Dark ring in southwestern Orientale Basin:
Origin as a single pyroclastic eruption

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[1] A large, 154 km diameter dark annular ring, located in the southwestern part of the Orientale Basin along the Montes Rook ring, was first discovered and documented in Soviet Zond 8 images. Clementine UV-visible multispectral data indicate that the dark ring consists of material similar to the pyroclastic glasses collected by the Apollo astronauts, with the glasses being more closely related to the orange glass beads that comprise the Aristarchus Plateau than to the crystallized black beads typical of Taurus-Littrow. This implies relatively rapid cooling times for the eruption products. We propose that the dark ring is the manifestation of a pyroclastic eruption originating at a fissure vent, an elongate 7.5 km by 16 km depression, located near the center of the ring. The event producing the eruption began with a dike rapidly emplaced from subcrustal depths to within ~3–4 km of the surface. The dike stabilized and degassed over ~1.7 years to form an upper foam layer which then penetrated to the surface to cause an eruption, lasting ~1–2 weeks. The eruption produced a ~38 km high symmetrical spray of pyroclasts into the lunar vacuum at velocities of ~350 to ~420 m/s, and the pyroclastic material accumulated in a symmetrical ring around the vent. The geometry of eruption caused the deposits to accumulate preferentially in a ring representing the material ejected at 45°. The paucity of pyroclastic rings of this type on the Moon can be attributed to the low probability of a dike stalling at just the right depth (~3–4 km) to create these eruption conditions. The detailed characteristics of this ring provide important new insight into the emplacement of pyroclastics in the more regionally continuous lunar dark mantle deposits, suggesting that their sources are dominated by effusive and Hawaiian-style eruptions. The Orientale dark ring deposit has similarities to the pyroclastic rings on the Galilean satellite Io. Both the Orientale dark ring and Ionian eruptions involve acceleration of small particles to a high velocity in an expanding gas stream, silicate pyroclasts derived from the disrupted foam layer in the lunar case, and a mixture of silicate pyroclasts and “snowflakes” of condensing SO2 solids on Io. In both cases the silicate pyroclasts ejected from the vent are accelerated by the gas until they become decoupled and continue on ballistic trajectories controlled only by gravity. Differences in mass flux and cloud opacity between the Moon and Io are the direct result of the origin and mass fraction of the volatile phase: the accumulation of a magmatic foam layer on top of a silicate intrusion in the lunar case and the intimate mixing on Io of a steadily erupting magma and liquid SO2. The interpretation of the Orientale dark mantle ring as a pyroclastic eruption provides an alternative to the hypothesis that the dark ring represents the presence of an ancient pre-Orientale impact structure and that the Orientale Basin cavity of excavation must therefore lie within the Outer Rook Mountain ring.

KEYWORDS: Orientale Basin, Moon, pyroclastics, dark mantle, Io, dike

1. Introduction and Background

[2] The Orientale dark ring was originally detected in Soviet Zond 8 images taken under lighting and viewing conditions which emphasize the albedo of surface deposits [Lipsky, 1975; Rodionov et al., 1976]. In these images (Figure 1a) this feature appears as a dark circular deposit about 175 km in diameter that mantles some of the massifs of the outer part of the Rook Mountain ring of the Orientale Basin. The deposit consists of two arcs of dark material that appear to be bisected by the bright inward facing massifs of the Orientale Basin Rook Mountains (Figures 1 and 2). Individual massifs and hills of
the Orientale Basin deposits protrude through the dark ring or are incompletely covered by ring deposits. Lunar Orbiter images (Figure 1b) complemented the Zond view, and together these data led to the interpretation that the ring mantled both the hummocky inter-ring deposits and the fissured and corrugated Maunder Formation of the Orientale Basin [Scott et al., 1977]. Schultz and Spudis [1978] noted that the materials of the ring were not evenly distributed and that they were often locally concentrated. They further pointed out that Montes Rook were discontinuous in the area of the ring (Figure 2) and that a part of the Montes Rook extends to the north into the basin for ~60 km along the western part of the dark ring; they suggested that the unusual massif topography could represent the presence of a large, pre-Orientale Basin impact crater spanning the Rook Mountain ring. Citing the common occurrence of annular volcanic vents in craters that have been modified volcanically [e.g., Schultz, 1976], Schultz and Spudis [1978] suggested that the dark halo deposit was produced by several vents aligned along preexisting zones of weakness associated with the postulated pre-Orientale 175 km basin. Thus, in this interpretation the deposit was produced by a large number of individual post-Orientale volcanic eruptions localized into a circular annulus by the influence of preexisting (and pre-Orientale) impact crater/basin structure. This interpretation, if correct, places important constraints on the location of the Orientale transient cavity. The presence of a buried crater spanning the outer Rook Mountain ring would place the cavity inward of that area, in contrast to the ring location of several other hypotheses [e.g., Head, 1974b;
Moore et al., 1974; Howard et al., 1974] (see review by Head [1977]).

[3] More recent Galileo images provided the first multispectral images of this region and the dark halo deposit itself. Greeley et al. [1993] and Kadel [1993] examined the geology and setting of the deposit and its characteristics in multispectral images. These analyses reveal a moderate 1 μm absorption, implying a pyroclastic deposit contaminated with the underlying highlands. Greeley et al. [1993] concluded that the deposit was emplaced as a series of pyroclastic eruptions using reactivated fractures associated with the ancient basin proposed by Schultz and Spudis [1978] as conduits. Using the Galileo data, Pieters et al. [1993] showed that the ring deposit is somewhat brighter than nearside mantling deposits, has a very weak 1 μm absorption band, and has a ultraviolet-visible (UV/VIS) ratio that is relatively high but lower than that seen in nearside deposits of black spheres. Pieters et al. [1993] concluded that these characteristics could be consistent with contamination of the deposit by highland material as suggested by Greeley et al. [1993] but that another possibility was that the spectral properties of the deposit itself dominate the measured properties. In the former case the deposits could be interpreted as similar to ilmenite-rich dark mantling material such as the black beads sampled at Apollo 17 [e.g., Pieters et al., 1974]. In the latter case the weak 1 μm band may indicate that the deposits are not homogeneous glass but are in a crystallized form. The low UV/VIS ratio, relative to black beads seen elsewhere, could be due to the lower abundance of ilmenite, the opaque component that causes darkening. In this case the Orientale ring deposit might have affinities with the local medium-TiO₂ deposits identified in the Orientale region [Greeley et al., 1993].

[4] In this paper we first analyze Clementine data of the dark halo region and produce a geologic map of the region (Figure 2).

Figure 2. (a) Clementine images and (b) geologic sketch map of the region of the Orientale dark ring deposit (DRD). Clementine 750 nm images from orbits 203-204-205.
using these Clementine data combined with Zond and Lunar Orbiter images (Figure 1). We then analyze Clementine UV/VIS data taken of the Orientale ring deposit and its surroundings in order to characterize and map out the nature of the deposit and the distribution of glasses within it (Figures 3 and 4). Using this information to constrain deposit characteristics and thicknesses, we then explore the possibility that the depression shown in Figure 3 is, in fact, the vent for this deposit and test the hypothesis that the ring could have been produced by an eruption from this vent. We model the eruption conditions that could have existed as a basaltic dike ascended to the vicinity of the surface (Figure 5) and the conditions that would exist within the volcanic plume required to have emplaced this deposit. We find that a good case can be made for the dark ring having formed as a result of a pyroclastic eruption from the near-surface emplacement of a basaltic dike (Figure 6).

2. Dark Mantle Deposits: Background and Characteristics

Regional and localized dark mantle deposits have been recognized on the Moon, generally in association with sinuous rilles or irregular depressions [Head, 1974a; Head and Wilson, 1980; Wilson and Head, 1980; Guddis et al., 1985; Hawke et al., 1989; Weitz and Head, 1999; Weitz et al., 1998]. Regional deposits are more extensive, with areas >1000 km², and are located on the uplands adjacent to younger mare [e.g., Head, 1974a; Weitz et al., 1998]. In contrast, localized deposits are smaller in extent and are more widely dispersed across the lunar surface [Head, 1976; Hawke et al., 1989; Coombs et al., 1990]. Surface geologic exploration [Schmitt and Cernan, 1973; Muehlberger et al., 1975; Pieters et al., 1974] and multispectral telescopic data on the dark mantle deposits (DMDs) indicate that they are composed of submillimeter volcanic glasses or crystallized beads. The Taurus-Littrow DMD spectrally matches the crystallized black beads collected at the Apollo 17 landing site [Pieters et al., 1974], while the Aristarchus Plateau DMD is dominated by orange/red glasses [Zisk et al., 1977]. Analyses of remote-sensing data on several localized DMDs by Hawke et al. [1989] indicate three compositional groups, each reflecting variations in the eruption conditions that emplaced them. [6] DMDs are thought to have originated by explosive volcanic activity [McGetchin and Head, 1973; McGetchin et al., 1974; Heiken et al., 1974; Head and Wilson, 1979, 1992; Wilson and Head, 1981a]. The low gravity and lack of an atmosphere on the Moon caused submillimeter gas bubbles in the magma to burst at the surface [Wilson and Head, 1981a], resulting in the submillimeter-sized volcanic glass particles identified in all the Apollo soils [Delano, 1986]. The regional DMDs may reflect larger, Hawaiian-style fire fountain eruptions, whereas the localized deposits may have been emplaced by strombolian or vulcanian eruptions [Head, 1975; Wilson and Head, 1981a; Hawke et al., 1989; Head and Wilson, 1992; Weitz et al., 1997, 1998]. The extent of crystallinity in the beads depends on their cooling history in the fire fountain. Glasses tend to be smaller in size and probably experienced a shorter time suspended in an optically thin fire fountain [e.g., Head and Wilson, 1989]. In contrast, the crystallized beads reflect larger cooling times either because of their larger sizes, because of their longer residence times in contact with hot gas, or because they erupted inside the optically dense portion of a fire fountain, where cooling was inhibited. Experiments by Arndt and von Engelhardt [1987] suggest that the crystallized black beads from the Apollo 17 landing site had cooling rates of 100 K s⁻¹, which is much slower than blackbody cooling expected in a vacuum. We use these characteristics and interpretations to aid in the mapping and analysis of the Orientale dark ring deposit.

3. Geology of the Orientale Dark Mantle Ring Deposit

The Orientale dark mantle ring deposit (DMRD) is centered on the Montes Rook in southwestern Orientale (Figure 1) at 29°S
Figure 4. Clementine UV/VIS spectra. (a) Spectra obtained from the region of the Orientale ring deposit and its surrounding (see Figure 2 for unit locations). Spectra represent an average of 4 × 4 pixel boxes at each location. The dark mantle deposit (DMD) has a relatively high reflectance compared with other dark mantle deposits [see Weitz et al., 1998] and a strong glass band absorption. The hilly terrain is characterized by an anorthositic signature. The vent area and several of the impact craters have noritic signatures. The mafic absorption distinguishes the dark streak (mafic debris in Figure 2b) from the DMD. (b) Spectra for regional dark mantle deposits for comparison to Orientale DMD. Spectra are shown only for mature soils to avoid maturity difference variations. From Weitz et al. [1998].
and 263°C. Clementine data provide detailed information necessary to further characterize the ring and its relationship to adjacent and underlying topography (Figures 2 and 3). The DMRD lies astride the Outer Rook Mountain ring. To the east the mountain ring is linear for a distance of almost 200 km along the southern part of the basin. In the vicinity of the DMRD it becomes extremely sinuous, breaking up into a series of valleys and broad ridges 15–25 km wide and oriented radial and subradial to the basin. Farther to the west the Orientale Basin rings are more poorly defined and lose much of the distinctive continuity seen in the eastern part of the basin [e.g., Head, 1974b; Scott et al., 1977].

[8] Scarpas associated with the Outer Rook Mountain ring are up to several kilometers high [Head, 1974b; Zuber et al., 1994; Neumann et al., 1996] and commonly appear bright owing to the shedding of optically mature soils from their relatively active

Figure 5. Interpreted sequence of events leading to the formation of the Orientale dark mantle ring deposit: (a) ascent of a basaltic dike from subcrustal depths to near the surface and the buildup of gas and (b) the eruption to the surface to produce the volcanic plume and deposit.
slopes [e.g., Head et al., 1993; Pieters et al., 1993]. Clementine images reveal that the DMRD has locally been shed off these slopes and has accumulated in adjacent lows (see Figure 2 at 3 o’clock and 10:30 o’clock). The scarps and massifs of the mountain range front are associated with hummocky terrain developed distally from the basin; this terrain is characterized by 2–8 km diameter closely packed domes and massifs forming a continuous deposit adjacent to the main topographic ring and mapped as “basin ring material” in Figure 2 (the massif facies of the Montes Rook Formation of Scott et al. [1977]). Here, too, when the dark ring material is superposed on these deposits, much of the material has been shed off the steeper slopes of the domes and massifs into the low-lying immediately adjacent areas to give the ring a mottled appearance (e.g., 8:30–10:30 o’clock and 3–4 o’clock in Figure 2).

[9] Another type of basin deposit is seen between the Outer Rook and Cordillera Mountain ring to the south. This consists of compact closely spaced small domes and hills less than ~3 km in diameter and is the “domical facies” of Head [1974b] and the knobby facies of the Montes Rook Formation of Scott et al. [1977]. Here, too (Figure 2, between 4:30 and 8:30 o’clock), the DMRD is relatively patchy, primarily because the material appears to have been shed off the small domes and concentrated in the low-lying areas or to have been preserved preferentially in local areas where hummocks are not prominent (e.g., Figure 2, in the outer parts of the deposit at about 4:30 and 6 o’clock).

[10] Inward from the Outer Rook Mountain front the basin is characterized by distinctly different deposits in the form of a “corrugated” facies [Head, 1974b] consisting of cracked and fissured deposits that have been interpreted to be an impact melt sheet associated with the Orientale impact basin event (the Maunzer Formation of Scott et al. [1977]). Interspersed among these facies are patches of relatively smooth light plains (Figure 2, within the DMRD) that have been interpreted to be smoother facies of impact melt [Head, 1974b; Head et al., 1993]. The corrugated and plains deposits are much smoother and have many fewer steep slopes than the other Orientale Basin deposits and thus have preserved the original distribution and characteristics of the deposits more clearly and with less subsequent downslope movement (e.g., Figure 2, between 11 and 2:30 o’clock and between 10 and 11 inward of the massifs). Even here, however, at highest resolution the local massifs and small hummocks have shed dark mantle deposits into adjacent lows (Figure 2).

[11] Several impact craters ranging from about 4 to 8 km in diameter have been superposed on the DMRD and have excavated underlying highlands material, causing disruption of the continuity of the DMRD. The four largest ones are located at 12:30, 3:30, 6:30, and 8:30 o’clock (Figure 2). The closest mare deposits are located ~100 km to the north of the DMRD, and none have been mapped within or immediately adjacent to the DMRD [Scott et al., 1977].

[12] At the approximate center of the ring is an elongate depression 7.5 km wide by 16 km long discovered in the Clementine images (Figures 2 and 3). The elongate depression was largely obscured in the Lunar Orbiter images (Figure 2) by shadows cast by the massifs due to the low solar elevation. The depression trends slightly west of north and is nested within one of the reentrants of the Outer Rook Mountain ring. It is slightly narrower toward its center, and the southern half of it has been somewhat disrupted by a nearby crater and its ejecta. The interior of the elongate depression has an apparently flat floor which makes up only a small part of the total interior; most of the interior is dominated by talus slopes which extend toward the interior to form the depression and sometimes meet at the base. In the absence of any other way of estimating the depth of the depression we assume that its slopes are ~30°, in which case the depth of the depression is ~2 km. Since the slopes could be less but are unlikely to be more, we note that this implies that we may be overestimating all of the subsequent parameters derived from this depth, perhaps by as much as ~20–30%. This elongate depression is similar to features seen in association with other dark mantle deposits (e.g., Sulpicius Gallus [Lucchitta and Schmitt, 1974]), but it has no obvious adjacent deposits of dark mantle or mare. On the basis of its location in the approximate center of the DMRD, we investigate the possibility that it is a source crater for an eruption producing the DMRD.

[13] Although prominent in the Zond and Lunar Orbiter images (Figure 1), the DMRD as seen in the Clementine images (Figures 1 and 2), and as described above, is not uniform in albedo or shape. Analysis of its distribution and relation to preexisting (Orientale Basin) and later (impact crater) deposits provides the basis to assess its original configuration immediately following its emplacement. The DMRD is darkest and widest (47 km) to the north (Figure 2), where it is superposed on the corrugated and relatively smooth Maunzer Formation. It is narrowest and brightest in the west, where it is superposed on the massif facies of the Montes Rook Formation and where the Rook Mountain scarp parallels its arc.

[14] Is the deposit symmetrical around the elongate depression? The closest distance from the elongate vent to the inner edge of the deposit is 49.4 km to the west. The farthest distance to the outer edge of the deposit is 108.3 km to the north. Even though the DMRD is darkest and widest to the north, an 8.3 km diameter crater caused some mixing with substrate and consequent brightening of the DMRD here. In the south a 9.7 km diameter crater impacted into the middle of the DMRD, causing significant mixing with brighter material and burial of the DMRD. In the west, highland downslope movement has shed dark mantle off slopes, making it difficult to map out the distribution of dark mantling material margins in detail. To the east, similar processes have caused an increase in the DMRD albedo; however, the increase is not as large as in the west, and more confident boundaries can be drawn. The average radius of the DMRD from the center of the vent is 77 km, and when a circle of this radius is centered on the elongate depression (Figure 3), the deposit seems generally symmetrical; only in the west does the deposit appear closer to the vent than this average radius. In summary, on the basis of these data, we adopt an average radius of 77 km for the DMRD and note that the elongate depression is slightly off-center to the east of the center of the DMRD.

4. Clementine UVVIS Data

4.1. Calibration and Procedure

[15] We have used remote-sensing data from the Clementine UV/VIS camera to characterize the nature and detailed distribution of dark mantle in the Orientale deposit. The UV/VIS data consist of 5 spectral channels: 0.415, 0.75, 0.9, 0.95, and 1.0 μm. The data were calibrated by: (1) removing electronic noise and correcting for frame transfer; (2) subtracting the flat fields derived for each channel from preflight and in-flight data; (3) photometrically correcting using standard RELAB bidirectional viewing geometry to account for differences in viewing geometry from scene to scene; and (4) spectrally calibrating to transform the calibrated DN data into bidirectional reflectance data using Apollo 16 soil as ground truth [Pieters et al., 1996]. However, calibration errors still exist, especially in the 0.415 and 1.0 μm channels, which were taken at longer exposure times. Images from orbits 203, 204, and 205 taken at 181 m/pixel resolution and 29.2°–33.9° phase angle cover almost all of the deposit, while orbits 71 and 72 (111 m/pixel and 32°–37° phase) cover the small gaps in the north not imaged by the latter orbits. In order to obtain spectra from both the DMD and the highland material, we used only the short exposure images.
5. Implications for the Eruption Conditions

[20] Because the elongate depression is located at the approximate center of the DMRD ring, we interpret it to represent a volcanic vent. In contrast to the interpretation of Schultz and Spudis [1978] that the DMRD ring represents deposits from a remnant system of multiple vents located along the deposit itself, we test the hypothesis that the DMRD could form as a deposit resulting from an eruption or eruptions from a single site, the central elongate vent. In this case the Orientale DMRD would be a 77 km radius annular deposit with the ejecta concentrated at the maximum distance from the elongate vent [e.g., Wilson and Head, 1981a].

[21] Wilson and Head [1981a] have shown theoretically that eruptions from a fissure source will concentrate the ejecta in the outer part of the deposit, while eruptions from a circular vent concentrate ejecta both near the vent and at the maximum range. The fact that the Orientale DMRD erupted from an elongate vent producing a deposit at the maximum distance is therefore in agreement with the theoretical studies. Elongate, fracture-bounded subsidence depressions become more circular as they are enlarged by further collapse, but the reverse is never seen; hence a large, elongate depression is extremely likely to imply that the original feature was even more elongate. Additionally, Wilson and Head [1981a] have shown that the maximum range of submillimeter ejecta is a function of their mass fraction in the total amount of ejecta and of the gas mass fraction in the erupted material. If we assume that the deposit is composed of the same submillimeter glasses sampled at all the Apollo landing sites, then to be ejected to more than 50 km, either (1) the glasses must represent only a few wt % of the total ejecta and the mass eruption rate and gas content must be high (i.e., >10⁶ kg s⁻¹ mass eruption rate and >750 ppm gas content) [Wilson and Head, 1981a] or (2) some process must have operated to concentrate a significant amount of magmatic gas into the erupted material.

[22] In a preliminary step we modeled the eruption that produced the Orientale DMRD using simple ballistic trajectories from a central vent, similar to models developed for eruptions on Io [Strom and Schneider, 1982]. Using the average radius of 77 km and assuming that material reaching the maximum range is ejected with an elevation of 45°, we calculate an eruption velocity of 354 m s⁻¹ and a flight time of 5.1 min. The greatest height above the surface for particles on this trajectory is ~19.2 km, and the greatest height reached by particles projected vertically from the vent is ~38.4 km. The farthest deposit is located at 108 km radius, which yields an ejection velocity of ~420 m s⁻¹ and a flight time of 4.1 min. Arndt et al. [1984] calculated that the Apollo 15 green glasses must have been suspended in a hot environment for ~10 min to produce the cooling rates required to form the mineral textures in the glasses. This value is considerably longer than the times calculated for simple ballistic transport in the present case [Weitz et al., 1998] and implies that the DMRD particles should be dominated by glasses rather than crystallized beads, consistent with the Clementine data.

[23] Unlike all the other observable regional DMDs on the Moon, the Orientale DMRD is unique in its circular planform and annular distribution around a candidate elongate depression and its lack of associated or nearby mare deposits. Other regional DMDs cover larger areas and appear homogenous in their distribution, making it difficult to determine the location of the vent. If emplaced as we propose, the volcanic plume associated with the Orientale DMRD may have resembled the umbrella-shaped plumes seen on Io that also produce an annular deposit around the vent [Strom and Schneider, 1982; McEwen et al., 1998, 2000] (Figure 8). In contrast, the other regional DMDs on the Moon may have been produced from plumes similar to those seen in Hawaii, where a large range of clast sizes are ejected in a broad plume, and the deposition distance of the clasts from the vent is a function of size. Alternatively, some regional deposits may have been emplaced in an annular mode around a vent, but deposits between the vent and...
the edge of the annulus may have been thicker, or subsequent mare deposits may have embayed the deposits, obscured symmetry, and made recognition difficult. We will return to this question in section 6.2.

6. Detailed Theoretical Analysis

6.1. Subsurface Magma Accumulation and Eruption

6.1.1. Basic model of magma intrusion. [24] Magma generated at depth in the lunar mantle is neutrally buoyant at the base of the anorthositic highlands crust and may stall there [Solomon, 1975; Head and Wilson, 1992]. Overpressurization of source regions causes rupture and propagation of magma-filled cracks (dikes) toward the surface. As they approach the surface, they may stall at various depths, producing a variety of conditions, or proceed to the surface to produce an eruption (Figure 5a). In the near-surface environment, stresses associated with the shallow stalling of the dike may cause graben formation, and the subsequent exsolution of gas from the magma after the dike has stalled may cause volcanic eruptions [Head and Wilson, 1979] or spatter cones [Head and Wilson, 1994]. The large size of the elongate depression at the center of the DMRD and its lack of an adjacent raised rim built up of pyroclastic deposits suggest that it may have formed by collapse. On the basis of the nature of the DMRD and its central elongate depression, the assumption that we make here is that the observed depression formed by the collapse of surface layers into a void space. This void space was formed from a zone of vesicular magma at the top of a dike that was erupted explosively to form the observed dark mantle deposit. We shall shortly show that the mass fraction of gas in the erupted mixture of fragmented magma (pyroclastic droplets) and free gas must have been much greater than the amount which could plausibly have been generated by bringing that magma to the low-pressure, near-surface environment. This implies that gas was concentrated into the erupted magma (prior to its fragmentation) by the rise of bubbles from below to form a foam layer at the top of the dike.

[25] We first estimate the total mass of erupted material. The apparent source depression measures $L = 16$ km along strike and is $W = 7.5$ km wide. If we assume that it has a triangular cross section and that its depth is $D = 2$ km as estimated earlier, then its volume is $(16 \times 7.5 \times 2 \times 0.5) = 120$ km$^3$, or $1.2 \times 10^7$ m$^3$. This must be equal to the volume $V_0$ of vesicular magma that has been erupted as pyroclasts. Assume that this magma (which we shall see shortly is likely to be very vesicular) has a bulk density $\beta$; then its mass is $M$, where

$$ M = V_0 \beta; \quad (1) $$

hence $M = (1.2 \times 10^{11} \beta)$ kg. The dense rock equivalent volume that this represents is $(M/\rho)$, where $\rho$ is the unvesiculated density of the magma, which we take as $3000$ kg m$^{-3}$ for a lunar basaltic melt. If we assume that the vesicular magma is converted, during the ejection process, into small pyroclastic glass spheres [Heiken et al., 1974] which have lost essentially all trapped gas, then these spheres will have a bulk density (assuming they form an offset cubic close packing array on landing) of $\rho' = (0.74 \rho)$, and so the deposit will have a total bulk volume $V_{dep}$ given by $M/(0.74 \rho)$, i.e.,

$$ V_{dep} = 1.351(M/\rho); \quad (2) $$

and using (1), $V_{dep} = 1.62 \times 10^{11} (\beta/\rho)$ m$^3$.

[26] We now investigate the likely value of $\beta$, the density of the preereption magma, as follows. Let the magma contain accumulated gas bubbles which occupy a volume fraction $f$, and let the excess pressure in the gas (defined as the pressure in excess of the hydrostatic load of the magma) be $P_e$. This excess pressure is defined as the pressure in excess of the hydrostatic load of the magma. An excess pressure, inherited from the magma source zone at depth, is always required to keep a dike open as it propagates toward the surface [Pollard, 1987]. For the moment we neglect the hydrostatic pressure, a valid approximation at sufficiently shallow depths in the vesicular magma, and just consider the effect of the excess pressure $P_e$ (we deal with the more general case later in section 6.1.3). Applying the perfect gas law to the gas (a reasonable approximation at the pressures and temperatures involved), its density is

$$ \rho_e = \frac{(mP_e)}{(Q)}; \quad (3) $$

where $m$ is the molecular weight of the gas ($28$ kg kmol$^{-1}$ for carbon monoxide, the most likely candidate [Sato, 1979]), $0$ is the temperature (say, $1600$ K [Witzel et al., 1977]), and $Q$ is the universal gas constant ($8.314$ kJ mol$^{-1}$ K$^{-1}$). The bulk density of the magma is given by

$$ \beta^{-1} = \left( \frac{n}{\rho_e} \right) + \left( \frac{1-n}{\rho} \right), \quad (4) $$

where $n$ is the mass fraction of gas in the erupted magma, and the gas volume fraction $f$ is given by

$$ f = (n \rho)/[n \rho + (1-n)\rho_e]. \quad (5) $$

6.1.2. Estimate of gas content and vertical extent of foam layer. [27] To find $n$, we note that this eruption projected pyroclasts to a maximum range $R = 108$ km. The minimum velocity needed to do this is $U = (g R)^{1/2}$, where $g$ is the acceleration due to gravity, and for a deposit of this size we can neglect the effect of the curvature of the Moon’s surface on the range. Using $g = 1.63$ m s$^{-2}$ and $R = 108$ km, we find $U = 420$ m s$^{-1}$. This ejection velocity is a result of the kinetic energy resulting from decompression of the gas contained in the bubbles dispersed through the magma from an initial pressure $P_e$ to a hard vacuum. This kinetic energy is shared between the gas itself and the entrained pyroclasts via fluid-dynamic drag forces which cause the particles to be accelerated along with the gas. The particles always have a lower velocity in the vertical direction than the gas because their gravitational weight causes them to fall through the gas at a finite terminal velocity. However, for the submillimeter-sized particles typical of lunar pyroclasts these terminal velocities will be $<1$ m s$^{-1}$ in the early stages of the gas expansion. Small energy losses due to friction with the vent walls can be neglected when the magma is released from shallow depth as in this case, as can the potential energy needed to raise the magma to the surface as long as that depth is only a few kilometers. Wilson and Head [1981a] discuss this process in detail, drawing attention to the fact that while pyroclasts are closely spaced just above the vent, the decompression process may be nearly isothermal, but it must eventually become adiabatic as the efficiency of thermal coupling between clasts and gas decreases. There are two reasons for this. First, the expansion of the gas causes the clasts to become widely separated, which both reduces the efficiency of conduction and convection in the gas and reduces the amount by which clasts in the outer part of the cloud prevent clasts in the inner part from radiating freely to space. Second, as the gas density decreases, the mean free path of the molecules eventually becomes greater than the typical sizes of the pyroclasts, and the drag forces become negligible [Wilson and Keil, 1997]. Over a small range of pressures which, for carbon monoxide gas at $1600$ K is centered on $P_e = 10$ kPa, the pyroclasts become decoupled from the gas motion and continue with their current speeds and directions on free ballistic trajectories. The speed $U$ which they will have
rachieved during decompression from \( P_e \) to \( P_d \) is related to the other variables by the energy equation [Wilson, 1980]

\[
(1/2) U^2 = \left[ (n Q 0)/m \right] \ln(P_e/P_d) + \left[ (1 - n)/\rho \right] (P_e - P_d). \tag{6}
\]

[28] Inserting \( U = 420 \) m s\(^{-1}\) and the values of \( Q, m, \rho, \) and \( \theta \) quoted above, we can find \( n \) as a function of the value assumed for \( P_e \) (first two columns of Table 1). We have no independent way of estimating \( P_e \) other than to note that this eruption required the buildup of a vesicular layer in the magma just below the surface prior to the eruption and therefore probably represents a case where a dike penetrated close to the surface, chilled at its upper contact with the country rocks, and finally fractured a pathway to the surface only when a sufficient excess pressure had built up to overcome the tensile strength of the country rocks (Figure 5a). In this case, \( P_e \) is theoretically equal to twice the tensile strength [Tait et al., 1989]. The typical tensile strengths of coherent basalts are \( \approx 8 \) MPa [Touloukian et al., 1980], and so \( P_e \) should be \( \approx 15 \) MPa. Table 1 then shows that \( n \) must be \( \approx 0.025 \), and since lunar magmas are expected to have been capable of producing \( \approx 1000 \) ppm of CO by oxidation-reduction reactions between carbon and metal oxides on nearing the surface [Sato, 1979; Fogel and Rutherford, 1995], this implies a roughly 25-fold gas concentration at the top of the dike. For any pair of values of \( P_e \) and \( n \) we can also find \( \rho_g \) from (3), \( f \) from (5), \( \beta \) from (4), \( M \) from (1), and \( V_{\text{dep}} \) from (2). The results are shown in Table 1, where the symbol \( \beta_0 \) has been used for the density to emphasize that we are assuming it to be a constant throughout the foam layer.

[29] The values of the gas volume fraction \( f \) in Table 1 allow us to obtain a more constrained model of the geometry of this eruptive system. If \( P_e \) is much less than \( 5 \) MPa, \( f \) is \( > 0.9 \): vesicularity this high represent unstable foams which drain and collapse quickly [Jaupart and Vergniolle, 1989], and so we do not expect values of \( P_e \) this low to be appropriate to this eruption. However, we have just argued that \( P_e \) will not be greater than \( \approx 15 \) MPa. Table 1 shows that if \( P_e \) lies in the range 5–15 MPa, the total mass of the erupted magma probably lies between about 4 and \( 11 \times 10^{13} \) kg and the total volume of pyroclastics lies between 17 and 50 km\(^3\).

[30] We can find the vertical extent of the vesicular layer at the top of the dike provided we assume a width and horizontal length for the dike. Head and Wilson [1992] showed that dikes penetrating to the surface from the base of the lunar crust should have widths \( W \) in the range 150–200 m; if their source regions are within the mantle, dikes reaching the surface can have even greater widths, up to 800 m. We adopt a plausible value of \( W = 500 \) m for the present purpose and also assume that the horizontal extent of the dike is equal to the \( 16 \) km length of the depression. The vertical extent of the vesicular layer is then \( H \), where \( (H W L) = V_e = 1.2 \times 10^{12} \) m\(^3\), implying \( H = 15 \) km. It is possible that collapse of near-surface rocks into the part of the dike vacated by the erupting magmatic foam was not complete, so that the horizontal extent of the dike was \( > 16 \) km. Doubting the assumed extent would halve the estimate of \( H \) to 7.5 km. However, there seems little alternative to assuming that \( H \) is of order 10 km, and so we cannot neglect the fact that the pressure gradient within a vesicular layer this thick will have been significant, implying that the density of the foam and the pressure within it cannot be assumed to be uniform with depth. The pressure and density must increase with depth, implying that the true total mass of magma in the foam layer must be greater than the values given in Table 1. We deal with these issues as follows.

### 6.1.3. Elaboration of conditions in the foam layer.

[31] Let the (now variable) pressure in the vesicular layer be \( P_{\text{tot}} \) at a general depth \( y \), so that \( P_{\text{tot}} = P_e \) at \( y = 0 \) and, in general,

\[
P_{\text{tot}} = P_e + \int_0^y \beta g \, dy. \tag{7}
\]

This implies that \( dP_{\text{tot}} = \beta g \, dy \), where \( \beta \) is now allowed to vary with depth in the foam layer, and if we substitute this expression for \( \beta \) in (4) and then substitute (3) for \( \rho_g \) (noting that what was the constant pressure \( P_e \) in (3) has now become the variable pressure \( P_{\text{tot}} \)), we obtain an expression relating \( P_{\text{tot}} \) to \( y \) which is readily integrated to give

\[
[\ln(P_{\text{tot}}/P_e) + (1 - n)(P_{\text{tot}} - P_e)]/\rho = g y. \tag{8}
\]

which means that we can find \( y \) as a function of \( P_{\text{tot}} \) analytically (though not vice versa). We next note that the total mass \( M(y) \) of vesicular magma between the top of the foam layer and a depth \( y \) within it is

\[
M(y) = \int_0^y \beta L W \, dy, \tag{9}
\]

and since \( dP_{\text{tot}} = \beta g \, dy \), the total mass is just

\[
M(y) = [(L W)/g][P_{\text{tot}} - P_e]. \tag{10}
\]

In order to find the total mass of magma within the foam layer of depth \( H \) therefore we put \( y = H \) in (8) and solve the equation

### Table 1. Values of the Magmatic Gas Mass Fraction \( n \), Gas Density \( \rho_g \), Gas Volume Fraction \( f \), Bulk Magma Density \( \beta_0 \), Total Mass of Erupted Magma \( M \), and Bulk Volume of Pyroclasts Erupted \( V_{\text{dep}} \) for a Series of Values of the Assumed Excess Pressure \( P_e \) in the Gas in the Foam Layer Prior to Eruption

<table>
<thead>
<tr>
<th>( P_e ), MPa</th>
<th>( n )</th>
<th>( \rho_g ), kg m(^{-3})</th>
<th>( f )</th>
<th>( \beta_0 ), kg m(^{-3})</th>
<th>( M ), ( 10^{13} ) kg</th>
<th>( V_{\text{dep}} ), km(^3)</th>
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<tbody>
<tr>
<td>100</td>
<td>0.0125</td>
<td>210.5</td>
<td>0.153</td>
<td>2573</td>
<td>30.9</td>
<td>139</td>
</tr>
<tr>
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<td>0.339</td>
<td>2017</td>
<td>24.2</td>
<td>109</td>
</tr>
<tr>
<td>40</td>
<td>0.019</td>
<td>84.2</td>
<td>0.408</td>
<td>1809</td>
<td>21.7</td>
<td>98</td>
</tr>
<tr>
<td>30</td>
<td>0.0206</td>
<td>63.1</td>
<td>0.500</td>
<td>1531</td>
<td>18.4</td>
<td>84</td>
</tr>
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<td>42.1</td>
<td>0.622</td>
<td>1159</td>
<td>13.9</td>
<td>63</td>
</tr>
<tr>
<td>15</td>
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<td>0.699</td>
<td>924</td>
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<td>50</td>
</tr>
<tr>
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<td>0.792</td>
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<td>2.1</td>
<td>0.992</td>
<td>51</td>
<td>0.6</td>
<td>3</td>
</tr>
</tbody>
</table>
recursively for $P_{\text{tot}}$. This value of $P_{\text{tot}}$ is then inserted in (10) to find the required mass $M_G$. If we take the maximum value of $H = 15$ km inferred above, then for $P_r = 5$ MPa we find $P_{\text{tot}} = 20.3$ MPa and $M_G = 7.5 \times 10^8$ kg, to be compared to the value of $3.9 \times 10^9$ kg found in Table 1 assuming a uniform foam density. Similarly, using $P_r = 15$ MPa, we find $P_{\text{tot}} = 48.9$ MPa and $M_G = 16.6 \times 10^8$ kg, to be compared to the uniform density estimate of $11.1 \times 10^8$ kg. These results imply that the total magma masses, and hence also the bulk pyroclast volumes, given in Table 1 should be increased: by a factor of ~2 for small values of $P_r$ and by a smaller factor at larger pressures. The revised estimates of the total magma mass lie between 7 and $17 \times 10^3$ kg, and those of the pyroclast volume lie between 33 and 75 km$^3$.

[32] Next we recall that we found earlier that the vesicular magma layer must have concentrated gas from a 25-fold greater mass of magma. If 15 km thick vesicular layers corresponding to $P_r = 5$ and 10 MPa were collapsed to the equivalent depth of gas-free magma, they would occupy vertical extents of about 1.5 and 3 km, respectively. Thus a 25-fold concentration of gas could have been obtained by scavenging all of the 1000 ppm available gas from the magma in a 40-75 km vertical extent of the dike. It should be noted that the release of this gas is possible only if the pressure in the magma at the time of gas release is less than ~20 MPa. However, in the above models the pressure at the base of the foam layer was found to be 20.3 MPa even at the lowest plausible pressure at the top of the foam layer (5 MPa). Thus, if the gas were being produced in the topmost part of a large convection cell that involved circulation of all of the magma in the dike after the dike came to rest, it is not clear that a sufficiently thick foam layer could be produced before the increasing pressure at the base of the foam layer turned off the gas production process.

[33] However, it is well understood that while a dike is propagating, the need to maximize the pressure gradient acting to overcome wall friction means that the pressure at the dike tip is very low [Rubin, 1993]. The pressure gradient in this region will be somewhat greater than the mean gradient over the entire length of the dike, which must be of the same order as the lithostatic stress gradient in the surrounding rocks (~5000 Pa m$^{-1}$ in the case of the Moon). Thus a pressure of 20 MPa will be reached ~4 km below the dike tip. We infer therefore that magma passing through the region within at least 1 or 2 km of the tip while the dike is propagating upward will have generated gas bubbles (Figure 5a). Some of these bubbles will have been carried upward with the magma in the center of the dike, and others will have been swept near to the dike walls by the lateral component of magma motion, which accommodates the fact that at any given depth the dike must widen as its tip moves upward. Thus there will be a population of gas bubbles throughout the dike as it comes to rest. Most of these bubbles will be exposed to a higher pressure than that at which they were nucleated, but they will not be resorbed into the magma, either chemically or by solution, on the timescale of subsequent convective overturn of the magma.

[34] This timescale can be estimated by assuming a small density contrast $\Delta \rho$ due to cooling of magma at the dike walls and equating the resulting buoyancy to the pressure gradient required to drive magma motion at a given speed $V$. Since the rising and falling convection currents each occupy about half of the dike width $W$ (the rising limb in the center), a slight modification of the formula for magma rise at high Reynolds numbers given by Wilson and Head [1981a] gives

$$V = \left[ \frac{g \Delta \rho W}{(2 \ K \ P)} \right]^{1/2},$$

where $K$ is a friction factor of order 0.01. For density contrasts of 100, 10, and 1 kg m$^{-3}$ the implied speeds are 33, 10, and 3.3 m s$^{-1}$, respectively, and the corresponding Reynolds numbers are $7.5 \times 10^5$, $2.2 \times 10^6$, and $7.5 \times 10^7$ for a magma with viscosity 3 Pa s, thus justifying the use of the above formula rather than one appropriate to laminar magma motion. The timescales for convective overturn of a 75 km vertical extent of dike (the larger of the earlier estimates) are then 0.6, 2.1, and 6.3 hours. The buildup of the foam layer will take much longer than this, however. Given the large absolute pressure change and the large rates of pressure change to which gas bubbles in lunar magmas near the surface are subjected, fragmentation is very thorough, and there is little opportunity for gas bubble coalescence. As a result, the median sizes of the lunar pyroclastic spheres, ~100 mm, imply that the gas bubbles in the foam layer are also of this size [Wilson and Head, 1981a]. Bubbles with this diameter will rise only at ~1 mm s$^{-1}$ and so will ascend only 200 mm in the ~200 s that they spend at the top of the convection cell. An average of 2 layers of bubbles will be accreted to the base of the foam layer in this time, and a 10 km accumulation of these bubbles will therefore require ~1.7 years. During this time, only the outer 10 m of the dike will have been significantly affected by cooling, so there is no significant thermal constraint on the accumulation of the foam taking this long.

[35] A final issue concerns the pressure rise in the foam layer. We pointed out earlier that dikes rising to the surface must always contain an excess pressure which is needed to overcome the elastic stresses exerted by the surrounding rocks. This initial excess pressure will be present in the foam as it begins to accumulate but must, of course, not be too large; otherwise, fracturing of the overlying rocks leading to an eruption will occur before a significant foam layer has built up. An increase in the excess pressure during the late stages of foam accumulation is required to explain the eventual onset of an eruption. This increase can be explained partly in terms of the transfer of gas bubbles from the deeper parts of the dike to the shallower parts [Pyle and Pyle, 1995] and partly in terms of the continued release of gas from batches of magma being cycled through the shallow, low-pressure part of the system (Figure 5a). Tait et al. [1989] give formulae for excess pressure increases as part of a more general treatment of magma chamber evolution on Earth. The factors important here can be represented by

$$\Delta P = \Delta \rho g_b \left[ \left( \frac{b^2}{g} \right)^{1/2} \right]^2,$$

where $P_b$ and $V_b$ are the current total volumes of liquid magma and gas, respectively, $P$ is the current total pressure, $\beta$ is the bulk modulus of the liquid magma, and $\mu$ is the rigidity of the country rocks, both ~10 GPa, $\rho_b$ is the density of the gas at the current pressure, and $\Delta P$ is the pressure rise to an incremental release of a mass $\Delta m_g$ of gas. Using the dike dimension estimates given earlier, the magma volume $V_1$ down to a depth of 75 km is $6 \times 10^{15}$ m$^3$, and its mass is $1.8 \times 10^{21}$ kg. The mean gas density in a 10 km deep foam layer which has a pressure $P_r = 10$ MPa at its top is ~30 kg m$^{-3}$ (see Table 1), and the volume of this gas is about $5 \times 10^{10}$ m$^3$, allowing for the fact that the foam is on average ~70% gas and ~30% liquid. The release of an extra 100 ppm gas from the 1.8 $\times 10^{15}$ kg of magma, i.e., $1.8 \times 10^{11}$ kg, then produces a pressure rise of ~1.2 MPa, a more than 10% change. We assume that the eruption was triggered during the late stages of this pressure rise.

6.2. Deposit Characteristics

6.2.1. Deposit thickness and volume. [36] The pyroclastic layer is assumed to be distributed around the vent in a deposit whose thickness varies with radial distance. Figure 18 of Wilson and Head [1981a] shows the thickness profiles expected for a
point source and a line source (fissure vent). In this case, since we have an extended source at the center of a nearly circular deposit, we assume that the profile for a linear source applies to the inner part of the deposit and that the profile for a point source applies to the outer part. We therefore approximate the layer as having a uniform thickness \(T\) out to 80% of the maximum range \(R\) and then thickening rapidly to a maximum of 3.6\(T\) at the outer edge. Numerical integration of the volume \(V_{\text{dep}}\) under this profile shows it to be about

\[
V_{\text{dep}} = 4.31 \ T \ R^2, \quad (13)
\]
to be compared with the volume \(\pi \ T \ R^2 = 3.14 \ T \ R^2\) of a circular deposit of uniform thickness \(T\) and radius \(R\).

[37] We can investigate the thickness of the layer of pyroclasts by using the spatial variation of the albedo of the deposit, as follows. Assume that the regolith thickness which has accumulated in this area since the time of the eruption is \(Z\) (where \(Z\) is greater than the maximum thickness of the pyroclast layer, 3.6\(T\)). The regolith clasts will have about the same bulk density and packing state as the pyroclasts, so that we can assume that the inner part of the deposit will be a uniform mixture of pyroclasts of albedo \(A_p\) from a layer of thickness \(T\) and bedrock clasts of albedo \(A_o\) from a layer of thickness \((Z-T)\). Its albedo will therefore be \(A_b\), where

\[
A_b = \left[ T \ A_p + (Z-T) \ A_o \right] / Z. \quad (14)
\]
Here we assume a random mosaic of clasts at the surface and neglect multiple scattering of light between the clasts. Similarly, the albedo of the outermost edge of the deposit will be the result of a layer of pyroclasts of thickness 3.6\(T\) mixing with a thickness \((Z-3.6T)\) of bedrock clasts; the albedo of this outer zone will therefore be \(A_o\), where

\[
A_o = \left[ 3.6 \ T \ A_p + (Z-3.6T) \ A_o \right] / Z. \quad (15)
\]
Thus we have

\[
A_b/A_o = \left[ T \ A_p + (Z-T) A_o \right] / \left[ 3.6 T A_p + (Z-3.6T) A_o \right], \quad (16)
\]
and if we now define \(F_{\text{bp}} = A_b/A_o\) and \(F_{\text{io}} = A_i/A_o\), we can invert the above relationship to give

\[
T/Z = \left[ F_{\text{bp}}(F_{\text{io}} - 1) \right] / \left[ (F_{\text{bp}} - 1)(3.6F_{\text{io}} - 1) \right]. \quad (17)
\]
Thus, if we can measure \(F_{\text{io}}\) from calibrated Clementine images and make an estimate of \(F_{\text{bp}}\) from published measurements of the albedoes of typical pyroclastics and the bedrock in the area around the vent, we can find a value for \(T/Z\) and hence, given a value for \(Z\), for \(T\).

[38] Since it is only the ratios of the various albedos that appear in this expression, we have used our measured reflectances as substitutes for the albedos. The average reflectances corresponding to \(A_b\), \(A_i\), and \(A_o\) are 2000, 1950, and 1400 units, respectively. We have no reflectance data for pyroclasts uncontaminated with regolith, of course, but use the lowest reflectance in the darkest, northern part of the deposit as an upper limit, giving \(~1200\) units as the reflectance corresponding to \(A_b\). We then find \(F_{\text{bp}} = 1.67\) and \(F_{\text{io}} = 1.39\), leading to \(T/Z = 0.25\). The typical post-DMRD-deposit-emplacement regolith thickness, which is most likely Imbrian in age, should be about \(Z = 8\ m\) [Oberbeck and Quaide, 1968], implying that \(T = ~2\ m\). The total deposit volume given by (13) with \(R = 77\ km\) and \(T = 2\ m\) will then be \(51\ km^3\), which lies near the middle of the 33–75 km\(^3\) range deduced earlier from the properties of the foam layer and the size of the vent depression. These results are not excessively sensitive to the value assumed for the reflectance of uncontaminated pyroclasts; if the value used were 1000 instead of the 1200 used above, \(T\) would be \(~1.5\ m\) and the volume would be \(~41\ km^3\); a value of 500 would yield \(T = 1\ m\) and a volume of \(~27\ km^3\).

### 6.2.2. Constraints based on cooling during flight. [19] We noted earlier that the pyroclasts ejected in this eruption must have cooled at rates of at least 100 K s\(^{-1}\) somewhere along their flight paths to ensure that they were glassy rather than crystalline. We can use this fact to place limitations on the mass eruption rate from the vent. Cooling through the several hundred degrees from the eruption temperature to below the glass transition temperature would require only \(~5\ s\), and since the particle velocity to reach the mean range of 77 km is close to 350 m s\(^{-1}\), this would correspond to a travel distance of only \(~2\ km\). In principle, this 2 km cooling zone could be mutually obscured by any radial distance between the outer envelope of the cloud of pyroclasts. We explore first the case in which the cooling zone is at the outer edge of the system. Wilson and Head [1981a] showed that only those particles located at distances less than some critical value \(\lambda\) from the outer edge of an eruption cloud would be able to radiate heat to their surroundings; pyroclasts deeper within the cloud would mutually obscure one another. The relationship between \(\lambda\) and the other variables is a function of the vent and deposit geometry. In the present case we are dealing with a nearly circular deposit and an elongate vent system. However, the radius \(R\) of the deposit (77 to \(~100\ km\)) is very much greater than the long axis of the vent \((L = 16\ km\)), which implies that most of the motion of the pyroclasts in the eruption cloud will be radial. For this case, equation (66) of Wilson and Head [1981a] can be inverted to give

\[
\Phi = (5 \ \varphi \ g^{1/2} \rho \ R^2 s^4) / (6 \ ?^2 \lambda), \quad (18)
\]
where \(\Phi\) is the total mass flux from the vent, \(\varphi\) is the mean pyroclast diameter, and the other parameters are as defined earlier. We now identify \(\lambda\) to be equal to the 2 km linear extent of the zone within which the pyroclasts in the DMRD cooled to their final glassy states; with \(\lambda\) equal to its mean value, 77 km, and \(\varphi = 100\ \mu m\), we find \(\Phi = 1.85 \times 10^8\ kg\ s^{-1}\). Our best estimates of the total mass of the deposit derived earlier were \(7–17 \times 10^{13}\ kg\). Dividing the mass by the eruption rate implies an eruption duration of \(3.8–9.2 \times 10^4\ s\), i.e., \(4.4–10.6\ days\).

[40] We now treat the extreme alternative case where the region in which the pyroclasts cool is located not at the outer edge of the eruption cloud but very near the vent. In this case the vent must be treated as a line source and the geometry of the system is somewhat different. We can use a development analogous to that employed by Wilson and Head [1981a] in deriving (18). Pyroclasts are projected at speed \(U\) into the region on either side of the fissure vent and cool while traveling the distance \(\lambda\), which requires a time \(\tau\) equal to \((U/\lambda)\). Since the mass flux is \(\Phi\) and the mean pyroclast diameter is \(\varphi\), the number \(N\) of clasts ejected per unit time is the flux divided by the typical mass, \(\Phi/[(\pi/6) \varphi \rho]\). These clasts occupy a region which extends the length of the fissure, \(Y\), and has a cross sectional area perpendicular to the strike equal to \((0.5 \pi \lambda^2)\), so that the volume is \((0.5 \pi \lambda^2 Y)\). The number of clasts within this region is \((N \pi \tau)\), and so their number per unit volume is \((N \rho / (0.5 \pi \lambda^2 Y))\); hence their mean separation is the cube root of the reciprocal of this quantity, \(\sigma = [(0.5 \pi \lambda Y)/(N \pi \tau)]^{1/3}\). Wilson and Head...
Values of the Fissure Vent Width $X$ and Length $Y$ and of the Foam Rise Speed Into the Base of the Fissure, $S$, for a Range of Values of the Fissure Length to Width Ratio $q$ When the Mass Eruption Rate is $1.85 \times 10^8$ kg s$^{-1}$

<table>
<thead>
<tr>
<th>$q$</th>
<th>$X$, m</th>
<th>$Y$, km</th>
<th>$S$, m s$^{-1}$</th>
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<tr>
<td>1,000</td>
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<td>2.8</td>
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</tr>
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<td>30,000</td>
<td>1.19</td>
<td>35.6</td>
<td>2.9</td>
</tr>
</tbody>
</table>

[1981a] show that the distance $\lambda$ over which the cloud of particles will become optically dense is given to a good approximation by $\lambda = [(5 \sigma \gamma) / (\pi \varphi)]$. Substituting the above expressions for $\sigma$, $N$, and $\tau$ and solving for $\Phi$, we find

$$\Phi = (5 \pi / 12) \varphi \rho U Y,$$

which, for this geometry, does not depend explicitly on the value of $\lambda$. Substituting $\varphi = 100 \mu m$, $\rho = \sim 3000$ kg m$^{-3}$, $U = \sim 350$ m s$^{-1}$ to reach the mean range of 77 km, and using $Y = \sim 10$ km, somewhat less than the 16 km length of the vent depression, to allow for some enlargement of the length during wall subsidence, we find $\Phi = 1.4 \times 10^8$ kg s$^{-1}$. This value is $\sim 130$ times smaller than the flux found by assuming that cooling took place in the outer part of the eruption cloud and corresponds to an eruption duration in the range 1.7–4.1 years.

[41] There is, of course, a complete spectrum of options between the extremes just considered for the location of the zone of cooling of the pyroclasts. If we assume the zone to be located at a distance $r$ which is farther from the vent than a few times the along-strike length of the vent, so that the motion of the clasts can be considered radial, but not so far from the vent that many pyroclasts are already falling back to the surface, we can derive the equivalent of (19) as

$$\Phi = (5 \pi \varphi \rho U r^2) / (9 \chi).$$

Then, if we consider the zone to be located about one third of the way from the vent to the maximum range, at $r = 30$ km, we have $\Phi = 8.2 \times 10^7$ kg s$^{-1}$. This value is already nearly half of the eruption rate required to have the zone of cooling at the outermost edge of the deposit, which suggests that rates within a factor of 2 of $10^8$ kg s$^{-1}$, corresponding to eruption durations less than a month, are much more likely than rates of order $10^6$ kg s$^{-1}$ and durations of years.

6.2.3. Geometry of eruptive vent. [42] The mass flux can also be used to estimate the geometry of the vent system through which the foam expanded on eruption. Since we do not know the geometry of the fracture through which the eruption took place, we assume that it had a length $Y$ and width $X$ such that $Y = q X$, where $q$ is the fissure aspect ratio. Its area was therefore $q X^2$. The mass flux $\Phi$ must be equal to the product of the vent area, the foam density, and the mean rise speed $S$ of the foam into the bottom of the fracture ($S$ is expected to be orders of magnitude smaller than $U$, the speed of the eruption products after disruption of the foam and gas expansion). We therefore have

$$\Phi = S q X^2 \beta.$$

$S$ can be related to $X$ by a relationship similar to that underlying (11), but with the modification that we now expect the foam to flow in the laminar, rather than turbulent, regime. The laminar rise speed $S$, though a fissure of width $X$, of a fluid with density $\beta$ acted on by a pressure gradient $dP/dz$ is

$$S = (X^2 dP/dz) / (12 \mu),$$

where $\mu$ is the magma viscosity, and so $\Phi$ is related to $X$ by

$$\Phi = (q X^4 \beta dP/dz) / (12 \mu).$$

[43] In the present case we can assume that $dP/dz$ is approximately equal to the excess pressure in the foam divided by the average distance between the foam layer and the surface. Since the intrusion of the dike containing the foam did not lead to the formation of a surface graben, we infer that its top was located at least 3 km below the surface [Head and Wilson, 1994]. Using $P_c = 10$ MPa and a dike top depth of 4 km leads to $dP/dz \sim 2.5 \times 10^6$ Pa m$^{-1}$. We adopt $\mu = 100$ Pa s (a value somewhat greater than the 10–30 Pa s that might be appropriate to pristine lunar basalts to allow for the bubble interactions in the foam), and $\beta = 1500$ kg m$^{-3}$, a representative value for pressures of a few tens of MPa in the body of the foam (Table 1). We can then find $X$ for any chosen value of $\Phi$ for a range of assumed values of the fissure aspect ratio $q$. $X$ can be used to find $S$ from (21) and $Y$ from the definition $Y = q X$.

[44] Table 2 shows the results obtained for $\Phi = 1.85 \times 10^8$ kg s$^{-1}$, the eruption rate needed to allow the zone of cooling to be at the outer edge of the eruption cloud. We expect $Y$, the length of the active fissure, to be somewhat less than the $L = 16$ km length of the depression produced by post-eruption collapse around the vent, and so $q$ must be close to 10,000, implying that the fracture width $X$ is $\sim 1.5$ m and the magma rise speed $S$ is $\sim 5$ m s$^{-1}$. Evaluation of the Reynolds number corresponding to these values confirms that the magma motion has correctly been treated as laminar. Table 3 shows the equivalent results for $\Phi = 1.4 \times 10^8$ kg s$^{-1}$, the eruption rate needed to allow the zone of cooling to be very close to the vent. In this case we need to select $q$ close to 30,000 to have $Y$ a little less than 16 km and find that the fracture width is now only 0.35 m and the magma rise speed is $\sim 0.25$ m s$^{-1}$.

6.2.4. Model constraints based on clast cooling during eruption. [45] We can obtain an indication of which of these extreme scenarios is more likely by considering the loss of heat from the foam as it rises through each of these different geometries. Wilson and Head [1981a] show that the maximum vertical distance that magma with viscosity $\mu$ can rise through a fissure of width $X$ under a pressure gradient $dP/dz$ without losing so much heat that it ceases to be able to flow is $H$, where

$$H = (X^4 dP/dz) / (48 \chi \mu).$$

and $\chi$ is a constant related to the thermal diffusivity of the magma with the value $1.25 \times 10^{-6}$ m$^2$ s$^{-1}$. Using $\mu = 100$ Pa s the results are as follows:

<table>
<thead>
<tr>
<th>$q$</th>
<th>$X$, m</th>
<th>$Y$, km</th>
<th>$S$, m s$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1,000</td>
<td>0.82</td>
<td>0.8</td>
<td>1.39</td>
</tr>
<tr>
<td>3,000</td>
<td>0.62</td>
<td>1.9</td>
<td>0.81</td>
</tr>
<tr>
<td>10,000</td>
<td>0.46</td>
<td>4.6</td>
<td>0.44</td>
</tr>
<tr>
<td>30,000</td>
<td>0.35</td>
<td>10.5</td>
<td>0.25</td>
</tr>
<tr>
<td>100,000</td>
<td>0.26</td>
<td>25.9</td>
<td>0.14</td>
</tr>
</tbody>
</table>
s and $dP/ \text{d} \beta = 2500 \text{ Pa m}^{-1}$ as before, we find that the maximum vertical distance which the magma erupting under the conditions shown in Table 2 could be expected to travel before excessive cooling is 2470 km, vastly greater than the 4 km which we require it to travel. In contrast, magma erupting under the conditions shown in Table 3 would nominally be able to travel at most 6.25 km. This distance is so close to the 4 km required that we find it extremely unlikely that the narrow pathway would be able to survive thermally for the 2–4 years needed to complete the eruption. We therefore conclude that scenarios in which most of the pyroclast cooling occurs somewhere in the outer half of the eruption cloud are very much more likely than those in which it occurs very near the vent. This corresponds to eruption rates between about 0.8 and $1.8 \times 10^8 \text{ kg s}^{-1}$ and eruption durations between 1 and 2 weeks (Figure 7).

7. Discussion and Conclusions

[46] Clementine data have revealed the presence of an elongate depression in the center of the 154 km diameter dark ring deposit located in the southwestern part of the Orientale Basin (Figures 1–3). Mapping with Clementine data shows the detailed morphological and spectral characteristics of this deposit (Figure 2). We interpret the elongate depression seen in the middle of the DMRD (Figure 3) to be the source vent for the ring. We model the dark mantle deposit as an eruption from the central, elongate vent, produced by the shallow intrusion of a dike, degassing of the magmatic material in the dike (Figure 5a), eruption of this gas-rich foam to the surface to produce an umbrella-shaped plume $\sim 38 \text{ km high}$ (Figure 5b and 6), and the emplacement of this ejecta to produce the ring-like dark mantle deposit. On the basis of our analysis, the dike emplacement event occurred over a period of $\sim 30 \text{ min to 1 hour}$ [Head and Wilson, 1992], the exsolution of gas and its buildup took place over a period of $\sim 1.7 \text{ years}$, the eruption had a duration of $\sim 1–2 \text{ weeks}$, and it took an additional 300 years for the dike to cool completely [e.g., Wilson and Head, 1981a] (Figure 7).

[47] On the basis of this analysis we find this model to be self-consistent and plausible and an alternative to those models calling on multiple eruptions along the ring to produce the dark mantle. The interpretation of the Orientale DMRD as a pyroclastic eruption argues against the hypothesis that the dark ring represents the presence of an ancient pre-Orientale impact structure [Schultz and Spudis, 1978] and implies that the Orientale Basin cavity of excavation must therefore lie within the Outer Rook Mountain ring.

[48] This model also accounts for the differences in morphology between the Orientale DRMD deposit and other dark mantle and pyroclastic deposits on the Moon [e.g., Weitz et al., 1998; Weitz...
behavior of dikes stalling in this near-surface environment [Head and Wilson, 1996] can be interpreted in terms of different stalling depths, with those forming linear graben among the shallowest and those forming collapse pits among the deepest.

[49] At first sight (Figure 8) the morphology of the dark ring deposit we have analyzed has much in common with the bright, annular deposits seen around some of the volcanic vents on Io [e.g., Strom and Schneider, 1982; Johnson and Soderblom, 1982; Kieffer, 1982; McEwen et al., 1998, 2000]. There are, however, both similarities and differences between the formation mechanisms of the two types of deposit. Both involve the acceleration of small particles to a high velocity in an expanding gas stream. In the lunar case the particles were all silicate pyroclasts derived from the disrupted foam layer. In the Io case they are a mixture of silicate pyroclasts and “snowflakes” of condensing SO$_2$ solids [e.g., Strom and Schneider, 1982], the snowflakes forming in flight and surviving on the ground only beyond a critical distance from the vent as the temperature and pressure of the gas phase decrease [e.g., Cataldo, 1999; Cataldo and Wilson, 1999]. In both cases the silicate pyroclasts ejected from the vent continue to be accelerated by the gas only until such time as the gas pressure decreases to the point where the mean free paths of the gas molecules are much greater than the typical pyroclast sizes [Wilson and Keil, 1997]; beyond this point the clasts continue on ballistic trajectories controlled only by gravity [Wilson and Head, 1981a, 1983].

[50] For many eruptions on Io, the mass flux is inferred to be large enough that there is a region of high opacity around the vent within which the gas and clasts cool by a negligible amount [Davies, 1996; James and Wilson, 1998; Cataldo and Wilson, 1999], conditions similar to those we have inferred for our lunar example. However, the relative size of this region in the Io eruptions documented so far is much smaller than it was in the lunar case. This is partly because of differences in the mass fluxes, though the Io fluxes [Wilson and Head, 1981b; Cataldo and Wilson, 1999] are probably within a factor of 3 of that found here, and partly because the velocities reached by the pyroclasts on Io are greater than those forming the DMRD by a factor of up to 2. The latter difference is the direct result of the origin and mass fraction of the volatile phase: the accumulation of a foam layer at magmatic temperature containing 2–3 wt % CO on top of a silicate intrusion in the lunar case and the intimate mixing on Io of a steadily erupting magma and up to 30 wt % of liquid SO$_2$. The SO$_2$ is probably fed into the vent system from a shallow SO$_2$ aquifer [Davies and Wilson, 1988; Leone and Wilson, 1997], and the large mass fraction ensures that the temperature of the mixture is significantly less than that of the magma alone. A final difference concerns the appearance of the resulting deposits: our ability to observe evidence of the presence of the pyroclasts in the lunar deposit depends on their contrast with, and degree of mixing with, regolith. This requires a significant thickness to be deposited relative to the regolith thickness before visibility is ensured. On Io, SO$_2$ has such a high albedo that it need only condense as a monomolecular layer to be readily visible, at least during and soon after an eruption.

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**References**


