thermal development proposed by
 794 Solomon and Head (1980, their Fig. 21) suggests that partial melting at depths less than ~ 250 km
 795 was confined to the first ~ 500 Ma of lunar history, and that melting at greater depths was not
 796 possible in this period. The models of Spohn et al. (2001) and Ziethe et al. (2009) allow for
 797 melting at depths between ~ 200 km and ~ 600 km during this period. Figure 6 shows a scenario
 798 in which early mare basalts are generated by partial melting within a finite region in the upper
 799 mantle of vertical extent E . The level at which the stresses combine to initiate a dike is at a
 800 depth Z below the surface. The positive buoyancy of the magma in the mantle diapir leads to an
 801 excess pressure at the dike inlet, and this pressure is available to support the column of magma in
 802 the dike. If the excess pressure is great enough, the column of magma can be supported all the
 803 way to the surface and an eruption can occur. If the pressure is not great enough, the dike will
 804 stall with its top at some depth H below the surface (Figure 6). In that case, the balance of
 805 stresses is

806

$$807 \quad g \rho_l (E + Z - H) = g \rho_c C + g \rho_m (E + Z - C) \quad (23a).$$

808

809 so that

810

$$811 \quad H = [C (\rho_m - \rho_c) - (Z + E) (\rho_m - \rho_l)] / \rho_l \quad (23b).$$

812

813 If H given by this expression is negative, a column of magma could in principle extend above the
 814 lunar surface. In practice, the excess pressure represented by the weight of this magma, $(g H \rho_l)$,
 815 is used to drive the magma motion against wall friction up the eruption pathway of length Z . The
 816 equivalent pressure gradient is then given by

817

$$818 \quad dP/dz = (g/Z) [(Z + E) (\rho_m - \rho_l) - C (\rho_m - \rho_c)] \quad (24).$$

819

820 The mean thickness W_{av} of the dike is again calculated using the model of Rubin and
 821 Pollard (1987). The thickness found is a realistic estimate of the intrusion thickness if no
 822 eruption occurs and is an estimate of the initial thickness, subject to later relaxation, when an
 823 eruption does occur. In that case the initial magma rise speed U_i is again found as the smaller of
 824 the values given by equations (20) and (21) and the initial volume flux, V_i , from a 1600 m-long
 825 fissure is given. Note that we do not specify any non-hydrostatic stress in the crust because we

826 are considering volcanism occurring at a time before large extensional or compressive stresses
 827 are likely to have built up in the lunar crust due to thermal effects (Solomon and Head, 1980).
 828 Also, we tacitly assume that the magma source region contains a great enough volume of magma
 829 to fill a dike extending to the surface, and that when eruptions occurred, the dike geometry and
 830 magma flow rate were such that the magma did not cool excessively while traveling to the
 831 surface. This was checked by evaluating the minimum rise speed, U_{\min} , to avoid significant
 832 cooling given by Wilson and Head (1981) as

$$834 \quad U_{\min} = 5 \kappa Z / W_{\text{av}}^2 \quad (25)$$

835
 836 where κ is the thermal diffusivity of the magma, $\sim 7 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$. In all cases U_i was found to be
 837 much greater than U_{\min} , so that cooling was never a problem.

838 The simplest result of this analysis can be illustrated by setting the vertical extent of the
 839 mantle melting zone, E , to a very small value, essentially zero. In that case an eruption will
 840 occur if the magma source is at a depth below the surface greater than a critical value Z_{crit} such
 841 that

$$842 \quad g \rho_l Z_{\text{crit}} = g \rho_c C + g (Z_{\text{crit}} - C) \rho_m \quad (26)$$

843
 844 Using $C = 30$ km on the lunar nearside and 50 km on the farside we find that if the magma
 845 density is 2900 kg m^{-3} , Z_{crit} for the nearside is 59.2 km and the value for the farside is 98.6 km.
 846 For a magma density of 3010 kg m^{-3} , Z_{crit} is 85.2 km for the nearside and 142.0 km for the
 847 farside. Thus if the tops of melt zones feeding dikes connected continuously to the surface had
 848 been at depths greater than a critical value in the range ~ 99 to ~ 140 km, eruptions should have
 849 taken place on the farside. If the vertical extent of the partial melting zone is increased to a finite
 850 value greater than zero, an excess pressure is generated at the dike inlet by the positive buoyancy
 851 of the magma below this level and it is possible for a dike to reach the surface from a source
 852 region having its top at a greater depth than the critical values given above.

853
 854 To illustrate this we first set the top of the mantle partial melting zone to be at a rheological
 855 boundary at an absolute depth of 50 km below the surface everywhere on the Moon, putting it at
 856 a depth of 20 km below the nearside crust, and at the exact base of the farside crust. We assign
 857 the same depth on both the near and far sides to the top of the diapir because the rheological
 858 boundary defining its location will probably be controlled more by the temperature than the
 859 pressure distribution in the lithosphere. We then explore the consequences of increasing the
 860 extent of the zone of partial melting, E , from zero to at least many tens of km based on the
 861 arguments in Section 2. Table 3 shows the results for a magma density of 2900 kg m^{-3} . First
 862 compare parts (a) and (b) of the table. For all vertical extents of the partial melt zone greater
 863 than ~ 5 km there is a great enough net buoyancy to ensure that eruptions must occur on the lunar
 864 nearside. In contrast, only if the vertical extent of partial melting within the diapir is greater than
 865 ~ 45 km is it possible for eruptions to occur on the farside; all smaller extents of melting lead to
 866 intrusions stalled at depths up to ~ 6 km below the surface. Thus a simple explanation for the
 867 paucity of mare basalt eruptions on the lunar farside is that the vertical extent of melting in very
 868 shallow mantle melt zones was less than 45 km. If the assumed depth to the top of the melt
 869 source zone is increased, for example to 60 km below the surface, eruptions now take place on
 870 the nearside for all extents of mantle melting, and the effective density contrast driving eruptions
 871 is somewhat increased because of the greater contribution from magma buoyancy in the mantle.

872 However, now eruptions can only occur on the farside if the vertical extent of the melting zone is
 873 at least ~35 km. Thus the range of melt-zone depths dictating the distinction between common
 874 eruptions on the nearside and rare eruptions on the farside decreases as the depth to the top of the
 875 melting zone increases. If the depth to the top of the melt source zone is increased further, to 70
 876 km, eruptions on the nearside again occur for all source extents whereas eruptions on the farside
 877 require source sizes greater than ~25 km. The other important trend show by Table 3 is that as
 878 both the depth and the vertical extent of the partial melt zone increase, the magma rise speeds (all
 879 turbulent) and the erupted volume fluxes also increase. Many of the largest values in the table
 880 are greater than any inferred in the literature for actual eruptions on the Moon. This strongly
 881 suggests that if partial melt zones existed in the shallow mantle in early lunar history, they did
 882 not have great vertical extents.

883 The above analysis was repeated for a magma density of 3260 kg m^{-3} . The trends (not
 884 shown) are the same as those in Table 3 but the greater magma density leads to some systematic
 885 differences. Greater vertical extents of partial melt zones are needed to ensure that eruptions
 886 occur, on both the near- and farsides. Intruded dike widths are less by a factor of 2 to 3, magma
 887 rise speeds (still turbulent) are smaller by a factor of ~2, and eruption volume fluxes are less by
 888 up to an order of magnitude than the values for the lower density melt. None of these differences
 889 change the major conclusion that shallow partial melt zones must not have had great vertical
 890 extents.

891

892 **3.F Dike intrusions and sills**

893 Examples of dikes from deep sources whose tops intrude a distance A_u into the crust are
 894 shown in Table 2. The allowed range of values of A_u , assuming that melt zone extents can range
 895 up to tens of km, is so large that the tops of those dikes that do not erupt could be located at any
 896 depth below the surface, in both the nearside and farside crusts. The widths of dikes stalling at a
 897 few km depth would be ~35 m on the nearside and ~50 m on the farside. In contrast, Table 2
 898 shows the properties of dikes from sources at shallow depths in the upper mantle, and indicates
 899 that the range of values of depths of dike tops, when eruptions do not occur, is much more
 900 restricted, especially on the lunar nearside. For partial melt zones with their tops at 50 km,
 901 intruded dikes should have widths up to ~40 m and have their tops at up to ~1 km below the
 902 surface.

903 Dikes intruding to shallow depths have the potential to induce surface graben if they
 904 generate stresses causing major fractures in the overlying crust (Head and Wilson, 1994; Petrycki
 905 et al., 2004; Klimczak, 2013). Petrycki et al. (2004) assessed the morphologies and measured the
 906 geometries of 248 lunar graben and found that 72 of them had associated minor volcanic
 907 features. The widths of these graben averaged 1.2 ± 0.6 km. Assuming that these graben were
 908 produced in response to the stresses associated with dike intrusion, the depths to the dike tops
 909 were inferred to lie in the range 0.5-1.6 km. An additional 176 graben not having easily
 910 recognized volcanic associations averaged 1.8 ± 0.8 km in width, implying possible dike top
 911 depths of 0.9-2.3 km. The depths of the graben with associated volcanic features were
 912 systematically greater than the depths of those without such associations, and the difference was
 913 inferred to imply a mean dike width of ~50 m, with a few examples implying greater widths in
 914 excess of ~150 m (Head and Wilson, 1994). The results presented here seem entirely consistent
 915 with these inferences.

916 The low mean flux of lunar magma (Head and Wilson, 1992a), the small percentage of the
 917 lunar crust formed of mare basaltic magma (Head, 1976), and the consequent infrequency of dike

918 emplacement events, all conspire to limit large shallow magma reservoirs and large Hawaii-like
 919 shield volcanoes on the Moon (Head and Wilson, 1991). Repeated intrusions of dikes over more
 920 extended geologic time, however, will have increased the bulk density of the crust, somewhat
 921 reducing the negative buoyancy of magmas. Intrusions will also have reinforced the trend,
 922 controlled by global cooling (Solomon and Head, 1980), of increasing compressive stress in the
 923 crust with time. This in turn will have led to the least compressive stress becoming vertical, thus
 924 favoring the formation of sills if opportunities exist. The clear example of such opportunities is
 925 present in the form of the breccia lenses beneath impact craters. Several authors have proposed
 926 magma injection as a possible origin of distinctive impact craters in which the floor is uplifted
 927 and fractured (Brennan, 1975; Schultz, 1976b; Wichman and Schultz, 1995; Jozwiak et al., 2012,
 928 2015) (Figure 8). The diameters of craters with floors modified in this way range from ~10 to
 929 ~300 km, so that the breccia lenses beneath them may have extended to depths of order two to a
 930 few tens of kilometers.

931 Tables 2 and 3 show that there are a wide range of circumstances that can lead to dikes
 932 ceasing to propagate upward when their tops reach depths of ~2-4 km. If one of these dikes
 933 encounters a breccia lens before it has reached its equilibrium height, it will initially invade the
 934 fractures between crustal blocks. This branching of the magma transport system will lead to
 935 magma cooling and reduce the chances of continued magma rise. Instead, the low density of the
 936 crustal material relative to the magma will create a tendency for a sill to form at the base of the
 937 brecciated zone. This sill will in principle be inflated until the top of the sill lies at the level that
 938 the magma would have reached if the crater were not present, so that sill thicknesses, and extents
 939 of crater floor uplift, may be at least of order a few km. Jozwiak et al. (2012, 2015) measured
 940 uplifts of up to 2 km in the small sample of floor-fractured craters that they examined in detail.
 941 The progressive enlargement of the growing intrusive body at the base of the breccia lens must
 942 have much in common with the growth of a laccolith as modeled by Michaut (2011), implying
 943 that small floor-fractured craters might display a domical uplift, largest in the crater center,
 944 whereas the largest diameter craters should have intrusions of more nearly uniform thickness and
 945 flatter floors (Jozwiak et al., 2012, 2015).

946 A second potentially attractive location for sills to form is at the density discontinuity at the
 947 base of the crust. However, two criteria must be satisfied for such intrusions to form when dikes
 948 arrive at the density boundary: the state of stress in the lithosphere must be such that the least
 949 principle stress is vertical and the excess pressure at the upper tip of the dike must be greater than
 950 the weight of the overlying crust. Favorable circumstances for this configuration would include
 951 (i) a dike generated by a very vertically extensive partial melt zone in the mantle, (ii) the dike
 952 arriving under a part of the crust that had been thinned by a basin-forming event, or (iii) the
 953 event taking place in the second half of lunar history when planetary cooling had induced a
 954 global horizontal compressive stress in the lithosphere. The weight of a 30 km thick crust
 955 thinned by a 3 km deep basin is ~112 MPa. Table 1 gives the excess pressures at the base, P_d ,
 956 and the middle, P_0 , of dikes from deep sources. The excess pressure at the top of such a dike is $(2$
 957 $P_0 - P_d)$ and the table therefore implies that a dike from a source of vertical extent ~46 km or
 958 greater would have the potential to form a sill as its upper tip arrived at the base of the crust. The
 959 magma volume involved in a single event could be as great as 5000 km³, forming, for example, a
 960 ~40 km radius sill of thickness 1 km. Without more detailed models of mantle melting and
 961 better information on the history of the stress state of the lithosphere it is hard to comment on the
 962 likely frequency of such events.

963 We now turn to the consequences of dikes breaching the surface to cause eruptions. Since
 964 all lunar magmas contained some, albeit small, amounts of volatiles and were erupted into what
 965 is essentially the interplanetary vacuum, some element of explosive activity should always have
 966 been involved. In such cases we need to distinguish three phases: a first phase when the dike is
 967 in the process of growing from its source but has not yet reached the surface; a second phase
 968 when the dike tip has broken through the surface but the pressure distribution in the dike magma
 969 has not yet reached an equilibrium configuration; and a final phase when the pressure
 970 distribution has stabilized to one that maximizes the magma discharge rate. The amounts and the
 971 release conditions of volatiles and the consequent styles of explosive activity can be very
 972 different in these three phases. They can also differ significantly from the consequences of the
 973 accumulation of gas at the shallow top of a dike that has initially failed to erupt at the surface, or
 974 at the top of a shallow sill growing from such a dike. In these cases both explosive eruptions of
 975 juvenile magma and simple surface collapse due to gas release can occur. We first consider the
 976 transient processes associated with dike propagation to the surface.

978 **4. Transient eruptions associated with dike emplacement**

979 **4.A Conditions in a dike propagating toward the surface**

980 While a dike is still propagating, the pressure distribution within it adjusts to maximize the
 981 flow rate (Lister and Kerr, 1991; Rubin, 1993, 1995; Detournay et al., 2003). The pressure at the
 982 base of the dike is fixed by the pressure in the magma source zone, and so the pressure at the
 983 propagating upper tip must decrease to the lowest possible value. Lister and Kerr (1991) and
 984 Rubin (1993) suggested that this minimum value should be the pressure at which the most
 985 soluble magmatic volatile present (commonly water on Earth) is just saturated, but did not
 986 consider the dynamics of volatile exsolution into small gas bubbles and the growth and eventual
 987 bursting of these bubbles to transfer free gas into the narrow, elongate tip cavity. Wilson and
 988 Head (2003a) suggested that beneath the gas-filled tip cavity there must be a zone of magmatic
 989 foam, with the pressure at the base of the foam layer being the saturation pressure at which gas
 990 exsolution starts.

991 The pressure at the base of the foam layer is controlled by the first appearance of exsolved
 992 magmatic volatiles. Hauri et al. (2011, 2015) showed that at least some lunar magmas may have
 993 contained small amounts (up to 1000 ppm) of water, with small amounts of sulfur and chlorine
 994 also present. The solubility of water in terrestrial mafic magmas can be approximated by $n_d =$
 995 $6.8 \times 10^{-8} P^{0.7}$ where n_d is given as a mass fraction and the ambient pressure P is expressed in
 996 Pascals (Dixon, 1997), so 1000 ppm water would saturate at a pressure of 0.9 MPa; detailed
 997 solubility data for S and Cl are not available but their vapor pressures in lunar magmas are
 998 expected to be much less than 0.1 MPa (Sato, 1976). However, pressure at the onset of gas
 999 release in a lunar dike is expected to be much greater than any of these values. This is because
 1000 the dominant lunar volatile is expected to be a mixture of CO and CO₂ (with CO comprising
 1001 ~90% of the mixture) produced in amounts up to ~2000 ppm by a smelting reaction between
 1002 elemental carbon (graphite) and various metal oxides (Sato, 1976; Fogel and Rutherford, 1995;
 1003 Nicholis & Rutherford, 2006; Wetzell et al., 2015). Gas production will begin when the ambient
 1004 pressure in the magma decreases below $P_{sm} = \sim 40$ MPa (Fogel and Rutherford, 1995) and we
 1005 take this to be the pressure at the base of the foam layer. Nicholis and Rutherford (2006) show
 1006 that the smelting reaction proceeds very rapidly with decreasing pressure, at a rate of ~0.43 MPa
 1007 change per 1000 ppm of CO produced for typical lunar magmas. Thus the production of ~2000
 1008 ppm of CO-dominated gas will be complete when the pressure has decreased from ~40 MPa to

1009 ~39 MPa. The vertical extent, ΔZ , of the foam layer, over which the pressure decreases from the
 1010 ~40 MPa level of onset of CO production at the base to ~0.5 MPa at the top, can be estimated
 1011 from the fact that the average pressure gradient, dP/dz , in the magma as the dike top nears the
 1012 surface must be approximately equal to the average gradient of the lithostatic load in the host
 1013 rocks, i.e.

$$1014 \quad dP/dz = \rho_c g \quad (27)$$

1015
 1016 A pressure decrease of $(40 - 0.5 =) 39.5$ MPa then implies that the foam layer extends vertically
 1017 for $\Delta Z = \sim 9.6$ km.

1018 The pressure at the top of the foam layer, P_i , marking the interface between the foam and
 1019 free gas, will be controlled by the mechanism determining the stability of the foam. The pressure
 1020 at the point of foam disruption can be found from either a critical gas bubble volume fraction
 1021 criterion (Sparks, 1978) or a critical strain rate criterion (Papale, 1999). Rutherford and Papale
 1022 (2009) found that, for mafic magmas, adopting the strain rate criterion did not predict eruption
 1023 conditions very different from the bubble volume fraction criterion. We therefore adopt the
 1024 simplest possible criterion, that the pressure at the interface between foam and free gas is the
 1025 pressure at which the gas bubble volume fraction reaches a critical value at which the foam
 1026 becomes unstable. Jaupart and Vergnolle (1989) suggest a critical value of 0.85, a little larger
 1027 than the ~0.75 value suggested by Sparks (1978). Approximating the gas properties by the
 1028 perfect gas law, the partial volumes of gas, v_g , and liquid, v_l , in the foam are

$$1029 \quad v_g = (n Q_u T_m) / (m P) \quad (28)$$

$$1030 \quad v_l = (1 - n) / \rho_l \quad (29)$$

1031 Here m is the molecular mass of the volatile, n the mass fraction of the volatile exsolved from the
 1032 magma, Q_u the universal gas constant ($8.314 \text{ kJ kmol}^{-1} \text{ K}^{-1}$), T_m the (assumed constant)
 1033 temperature of the magma and ρ the density of the magmatic liquid. Thus the criterion $[v_g / (v_g +$
 1034 $v_l)] = 0.85$ implies that the magma disruption pressure P_i is given by

$$1035 \quad P_i = (0.15 n Q_u T \rho_l) / [0.85 (1 - n) m] \quad (30)$$

1036 Assuming a 90% CO, 10% CO₂ mixture with $m = 29.6 \text{ kg kmol}^{-1}$, $T_m = 1623 \text{ K}$ (i.e. the 1350 °C
 1037 liquidus of the Apollo 17 orange glass magma) and $\rho_l = 2900 \text{ kg m}^{-3}$, Figure 9 shows how P_i
 1038 varies with the total amount of released gas, n . This pressure must also be the pressure in the gas
 1039 in the tip cavity above the interface. Values are less than 1 MPa for likely volatile amounts.

1040 The vertical extent of the free gas cavity cannot be found analytically because, as discussed
 1041 by Lister and Kerr (1991), Rubin (1993, 1995) and Detournay et al. (2003), it depends on the
 1042 detailed motion of the magma in the region between the onset of gas generation and bubble
 1043 bursting. For dikes in the Earth's crust estimates are of order hundreds of meters (Lister and
 1044 Kerr, 1991). We approach the problem as follows. Magma rising through the foam layer toward
 1045 the dike tip moves most quickly along the center-line of the dike and migrates to the dike walls
 1046 where it stagnates (a no-slip boundary condition requires the magma speed to be zero at the
 1047 wall). If gas bubbles nucleate at the base of the foam layer with diameters of ~20 microns
 1048 (Sparks, 1978; Larsen and Gardner, 2004; Yamada et al., 2005, 2008; Bai et al., 2008) and
 1049

1055 decompress from the 40 MPa pressure at nucleation to the gas pressure in the dike tip as they are
 1056 carried up by the magma flow, they will have expanded isothermally to the sizes shown in Figure
 1057 9, of order 100 microns. The rise speed of bubbles of these sizes in the closely packed foam will
 1058 be very small. The bulk viscosity of the foam will be much greater than that of the liquid magma
 1059 alone. Jaupart and Vergnolle (1989) suggest that the effective viscosity increase is by a factor
 1060 of $(1 - \varepsilon)^{5/2}$ where ε is the bubble volume fraction, 0.85; with a melt viscosity of 1 Pa s this
 1061 suggests a foam viscosity of 115 Pa s. Equating the buoyancy to the viscous drag shows that the
 1062 rise speed of a 100 micron diameter bubble in lunar magma will be $\sim 30 \text{ nm s}^{-1}$. During the $5 \times$
 1063 10^3 to 5×10^4 s needed for a magma to rise from a depth of 50 to 500 km at $\sim 10 \text{ m s}^{-1}$, bubbles at
 1064 the top of the foam will have migrated at most 50 to 1500 microns, no more than 15 bubble
 1065 diameters. Thus gas addition to the tip cavity by this mechanism is minimal.

1066 More important will be the shearing of gas bubbles as magma approaches the dike walls.
 1067 If we assume that a single layer of bubbles collapses and delivers gas to the cavity as it migrates
 1068 to the wall, we can track numerically the amount of gas delivered as the dike tip propagates
 1069 upward by multiplying the current width of the gas/foam interface, initially assumed to be
 1070 vanishingly small, by the ~ 100 micron diameter of the bubbles. The detailed shape of the cavity
 1071 depends on the stress distribution around the dike tip. The width/length aspect ratios of dikes are
 1072 essentially equal to the ratio of the shear modulus of the host rocks ($\sim 4 \text{ GPa}$, Rubin, 1995) to the
 1073 dike inlet pressure. In the case of the dikes illustrated in Tables 1 and 3 the ratio would be $\sim (4$
 1074 $\text{GPa}/10 \text{ MPa})$, i.e. 400, and using this value implies the gas cavity heights and widths shown in
 1075 Figure 10 for a range of dike source depths encompassing mare basalts and deep-sourced
 1076 picrites. If multiple layers of bubbles shear at the wall the values in Figure 10 would increase.
 1077 The relationship involves the square root of the number of bubble layers; thus if 100 layers of
 1078 bubbles collapsed near the dike wall the heights and widths in Figure 10 would increase by a
 1079 factor of 10. Thus it seems likely that the vertical extents of dike tip gas cavities associated with
 1080 the eruptions of mare basalts will have been a few tens to a few hundreds of meters. Deep-
 1081 sourced picritic dikes may have had gas cavities extending for as much as 1 to 2 km.

1083 **4.B Transient eruption phenomena as dikes first breach the surface**

1084 **4.B.1 Release of gas from the dike tip cavity**

1085 The first consequence of a dike breaking through to the surface will be the very rapid
 1086 release of the gas that has accumulated in the cavity in the dike tip. Given the likely vertical
 1087 length of the cavity, at least tens to hundreds of meters, this gas should contain almost no
 1088 entrained magma droplets from the disrupting gas-magma interface below it. The gas may,
 1089 however, entrain regolith clasts as it emerges, and may also have entrained rock fragments from
 1090 the walls of the dike. The latter is likely because the decompression of the gas causes inward
 1091 stresses across the dike walls that may exceed the tensile strength of the crustal rocks if the gas
 1092 cavity is more than several hundred meters deep. We therefore define the gas to represent a mass
 1093 fraction N of the expelled gas-clast mixture and expect N to range from 1.0 (no entrained clasts)
 1094 to perhaps 0.5 if a great deal of dike wall disruption occurs. The gas will probably be at a
 1095 temperature close to that of the magma from which it has been released, though heat loss to the
 1096 cavity walls may be non-trivial if the cavity is very long. The ultimate velocity U_u reached by a
 1097 parcel of gas in expanding into the vacuum above the lunar surface from a depth z and pressure
 1098 P_i is given by

$$1100 \quad 0.5 U_u^2 = [\gamma / (\gamma - 1)] [(N Q_u T_m) / m] + [(1 - N) / \rho_l] P_i - g z \quad (31)$$

1101 where γ is the ratio of the specific heats of the gas at constant pressure and constant volume, very
 1102 close to 1.3 for dominantly CO, and it is assumed that the gas receives no additional heat from
 1103 the underlying magma during its very rapid expansion. Insertion of $P_i = \sim 0.5$ MPa, the value
 1104 found in Section 4.A for a magma producing 2000 ppm of CO, and values of z as large as 10 km,
 1105 we find that the last two terms in equation (31) are very small compared with the first term, and
 1106 using $T_m = 1623$ K for a picritic melt we have $U_u = 2.0$ km s⁻¹ when $N = 1.0$ and $U_u = 1.4$ km s⁻¹
 1107 when $N = 0.5$. These values are less than the 2.38 km s⁻¹ escape velocity from the Moon, but
 1108 lead to extremely wide dispersal of all ejected clasts small enough to acquire an appreciable
 1109 fraction of the gas speed. The distance D_u measured along the surface of a planet radius R
 1110 traveled by a clast ejected at speed U_u and at an elevation angle from the horizontal θ is
 1111

$$1112 \quad D_u = 2 R \tan^{-1}[(U_u^2 \sin \theta \cos \theta)/(R g - U_u^2 \cos^2 \theta)] \quad (32)$$

1113 The maximum distance is not in general given by $\theta = 45^\circ$, the maximum range on a horizontal
 1114 surface, and is most simply found by trial and error. For $U_u = 1.4$ km s⁻¹ the maximum travel
 1115 distance is ~ 1950 km when $\theta = \sim 30^\circ$, and for $U_u = 2$ km s⁻¹ it is ~ 5210 km (almost to the middle
 1116 of the opposite hemisphere) when $\theta = \sim 35^\circ$. Thus while this kind of event might qualify as the
 1117 lunar equivalent of a terrestrial ultraplinian eruption, it would produce a deposit of very limited
 1118 volume that was so widespread that it would almost certainly never be recognizable. The
 1119 duration of such an event would be determined by the passage of an expansion wave through the
 1120 trapped gas; with a typical wave speed of order half of the 765 m s⁻¹ speed of sound in CO at
 1121 magmatic temperature, gas cavities with lengths of 200 m and 2 km would be emptied on time
 1122 scales of 0.25 and 2.5 seconds, respectively. A likely consequence of this gas release process
 1123 would be the severe disturbance of the fine-grained regolith in the immediate vicinity of the vent.
 1124
 1125

1126 **4.B.2 Release of gas and magma from the foam beneath the dike tip**

1127 After all of the gas trapped in the dike tip cavity has been released, the expansion wave
 1128 continues to propagate, now into the underlying foam. A simplifying characteristic of the foam
 1129 layer is that, although the pressure within it will increase with depth, all of the magma within it
 1130 will have passed through the ~ 40 MPa pressure level at which the smelting reaction occurs and
 1131 so will contain the same amount of released dominantly CO gas. Section 4.A showed that the
 1132 most likely vertical extent of the foam layer is $\Delta Z = 9.6$ km. Section 4.B.1 provided a range of
 1133 estimates of the vertical length of the gas cavity from which 300 m can be selected as typical.
 1134 Thus a plausible scenario is one where the pressure P_{foam} in the foam layer varies from $P_i = 0.5$
 1135 MPa at 0.3 km depth to $P_{\text{sm}} = \sim 40$ MPa at 9.6 km depth. As the expansion wave passes down the
 1136 foam layer the foam disaggregates into a mixture of gas and pyroclasts which expands to a
 1137 pressure P_f at which the clasts and gas decouple as the system reaches the Knudsen regime where
 1138 the mean free path of the gas molecules exceeds the typical pyroclast size, d . The pressure at
 1139 which this takes place is given by Wilson and Keil (2010) as
 1140

$$1141 \quad P_f = (2^{1/2} Q_u T)/(3 \pi \phi^2 N_a d) \quad (33)$$

1142 where ϕ is the effective diameter of the gas molecules, 3.4×10^{-10} m for CO, and N_a is
 1143 Avogadro's number, 6.0225×10^{26} kmol⁻¹. For typical $d = 300$ μm sized pyroclasts, P_f is ~ 90 Pa.
 1144
 1145

1146 The speed U_m reached by the mixture of gas and pyroclasts as it decompresses from the pressure
 1147 P_{foam} at depth z to its final pressure $P_f = 90$ Pa, is

$$1148 \quad 0.5 U_m^2 = [(n Q_u T)/m] \ln(P_{\text{foam}}/P_f) + [(1 - n)/\rho_1] (P_{\text{foam}} - P_f) - g z \quad (34)$$

1150
 1151 Table 4 then shows how the eruption speed U_m changes as the foam layer is progressively
 1152 erupted to the surface and also gives the corresponding maximum pyroclast ranges R_m . The
 1153 maximum range increases from ~ 6 to ~ 10 km as the foam is discharged. Adding the effects of
 1154 the exsolution of 1000 ppm H_2O from a very water-rich lunar magma would approximately
 1155 double these dispersal distances to ~ 12 to ~ 20 km.

1156 **4.C Dikes that approach the surface but do not erupt large magma volumes**

1157 An upward-propagating dike may fail to reach the surface for a number of reasons, e.g.,
 1158 insufficient magma volume and pressure in the source region or inappropriate combinations of
 1159 lithosphere and magma density. If such a dike has a small width or stops with its upper tip
 1160 sufficiently far below the surface, there will never be any surface indication of its presence
 1161 (Figure 8), though it might be detectable by geophysical techniques and it will contribute to
 1162 increasing the mean density of the crust - see calculations in Head and Wilson (1992a).
 1163 However, if the dikes induce stresses that allow fractures to form between the dike tip and the
 1164 surface, graben formation is possible (Figure 8), as discussed in Section 3.F, and minor eruption
 1165 of juvenile material or surface collapse due to gas release may take place, on both short and long
 1166 time scales.

1167 **4.C.1 Short-term effects of near-surface dikes**

1168 As the top of a dike approaches its final configuration (Figure 8), the component of the
 1169 vertical pressure gradient driving magma upward must decrease smoothly to zero. This implies
 1170 that the absolute pressures in the gas in the dike tip cavity and in the underlying foam layer will
 1171 be significantly greater than their values during most of the vertical rise of the dike. If the
 1172 pressure in the gas becomes greater than ~ 10 MPa, the stresses on the crustal rocks overlying the
 1173 dike top may exceed their tensile strength and fractures may open to the surface allowing the gas
 1174 to vent. Given that the pressure prior to this adjustment was probably ~ 0.5 MPa, an
 1175 approximately $(10/0.5 =) 20$ -fold compression of the gas would occur during the build-up to this
 1176 gas release, reducing the vertical extent of the gas cavity by a similar factor and bringing the
 1177 underlying foam layer closer to the surface. There is clearly the potential for the propagation of
 1178 an expansion wave into the foam, causing a restricted but energetic pyroclastic eruption through
 1179 the crustal fractures until the foam is exhausted. All of the magma immediately beneath the
 1180 foam layer will have passed through the critical 40 MPa pressure level during the emplacement
 1181 of the dike and so will have produced all of the CO that it is capable of producing by the
 1182 smelting reaction, but if the fractures to the surface remain open, exposure of the magma at the
 1183 top of the melt column to the essentially zero pressure above the surface may cause some
 1184 exsolution of dissolved species like water and sulphur, increasing the total volatile budget. A
 1185 calculation using equation (34) but now allowing for the compaction of the foam layer and hence
 1186 the reduction in z implies ranges for CO-dominated but also water-rich magma pyroclasts of up
 1187 to ~ 25 km.

1188 **4.C.2 Longer-term effects of near-surface dikes**

1192 For the widest dikes likely to be emplaced with their tops near the surface, formation of
 1193 collapse craters, as well as graben, is a possible consequence of gas accumulation and subsequent
 1194 venting to the surface. However, features like these are rare on the Moon and only two cases
 1195 have been described and analyzed. Head et al. (2002) showed that an ~75 km radius dark
 1196 pyroclastic deposit superposed on the southern part of the Orientale basin interior could have
 1197 been the consequence of gas accumulation at the top of an unusually wide (~500 m) dike with
 1198 eventual surface collapse producing a 7.5 km by 16 km depression. Wilson et al. (2011) found
 1199 that the ~30 km radius pyroclastic deposit surrounding the Hyginus crater complex could be
 1200 explained by smaller amounts of gas accumulation at the top of a ~240 m wide dike that also
 1201 intruded a small sill at shallow depth and produced Hyginus crater and the graben and associated
 1202 minor collapse pits of Rima Hyginus. In both of these cases, convective overturn of the magma
 1203 in the cooling dike was invoked to enhance the amount of gas accumulation at the top of the
 1204 intrusion. Clearly this process is only likely to have been important for a few unusually wide
 1205 dikes where the long dike cooling time allows for many cycles of convective overturn and
 1206 upward gas segregation.

1207

1208 **5. Steady eruptions from dikes erupting at the surface**

1209 Section 3 provided estimates of magma rise speeds up to tens of m s^{-1} and erupted volume
 1210 fluxes mainly in the range 10^4 to $10^6 \text{ m}^3 \text{ s}^{-1}$ through dikes feeding surface eruptions, and Section
 1211 4 introduced the concept of volatile release from lunar magmas, implying that explosive activity
 1212 was common on the Moon. We now show how the resulting pyroclastic deposits are related to
 1213 the fire fountains produced by steady explosive eruptions.

1214

1215 **5.A Near-surface processes of gas release in steady explosive eruptions**

1216 Local and regional dark mantle deposits (DMD) interpreted to be of pyroclastic origin have
 1217 long been recognized on the Moon (Wilhelms and McCauley, 1971), generally in association
 1218 with irregular depressions and sinuous rilles (Head, 1974; Head and Wilson, 1980; Wilson and
 1219 Head, 1980; Gaddis et al., 1985, 2003; Hawke et al., 1989; Weitz et al., 1998; Weitz and Head,
 1220 1999; Gustafson et al., 2012). Regional deposits are the most extensive (equivalent radii mostly
 1221 in the range 20–40 km) and are commonly located on uplands adjacent to younger mare deposits
 1222 (e.g., Head, 1974; Weitz et al., 1998; Gaddis et al., 2003). Localized deposits, in contrast, are
 1223 smaller in extent and more widely distributed across the lunar surface (Head 1976; Hawke et al.,
 1224 1989; Coombs et al., 1990; Gaddis et al., 2003), and we consider these first.

1225 Shortly after the upper tip of a dike breaches the surface, the pressure distribution with
 1226 depth in the magma develops into whatever pattern maximizes the volume flux flowing through
 1227 the system. This adjustment takes place via the passage of pressure waves through the dike
 1228 system at speeds comparable to the speed of sound in the magma, $\sim 1 \text{ km s}^{-1}$. The time needed
 1229 for such a wave to propagate from a magma source at 400 km depth in the mantle would be 400 s
 1230 and so even if the adjustment required the passage of waves back and forth between the source
 1231 and surface several times the process would be complete in less than an hour. This is a very
 1232 small fraction of the durations that we shall infer later for many lunar eruptions.

1233 Numerous models of the key aspects of explosive volcanic processes on the terrestrial
 1234 planets have been published (McGetchin and Ulrich, 1973; Wilson, 1980; Wilson and Head,
 1235 1981; Valentine and Wohletz, 1989; Giberti and Wilson, 1990; Dobran et al., 1993; Papale and
 1236 Dobran, 1993; Wilson and Keil, 1997; Kaminski and Jaupart, 1998; Neri et al., 1998, 2003;
 1237 Papale et al., 1999; Wilson, 1999; Cataldo et al., 2002; Wilson and Head, 2001, 2003; Mitchell,

1238 2005; Wilson and Head, 2007b; Rutherford and Papale, 2009). In the simplest possible scenario,
 1239 the equilibrium pressure in the magma emerging through the surface vent is equal to the local
 1240 atmospheric pressure. However, when the atmospheric pressure is essentially zero, as on the
 1241 Moon, this implies an infinitely wide vent, clearly physically impossible. In practice, the
 1242 presence of even extremely small amounts of volatiles intervenes to dictate a finite pressure in
 1243 the vent. Volatiles dissolved in the magma, or produced by pressure-dependent chemical
 1244 processes, will be released as the magma ascends and the pressure decreases toward the surface
 1245 in the shallow part of the conduit system. Whatever the source, the volatiles form gas bubbles in
 1246 the liquid melt, and expansion of the bubbles as the magma rises and the pressure decreases
 1247 accelerates the magma. The bubble volume fraction may become large enough that, combined
 1248 with the strain rates to which the liquid-bubble foam is subjected, the liquid is disrupted into a
 1249 free gas phase entraining pyroclasts. However, such disruption does not necessarily take place
 1250 below the surface in all cases on the Moon (*Rutherford and Papale, 2009*); when it does not
 1251 occur, magma disruption must occur immediately above the vent at the base of the system of
 1252 shocks that decompresses the gas phase into the vacuum.

1253 The speed of sound in a 2-phase (3-phase if crystals are also present) fluid, whether liquid
 1254 plus bubbles or gas plus pyroclasts, is much less than the speed of sound in a single-phase liquid
 1255 (Kieffer, 1977). Thus as magma approaches the surface it is possible for the steadily increasing
 1256 magma rise speed to become equal to the rapidly decreasing sound speed. If this condition is
 1257 reached in a parallel-sided or converging conduit system, there can be little further acceleration;
 1258 the magma speed stays equal to the sound speed and the system is said to be choked. There may
 1259 in fact be some change in speed, because if the pressure decreases and more volatiles exsolve,
 1260 the sound speed will change and hence so will the flow speed; however, the Mach number must
 1261 stay equal to unity. As shown by Giberti and Wilson (1990) and Mitchell (2005), the total mass
 1262 flux through the volcanic system is maximized when the sonic condition is reached exactly at the
 1263 surface vent, and it is likely that the system will rapidly adjust to this condition. Decompression
 1264 to atmospheric pressure, zero pressure in the case of the Moon, and acceleration to supersonic
 1265 speeds, is then accomplished immediately above the vent through a system of shocks (Kieffer,
 1266 1982, 1989).

1267 The only way that a subsonic to supersonic transition can occur beneath the surface is for
 1268 the conduit system to flare outward toward the surface by more than a critical amount. The
 1269 combinations of conduit shapes and volatile contents of both mafic and silicic magmas on Earth
 1270 ensure that both choked flows, where the vent pressure is greater than atmospheric, and
 1271 supersonic flows, where the vent pressure is equal to the atmospheric pressure, may occur in
 1272 eruptions. Wilson and Head (1981) showed that to ensure that the vent pressure can be equal to
 1273 the atmospheric pressure in mafic eruptions on Earth it is typically necessary for the conduit to
 1274 increase in width by a factor of 2 to 3 over the uppermost ~100 m of the conduit system. The
 1275 equivalent expansion factor for the Moon, where the atmospheric pressure is essentially zero,
 1276 was shown to be in the range 10 to 30. It is not likely that the stresses inducing dike propagation,
 1277 even if there were large near-surface tensile stresses in the lithosphere, would lead to dikes with
 1278 these near-surface shapes. Thus whereas some but not all explosive eruptions on Earth may be
 1279 choked, all explosive eruptions on the Moon are expected to be choked. This is true even though
 1280 lunar magma volatile amounts are much less than typical terrestrial values.

1281 The vent pressure implied by imposing choked flow can be found by evaluating the rise
 1282 speed, U , of the magma and the sound speed, S , within it as a function of pressure, P . This is
 1283 particularly straightforward for lunar magmas where the dominant volatile is the CO-dominated

1284 mixture generated by smelting (Fogel and Rutherford, 1995), because formation of gas bubbles
 1285 in the magma will have been completed very quickly as the pressure in the magma fell below 40
 1286 MPa (Nicholis and Rutherford, 2005), in contrast to conditions in terrestrial magmas where
 1287 pressure-dependent gas release will in general still be ongoing. Wilson and Head (1981) showed
 1288 that for the ranges of pressures and volatile contents relevant to volcanic systems the formula
 1289

$$1290 \quad S = P [m / (n Q_u T_m)]^{1/2} \{[(n Q_u T_m) / (m P)] + [(1 - n) / \rho_l]\} \quad (35)$$

1291
 1292 gives values of the sound speed within a few percent of those obtained from more complex
 1293 treatments (e.g. Hsieh and Plesset, 1961; Soo, 1961, 1967; Kliegel, 1963; Rudinger, 1964; Cole
 1294 et al., 1970; Kieffer, 1977; Pai et al., 1978). To illustrate likely values for mafic melts we adopt
 1295 a density of 2900 kg m⁻³ and a temperature of 1500 K. On Earth mafic magma volatiles are
 1296 dominated by H₂O ($m = 18 \text{ kg kmol}^{-1}$) and CO₂ ($m = 44 \text{ kg kmol}^{-1}$) in roughly equal proportions
 1297 whereas lunar magmas mainly produced a 90% CO, 10% CO₂ mixture with $m = 29.6$ via a
 1298 smelting reaction (Fogel and Rutherford, 1995) and exsolved smaller amounts of H₂O (Hauri et
 1299 al., 2011) and traces of S₂ ($m = 64 \text{ kg kmol}^{-1}$) (Saal et al., 2008). In both cases, therefore, a value
 1300 of $m = \sim 30 \text{ kg kmol}^{-1}$ seems an adequate approximation for comparisons.

1301 The variation with pressure of the eruption speed though the vent in an explosive eruption
 1302 is not easy to evaluate, choked or otherwise. When significant amounts of volatiles are exsolved
 1303 it is generally the case (e.g. see examples in Wilson and Head, 1981) that the magma rise speed
 1304 before exsolution starts is very much less than the eruption speed, and so can be neglected. Also,
 1305 the motion after the onset of volatile release but prior to magma disruption into pyroclasts is
 1306 limited by friction between the magmatic liquid and the conduit walls and so the speed increase
 1307 is not large. It is after magma disruption into a continuous gas phase with entrained pyroclasts
 1308 that most of the acceleration occurs. If the pressure at the point of disruption is P_{dis} , and the
 1309 magma accelerates to reach speed U when the pressure is some smaller value P , then to a good
 1310 approximation (Wilson, 1980)

$$1311 \quad 0.5 U^2 = [(n Q_u T) / m] \ln(P_{\text{dis}} / P) + [(1 - n) / \rho_l] (P_{\text{dis}} - P) - g Z_{\text{diff}} \quad (36)$$

1312
 1313 where Z_{diff} is the distance over which the pressure change occurs, estimated by assuming that the
 1314 pressure in the erupting magma is close to lithostatic, so that $Z_{\text{diff}} = (P_{\text{dis}} - P) / (g \rho_c)$. Two
 1315 energy terms have been neglected here, the initial kinetic energy of the rising magma before the
 1316 smelting reaction begins and the work done against wall friction; detailed numerical simulations
 1317 like those in Wilson and Head (1981) show that both of these terms are small compared with the
 1318 terms listed in equation (36) and, being of opposite signs, they partially compensate for one
 1319 another.
 1320

1321 The pressure, P_{dis} , at which magma disruption takes place is controlled by the same foam
 1322 stability criterion discussed in the previous section, and so P_{dis} is the equivalent in a steady
 1323 eruption of P_i given by equation (30). Substitution of equation (30) into equation (36) allows U
 1324 to be found as a function of P , and equation (35) gives S as a function of P . Thus the pressure
 1325 P_{ch} at which the choked sonic condition $U = S$ exists at the vent will be the pressure
 1326 simultaneously satisfying the expressions for U and S from equations (35) and (36), given by
 1327

$$1328 \quad [(n Q_u T_m) / m] \ln(P_{\text{dis}} / P_{\text{ch}}) + [(1 - n) / \rho_l] (P_{\text{dis}} - P_{\text{ch}}) =$$

$$1329 \quad P_{\text{ch}}^2 [m / (2 n Q_u T_m)] \{[(n Q_u T_m) / (m P_{\text{ch}})] + [(1 - n) / \rho_l]\}^2 \quad (37)$$

1330

1331 This equation involves P_{ch} in both logarithmic and algebraic terms, so there is no analytical
 1332 solution and the value of P_{ch} must be found by an iterative method. For equations of this kind
 1333 we can calculate a new approximation to P , P_{new} , from an older approximation, P_{old} , using

1334

$$1335 \quad [(n Q_u T_m) / m] \ln(P_{\text{dis}} / P_{\text{new}}) + [(1 - n) / \rho_l] (P_{\text{dis}} - P_{\text{old}}) =$$

$$1336 \quad P_{\text{old}}^2 [m / (2 n Q_u T_m)] \{ [(n Q_u T_m) / (m P_{\text{old}})] + [(1 - n) / \rho_l] \}^2 \quad (38)$$

1337

1338 which can, of course, be solved analytically; if the value $(0.5 P_{\text{dis}})$ is used as the first value of P_{old}
 1339 the solution converges to better than 1% after 4 iterations and better than 1 part in 10^6 after 12
 1340 iterations.

1341

1342 Figure 11 gives the values of P_{dis} and P_{ch} found in this way for a wide range of value of n
 1343 for $m = 30 \text{ kg kmol}^{-1}$. In all cases P_{dis} is greater than P_{ch} , i.e. magma is disrupted into pyroclasts
 1344 before emerging from the vent. For the Moon, with total released amounts of CO probably in the
 1345 range 250-2000 ppm (Fogel and Rutherford, 1995; Saal et al., 2008) and up to ~1000 ppm H₂O
 1346 (Hauri et al., 2011), P_{ch} is predicted to be in the range 0.04-0.4 MPa. For comparison on Earth,
 1347 mafic magmas commonly exsolve from 0.2 to 1.0 mass % total volatiles (Wallace and Anderson,
 1348 2000) implying that P_{ch} is at least ~1 MPa for eruptions where the vent shape does not become
 1349 wide enough to prevent choking. For the Moon, a key issue is what these combinations of P_{ch}
 1350 and n imply about the ranges of pyroclasts in steady eruptions. As discussed by Wilson and
 1351 Head (1981), the dispersion of pyroclasts into a vacuum is determined mainly by the speed with
 1352 which the pyroclasts emerge from the vent and in part by the size distribution of the liquid
 1353 droplets into which the magma is disrupted. The pyroclastic glass beads returned by the Apollo
 1354 missions generally have sizes in the 100-1000 micrometer range (Weitz et al., 1998); if these are
 1355 typical of all lunar pyroclasts then almost all of the pyroclasts in any of the eruptions modeled
 1356 here would stay locked to the expanding and accelerating gas cloud for long enough that they
 1357 acquired a very large fraction of the ultimate gas speed. However, we have no direct evidence of
 1358 how far from their source vents the Apollo sample pyroclasts were collected, and it is possible
 1359 that larger, unsampled clasts may have been produced.

1360

1361 The speed with which gas and small pyroclasts emerge through the vent into a lunar fire-
 1362 fountain-like eruption, U_v , is obtained by substituting the value of P_{ch} for P in equation (36).
 1363 Next, the gas-pyroclast mixture is allowed to expand above the vent to the final pressure P_f at
 1364 which gas and clasts decouple, which we saw earlier is ~90 Pa. In the simplest case, therefore,
 1365 where all of the pyroclastic droplets stay locked to the expanding gas until this decoupling
 1366 pressure is reached, this allows their final common velocity U_b to be found from

1365

$$1366 \quad 0.5 U_b^2 = 0.5 U_v^2 + [(n Q_u T_m) / m] \ln(P_{\text{ch}} / P_f) + [(1 - n) / \rho_l] (P_{\text{ch}} - P_f) \quad (39)$$

1367

1368 Figure 12 shows the resulting values of U_b for a range of values of n , and also the implied
 1369 maximum ranges R_f that pyroclasts would reach assuming ballistic trajectories. For total
 1370 magmatic CO and H₂O contents up to ~3000 ppm, maximum dispersal distances of sub-mm
 1371 sized pyroclasts from their vents will be up to ~10 km.

1372

1373 It is extremely important to consider the structure of the fire fountain that forms over a
 1374 lunar vent. Some combination of limited range, small pyroclast size and large volume flux can
 1375 lead to conditions in which the fire fountain is optically dense, in the sense that pyroclasts in the
 interior of the fountain cannot radiate heat into space because they are shielded by other

1376 pyroclasts. The result is that essentially all of the pyroclasts reach the ground at magmatic
 1377 temperature and coalesce into a lava pond, which in turn feeds a lava flow. Treatments of this
 1378 issue have been given by Wilson and Keil (1997), Wilson and Head (2001) and Wilson and Keil
 1379 (2012) who found that significant heat can only be lost from within an outer shell extending
 1380 inward from the edge of the fountain by a critical distance X , which may be termed the opacity
 1381 depth. All of these treatments assumed that pyroclasts were distributed uniformly in the fire
 1382 fountains, and we have now extended the analysis by relaxing this assumption. The detailed
 1383 distribution is found by numerically following the paths of a large number of pyroclasts ejected
 1384 at a given speed and over a given range of elevation angles for a great enough time that all of the
 1385 pyroclast reach the surface. The space around the vent is divided radially into 5000 discrete,
 1386 equal-sized cells and a record is kept of the cell in which each pyroclast is located at each of
 1387 1000 finite time intervals during its flight. We find that for a 2-dimensional fountain produced
 1388 by an elongate fissure vent the mean number of pyroclasts per unit volume in the outer part of
 1389 the fountain is about double the mean value. The situation is quite different for a point-source
 1390 vent producing a circularly-symmetric fountain because pyroclasts are distributed into ever-
 1391 larger annular zones as their horizontal distance from the vent increases. This causes the mean
 1392 number of pyroclasts per unit volume in the outer part of the fountain to be about one fifth of the
 1393 mean value for the entire fountain. With these geometric corrections to the treatment of Wilson
 1394 and Keil (2012) we find that for a point-source vent forming a circularly-symmetric fountain

$$1395 \quad X = (6.17 d g^{1/2} R_f^{5/2}) / F_e \quad (40)$$

1396 where F_e is the total erupted volume flux, and for a fissure vent erupting actively for a distance
 1397 L_e along strike

$$1400 \quad X = (0.52 d g^{1/2} R_f^{3/2} L_e) / F_e \quad (41)$$

1403 We now combine these results with the data in Figure 12 to show how the fraction of
 1404 pyroclasts falling back to the surface at magmatic temperatures is related to the volume flux and
 1405 volatile content of the magma. We treat the more conservative case of point-source vent feeding
 1406 a circularly-symmetric fire fountain. The fraction of pyroclasts landing hot to form a lava pond
 1407 is then $F = [(R_f - X) / R_f]^2$. This quantity is shown as a percentage in Table 5 as a function of
 1408 magma CO content n for a range of values of magma volume flux F_e . Wilson and Head (1980)
 1409 suggested that lava ponds of this kind formed around vents feeding high effusion rate eruptions
 1410 of turbulent lavas that thermally eroded their substrates to erode sinuous rilles (Hulme, 1973;
 1411 Carr, 1974). Wilson and Head (1981) showed that the motion of the lava in the ponds would
 1412 also be turbulent, so that the pond floors should also be eroded. Head and Wilson (1980)
 1413 measured the radii and smaller half-widths of several sinuous rille source depressions, finding
 1414 values between 1.1 and 2.4 km, entirely consistent with the present predicted pyroclast ranges,
 1415 especially for the smaller values of n .

1416 A final issue deserves attention for steady explosive eruptions. We have assumed in Figure
 1417 12 that all pyroclasts are small enough to stay locked to the gas phase during its expansion and
 1418 thus to acquire most of the gas speed. We have no pyroclast samples from the Moon that are
 1419 known to be collected very close a vent, and so we cannot rule out the possibility that
 1420 coalescence of gas bubbles, perhaps encouraged by shearing forces at the edge of the conduit,
 1421 may sometimes lead to magmas being erupted with a wider range of clast sizes, including clasts

1422 significantly coarser than the ~1 mm typical of the Apollo samples. In Figure 13 we simulate an
 1423 eruption in which a large fraction, in this case 80%, of the clasts are so coarse that they acquire
 1424 only 50% of the gas speed and fall out of the ejecta cloud near the vent. This implies that the
 1425 effective gas mass fraction accelerating the remaining 20% consisting of small clasts is increased
 1426 by a factor of $(80/20 =) 4$, causing their incremental speeds to increase by a factor of up to 2 and
 1427 their ranges by a factor of up to 4. Figure 13 gives the ranges of the largest, R_{coarse} , and smallest,
 1428 R_{fine} , clasts predicted by this simple model for the same values of n as Figure 12 and compares
 1429 these ranges with the range R_{mono} of the monodisperse size distribution listed in Figure 12.
 1430 Consider the case for $n = 2000$ ppm. Whereas Figure 12 would have predicted that pyroclasts
 1431 would reach the surface fairly uniformly distributed over an area having a radius of 6.5 km, we
 1432 now expect 80% of the erupted mass to fall within a radius of 1.6 km, covering an area $(5.1/1.6)^2$
 1433 = 10 times smaller and thus forming a layer $(0.8 \times 10 =) 8$ times deeper than before. This layer
 1434 could, of course, take the form of a cinder- or spatter-cone around the vent, detectable using
 1435 Lunar Orbiter Laser Altimeter (LOLA) data (Head and Wilson, 2015). The remaining 20% of
 1436 the pyroclasts are distributed out to a radius of 19.3 km, covering an area 14.3 times larger than
 1437 before, and forming a layer $(14.3/0.2 =) \sim 72$ times thinner than before. Thus some small cinder
 1438 or spatter cones might be expected to be surrounded by a 10-30 km radius aureole of thinly
 1439 spread pyroclasts (Head and Wilson, 1994), no doubt at least partly disguised by regolith
 1440 gardening, but possibly detectable using multispectral data (see Head and Wilson, 2015).

1441

1442 **5.B Consequences of steady magma eruption (1): lava flows**

1443 Although some element of explosive activity is expected to have been present in all lunar
 1444 basaltic eruptions, lava ponds formed by the accumulation of hot pyroclasts will have been
 1445 common and will have drained down-slope to feed lava flows. The speed U_f of a lava flow of
 1446 density ρ and thickness D_f will depend on whether the lava motion is laminar or turbulent. In
 1447 laminar flow the speed is given by

1448

$$1449 U_f = (\rho g D_f^2 \sin \alpha_f) / (3 \eta) \quad (42)$$

1450

1451 where η is the bulk viscosity and in turbulent flow by

1452

$$1453 U_f = [(8 g D_f \sin \alpha_f) / \lambda]^{1/2} \quad (43)$$

1454

1455 where a convenient formulation of the friction factor λ in terms of the Reynolds number Re_f is

1456

$$1457 \lambda = [0.79 \ln Re_f - 1.64]^{-2} \quad (44)$$

1458

1459 and Re_f is defined by

1460

$$1461 Re_f = (4 U_f D_f \rho) / \eta \quad (45)$$

1462

1463 The dependence of Re_f on U_f and the presence of the $(\ln Re_f)$ term in the definition of λ mean
 1464 that in turbulent flow U_f must be obtained from equation (43) by a recursive method from an
 1465 initial estimate. Further, the decision as to whether the flow motion is laminar or turbulent must
 1466 also be made retrospectively after evaluating U_f from both of equations (42) and (43); the
 1467 relative dependencies of friction on Reynolds number in laminar and turbulent flow are such that

1468 taking the smaller value of U_f is always the correct solution (Wilson and Head, 1981). These
 1469 operations are readily programmed as a spreadsheet.

1470 Although the boundaries of many mare lava flows have been blurred by regolith formation
 1471 after their emplacement, it is possible to measure the lengths and thicknesses of a few of the
 1472 flows in Mare Imbrium (Schaber, 1973; Bugiolacchi and Guest, 2008; Campbell et al., 2009) and
 1473 to estimate the topographic slopes of the surfaces on which they flowed (Rosenburg et al., 2011;
 1474 Kreslavsky et al., 2013). We take as representative measured values a thickness D_f of 20 m, a
 1475 width W_f of 20 km, and a slope α_f such that $\sin \alpha_f = 1 \times 10^{-3}$. The largest flow length X_f
 1476 described by Schaber (1973) was 1200 km. With a plausible mare lava viscosity of 1 Pa s we
 1477 find $U_f = 4.8 \text{ m s}^{-1}$; the flow motion is fully turbulent with $Re_f = 1.15 \times 10^6$. A flow length of X_f
 1478 = 1200 km would require an emplacement time, t_f , of ~69 hours and the volume flux feeding a
 1479 $W_{\text{flow}} = 20 \text{ km}$ wide flow would be $F_f = (U_f W_{\text{flow}} D_f) = 1.9 \times 10^6 \text{ m}^3 \text{ s}^{-1}$. Increasing the viscosity
 1480 by a factor of 10 to 10 Pa s decreases the implied speed to 3.75 m s^{-1} ; the volume flux decreases
 1481 to $1.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ and the emplacement time increases to ~89 hours. The most recent Lunar
 1482 Reconnaissance Orbiter images show outcrops on steep slopes of what may be lava flows having
 1483 thicknesses of 3-14 m (Ashley et al., 2012). These are comparable to flow thickness estimates of
 1484 10-20 m for outcrops in the walls of Rima Hadley (Howard and Head, 1972; Spudis et al., 1988).
 1485 To illustrate the conditions that may have produced these deposits we show in Table 13 the flow
 1486 speeds, Reynolds numbers, emplacement times and volume fluxes for 20 km wide flows with
 1487 viscosity 1 Pa s emplaced on a slope of $\sin \alpha_f = 1 \times 10^{-3}$ with thicknesses between 1 and 30 m.
 1488 The emplacement times assume a more conservative flow length of 600 km. All of these flows
 1489 are fully turbulent.

1490 We have explored the possibility that these large volume fluxes are overestimates. Thus it
 1491 is possible that isostatic subsidence of the centers of mare basins has caused present-day slopes
 1492 to be greater than those at the time of eruptive activity; also large-volume lava flows on Earth
 1493 often exhibit inflation (Hon et al., 1994; Self et al., 1996, 1998; Thordarson and Self, 1998). To
 1494 explore the consequence of such changes we reduce α_f by a factor of 3 in the example given
 1495 above so that $\sin \alpha_f = 0.3 \times 10^{-3}$ and we decrease D_f by a factor of ~3 from 20 m to 7 m. The
 1496 result is $U_f = 0.89 \text{ m s}^{-1}$; the flow motion is still fully turbulent with $Re_f = 7.5 \times 10^3$ and the
 1497 volume flux feeding a 20 km wide flow is $1.25 \times 10^5 \text{ m}^3 \text{ s}^{-1}$. It is thus very difficult to avoid the
 1498 conclusion that mare lava flows having thicknesses of at least ~10 m were emplaced in eruptions
 1499 having volume eruption rates of at least 10^4 and more commonly 10^5 to $10^6 \text{ m}^3 \text{ s}^{-1}$. We note that
 1500 this result, along with all of the cases shown in Table 13, is entirely consistent with our
 1501 conclusions in Section 3 regarding the range of volume eruption rates expected for a magma
 1502 rising from great depths in the Moon.

1503 It is of interest to explore whether the sizes of mare lava flows were typically limited by
 1504 the available volume of magma in the deep source zone or by the environmental conditions -
 1505 specifically were they volume-limited or cooling-limited. By definition a volume-limited flow
 1506 stops advancing when the source region can no longer supply magma to the vent. A cooling-
 1507 limited flow, in contrast, stops advancing when cooling at the margins of the flow penetrates far
 1508 enough into the interior. Pinkerton and Wilson (1994) showed that cooling limited flows stop
 1509 when the Grätz number for the flow, G_z , defined by

$$1510 \quad G_z = [(16 D_f^2)/(\kappa t_f)] \quad (46)$$

1511
 1512

1513 decreases from an initially large value to a critical limiting value, G_{zc} , equal to ~ 300 . Here $\kappa = \sim 7$
 1514 $\times 10^{-7} \text{ m}^2 \text{ s}^{-1}$ is the thermal diffusivity of silicate lava and t_f is the time after which lava motion
 1515 ceases. Eliminating t_f by assuming a constant flow speed U_f , so that $t_f = (X_f / U_f)$, and writing U_f
 1516 in terms of the volume flux $F_f = (U_f W_f D_f)$ we can re-order equation (U) in terms of F_f and
 1517 measurable quantities as

$$1518 \quad F_f = (18.75 \kappa X_f W_f) / D_f \quad (47)$$

1520
 1521 which with $X_f = 1200 \text{ km}$, $W_f = 20 \text{ km}$ and $D_f = 20 \text{ m}$ yields $F_f = 1.6 \times 10^4 \text{ m}^3 \text{ s}^{-1}$. Thus only if
 1522 the volume flux feeding this flow had been this small would the flow unit have stopped growing
 1523 due to cooling. All volume fluxes larger than this value (which the cooling constraint on magma
 1524 rise from the mantle suggests should be common) feeding a lava flow with this thickness and
 1525 width would have been capable of generating a flow unit longer than 1200 km. The clear
 1526 inference is that the typical lava flow units observed on the Moon, which are shorter than 1200
 1527 km, were limited in their extents by the volumes of magma available for eruption and not by
 1528 cooling. Given that a flow 300 km long, 20 km wide and 20 m thick has a volume of $\sim 120 \text{ km}^3$,
 1529 this suggests that magma batches with volumes of a few hundred km^3 were commonly generated
 1530 in and extracted from the mantle. This result, combined with the data in Table 1, suggests that
 1531 the vertical extents of deep mantle partial melt zones were of order 20 to 25 km.

1532 **5.C Consequences of steady magma eruption (2): sinuous rilles**

1533
 1534 Hulme (1973) proposed that lunar sinuous rilles were the products of surface erosion by
 1535 turbulent flowing lava and Carr (1974) estimated erosion rates that supported this idea.
 1536 Subsequently Hulme and Fielder (1977), using Hulme's (1974) model of non-Newtonian lava
 1537 rheology, suggested that the low viscosity of lunar lavas, combined with the shallow slopes of
 1538 pre-existing lava surfaces within mare basins, meant that small differences in slope or effusion
 1539 rate could determine whether lava flows were turbulent or laminar, and hence more or less likely
 1540 to erode sinuous rilles. The efficiency of thermal erosion was discussed by Hulme (1982) and
 1541 Fagents and Greeley (2001). Detailed models of thermal erosion using explicit thermal and
 1542 mechanical properties of volcanic rocks known or inferred to be present on planetary surfaces
 1543 were developed for eruptions on Earth (Williams et al., 1998, 1999), Io (Williams et al., 2000a),
 1544 the Moon (Williams et al., 2000b) and Mars (Williams et al., 2005) and have been applied
 1545 specifically to the formation of the major Rima Prinz rille on the Moon (Hurwitz et al., 2012).
 1546 These newer models concur with the earlier work in requiring eruptions lasting typically a few
 1547 months to explain the observed depths of the rille channels.

1548 Our focus is on relating sinuous rille formation to lava eruption rates. We therefore use
 1549 arguments developed by Head and Wilson (1980, 1981) and Wilson and Head (1980) that utilize
 1550 the observed widths of sinuous rille channels, W_r , and the geometries of the source depressions
 1551 that feed the rilles. In the case of a rille, let the volume flow rate of lava in the channel be F_r ;
 1552 the depth of flowing lava (which in general will not fill the channel) is D_r and the speed is U_r ; then
 1553 by definition

$$1554 \quad F_r = (U_r D_r W_r) \quad (48)$$

1555
 1556
 1557 The Reynolds number for the flow motion is

1558

1559 $Re_r = (4 U_r D_r \rho) / \eta$ (49)

1560

1561 and eliminating the product $(U_r D_r)$ between the equations gives

1562

1563 $F_r = (W_r Re_r \eta) / (4 \rho)$ (50)

1564

1565 We postulate that for efficient thermal erosion the motion must be fully turbulent, so that Re_r
 1566 must be at least ~ 2000 ; this implies that the minimum volume flow rate through the rille channel
 1567 must be F_{\min} given by

1568

1569 $F_{\min} = (500 W_r \eta) / \rho$ (51)

1570

1571 Typically rille channels have widths in the range 1000-3000 m (*Schubert et al.*, 1970; Hurwitz et
 1572 al, 2012; 2013) and so using $D_r = 2000$ m, $\eta = 1$ Pa s and $\rho = 2900$ kg m⁻³ we find $F_{\min} = \sim 300$
 1573 m³ s⁻¹. More stringent limits can be set by considering the turbulent lava ponds that feed the
 1574 rilles. Wilson and Head (1980) showed that the equivalent of equation (51) for such a pond is

1575

1576 $F_{\min} = (2000 R_p \eta) / \rho$ (52)

1577

1578 where R_p is the pond radius. Measurements of the source ponds for the rilles Prinz, Vera, Ivan,
 1579 Beethoven and Handel (Head and Wilson, 1980) give an average of $R_p = 1860$ m, implying F_{\min}
 1580 $= \sim 1200$ m³ s⁻¹.

1581 Note, however, that both of these values of F_{\min} are very much lower limits because we
 1582 expect Re_r to be much greater than the limiting value of ~ 2000 . Thus by applying Hulme's
 1583 (1973) model of lava flow in rille channels to the rilles numbered 2, 3, 4, 5, 6, 7, and 18 in the
 1584 catalog of Oberbeck et al. (1971), Head and Wilson (1981) found Reynolds number of order 10^5 .
 1585 Lava flow depths were inferred to be ~ 10 m in channels measured to be 100-300 m deep, flow
 1586 speeds were within a factor of two of 6 m s⁻¹, channel floor erosion rates were within 50% of 1
 1587 meter per day, and eruption durations were 100-300 days. The implied volume eruption rates
 1588 were in the range 10^4 to 10^6 m³ s⁻¹. The durations of the eruptions were found by dividing the
 1589 rille channel depths by the thermal erosion rates, and multiplying the durations by the volume
 1590 rates implied erupted volumes of ~ 100 to nearly 2000 km³. Volumes this large would imply
 1591 mantle partial melt source regions of up to ~ 35 km in vertical extent.

1592

1593 **5.D Lava flows and sinuous rilles compared**

1594 In Section 5.B we found that the smallest volume flux likely to be associated with an
 1595 eruption feeding a typical mare lava flow was $\sim 10^4$ m³ s⁻¹, and that typical flows were fed by
 1596 eruption rates in the range 10^5 to 10^6 m³ s⁻¹ producing flow volumes of order 100 km³. The
 1597 analyses described in Section 5.C show that sinuous rille-forming eruptions have similar
 1598 minimum and maximum magma discharge rates and minimum magma volumes. However, the
 1599 rille-forming eruptions commonly involved greater magma volumes erupted over much longer
 1600 periods of time. The longer durations, rather than any subtle topographic slope effects dictating
 1601 laminar or turbulent flow, appear to be the key to the ability of these flows to erode rille
 1602 channels.

1603

1604 Additional distinctive properties include narrower precursor lava flows erupted from vents
 with much smaller horizontal extents than those feeding the more common sheet-like flows. The

1605 greatest length of a fissure feeding a sinuous rille appears to be the ~14 km long major part of the
 1606 source depression of Rima Hadley (Head and Wilson, 1981), but few other such fissures exceed
 1607 ~6 km in length (Oberbeck et al., 1971). Indeed, where the sources of the rilles are circular
 1608 depressions, it is not the actual vent geometry that defines the lava flow width but rather the size
 1609 of the overflowing lava pond feeding the flow, which in turn is dictated by the explosivity of the
 1610 eruption. Given that there is currently no available three-dimensional model of dike propagation
 1611 through a planetary crust that takes detailed account of the stress changes associated with the
 1612 dike reaching the surface, we cannot provide any detailed explanation of these observations in
 1613 terms of crustal stresses.

1614 There may, however, be an explanation in terms of the long durations of the eruptions. In
 1615 long-lasting fissure eruptions on Earth it is common for activity to focus progressively toward
 1616 the center of the active fissure, so that eventually just a short fissure segment or a single localized
 1617 vent is active. This trend is ascribed in part to preferential magma chilling at the thin dike tips
 1618 (Bruce and Huppert, 1987, 1990; Carrigan et al., 1992; Head and Wilson, 1992a). However, if
 1619 flow in a fissure continues for long enough, the walls of the feeder dike are heated to the point
 1620 where magma that has already chilled against the wall is re-melted and removed, and eventually
 1621 the initial dike width increases as the wall rocks are thermally eroded. This change from
 1622 narrowing to widening with time occurs preferentially at the widest part of the initial fissure, i.e.,
 1623 at or near its center. Magma transport then becomes concentrated in this central, widening part
 1624 of the dike (Bruce and Huppert, 1987) and the eventual blocking of the distal ends takes place
 1625 quickly. Bruce and Huppert (1990) provide examples of the behavior of mafic magma in dikes
 1626 propagating vertically for distances of 2 and 5 km under similar pressure gradients to those
 1627 inferred here for dikes penetrating the lunar crust. We have extrapolated these data to the ~30
 1628 km thickness of the nearside lunar crust. In the likely lunar case, where there is no pre-heating of
 1629 the crustal rocks above lunar ambient temperatures by immediately-preceding regional volcanic
 1630 activity, we find that, for a dike width in excess of ~2.5 m, which is a much smaller dike width
 1631 than any we have found, there will be a negligible initial period of magma chilling against the
 1632 dike walls in the widest part of the dike, and widening of this region, with consequent capture of
 1633 most of the volume flux, begins almost immediately. The rate of dike wall erosion will be
 1634 comparable to that found by Head and Wilson (1981) for the floors of sinuous rille channels, ~15
 1635 $\mu\text{m s}^{-1}$. For a range of eruption conditions, Table 7 shows how the magma rise speed at depth,
 1636 below the levels where volatile release is important, increases by ~50% as a 1600 m long fissure
 1637 vent evolves into the circular shape needed to accommodate the same volume flux. The time
 1638 required for the change ranges for 66 to 108 days. Given the likely 100-500 day durations of the
 1639 rille-forming eruptions (Hulme, 1973; Head and Wilson, 1981), it is not surprising, therefore,
 1640 that they appear to be fed by relatively short fissures.

1641 1642 *5.E Non-mare volcanism*

1643 The presence of the morphologically and spectroscopically (Head and McCord, 1978;
 1644 Glotch et al., 2011; Kusuma et al., 2012; Ivanov et al., 2015) distinctive Gruithuisen and Mairan
 1645 domes in N.E. Oceanus Procellarum and the domes between the craters Belkovich and Compton
 1646 (Jolliff et al., 2011; Chauhan et al., 2015) implies the localized eruption of unusually viscous,
 1647 probably rhyolitic, magma (Wilson and Head, 2003b), a very rare occurrence on the Moon.
 1648 Wilson and Head (2003b) used the morphologies of the Gruithuisen and Mairan domes to infer
 1649 the yield strengths and plastic viscosities of the magmas forming them assuming that they
 1650 behaved as Bingham plastics and to deduce the magma volume eruption rates, ~tens of $\text{m}^3 \text{s}^{-1}$,

1651 and durations, ~10-50 years. We have repeated the analysis, using the improved crustal density
 1652 estimates from GRAIL, and relaxing some of the assumptions about the feeder dike geometry.
 1653 Table 8 shows the original rheological parameters and the new estimates of dike width and
 1654 magma rise speed. Also shown are the minimum magma rise speeds needed to offset excessive
 1655 cooling during magma ascent from the base of the crust found using equation (25). In all cases
 1656 the eruptions are thermally viable.

1657 The origin of this highly silicic magma is uncertain; options include basal melting of the
 1658 lunar crust by large volumes of under-plating basalt or differentiation during cooling of large-
 1659 volume basaltic magma bodies, again most likely located at the crust-mantle boundary density
 1660 trap. In section 3.F it was shown that substantial volumes, ~5000 km³, of basaltic magma could
 1661 be emplaced as intrusions at the base of the crust under suitable circumstances. Such intrusions
 1662 are easiest to understand late in lunar history when horizontal compressive stresses in the
 1663 lithosphere make it likely that the least principle stress will have been vertical. However, the
 1664 silicic domes are inferred to have been formed ~ 3.8 Ga ago, and so crustal thinning and stress
 1665 modification due to basin-forming impacts in the early period before warming of the lunar
 1666 interior generated extensional stresses in the lithosphere are the more likely source of the
 1667 required stress conditions. The volumes of the larger domes are ~300-500 km³, (Wilson and
 1668 Head, 2003b), an order of magnitude smaller than a possible 5000 km³ basalt intrusion, and so
 1669 both partial melting of overlying crust and fractional crystallization of sill magma are viable
 1670 sources of the silicic melt on thermal grounds.

1671 If fractional crystallization were the source mechanism, concentration of volatiles into
 1672 residual melt could have enriched the melt in water, perhaps by a factor of ~10 over the ~1000
 1673 ppm found in some lunar samples by Hauri et al. (2011). The treatment of section 5.A shows
 1674 that the eruption of silicic melt with ~10,000 ppm, i.e. ~1 mass %, of water could have ejected
 1675 pyroclastic material in explosive phases of the eruptions to distances of ~30 km.

1676

1677 ***5.F Late-stage lunar volcanism***

1678 The thermal models of Solomon and Head (1982), Spohn et al. (2001) and Ziethe et al.
 1679 (2009) all suggest that the zone within which partial melting can occur in the Moon's mantle
 1680 must migrate deeper into the mantle with time and shrink in its vertical extent. The models differ
 1681 in their predictions of when melting should have ceased, mainly as a result of differing
 1682 assumptions about the solidification of the initial magma ocean. It is inevitable that the
 1683 progressive decay of radioactive heat sources must cause the rate of melt generation to decrease
 1684 with time. The rate of percolation of melt within partial melt zones is linked to the melt volume
 1685 fraction and melt viscosity. If melting is occurring at all, the melt viscosity will not change
 1686 significantly, but the percolation speed will decrease because the melt volume fraction will
 1687 decrease as the melt production rate decreases. Thus it will take longer for a given dike to grow
 1688 upward from a diapiric partial melt zone, and the vertical extent, and hence volume, of the dike
 1689 that eventually detaches from the melt zone will be less as a function of time because the vertical
 1690 extent of the melt zone decreases.

1691 These trends suggest that late in lunar volcanic history both the volumes of batches of melt
 1692 arriving at the crust-mantle boundary and the frequency with which they arrived will have been
 1693 less than in earlier times. Given that the horizontal compressive stress in the lithosphere will
 1694 have been increasing with time in late lunar history, it is difficult to anticipate with confidence
 1695 how these changes will have influenced the ability of magma to penetrate the crust. However,
 1696 the likely expectation is that large volumes of basaltic melt must have accumulated in sills at the

1697 base of the crust before conditions allowed dikes to penetrate the crust as a result of excess
 1698 pressures in the sills. When eruptions finally occurred, they would have involved larger volumes
 1699 of magma than in earlier times, with the intervals between eruptions being much greater than
 1700 before. The final stages of such activity might have involved dikes that penetrated part way
 1701 through the crust but did not erupt magma. Volatiles in the accumulation zones at the tops of
 1702 these dikes might, however, have made their way to the surface. It is tempting to speculate that
 1703 morphologically (Garry et al., 2012) and spectroscopically (Braden et al., 2014) enigmatic
 1704 features like Ina, which may have formed relatively recently (Schultz et al., 2006), may be linked
 1705 to this very late stage activity (Figure 8) (Head and Wilson, 2015).

1706

1707 **6. Summary and Conclusions**

1708 ***6.A General Setting for Mare Volcanism***

1709 Secondary planetary crusts are those derived from partial melting of the mantle, and the
 1710 consequent collection, ascent and eruption of the resulting magmas. The geologic record of
 1711 these plutonic and volcanic products represents the history of planetary crustal and thermal
 1712 evolution, and reflects the dominant mode of planetary lithospheric configuration and heat
 1713 transfer. Lunar mare volcanism is the primary manifestation of secondary crustal formation on
 1714 the Moon and provides key insights into lunar thermal evolution. We used new data on the
 1715 density and thickness of the crust, the petrologic properties and the geologic record of mare
 1716 basalt volcanism to assess: 1) the range of magma source depths, 2) modes of magma generation,
 1717 ascent and eruption, 3) the volumes and volume fluxes of magma, 4) the partitioning into
 1718 intrusive and extrusive deposits, 5) the role of primary lunar crustal formation and configuration
 1719 in modulating intrusion and eruption style, 6) the role of thermal evolution in controlling the
 1720 source depths and eruption frequencies, styles and fluxes, 7) the predicted relationship of these
 1721 properties to observed landforms and deposits, 8) the relationship of magmatic volatile
 1722 production to predicted explosive eruption style, and landform/deposit characteristics, 9) the
 1723 causes of patterns of mare basalt areal distribution (e.g., nearside/farside asymmetry) and styles
 1724 (e.g., long lava flows, sinuous rilles), and 10) the likelihood of recent and current mare basalt
 1725 plutonic and volcanic activity on the Moon. We use this basic setting and the following
 1726 considerations to assess the lunar geological record for consistency with these predictions (Head
 1727 and Wilson, 2015).

1728

1729 ***6.B Basic Configuration of Lunar Mare Basalt Genesis and Eruption***

1730 We find that the basic configuration of lunar mare magmatism is fundamentally controlled by
 1731 1) the formation of the low-density anorthositic primary crust, 2) the consequences of its
 1732 formation and aftermath for the nature of the mantle and the distribution of heat sources, and 3)
 1733 the resulting one-plate-planet tectonic structure characterized by conduction-dominated
 1734 lithospheric heat transfer, and a lithosphere that progressively thickened with time. The
 1735 formation of large multi-ringed basins, some of which date to the early lunar mare volcanism era,
 1736 regionally thinned the crust and introduced short-term perturbations in the thickness of the
 1737 lithosphere. These basic factors provided a density barrier (the low-density anorthositic crust)
 1738 fixed early in lunar history, and a mechanical barrier (the base of the lithosphere) that
 1739 progressively deepened with time. The thermal evolution of the Moon, characterized by the
 1740 evolving ratio of accretional heat and radiogenic heat sources, and continual lithospheric heat
 1741 loss to space, resulted in a change in the net state of stress in the lithosphere from extensional to
 1742 contractional in early-middle lunar history. This change was a key factor in the mare basalt

1743 surface volcanic flux and eruption style, progressively inhibiting the ascent and eruption of
 1744 magma, and changing eruption styles toward extremely voluminous individual eruptions, often
 1745 with accompanying sinuous rilles.

1746

1747 ***6.C Modeling the Generation, Ascent and Eruption of Magma***

1748 In modeling the generation, ascent and eruption of magma, we used new estimates of the
 1749 vertical extent of partial melting (up to ~150 km) in lunar mantle diapirs and of the depths of
 1750 density/rheological traps, and include excess magma source pressures as well as magma
 1751 buoyancy. We find that excess pressures in shallower magma reservoirs and buoyancy traps are
 1752 about an order of magnitude smaller than those in deep partial melt zones. Rates of melt removal
 1753 from the mantle source regions should be much lower on the Moon than Earth; lunar mantle
 1754 convection rates are lower by about an order of magnitude due to lunar gravity, so reservoir
 1755 overpressurization and melt extraction should be at much lower rates, implying that only a very
 1756 small amount of magma can be extracted rapidly from a deep lunar mantle source, and,
 1757 consequently, that large mantle source regions, of the order 10^5 to 10^6 km³, are required.

1758

1759 ***6.D Lunar Mare Basalt Magma Transport in Dikes***

1760 Transport of magma toward the surface is by brittle fracture in rocks overlying the melt
 1761 source and the consequent propagation of a dike. A dike containing magma everywhere buoyant
 1762 relative to its host rock would inevitably reach the surface and erupt until the magma supply is
 1763 exhausted. If magma is not positively buoyant at all depths, excess pressure in the source region
 1764 can assist in the vertical growth of a dike. Dikes can cease to grow due to: 1) lack of sufficient
 1765 buoyancy/overpressure, 2) excessive cooling, 3) lack of sufficient dike tip stress intensity, or 4)
 1766 exhaustion of magma supply in the source region. Unlike Earth, the great depth of lunar magma
 1767 source regions generally limits the role of volatiles in assisting magma ascent.

1768 Mare basalt magma dikes intruding into the anorthositic crust should be everywhere
 1769 negatively buoyant, and if the horizontal stress in the lithosphere is sufficiently compressive are
 1770 predicted to extend laterally to underplate and create secondary reservoirs at the crust-mantle
 1771 boundary. If the positive excess pressure of the portion of the dike in the mantle is great enough,
 1772 however, dikes containing magma that is negatively buoyant relative to the crust can still
 1773 penetrate into the crust and reach the surface to erupt. In the first quarter of lunar history, with
 1774 abundant mantle heating and mild global expansion inducing an extensional state of stress in the
 1775 lithosphere, such dike intrusions through the crust and consequent eruptions should have been
 1776 common. In later lunar history, when global cooling thickened the lithosphere and induced
 1777 lithospheric compressional stresses, eruptions should have been inhibited and crustal
 1778 underplating is predicted to be favored.

1779 With sufficiently deep melt source regions and slow growth, dikes can disconnect from their
 1780 source regions and rise as discrete blade-shaped diapirs of fixed volume. We find that from
 1781 source depths greater than about 500 km, it is implausible that continuous dike pathways can
 1782 exist between the deep mantle source regions and the surface. Volumes of magma in these
 1783 pinched-off dikes are of the order of a few thousand km³ (a fraction of which will reach the
 1784 surface) and dike widths are so large that magma motion is predicted to be turbulent and not
 1785 controlled by viscosity or influenced by heat loss to the host rocks; typical rates of ascent are 30
 1786 m s⁻¹, requiring only ~4.6 hours to reach the surface.

1787 For isolated dikes encountering the basal crustal density trap, what are the conditions by
 1788 which they reach the surface? The tops of typical intrusions (up to ~43 km) in thicker farside

1789 crust is much deeper than those in the thinner nearside crust (up to ~13 km), and the range of
 1790 values for the nearside indicates that nearside eruptions should be heavily favored over farside
 1791 eruptions. The predominance of lunar nearside eruptions (thinner crust) also implies that the
 1792 vertical extents of mantle diapiric source regions that could produce eruptions lie in the range of
 1793 17-36 km. When dike tips reach the surface, the volume of magma erupted is a function of the
 1794 magnitude of the horizontal extensional stresses and can range from a small percentage of, to the
 1795 vast majority of, the total dike volume (i.e., tens of km³ to more than 1600 km³). The implied
 1796 eruption volume fluxes are huge, ranging from 10⁵ to 10⁶ m³ s⁻¹.

1797 Dikes that remain connected to their melt source zones are generally required to be sourced
 1798 from the shallower mantle, and would be favored in the earlier period of mare history under
 1799 several global thermal evolution models, with shallow partial melt zones being limited in vertical
 1800 extent relative to their deeper counterparts. We show that a simple explanation for the paucity of
 1801 eruptions on the lunar farside is that the vertical extent of melting in relatively shallow mantle
 1802 melt zones was less than ~45 km.

1803 Dikes from deep mantle source regions could extrude to the surface or intrude to any depth in
 1804 the lunar crust and are predicted to have widths of 35-50 m, with rise speeds during emplacement
 1805 indicating turbulent flow behavior. Dikes from shallow mantle sources are more restricted in the
 1806 range of the depth to the top of the dike when eruptions do not occur. A 50 km deep mantle
 1807 source (ponded near the base of the crust) is predicted to produce ~50 m wide dikes, with the
 1808 tops of dike intrusion within ~1 km of the surface.

1809

1810 ***6.E Range of Behavior of Dikes Intruding the Crust***

1811 Dikes intruding into the lunar crust can have several fates and consequences (Figure 8): 1)
 1812 those intruded to more than 10-20 km below the surface will solidify; 2) those reaching
 1813 shallower depths will undergo gas exsolution and gas accumulation and potentially vent gasses
 1814 to the surface; 3) those reaching the upper several kilometers of the crust and stalling can
 1815 produce near-surface stress fields and graben; 4) those reaching near-surface very low density
 1816 regions (brecciated crater lenses) can intrude laterally and produce sills and floor-fractured
 1817 craters; 5) those that just reach the surface can extrude small amounts of lava and produce small
 1818 shield volcanoes and pyroclastic venting; and 6) those that reach and have the potential to
 1819 overshoot the surface can produce high-flux and high-volume effusive volcanic eruptions,
 1820 creating long lava flows and sinuous rilles.

1821

1822 ***6.F Explosive Activity Accompanying Mare Basalt Eruptions***

1823 Dikes that breach the surface and erupt should all be accompanied by some level of explosive
 1824 activity due to the presence of small amounts of mainly CO from the smelting reaction that
 1825 occurs in the upper few kilometers, and the venting of this gas into the vacuum. There are three
 1826 phases of gas production during dike ascent and eruption, each with consequences for pyroclastic
 1827 activity: 1) gas is generated in the low-pressure dike-tip during dike propagation from the source
 1828 toward the surface, and accumulates into gas-filled cavities with vertical extents of tens to
 1829 hundreds of meters, overlying a magmatic foam layer of up to 10 kilometers vertical extent
 1830 above the rising magma); 2) there is a very short period (tens of seconds) after the dike tip breaks
 1831 the surface during which the gas in the pure gas cavity vents to the surface at very high velocity
 1832 with few magmatic particles but some entrained regolith/wall rock fragments; this is a lunar
 1833 equivalent of a terrestrial ultraplinian eruption phase with a pyroclast dispersal maximum
 1834 approaching 2000 km); 3) the pressure distribution in the dike now evolves to maximize the

1835 magma discharge rate; an expansion wave initially propagates into the underlying foam,
 1836 disrupting it into gas and pyroclasts which are dispersed to a maximum range of 6-20 km from
 1837 the vent; complete stability and steady eruption conditions are reached after the passage of
 1838 pressure waves down and back up the dike, taking ~1 hour. Unusually wide dikes (250-500 m)
 1839 that stalled near the surface without initially erupting could experience further gas accumulation
 1840 at the top of the magma column due to convection in the underlying magma, eventually causing
 1841 a gas-rich eruption.

1842 Dikes producing steady effusive eruptions to the surface should be accompanied by steady
 1843 pyroclastic activity. Volatiles form gas bubbles in the rising melt and these undergo expansion,
 1844 increasing the speed of the rising magma, and ultimately disrupting it into a free gas phase
 1845 entraining pyroclasts. On Earth, two types of conditions in the conduit can evolve at this point:
 1846 choked flow (where the vent pressure is greater than atmospheric) and supersonic flow (where
 1847 the vent pressure is equal to the atmospheric pressure). All lunar explosive eruption are
 1848 predicted to be choked. The ensuing dispersal of pyroclasts into the vacuum above the vent is
 1849 controlled by the exit speed from the vent and the size distribution of the liquid droplets into
 1850 which the magma is disrupted. Liquid droplets similar in size to the pyroclastic beads collected
 1851 on the Moon (100-1000 micrometers) will stay locked to the expanding and accelerating gas
 1852 cloud sufficiently long to be accelerated to significant speeds, ensuring widespread dispersal
 1853 away from the vent up to about 10 km. Larger particles that are produced will be accelerated
 1854 much less efficiently and will collect nearer the vent, with the largest ones potentially forming
 1855 cinder or spatter cones.

1856 Despite the very rapid acceleration of magma droplets by the gas cloud expanding out into
 1857 the surface vacuum, combinations of factors (limited range, small pyroclast size and large
 1858 volume flux) can lead to parts of the fire fountain being optically dense, with some specific
 1859 consequences for deposits and landforms. A high optical density means that particles cannot
 1860 radiate heat efficiently, due to shielding by other particles, and they fall to the ground at
 1861 magmatic temperatures and coalesce into a lava pond, which typically feeds a lava flow. Large
 1862 volume flux eruptions are typically predicted to be surrounded by such a lava pond, in which the
 1863 flow is turbulent, and to have formed the source depressions surrounding many sinuous rilles by
 1864 thermal erosion. For lower volume fluxes and larger clast sizes (larger than the ~1 mm glass
 1865 beads collected by the Apollo astronauts), acceleration by the expanding gas cloud is much less
 1866 efficient, and pyroclasts will fall out of the cloud within a range typically less than about 2 km
 1867 from the vent to produce cinder and spatter cones. When the large particles fall out of the cloud,
 1868 the effective gas mass fraction is increased, and this can cause increased acceleration of the finer
 1869 droplets, propelling them to several tens of km.

1870

1871 ***6.G Effusive Activity Accompanying Mare Basalt Eruptions***

1872 The typical fate of dikes reaching the close vicinity of the surface is to penetrate to the
 1873 surface and form effusive eruptions. Dikes with magmatic pressures just sufficient to penetrate
 1874 the surface will form low effusion rate, low-volume eruptions, and produce small shield
 1875 volcanoes situated on or along the top of the dike. The spectrum of overpressurization values
 1876 required to propagate a dike to the vicinity of the lunar surface means that a portion of the dike
 1877 population will be characterized by sufficiently high values to “overshoot” the surface; these
 1878 dikes will be characterized by very high effusion rates and the magma they erupt will drain
 1879 downslope from the vent to feed extensive lava flows. The velocity of erupting lava flows will
 1880 control whether the motion in the flow is laminar or turbulent. Lava flow thicknesses of a few to

1881 ~30 m have been reported, and for typical slopes and flow widths, and lengths of ~600 km, all
 1882 flows are fully turbulent. Discrete lava flows with thicknesses in excess of ~10 m were
 1883 characterized by eruptions having volume eruption rates of at least 10^4 , and more likely, 10^5 to
 1884 $10^6 \text{ m}^3 \text{ s}^{-1}$, comparable to our predictions for magma rising from significant depths in the lunar
 1885 interior.

1886 Lava flows generally have one of two fates: the flows can stop due to sufficient cooling of
 1887 the lavas so that the flow front can no longer advance (cooling-limited flows; reaching a limiting
 1888 Gratz number of ~300), or alternatively, the source region no longer supplies magma to the vent,
 1889 and the advancing flow stops due to lack of new magma (volume-limited flows). Analysis of the
 1890 fluxes and cooling behavior of lunar lava flows strongly implies that typical lava flows shorter
 1891 than ~1200 km would be supply-limited, not cooling-limited. This, in turn, suggests that magma
 1892 batches with volumes of a few hundred km^3 were commonly generated in the mantle and
 1893 extracted through dike and lava flow emplacement.

1894 How are sinuous rilles, interpreted to be caused by thermal erosion, related to lava flows?
 1895 Flow in sinuous rilles, like that in long lava flows, is shown to be fully turbulent. Analysis of
 1896 sinuous rille morphologies suggests that typical sinuous rille eruptions were characterized by
 1897 volume eruption rates of 10^4 to $10^6 \text{ m}^3 \text{ s}^{-1}$, eruption volumes of 100-2000 km^3 , eruption
 1898 durations of 100-300 days, and thermal erosion rates of ~1 meter per day. Thus, eruptions
 1899 producing typical lunar lava flows (volume eruption rates $>10^4$ to $10^6 \text{ m}^3 \text{ s}^{-1}$, typically 10^5 to 10^6
 1900 $\text{m}^3 \text{ s}^{-1}$; eruption volumes of ~100 km^3) overlap on the lower end of, and have similar
 1901 characteristics to, those producing sinuous rilles. The major difference between lava flow-
 1902 producing eruptions and those producing sinuous rilles is the longer durations of the eruptions
 1903 and the generally greater volumes of lava erupted, both factors enhancing the role of thermal
 1904 erosion in creating the rille channels. A further distinction between lunar lava flows and sinuous
 1905 rilles is the nature of the typical source regions. Lava flows often emerge from linear fissures,
 1906 but sinuous rille sources are commonly circular or slightly elongate depressions less than a few
 1907 kilometers in diameter. These sinuous rille vent shapes strongly suggest that due to the high
 1908 magma flux and duration of sinuous rille eruptions, thermal erosion of the widest parts of fissure
 1909 vent walls together with cooling of magma in the thinnest parts of the underlying dikes acts to
 1910 centralize the effusion to a pipe-like conduit; the result is the capture of most of the mass flux in
 1911 the central pipe, more rapid cooling of the rest of the dike walls, and an increase of magma rise
 1912 speeds by ~50%. Thus, sinuous rilles appear to differ from lava flows due to thermal erosion of
 1913 both the vent region and the substrate below the vent.

1914

1915 ***6.H Mare Basalt Lunar Resurfacing***

1916 The fate of erupted lavas fed by both flows and sinuous rilles depends on local and regional
 1917 slopes and the nature of the range of topographic features existing prior to eruptions. Mare lava
 1918 flows in early lunar history are predicted to be focused in the interiors of impact craters and
 1919 basins. Later lava flows will spread out over larger areas, or down regional slopes related to
 1920 loading and flexure by the earlier lava emplacement and basin filling. Repeated dike intrusions
 1921 over the course of mare basalt magmatism will also increase the density of the crust, somewhat
 1922 reducing the negative buoyancy of the magmas. The trend in global cooling will increase
 1923 compressive stress in the lithosphere with time, a trend reinforced by the progressive intrusion of
 1924 dikes in the crust.

1925

1926 In summary, in this contribution we make specific predictions about the nature and
1927 distribution of the spectrum of lunar mare volcanic landforms and deposits. These predictions
1928 and guidelines are analyzed and tested using the comprehensive array of data obtained by the
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1930

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2414
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2417 **Notation**

2418

2419 **Symbol****Definition and Units**

2420	A_d	total vertical extent of dike, m
2421	A_l	vertical extent of dike below base of crust, m
2422	A_{lf}	vertical extent of lower part of dike that reaches the surface, m
2423	A_u	vertical extent of dike above base of crust, m
2424	C	thickness of planetary crust, m
2425	D_f	thickness of lava flow, m
2426	D_r	depth of flowing lava in sinuous rille channel, m
2427	D_u	distance along planetary surface traveled by pyroclast, m
2428	E	vertical extent of shallow mantle partial melt zone, m
2429	E_d	vertical extent of deep mantle partial melt zone, m
2430	F	dimensionless fraction of pyroclasts landing hot to form lava pond
2431	F_d	volume flux of viscous magma forming domes, $m^3 s^{-1}$
2432	F_e	erupted magma volume flux from fissure vent, $m^3 s^{-1}$
2433	F_f	volume flux of lava in lava flow, $m^3 s^{-1}$
2434	F_i	initial erupted volume flux from fissure, $m^3 s^{-1}$
2435	F_{min}	minimum volume flux in sinuous rille lava for channel erosion, $m^3 s^{-1}$
2436	F_r	volume flux of lava in sinuous rille channel, $m^3 s^{-1}$
2437	G_z	dimensionless Grätz number for lava flow
2438	H	depth below surface of intruded dike top, m
2439	K_{base}	stress intensity at lower tip of dike at crust-mantle boundary, $Pa m^{1/2}$
2440	K_{crit}	fracture toughness of host rocks, $Pa m^{1/2}$
2441	K_l	stress intensity at lower tip of deep mantle dike, $Pa m^{1/2}$
2442	K_{top}	stress intensity at upper tip of dike at crust-mantle boundary, $Pa m^{1/2}$
2443	K_u	stress intensity at upper tip of deep mantle dike, $Pa m^{1/2}$
2444	L	vertical extent of dike growing from deep mantle melt zone, m
2445	L_d	horizontal extent of surface fissure vent forming dome, m
2446	L_e	horizontal extent of surface fissure vent, m
2447	L_m	critical length at which dike disconnects from melt source, m
2448	N	mass fraction of gas in mixture of gas and entrained wall rocks
2449	N_a	Avogadro's number, equal to $6.0225 \times 10^{26} kmol^{-1}$
2450	P	ambient pressure, Pa
2451	P_b	pressure at base of dike, Pa
2452	P_c	pressure due to the weight of the crust, Pa
2453	P_{ch}	pressure when gas-pyroclast flow speed is choked at sonic speed, Pa
2454	P_d	driving pressure at inlet at base of dike, Pa
2455	P_{dis}	pressure at which magmatic foam disrupts, Pa
2456	P_f	pressure at which gas and clasts decouple in Knudsen regime, Pa
2457	P_{foam}	pressure in magmatic foam layer, Pa
2458	P_i	initial pressure of explosively erupting gas-pyroclast mixture, Pa
2459	P_m	pressure due to the weight of a magma column, Pa
2460	P_n	driving pressure at center of dike at base of crust, Pa
2461	P_{sm}	pressure below which smelting reaction occurs, Pa
2462	P_w	pressure due to weight of magma in dike, Pa

2463	P_0	driving pressure at center of deep mantle dike, Pa
2464	Q_u	universal gas constant, equal to $8.314 \text{ kJ kmol}^{-1} \text{ K}^{-1}$
2465	R	radius of planetary body, m
2466	R_{coarse}	range of large pyroclasts in polydisperse mixture, m
2467	R_f	maximum range of ballistic pyroclasts, m
2468	R_{fine}	range of small pyroclasts in polydisperse mixture, m
2469	R_{mono}	range of ballistic pyroclasts in monodisperse mixture, m
2470	R_p	radius of lava pond fed by opaque fire fountain, m
2471	Re_f	dimensionless Reynolds number for surface lava flow
2472	Re_r	dimensionless Reynolds number for lava in sinuous rille
2473	S	speed of sound in gas-pyroclast mixture, m s^{-1}
2474	T	horizontal tension, relative to hydrostatic stresses, in lithosphere, Pa
2475	T_m	magma temperature, equal to 1623 K
2476	U	flow speed of magma in deep mantle dike, m s^{-1}
2477	U_b	eruption speed of pyroclasts entering fire fountain, m s^{-1}
2478	U_d	rise speed of viscous magma forming domes, m s^{-1}
2479	U_e	rise speed of erupting basaltic magma at depth, m s^{-1}
2480	U_f	mean speed of lava flow, m s^{-1}
2481	U_i	initial rise speed of erupting magma from shallow source, m s^{-1}
2482	U_{lam}	laminar flow speed of magma, m s^{-1}
2483	U_m	final eruption speed of pyroclasts locked to gas motion, m s^{-1}
2484	U_{min}	minimum rise speed of erupting magma to avoid cooling, m s^{-1}
2485	U_{turb}	turbulent flow speed of magma, m s^{-1}
2486	U_u	ultimate velocity of gas expanding to a vacuum, m s^{-1}
2487	U_v	speed at which gas and small pyroclasts emerge through vent m s^{-1}
2488	V	volume of magma in deep mantle dike, m^3
2489	V_e	volume of magma erupted from dike, m^3
2490	V_f	volume of magma remaining in dike after eruption, m^3
2491	V_i	initial volume flux of erupting magma from shallow source, $\text{m}^3 \text{ s}^{-1}$
2492	W	mean thickness of deep mantle dike, m
2493	W_{av}	mean thickness of dike reaching surface from shallow depth, m
2494	W_d	mean thickness of dike feeding dome-forming eruption, m
2495	W_f	mean thickness of dike reaching surface from great depth, m
2496	W_{flow}	width of lava flow, m
2497	W_n	mean thickness of dike at crust-mantle boundary, m
2498	W_r	width of sinuous rille channel, m
2499	X	opacity depth of fire fountain, m
2500	X_f	length of lava flow, m
2501	Z	depth to top of melt zone in shallow mantle, m
2502	Z_{crit}	minimum depth to top of melt zone in mantle to ensure eruption, m
2503	Z_{diff}	distance over which pressure decreases in erupting material, m
2504	d	average diameter of molecules in magmatic gas, equal to $\sim 300 \text{ }\mu\text{m}$
2505	dP/dz	pressure gradient driving magma flow, Pa m^{-1}
2506	f	dimensionless friction factor at dike wall, equal to 0.02
2507	g	acceleration due to gravity, equal to 1.62 m s^{-2}
2508	m	average molecular mass of released magmatic volatiles, kg kmol^{-1}

2509	n	mass fraction of volatiles released from magma
2510	n_d	solubility of water in terrestrial mafic magma, as mass fraction
2511	t_f	time to emplace lava flow, s
2512	v_g	dimensionless partial volume of gas in magmatic foam
2513	v_l	dimensionless partial volume of liquid in magmatic foam
2514	z	depth from which gas-pyroclast mixture erupts explosively, m
2515	ΔZ	vertical extent of foam layer near top of propagating dike, m
2516	$\Delta\rho$	density difference between host mantle and magma, kg m^{-3}
2517	α_f	slope of ground on which lava flows
2518	γ	ratio of specific heats of gas at constant pressure and constant volume
2519	ε	dimensionless bubble volume fraction in magmatic foam, 0.85
2520	η	basaltic magma viscosity, equal to 1 Pa s
2521	η_d	plastic viscosity of magma in viscous domes, m
2522	θ	elevation angle from horizontal at which pyroclast is ejected
2523	κ	thermal diffusivity of magma, equal to $7 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$
2524	λ	dimensionless basal friction factor for flow of lava on surface
2525	μ	shear modulus of host rocks, Pa
2526	ν	Poisson's ratio for host rocks, dimensionless
2527	ρ	bulk density of lava flow, kg m^{-3}
2528	ρ_{\square}	average mantle density, equal to 3260 kg m^{-3}
2529	ρ_{\square}	average magma density, equal to 2900 or 3010 kg m^{-3}
2530	ρ_{\square}	average crust density, equal to 2550 kg m^{-3}
2531	τ_d	yield strength of magma in viscous domes
2532	ϕ	effective diameter of gas molecules, equal to $\sim 3.4 \times 10^{-10} \text{ m}$
2533	φ	diameter of largest bubbles in magmatic foam, m
2534		
2535		
2536		

2537 Table 1. Variation with vertical extent, E_d , of mantle partial melt zone of the driving pressure at
 2538 the dike base, P_d , the driving pressure at the dike center, P_0 , the length, L_m , the mean width, W ,
 2539 and the volume, V , of a dike that disconnects from the source and migrates as a discrete magma
 2540 body.

2541

E_d /km	P_d /MPa	P_0 /MPa	L_m /km	W /m	V /km ³
5	2.9	7.5	15.7	4.5	0.9
10	5.8	15	31.4	18.0	13.9
15	8.7	22	47.1	40.5	71
20	12	30	62.8	71.9	223
25	15	37	78.5	112.4	545
30	17	45	94.2	161.9	1129
35	20	52	110.0	220.3	2092
40	23	60	125.7	287.8	3569
45	26	67	141.4	364.2	5718
50	29	75	157.1	449.7	8714
75	44	112	235.6	1011.8	44117
100	58	150	314.2	1798.7	139430

2542

2543

2544 Table 2. Comparison of properties of isolated dikes as they disconnect from their mantle
 2545 sources and after they are emplaced as stable intrusions at the crust-mantle boundary. Values at
 2546 disconnection are: total vertical length, L_m ; mean width, W ; volume, V ; and central driving
 2547 pressure, P_0 . Values after intrusion are: extent above crust-mantle boundary, A_u ; extent below
 2548 boundary, A_l ; driving pressure at boundary, P_n ; and mean width, W_n . Values in italics in the
 2549 upper parts of the tables represent intrusions; values in the central parts of the tables represent
 2550 eruptions on the near-side of the Moon; values in italics in the lower parts of the tables represent
 2551 potential eruptions on the far-side, not observed.

2552

2553 (a) Values for magma density, $\rho_l = 2900 \text{ kg m}^{-3}$

2554

E_d /km	L_m /km	W /m	V /km ³	P_0 /MPa	A_u /km	A_l /km	P_n /MPa	W_n /m
<i>5</i>	<i>15.7</i>	<i>4.5</i>	<i>0.9</i>	<i>7.5</i>	<i>7.3</i>	<i>9.7</i>	<i>3.7</i>	<i>3.9</i>
<i>10</i>	<i>31.4</i>	<i>18.0</i>	<i>13.9</i>	<i>15.0</i>	<i>17.3</i>	<i>18.7</i>	<i>7.0</i>	<i>13.8</i>
<i>15</i>	<i>47.1</i>	<i>40.5</i>	<i>70.6</i>	<i>22.5</i>	<i>27.0</i>	<i>27.8</i>	<i>10.3</i>	<i>29.9</i>
16.59	52.1	49.5	105.5	24.9	30.0	30.8	11.4	36.4
20	62.8	71.9	223.1	30.0	36.5	37.1	13.7	52.5
25	78.5	112.4	544.6	37.5	46.1	46.3	17.1	81.3
27.07	85.0	131.8	748.6	40.6	50.0	50.1	18.5	95.1
<i>30</i>	<i>94.2</i>	<i>161.9</i>	<i>1129.4</i>	<i>45.0</i>	<i>55.6</i>	<i>55.5</i>	<i>20.5</i>	<i>116.6</i>
<i>35</i>	<i>110.0</i>	<i>220.3</i>	<i>2092.3</i>	<i>52.5</i>	<i>65.0</i>	<i>64.7</i>	<i>24.0</i>	<i>158.1</i>
<i>50</i>	<i>157.1</i>	<i>449.7</i>	<i>8714.4</i>	<i>75.0</i>	<i>93.4</i>	<i>92.4</i>	<i>34.2</i>	<i>321.1</i>
<i>100</i>	<i>314.2</i>	<i>1798.7</i>	<i>139430.2</i>	<i>149.9</i>	<i>187.8</i>	<i>184.8</i>	<i>68.3</i>	<i>1278.7</i>

2555

2556 (b) Values for magma density, $\rho_l = 3010 \text{ kg m}^{-3}$

2557

E_d /km	L_m /km	W /m	V /km ³	P_0 /MPa	A_u /km	A_l /km	P_n /MPa	W_n /m
<i>5</i>	<i>15.7</i>	<i>3.1</i>	<i>0.6</i>	<i>5.2</i>	<i>5.3</i>	<i>11.3</i>	<i>3.4</i>	<i>3.2</i>
<i>10</i>	<i>31.4</i>	<i>12.5</i>	<i>9.7</i>	<i>10.4</i>	<i>12.8</i>	<i>21.5</i>	<i>6.2</i>	<i>11.2</i>
<i>15</i>	<i>47.1</i>	<i>28.1</i>	<i>49.0</i>	<i>15.6</i>	<i>20.1</i>	<i>32.0</i>	<i>9.2</i>	<i>24.3</i>
<i>20</i>	<i>62.8</i>	<i>50.0</i>	<i>154.9</i>	<i>20.8</i>	<i>27.2</i>	<i>42.5</i>	<i>12.2</i>	<i>42.6</i>
21.93	68.9	60.1	223.8	22.8	30.0	46.6	13.3	51.0
25	78.5	78.1	378.2	26.0	34.4	53.1	15.2	66.0
30	94.2	112.4	784.3	31.2	41.5	63.6	18.2	94.5
35	110.0	153.0	1453.0	36.4	48.6	74.2	21.2	128.2
35.98	113.0	161.7	1623.0	37.5	50.0	76.3	21.8	135.5
<i>50</i>	<i>157.1</i>	<i>312.3</i>	<i>6051.7</i>	<i>52.1</i>	<i>69.9</i>	<i>105.9</i>	<i>30.3</i>	<i>260.3</i>
<i>100</i>	<i>314.2</i>	<i>1249.1</i>	<i>96826.5</i>	<i>104.1</i>	<i>140.5</i>	<i>211.7</i>	<i>60.4</i>	<i>1036.4</i>

2558

2559

2560 Table 3. Parameters controlling whether eruptions or intrusions occur on the Moon from dikes
 2561 connecting shallow mantle magma sources to the surface as a function of the vertical extent of
 2562 the zone of partial melting, E . If an eruption occurs, values are given for the initial magma rise
 2563 speed U_i and the volume flux F_i from a 1600 m long fissure. If no eruption can occur, the depth
 2564 H of the top of an intruded dike is given. In all cases the mean dike width W_{av} is shown; Z is the
 2565 depth of the top of the partial melt zone below the surface.

2566

(a) nearside, $Z = 50$ km.					(b) farside, $Z = 50$ km.				
E	H	W_{av}	U_i	F_i	E	H	W_{av}	U_i	F_i
/km	/km	/m	/(m s ⁻¹)	/(m ³ s ⁻¹)	/km	/km	/m	/(m s ⁻¹)	/(m ³ s ⁻¹)
0	1.1	38			0	6.0			
10		80	4	4.7×10^5	10	4.8			
20		122	16	3.2×10^6	20	3.6	6		
30		164	26	6.9×10^6	30	2.3	79		
40		206	36	1.2×10^7	40	1.1	162		
50		249	45	1.8×10^7	50		251	8.4	3.4×10^6

2567
2568

(c) nearside, $Z = 60$ km.					(d) farside, $Z = 60$ km.				
E	H	W_{av}	U_i	F_i	E	H	W_{av}	U_i	F_i
/km	/km	/m	/(m s ⁻¹)	/(m ³ s ⁻¹)	/km	/km	/m	/(m s ⁻¹)	/(m ³ s ⁻¹)
0		91	4	5.2×10^5	0	4.8			
10		142	16	3.7×10^6	10	3.6			
20		194	26	8.1×10^6	20	2.3	24		
30		245	36	1.4×10^7	30	1.1	62		
40		297	45	2.1×10^7	40		101	5	7.8×10^5
50		348	54	3.0×10^7	50		139	16	3.6×10^6

2569
2570

(e) nearside, $Z = 70$ km.					(f) farside, $Z = 70$ km.				
E	H	W_{av}	U_i	F_i	E	H	W_{av}	U_i	F_i
/km	/km	/m	/(m s ⁻¹)	/(m ³ s ⁻¹)	/km	/km	/m	/(m s ⁻¹)	/(m ³ s ⁻¹)
0		149	15	3.6×10^6	0	3.6	6		
10		208	25	8.3×10^6	10	2.3	61		
20		268	34	1.5×10^7	20	1.1	115		
30		327	44	2.3×10^7	30		169	6	1.6×10^6
40		387	53	3.3×10^7	40		224	19	6.8×10^6
50		446	62	4.5×10^7	50		278	29	1.3×10^7

2571
2572

2573

2574 Table 4. Values of the eruption speeds U_m and resulting maximum ranges R_m of pyroclasts
 2575 erupted from progressively greater depths below the surface and hence progressively greater
 2576 pressures P_{foam} in a decompressing foam layer underlying the dike tip gas cavity in a dike that
 2577 has just reached the surface.

2578

depth below surface/km	P_{foam} /MPa	U_m /(m s ⁻¹)	R_m /km
0.3	0.5	96	5.7
1.0	3.7	113	7.9
3.0	12.8	122	9.1
5.0	21.8	125	9.6
7.0	30.9	126	9.8
9.6	40.0	127	10.0

2579

Table 5. Values of opacity depth, X , and % of pyroclasts landing uncooled as a function of released magma volatile content n for each of a series of values of the erupting magma volume flux F_e . Maximum ranges, R_f , of pyroclasts are repeated for comparison with values of X .

n	R_f	$F_e/(m^3 s^{-1})$		$F_e/(m^3 s^{-1})$		$F_e/(m^3 s^{-1})$		$F_e/(m^3 s^{-1})$		$F_e/(m^3 s^{-1})$		$F_e/(m^3 s^{-1})$	
		$= 5 \times 10^3$		$= 1 \times 10^4$		$= 3 \times 10^4$		$= 1 \times 10^5$		$= 3 \times 10^5$		$= 1 \times 10^6$	
/ppm	/km	X	% hot	X	% hot	X	% hot	X	% hot	X	% hot	X	% hot
250	0.8	8.5 m	97.9	4.3 m	98.9	1.4 m	99.6	0.4 m	99.9	0.1 m	100.0	40 mm	100.0
500	1.7	54 m	93.7	27 m	96.8	8.9 m	98.9	2.7 m	99.7	0.9 m	99.9	0.3 m	100.0
750	2.5	140 m	88.9	70 m	94.4	23 m	98.1	7.0 m	99.4	2.3 m	99.8	0.7 m	99.9
1000	3.2	292 m	83.0	146 m	91.3	49 m	97.1	15 m	99.1	4.9 m	99.7	1.5 m	99.9
1250	4.1	492 m	77.2	246 m	88.2	82 m	96.0	25 m	98.8	8.2 m	99.6	2.5 m	99.9
1500	4.9	788 m	70.4	394 m	84.5	131 m	94.7	39 m	98.4	13 m	99.5	3.9 m	99.8
2000	16.5	1.6 km	56.3	818 m	76.6	273 m	91.9	82 m	97.5	27 m	99.2	8.2 m	99.8
3000	9.8	4.5 km	29.5	2.2 km	59.5	747 m	85.3	224 m	95.5	75 m	98.5	22 m	99.5
5000	16.0	15.3 km	0.2	7.6 km	27.4	2.5 km	70.7	763 m	90.7	254 m	96.9	76 m	99.1
10000	32.7	90.8 km	0.0	45.4 km	15.2	15.1 km	28.8	4.5 km	74.1	1.5 km	91.0	454 m	97.2

Table 6. Flow parameters speed, U_f , Reynolds number, Re_f , emplacement time, t_f , and volume flux, F_f , for 20 km wide, 600 km long mare lava flows with viscosity 1 Pa s emplaced on a slope $\sin \alpha_f = 1 \times 10^{-3}$ with thicknesses between 1 and 30 m.

D_f /m	U_f /(m s ⁻¹)	Re_f	t_f /days	F_f /(m ³ s ⁻¹)
1	0.6	7.4×10^3	11.6	1.2×10^4
2	1.0	2.5×10^4	6.9	4.1×10^4
5	1.9	1.2×10^5	3.7	1.9×10^5
10	3.1	3.7×10^5	2.2	6.1×10^5
20	4.8	1.2×10^6	1.4	1.9×10^6
30	6.2	2.2×10^6	1.1	3.7×10^6

Table 7. Comparison of magma rise conditions in a fissure and in a circular conduit carrying the same volume flux F_e . Values are for the range of eruption parameters shown in Table 2 for magmas of density 2900 kg m^{-3} erupting from mantle sources with the vertical extents E_d . The fissure is 1600 m long in each case. The average fissure width is W_n , the magma rise speed in the fissure is U_e , and the pressure gradient is dP/dz . Thermal erosion of the dike wall at a rate of $15 \text{ } \mu\text{m/s}$ widens the central part of the fissure by a total horizontal amount Y and magma chilling pinches off the distal parts of the dike until magma is rising in a circular conduit of diameter D_c with speed U_c . t_Y is the time needed to accomplish this.

E_d /m	F_e /(m ³ s ⁻¹)	W_n /m	U_e /(m s ⁻¹)	dP/dz /(Pa m ⁻¹)	D_c /m	U_c /(m s ⁻¹)	Y /m	t_Y /days
16.59	1.80×10^5	36	3.1	15.3	209	5.2	173	66
20	6.61×10^5	52	7.9	68.6	260	12.4	208	81
25	1.76×10^6	81	13.5	130.8	339	19.5	257	99
27.07	2.41×10^6	95	15.8	152.4	372	22.1	277	108

Table 8. Parameters relating to eruption of Gruithuisen and Mairan domes. Magma yield strength, τ_d , plastic viscosity, η_d , volume eruption rate, F_d , and surface fissure length estimate, L_d , are from *Wilson and Head (2003b)*. Dike width, W_d , magma rise speed, U_d , and minimum magma rise speed to avoid excessive cooling, U_{\min} , are derived here.

Dome name	τ_d /Pa	η_d /(Pa s)	F_d /(m ³ /s)	L_d /m	W_d /m	U_d /(m/s)	U_{\min} /(m/s)
Gruithuisen γ	7.7×10^4	3.2×10^8	119	7500	173	9.2×10^{-5}	5.0×10^{-6}
Gruithuisen δ	1.3×10^5	1.2×10^9	48	13500	296	1.2×10^{-5}	1.7×10^{-6}
Gruithuisen NW	1.2×10^5	9.9×10^8	24	2500	276	3.5×10^{-5}	2.0×10^{-6}
Gruithuisen NW+ γ	1.0×10^5	6.0×10^8	143	19000	224	3.4×10^{-5}	3.0×10^{-6}
Mairan T	1.3×10^5	1.2×10^9	24	1500	294	5.4×10^{-5}	1.7×10^{-6}
Mairan "middle"	6.9×10^4	2.5×10^8	52	2500	155	1.3×10^{-4}	6.3×10^{-6}
Mairan "south"	5.3×10^4	1.3×10^8	51	1000	119	4.3×10^{-4}	1.1×10^{-5}
Mairan "m"+"s"	6.1×10^4	1.8×10^8	52	9000	137	4.2×10^{-5}	8.0×10^{-6}

Figures and Captions

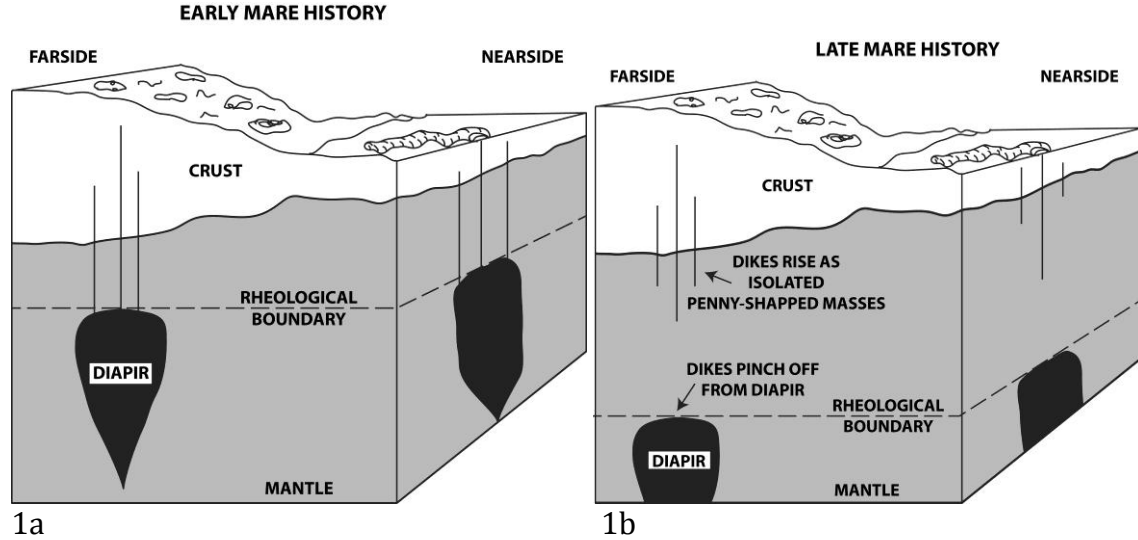


Figure 1. Block diagram perspective view of the lunar surface and interior showing the lunar anorthositic crust (white; thin nearside, thicker farside), the lunar mantle (gray), and basaltic source region diapirs (black) rising from deeper in the mantle. In the early history of mare volcanism (a), the lithosphere (thermal boundary layer) has thickened to greater than the thickness of the crust (compositional layer). Rising mantle diapirs rise, deforming plastically, and stall at the base of the lithosphere (rheological boundary; dashed line). Overpressurization causes the lithosphere to deform elastically and magma-filled cracks (dikes) are propagated toward the surface. For a typical range of overpressurization values, some dikes will reach the surface and form eruptions, while others will stall and solidify in the crust and mantle. The nearside-farside crustal thickness differences will favor eruptions on the lunar nearside, and intrusions on the farside. Later in mare history (b), the cooling of the Moon will thicken the lithosphere, driving the rheological boundary deeper into the interior. Diapir tops are so deep that magma supply is insufficient to permit dikes to be continuous from these depths to the surface; all dikes pinch off from their diapirs and rise buoyantly in the mantle as isolated penny-shaped entities. These dikes stall centered on the crust-mantle boundary; the largest dikes will cause surface eruptions, while others will solidify in the crust and upper mantle. Eventually, overall cooling of the Moon, decrease in sources of melting, thickening of the lithosphere, deepening of source regions, and increasingly contractional stresses in the lithosphere will all work to decrease magma generation, and minimize the likelihood of surface eruptions.

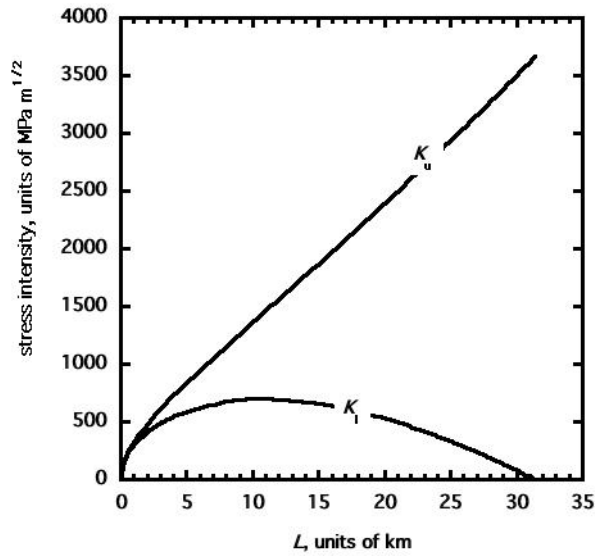


Figure 2. The variation with dike length, L , of K_u and K_l , the stress intensity at the upper and lower tips, respectively, of a growing vertical dike, when the driving pressure at the dike inlet, $P_d = 5.8$ MPa and the density difference between host mantle rocks and magma is $\Delta\rho = 360$ kg m $^{-3}$.

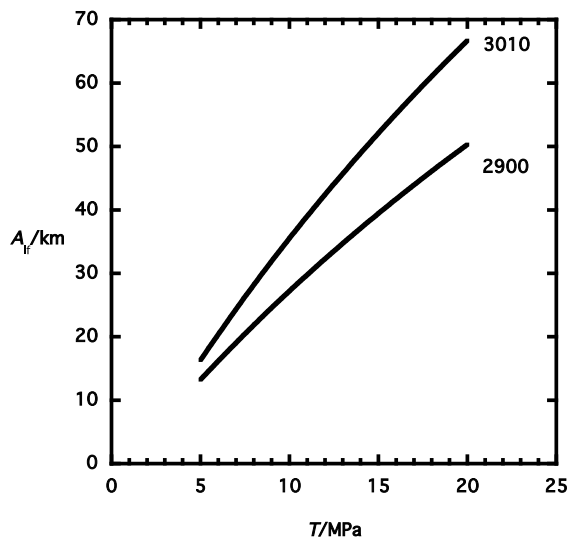


Figure 3. Variation of A_{if} , the vertical extent into the mantle of a rising isolated dike that has penetrated through the crust and initiated an eruption, with T , the lithospheric tensional stress allowing the dike to exist as a stable structure, for two values of the dike magma density in kg m $^{-3}$.

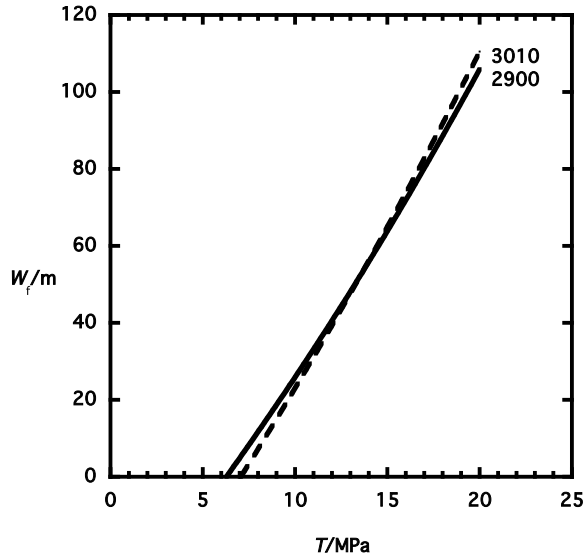


Figure 4. Variation of W_f , the mean width of a rising isolated dike that has penetrated through the crust and initiated an eruption, with T , the lithospheric tensional stress allowing the dike to exist as a stable structure, for two values of the dike magma density in kg m^{-3} . No stable residual intrusion can exist if T is less than 6-7 MPa.

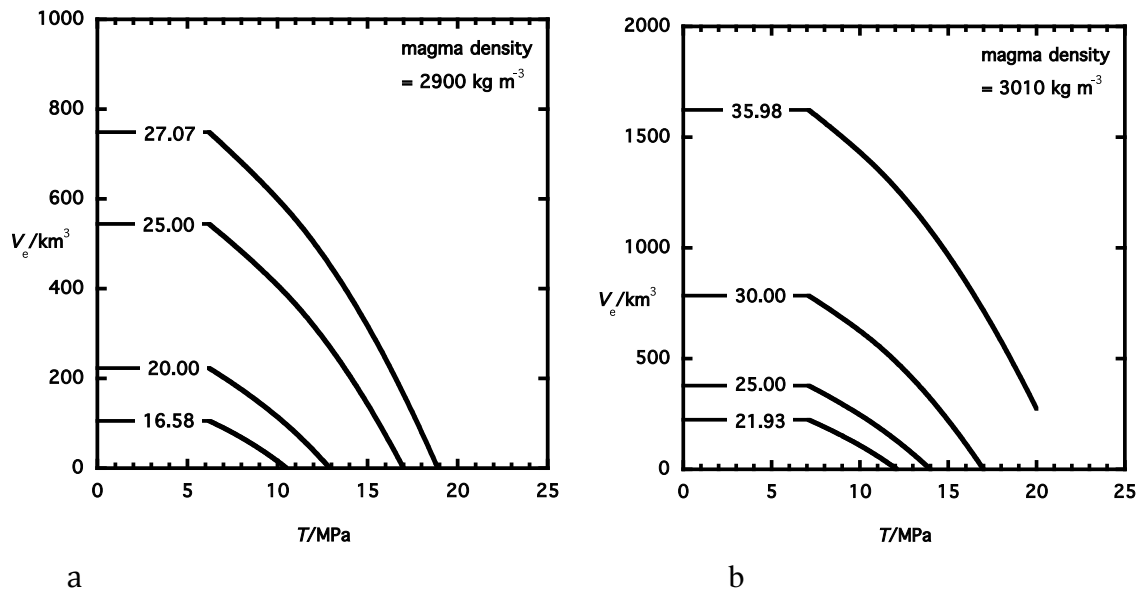


Figure 5. Variation of V_e , the volume of magma that is erupted, with T , the lithospheric tensional stress, after a dike with properties shown in Figs. 1 and 2 has reached its equilibrium configuration. Curves are labeled with the range of vertical extents of the deep mantle partial melt source region that can generate an erupting dike. Part (a), magma density 2900 kg m^{-3} . Part (b), magma density 3010 kg m^{-3} .

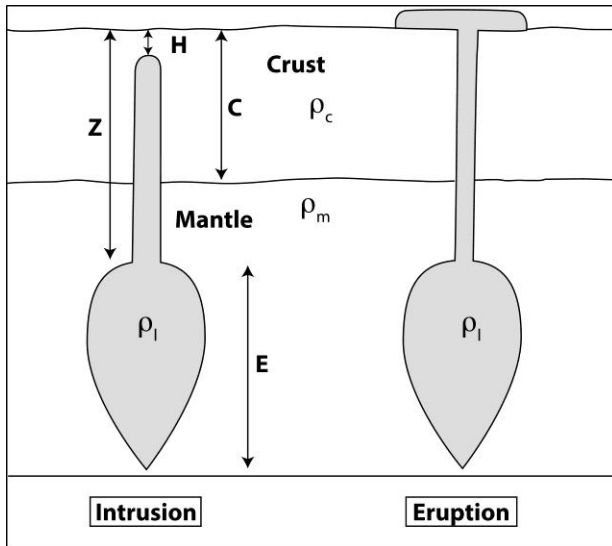


Figure 6. Intrusion and eruption of magma. Early mare basalts are generated by partial melting within a finite region in the upper mantle of vertical extent E , below a crust of thickness C . The level at which the stresses combine to initiate a dike is at a depth Z below the surface. The positive buoyancy of the magma in the mantle diapir leads to an excess pressure at the dike inlet, and this pressure is available to support the column of magma in the dike. If the excess pressure is great enough, the column of magma can be supported all the way to the surface and an eruption can occur. If the pressure is not great enough, the dike will stall with its top at some depth H below the surface. Density of the liquid magma, crust, and mantle are shown by ρ_l , ρ_c , ρ_m respectively.

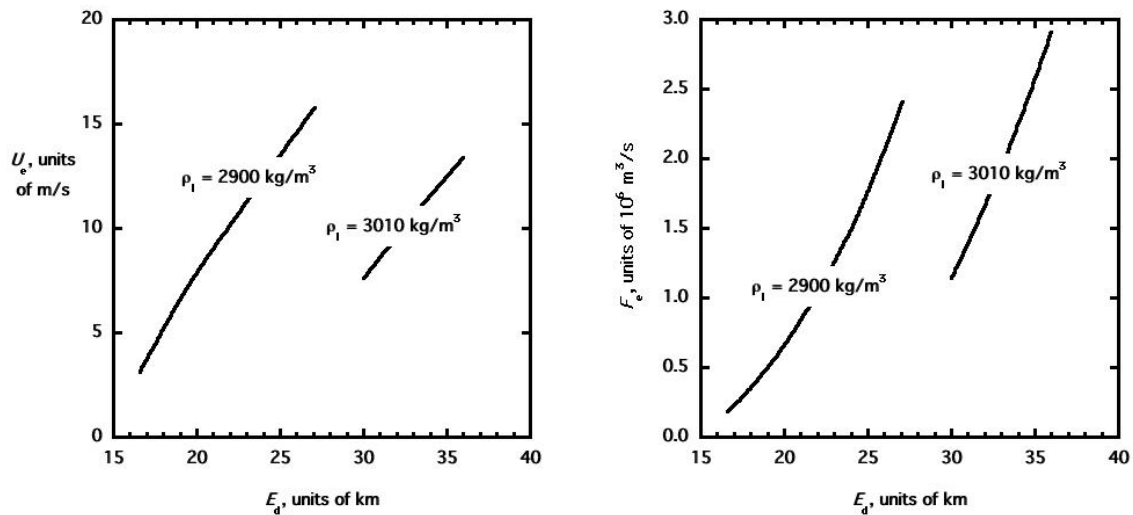


Figure 7. Parameters associated with the range of diapiric source zone extents E_d that allow the upper tips of isolated dikes to reach the surface and produce eruptions on the lunar near-side. (a) the rise speed, U_e , of magma erupting through the dike; (b) the erupted volume flux, F_e . In each case values are shown for magma densities, ρ_l , of 2900 and 3010 kg m^{-3} .

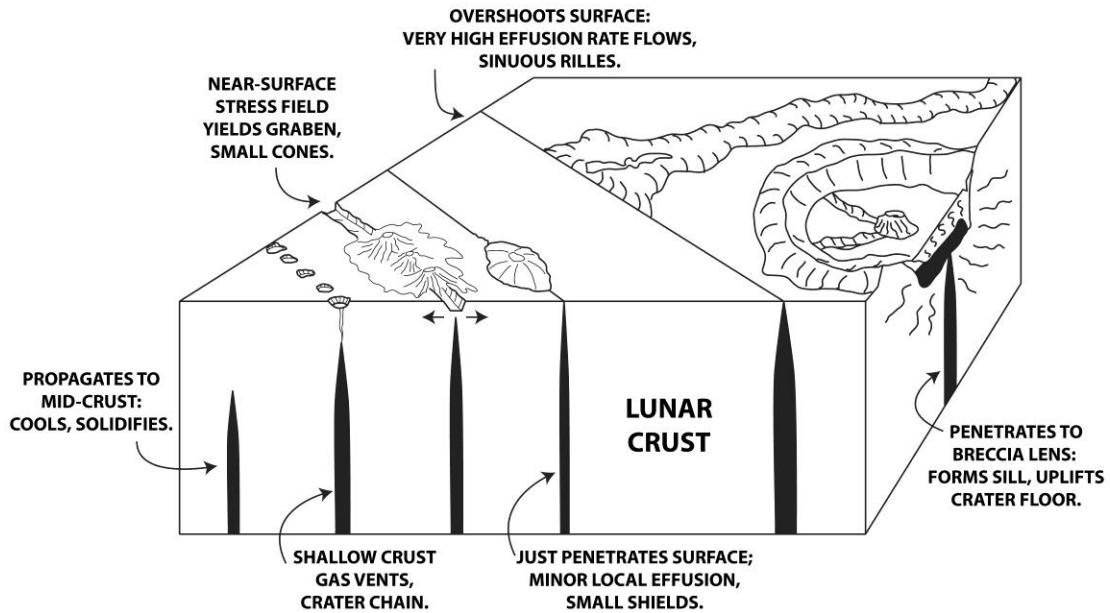


Figure 8. Perspective view block diagram illustrating the fate of dikes intruding into the crust to various levels. Left to right: Dikes propagating only to mid-crustal depths cool and solidify. Those nearing the surface, but not erupting, can vent gas to form a crater chain either abruptly (perhaps some pyroclasts and pulverized regolith), or passively (drainage). Those nearing the surface to form a near-surface extensional stress field will produce graben, and can vent gas and magma to produce an array of cones, domes and pyroclastic deposits. Dike just penetrating the surface, can produce small, low-volume eruptions that form small lava shields. Large dikes penetrating the surface will have very high effusion rates and form very long lava flows and if the eruption duration is sufficiently long to favor thermal erosion, sinuous rilles. Dikes approaching the surface, but encountering a low-density breccia lens below a crater floor, can intrude sills, uplifting the crater floor and forming floor-fractured craters.

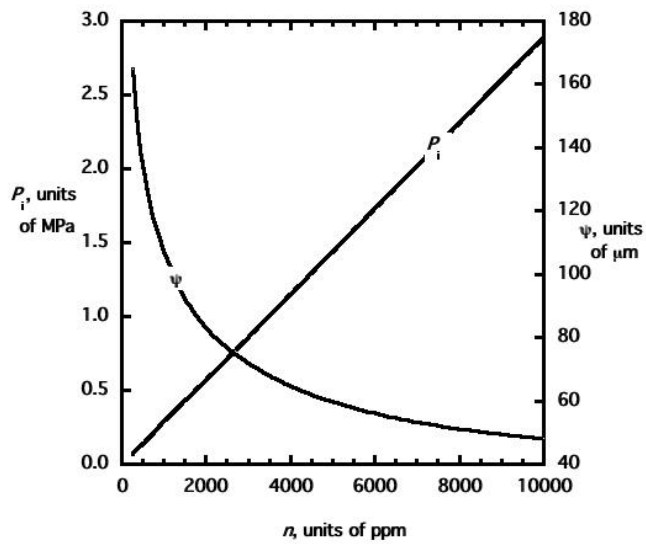


Figure 9. Variation of the pressure in the free gas at the top of the foam layer in a dike tip, P_i , and the diameters of the largest bubbles at the top of the foam, ψ , as a function of the total amount of CO-dominated gas released from the magma, n .

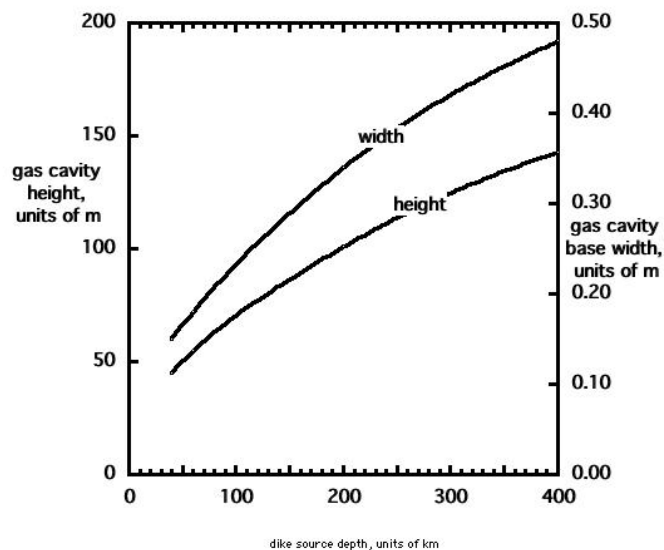


Figure 10. Minimum vertical extents and basal widths of gas cavities in the tips of dikes that have propagated from sources at the stated depths.

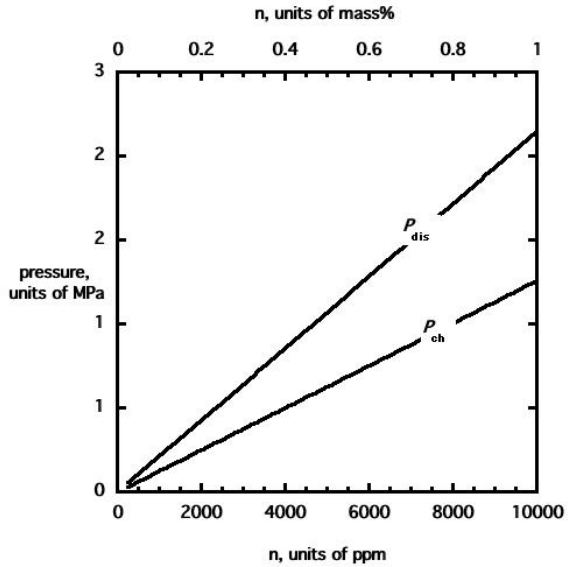


Figure 11. Values of pressure in magma at depth of disruption into pyroclasts, P_{dis} , and pressure in choked gas-pyroclast mixture exiting the vent, P_{ch} , for eruptions on the Moon and Earth as a function of released volatile contents n , given in both ppm and mass % for comparison with common usage. Values of n up to ~ 2000 ppm apply to most eruptions on the Moon and values greater than ~ 2000 ppm are relevant to most basaltic eruptions on Earth.

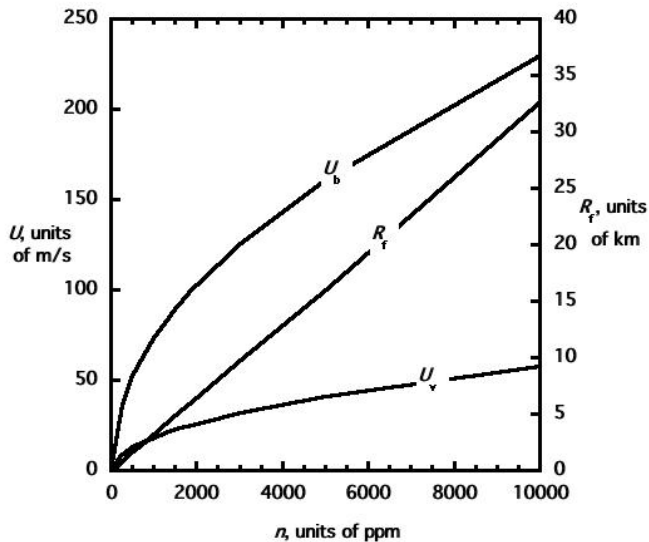


Figure 12. Implied speeds in the vent, U_v , speeds after decoupling from the gas expansion, U_b , and maximum ballistic ranges, R_f , of gas and 300 μ m diameter pyroclasts in steady explosive eruptions as a function of the CO content, n , of the erupting magma.

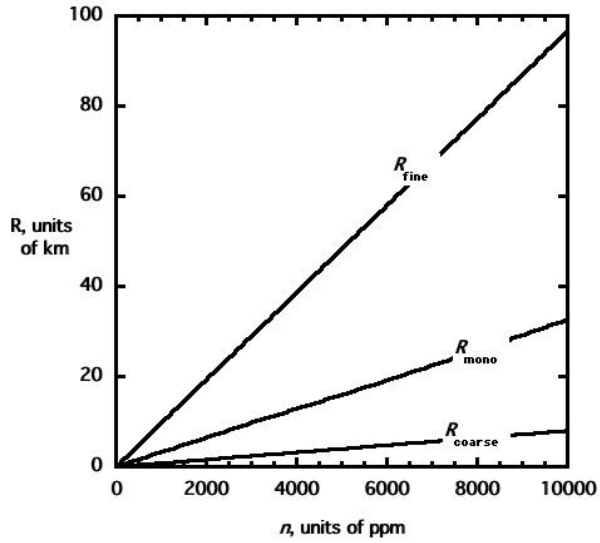


Figure 13. Pyroclast ranges R_{mono} in steady eruptions of monodisperse 300 μm diameter pyroclasts compared with the maximum ranges of the coarse and fine size fractions of a distribution in which 80% of clasts are much larger than ~ 1 mm and decouple rapidly from the expanding gas phase.