Explosive Volcanism on Hecates Tholus, Mars: Investigation of Eruption Conditions

PETER J. MOUGINIS-MARK, LIONEL WILSON,1 AND JAMES W. HEAD III

Department of Geological Sciences, Brown University, Providence, Rhode Island 02912

Evidence is presented supporting the identification of a discrete, well-preserved air fall deposit generated by explosive volcanism on Mars. The deposit is located immediately to the west of the summit caldera of Hecates Tholus (32°N, 209°W), the northernmost of the three volcanic constructs in the Elysium Planitia region. The absence of superposed impact craters larger than the resolution limit of the Viking images (40 m per pixel) indicates a very young age for this eruption, which evidently postdates even the most recent collapse episode of the Olympus Mons caldera (~300 m.y. B.P.; Neukum and Hoffer [1981]). From the distribution of craters smaller than 2 km in diameter, it appears that the air fall material mantles an area about 50 × 75 km in extent to an estimated thickness of about 70 km. In order to attain such an altitude, the mass eruption rate must have been ~10^7 kg/s and the volatile content either ~1 wt % for H_2O or ~2 wt % for CO_2. Although it is not possible to distinguish between silicic and basaltic volcanism on Hecates, we are the first detailed Mariner 9 images. In general, the large Tharsis volcanoes have received the most attention, due primarily to their great size and similarity (hence interpretable morphology) to terrestrial basaltic shield volcanoes [e.g., Greeley, 1973; Carr and Greeley, 1980]. Recent investigations of Martian surface features have, however, speculated on the existence of rhyolitic lava flows [Fink, 1980] and silicic domes [Plescia, 1981]. Explosive volcanic eruptions have also been postulated for Mars, particularly in association with Olympus Mons and the Elysium Planitia volcanoes. Deposits within Amazonis have been described as ignimbrites [Scott and Tanaka, 1980], while other forms of pyroclastic flow are thought to be found within the Olympus Mons aureole materials [Morris, 1980]. Other surface features have also been interpreted as explosively generated volcanic ash deposits [Ward, 1979], ash flows, and welded tuffs [Peterson, 1981] or the products of volcanic density currents [Reimers and Komar, 1979]. These analyses were, however, based on photogeological data and did not explore the styles of volcanism and magma chemistries that such interpretations would imply. Malin [1977] and Greeley and Spudis [1981] have described the petrological implications that explosive volcanism would have for Mars; they have argued that, by analogy with terrestrial volcanoes, such activity would suggest the existence of Martian silicic magmas (although Francis and Wood [1981] warn that only general inferences about magma chemistry can be made from such a comparison). All of the above mentioned candidate examples of explosive volcanism are, however, relatively old, with their source regions and modes of emplacement unidentified or poorly constrained.

We have reexamined the medium and high-resolution Viking images of Amazonian and Hesperian age volcanic centers on Mars [Scott and Carr, 1978] and believe that an excellent example of well-preserved explosive activity does indeed exist close to the summit of Hecates Tholus, the northernmost of the three large Elysium Planitia constructs. A mantled region to the west of the summit caldera appears to be an example of a geologically very recent plinian air fall ash deposit. We present here our morphological evidence for describing this as explosively generated material and also employ numerical models of magma ascent and eruption in the Martian environment [Wilson and Head, 1981a] to estimate the physical characteristics (eruption cloud height, magma discharge rate, magma volatile content, duration of activity, and vent size) for this event. We also discuss the implications of this eruptive style for the composition of the erupted magma.

MORPHOLOGY OF HECATES THOLUS

Regional Setting

The three Elysium volcanoes (Hecates Tholus, Elysium Mons, and Albor Tholus) surmount the broad Elysium Planitia Rise, which measures approximately 1700 × 2400 km (Figure 1). Hecates Tholus is the northernmost of these constructs and is elliptical in plan with axis lengths 160 × 175 km. The volcano is centered at 32°N, 209°W, and Mariner 9 UVS altimetry estimates [Hord et al., 1974] indicate that the volcano rises about 6 km above the surrounding plain. Both morphometric [Pike, 1978] and morphological [Greeley and Spudis, 1981] data indicate that Hecates resembles a low shield, which has prompted its comparison to the Tharsis

1 Lunar and Planetary Unit, Department of Environmental Sciences, University of Lancaster, Lancaster LA1 4YQ, United Kingdom.

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Fig. 1. Location map for Hecates Tholus and the other main volcanic constructs within Elysium Planitia. Hecates Tholus is centered at 32°N, 209°W.

shields. Regional mapping of Elysium Planitia [Malin, 1977; Brown, 1980] indicates that the northern boundary of the volcano merges smoothly with the surrounding plains materials but that the southern flanks are covered by lateral lava flows from Elysium Mons. This partial burial of the southern slopes has resulted in an apparent 30-km southward offset of the summit caldera from the preserved center of figure of the volcano.

The summit of Hecates Tholus (Figure 2) is characterized by a nested caldera complex 11.3 × 9.1 km in extent. At least

Fig. 2. Photomosaic of the summit area of Hecates Tholus. The marked difference in the frequency of occurrence of the channels and craters around the nested caldera is inferred to be the product of an air fall deposit which has mantled the area to the west of the caldera. Image resolution ~50 m per pixel, Viking frames 651A17–21.
Fig. 3. Distribution map of sinuous channels and large (>2-km diameter) craters on the flanks of Hecates Tholus. Locations mapped from orthographic versions of Viking frames 86A38-46 and 651A15-23.

four episodes of caldera collapse are evident, and from shadow measurements we estimate the maximum caldera depth to be ~470 m. The caldera volume is therefore about 28 km$^3$ for the outer elliptical depression and ~10 km$^3$ for the three inner craters. Unlike the large Tharsis shields, which show ample evidence of lobate lava flows on their flanks [Carr et al., 1977; Schaber et al., 1978], Hecates Tholus has no flow features of this kind identifiable at an image resolution of 40 m per pixel. In this respect, Hecates more closely resembles the volcanoes Uranus and Ceraunius tholi, which have numerous sinuous channels on their flanks [Reimers and Komar, 1979]. The possible mode of formation of the numerous channels on Hecates is discussed below, but previous analyses have described them as either the products of erosion by volcanic density currents [Reimers and Komar, 1979] or other styles of volcanism [Malin, 1977].

Numerous near-circular craters occur over the flanks of Hecates. Because these features possess raised rims, we believe that the vast majority of such craters are of impact origin. Utilizing these craters, size-frequency distribution curves give an approximate age for the volcano of 3600 ± 38 m.y. or 1680 ± 326 m.y., depending upon which model for the cratering flux is employed [Plescia and Saunders, 1979].

**Distribution and Origin of the Sinuous Channels**

The numerous sinuous channels which radiate from the summit area (Figure 3) are striking features on the flanks of Hecates Tholus, and indeed it was the existence of these channels which lead Reimers and Komar [1979] to originally propose that Hecates had experienced explosive activity. Reimers and Komar concluded that the erosion of the preexisting topography by volcanic density currents was the most likely mechanism for their origin. We disagree with this interpretation and favor a combination of thermally eroded
Fig. 4. Distribution of all the sinuous channel source regions; prepared from same base map as Figure 3.

lava channels (analogous to lunar sinuous rilles; cf. Head and Wilson [1981]) and fluvial channels. Specific aspects of the channel network and the locations of the channel source areas (Figure 4) argue against a density current origin. Many of the Hecates channels have coalescing (almost dendritic) patterns (Figures 3 and 5), with individual examples merging at increasing distances from the summit in a manner uncharacteristic of terrestrial pyroclastic flows [Fisher, 1977; Nairn and Self, 1978]. In addition, it is extremely unlikely that clasts large enough to have been responsible for channel erosion (≥1 m by analogy with terrestrial examples) could have been transported as much as a few kilometers from the vent [Wilson, 1976]. Even if these clasts represented the equivalent of a coignimbrite lag fall deposit resulting from a major ignimbrite eruption [e.g., Wright and Walker, 1981], erosion closer to the vent would be observed. Thus the Hecates channels, which commonly originate at distances in excess of 20–40 km from the summit caldera complex, cannot be directly associated with explosive eruptions from the summit caldera. On the contrary, each channel would require a separate vent in its immediate vicinity. We consider this to be an unrealistic situation, because it would imply that many violent explosive eruptions occurred on the volcano’s flanks (a feature which is not observed for terrestrial volcanoes) and that each eruption subsequently buried the vent area so that it is no longer visible on the Viking images.

An alternative method by which the channels could have been formed, which we also reject as a sole mechanism, is by the eruption of low-viscosity magmas, with the formation of lava channels by thermal erosion [Hulme, 1973; Carr, 1974]. Close examination of craters cut by the channels reveals crater infilling/emptying relationships which are not consistent with lava flooding [Theilig and Greeley, 1979]. Examples of craters with their upslope rim breached by channels can be found (Figures 6a and 6b), as well as craters where
Fig. 5. The channel networks present on the northern flanks of Hecates Tholus are similar to many of the fluvial systems described for other parts of Mars by Pieri [1986]. Although the characteristic amphitheater morphology of the source regions, typically indicative of groundwater origin, is absent, the undisturbed nature of the interchannel areas and the development of crude channel lines (arrowed) suggest a groundwater origin for these channels. Direction of channel flow is from the top to the bottom of the image. Image resolution is 40 m per pixel, Viking Frame 86A39-43.
channels have cut both the upslope and downslope rims (Figure 6c). By analogy with terrestrial lava flow eruptions [Walker, 1973a], the great lengths of the Hecates channels would require a high lava effusion rate, thus involving large volumes of magma. Craters with just their upslope rims breached by channels (Figures 6a and 6b) fail to show the expected infilling by this hypothesized lava. Furthermore, craters cut on both upslope and downslope rims (Figure 6c) demonstrate that subsequent drainage of material from within the crater occurred to a level below that of the outflow channel. This last example (Figure 6c) shows no evidence of lava terraces within the crater, and the perched egress
channel argues against a drained lava lake once existing within the crater as a result of flooding by the channel. We do not, however, completely discount the possibility that some of the channels on Hecates may be the central channels of lava flows, merely asserting that potential examples appear to have been modified after formation by erosive processes.

We conclude that the most plausible mechanism for formation of the majority of the channels is fluvial action. The observed branching pattern of the channels (Figures 3 and 5), the variable width and depth of each channel along its length (with the source region rarely the deepest point), and the apparent influence that local and regional topography have over channel directions all suggest a fluvial origin. Many of the channel networks also have digitate-to-parallel patterns which are similar to those associated with fluvial systems recognized elsewhere on Mars [Pieri, 1980]. Deciding between a pluvial or sapping mode of formation for the channels is, however, difficult. The existence of many undissected interchannel areas (Figure 5) suggests that a sapping process was responsible [Pieri, 1980], but, in the same areas, the amphitheater structure of the source regions which is normally associated with sapping [Pieri et al., 1980] is absent on Hecates. The occurrence of several channels along the same break of slope (Figures 2 and 3) may also corroborate a sapping mechanism, since this may represent a streamline formed where a subsurface water table has intersected the surface.

Although we infer that the occurrence of sinuous channels on Hecates does not in itself indicate explosive activity, nevertheless their existence may have a bearing on understanding the local style of volcanism. Few Martian volcanoes possess such channels [Reimers and Komar, 1979], so that their existence on Hecates must indicate atypical conditions at this locality. One possible explanation is that the channels were able to form owing to an easily erodable ash deposit being present on the volcano's flanks, while, for example, Olympus Mons lacks channels of this kind because its flanks are made of resistant lava flows.

**Small Crater Distribution—Evidence for a Resurfacing Event**

A close analysis of the distribution of craters smaller than 2 km in diameter reveals a remarkable asymmetry on Hecates Tholus. An isofrequency map of these small craters (Figure 7) indicates that the maximum number of craters is in excess of 0.4 craters per square kilometer. Typical values are between 0.2 and 0.3 craters per square kilometer, but to the immediate west of the caldera, the number of craters falls below 0.04 per square kilometer. Because no lava flow fronts or source vents for lava flows can be recognized in the area of low crater frequency, it is unlikely that this paucity of craters west of the caldera is the result of effusive activity.

Resurfacing by a mechanism other than the eruption of lava is also supported by the extensive mantling of large sinuous channels to the west of the caldera and the total absence of smaller channels (Figure 2). Pyroclastic flows can be discounted on the basis of the morphology of this mantle; there is an absence of flow patterns indicative of ground flow, there are no preserved flow boundaries, and there is an apparently gradual thinning of the mantle toward its edges. We believe, however, that these observational data are strong evidence in support of extensive blanketing of the terrain by a volcanic air fall deposit. This interpretation is strengthened by both the proximity of the deposit to the inferred vent area (the summit caldera complex) and the similarity in shape that this hypothesized air fall has to the areal distribution of known terrestrial tephra deposits [Walker, 1973b].

An estimate can be made of the volume and mass of this
mante deposit. From the distribution of the observed small craters (Figure 7), the outer edge of the substantially mantled region (taken to be the 0.1 crater contour) may be approximated as a circle of radius 25 km. Based on the assumption that the average size of craters in the unmantled area is about 500-m diameter, depth/diameter estimates for other Martian craters in this size range by Pike [1980] would indicate an average crater depth of about 100 m. If it is assumed that the required thickness of the mantle deposit needed to obscure each crater is equivalent to this crater depth and that the mantle deposit decreases linearly from the center to the edge, then the implied mantle volume is about 65 km$^3$. It should be noted that this volume estimate does not have to be too precise, since, as described below, its value will only affect the estimate of the eruption duration. Taking the bulk density of the deposit as 1000 kg/m$^3$ (a compromise between the likely densities of pumiceous and scoriaceous clasts), the equivalent dense rock (2800 kg/m$^3$) volume is about 23 km$^3$. This estimated volume for the Hecates air fall deposit corresponds to a mass of $\sim 7 \times 10^{13}$ kg, which in view of all the inherent approximations compares favorably with the masses of air fall deposits on earth, which range up to values of at least $3 \times 10^{14}$ kg [Walker, 1973b; Drexler et al., 1980].

Terrestrial eruptions [Smith, 1979] producing erupted volumes (of silicic magma) in the range 1 to 100 km$^3$ ($10^{12}$ to $10^{14}$ kg) are also very commonly associated with positive volcanioc constructs which have summit calderas like Hecates Tholus.

**Analysis of the Hecates Mantling Deposit**

*Factors Controlling the Formation of Air Fall Deposits on Mars*

Analyses of the factors controlling explosive eruptions [Wilson et al., 1980; Wilson and Head, 1981b] demonstrate that the only prerequisite for explosive activity on Mars is that the erupting magma should exsolve at least about 0.02 wt % of a volatile phase such as H$_2$O or CO$_2$, thus ensuring that the proportion of gas bubbles in the magma can become sufficiently great (≥75% by volume) that the magma disrupts into pyroclastic fragments [Sparks, 1978]. Gas bubbles will nucleate in Martian magmas at depths roughly 2.6 times greater than on earth due to the lower Martian gravity, thus allowing more time for diffusional growth of bubbles at a given magma rise rate [Sparks, 1978]. The low atmospheric pressure on Mars will cause decompositional bubble expansion to be up to six (the cube root of the atmospheric pressure ratio) times greater than on earth. In the absence of bubble coalescence (see below), therefore, these factors will combine to ensure that clasts produced in most types of Martian explosive eruption are in general substantially smaller than their terrestrial equivalents.

The rise speed through a crustal fissure of a magma erupting onto a planetary surface depends only weakly on the gravity and fissure width, being most strongly controlled by the magma rheology [Wilson and Head, 1981b]. At sufficiently low magma rise speeds through the crust (less than about 1 m/s) in magmas with viscosities similar to those associated with terrestrial basalts (less than about 1000 Pa s), coalescence of gas bubbles can lead to strombolian activity [Wilson and Head, 1981b]. Bubble coalescence in higher viscosity magmas is of negligible importance at all magma rise speeds [Sparks, 1978]. At extremely low magma rise rates in any type of magma, sufficient chilling of extruded or intruded magma may occur to ensure that activity consists of intermittent (plinian style) explosions [Self et al., 1979]. Otherwise, explosive eruptions involving the steady discharge of magma will be of the Hawaiian style (producing relatively coarse, scoriaceous ejecta) for low-viscosity, low volatile content magmas and of the plinian style (producing a relatively wide range of grain sizes) for intermediate to high-viscosity magmas (and low-viscosity magmas with high volatile contents). In both the latter cases, a high, convecting eruption cloud producing an air fall deposit is likely to be generated over the vent as atmospheric gases are entrained by the eruption products.

The mass fraction of volatiles exsolved from the magma in plinian (and Hawaiian) eruptions directly controls the upward velocity through the vent of the emerging gas and small pyroclasts (to be exact, all clasts whose terminal velocity in the gas stream is much less than the velocity of the gas stream relative to the planetary surface) [Wilson et al., 1980; Wilson and Head, 1981a]. The eruption velocity, together with the magmatic gas density and the acceleration due to gravity, in turn controls the size of the largest clast which can be carried out of the vent [Wilson, 1976]. The low Martian atmospheric pressure and (to a much smaller extent) the low Martian gravity allow more energy to be released by the decompressing magmatic gas in the vent than on earth so that eruption velocities are typically about 1.5 times greater on Mars than on earth, being in the range 100–800 m/s [Wilson and Head, 1981a]. The sizes of the largest clasts transported out of the vent on Mars will be much smaller than on earth (the lower gas density in the vent after decompression to Martian atmospheric pressure more than offsets the higher eruption velocity and lower gravity) and will be in the centimeter rather than the meter size range.

The height to which a convecting eruption cloud can rise from a point source is known to be proportional to the fourth root of the rate at which heat is injected into the cloud [Morton et al., 1956]. The heat release rate is proportional to the total mass eruption rate in volcanic events and the relationship has been well established for several terrestrial eruptions [Settle, 1978; Wilson et al., 1978]. The possibility of a more complicated relationship exists for eruptions from very long elongate, fissure-type vents; but provided that the height of the eruption cloud is much greater than the longest dimension of the vent system, the fourth-root dependence will still hold. A corresponding relationship applies to Martian eruption clouds, but the physical properties of the Martian atmosphere and the acceleration due to gravity are such that Martian eruption clouds rise about 5 times higher than terrestrial clouds for the same mass eruption rate [Wilson and Head, 1981a].

Under certain circumstances on earth (favored by low magma volatile content or wide vent radius and hence high mass eruption rate) it is not possible for a stable convecting eruption column to exist, and instead a fountain forms over the vent feeding pyroclastic flows [Sparks and Wilson, 1976; Sparks et al., 1978]. The conditions under which convecting plinian eruption clouds would not be stable on Mars can be calculated [Wilson and Head, 1981a]. These calculations show that, for a given magma volatile content, pyroclastic flow formation sets in at a mass eruption rate about 3 times smaller than on earth.

All of the above considerations relating to Martian erup-
tion clouds can be combined to show the eruption velocity in the vent as a function of mass eruption rate for several values of exsolved magma H₂O or CO₂ content (Figure 8). The scale at the top of each part of the figure gives the eruption cloud heights corresponding to the mass eruption rates at the bottom for the cases where a stable convecting cloud can form. The scale on the right-hand side gives the maximum sizes of pyroclasts with density 1000 kg/m³ which can be transported out of the vent by the gas flow speeds given on the left-hand side. The dashed line marks the boundary between stable convecting clouds and 'collapsed' clouds feeding pyroclastic flows: all combinations of magma volatile content and mass eruption rate lying to the right of the line correspond to the pyroclastic flow case.

**Eruption Conditions Producing the Proposed Hecates Air Fall Deposit**

Figure 8 demonstrates that the two principal relationships between the various parameters involved are that between the exsolved magma volatile content and the size of the largest erupted clast and that between eruption cloud height and mass eruption rate. It would be possible to estimate directly the exsolved volatile content of the erupted material if information were available on the sizes of pyroclasts in the deposit. Unfortunately, there are neither visual, thermal, nor radar data of adequate resolution to provide this information for the Hecates Tholus area. We can, however, estimate the maximum height of the eruption cloud which would have produced the mantling deposit and hence infer the eruption rate; this, in turn, can be used to place some constraints on the magma volatile content, the depth of origin of the magma, and the size of the vent structure.

A critical step in the analysis is the recognition of the fact that the maximum heights of plinian eruption clouds on the earth are approximately equal to the widths of the deposits which they produce [Wilson, 1978]. This result appears to hold for all eruptions which take place under atmospheric conditions in which a sufficiently strong wind profile exists to produce a deposit with appreciable downwind elongation and is a consequence of the detailed way in which the widths of eruption clouds increase with height [Wilson, 1976]. Two major factors influence the shape of an eruption cloud: the scale height of the atmosphere and the value of the entrainment constant which relates the horizontal inflow velocity of atmospheric gas into the cloud to the vertical velocity on the cloud axis. The entrainment constant is not expected to be a function of the atmospheric properties [Briggs, 1969] and the scale heights of the Martian and terrestrial atmospheres differ by only about 25%; calculations based on developments of models described by Wilson [1976] and Sparks and Wilson [1982] suggest that the shapes of Martian eruption clouds will be very similar to those of terrestrial examples.

Figure 7 shows that the Hecates deposit is significantly elongate; its maximum width at the 0.1 crater/km² contour is about 50 km which can certainly be taken as a lower limit to the maximum eruption cloud height and implies an eruption rate of 6 \times 10⁶ kg/s (see Figure 8). It is possible, however, based on the locations of the 0.2 crater/km² contours, that the width is slightly greater than 80 km, implying an eruption rate of more than 4 \times 10⁷ kg/s. A plausible compromise is a width of 70 km and an eruption rate of 3 \times 10⁷ kg/s. This rate lies well within the range of mass eruption rates observed for terrestrial explosive eruptions which can range up to 3 \times 10⁹ kg/s [Wilson et al., 1978; Walker, 1980] and implies that the time required to erupt the estimated deposit mass of 7 \times 10¹³ kg was of the order of 20 to 30 days. This interval is long in comparison with the durations of historic plinian eruptions on earth, which have commonly been less than 1 day, but is quite comparable to the 20–27 days estimated for a prehistoric eruption in Central America [Ledbetter and Sparks, 1979]. It is, of course, possible that the observed deposit is the result of the superposition of a number of eruptions, in which case the 20- to 30-day interval is a lower limit to the total duration of all of the events.

The position of the dashed line in Figure 8a demonstrates that if an air fall deposit is to be formed rather than a pyroclastic flow deposit from an event in which the eruption rate is 3 \times 10⁷ kg/s, the total magma volatile content must be greater than about 1.5 wt % and the eruption velocity must be greater than about 370 m/s if the volatile is H₂O. Equivalent figures corresponding to the lower limit eruption rate of 6 \times 10⁶ kg/s are about 0.5 wt % and 220 m/s. If the magma volatile were CO₂, similar calculations (Figure 8b) show that the corresponding volatile contents would have to be 2.8 and 1.3 wt % and the velocities 240 and 140 m/s, respectively. The solubility of CO₂ in silicate melts is so low [Mysen, 1977] that these weight fractions could only be dissolved in the erupting magma if it had originated at a depth of not less than 100 to 50 km. The corresponding depths if the volatile were H₂O range from 4.1 to 0.9 km for a basaltic magma and 1.3 to 0.1 km for a silicic magma.

The scale on the right-hand side of Figure 8a shows that, for magma water contents in the range 0.5 to 1.5 wt %, minimum estimates of the sizes of the largest clasts expected to be commonly present in the air fall deposit lie between 10 and 30 mm for clasts with a density of 1000 kg/m³ (the sizes corresponding to other densities can be calculated by noting that the clast size is inversely proportional to the assumed density). The equivalent sizes have been calculated for CO₂ as the volatile phase (Figure 8b) and are found to be almost identical. Coarser clasts may be present in the vicinity of the vent, especially if fluctuations in the eruption conditions occur [Wilson et al., 1980] but are not expected to form a volumetrically large proportion of the deposit.

The lower limits on the volatile contents given above can be used to estimate the minimum size of vent structure expected to have been formed in the eruption. The mass eruption rate M can be expressed with sufficient accuracy as [e.g., Wilson and Head, 1981b]

\[
M = \frac{U \pi D^2 P x}{4 N Q T}
\]

where U is the eruption velocity, D is the vent diameter, P is the atmospheric pressure, x is the molecular weight of the volatile species, N is the exsolved volatile weight fraction, Q is the universal gas constant (equal to 8.3 \times 10³ J kg⁻¹ K⁻¹), and T is the magma temperature. We use P = 250 Pa for the summit of Hecates Tholus and assume T = 1000 K. Substitution of the critical values of N and U given above for H₂O (x = 18) yields minimum vent diameters of D = 1.6 km for M = 3 \times 10⁷ kg/s and 1.2 km for M = 6 \times 10⁶ kg/s. The values are increased to 1.8 and 1.3 km, respectively, if the volatile is CO₂ (x = 44). It should be stressed that these calculations assume that the magmatic gases involved in Martian eruptions are completely decompressed to local atmospheric
Fig. 8. Shown separately for plinian eruptions on Mars are the velocity of gas and small pyroclasts in the vent (left side of diagram) as a function of mass eruption rate (lower edge) and exsolved magma volatile content. Figure 8a (left) shows conditions for H₂O; Figure 8b (right) shows conditions for CO₂ (curves labelled in weight percent). The upper edge of each diagram shows the eruption cloud heights corresponding to the mass eruption rates on the lower edge. The right side shows the size of the largest clast having a density of 1000 kg/m³ which can be carried out of the vent at the corresponding gas speed on the left side. The dashed line in each figure marks the boundary between eruption conditions leading to a stable convecting eruption cloud (to the left of each line) and a ‘collapsed’ fountain over the vent feeding pyroclastic flows.
pressure. It is not guaranteed that this will be the case within the vent (see Wilson et al. [1980] for a discussion of this matter in relation to terrestrial eruptions), but equilibrium within the atmosphere must be reached not far above the vent. Thus the above estimates of vent structure size apply to the coarse debris deposit built up around the vent beneath the base of the eruption cloud; the fissure or conduit through which the erupting magma emerges may well be much smaller than suggested by the above figures—by a factor of up to 100 if the rocks in which the vent forms can withstand stresses up to a few tens of bars. The detailed topography to be expected in the deposit built around the vent is hard to predict: it will depend, for example, on the size distribution of the coarsest component of the ejecta, since this will control the ejecta dispersal [Wilson, 1976]; it is quite possible, however, that the vent structure will take the form of a depression with a diameter within which no material was deposited from the eruption cloud. Several depressions with sizes of this order are visible within the summit caldera complex on Hecates Tholus, and one occurs on the flank of the volcano about 10 km west of the outer edge of the complex (Figure 2).

Alternative Models for the Hecates Tholus Eruption

The analysis presented so far has been based on the assumption that the observed Hecates deposit was produced by a relatively steady, plinian explosive eruption. A number of other scenarios are possible, although most of these can be rejected for various reasons. For example, the extreme alternative to a single, steady eruption is a series of discrete, vulcanian-style explosions, possibly involving groundwater produced by melting of subsurface permafrost. The dispersal of the air fall deposit would still imply that a typical eruption cloud reached a height of about 70 km. Eruption clouds from discrete explosions on earth rise to heights \( H \) given by [Morton et al., 1956]

\[
H = 1.87E^{0.25}
\]  

(2)

where \( H \) is in meters and \( E \) is the released thermal energy in joules and is proportional to the erupted mass. This formula requires modification for the fact that the Martian and terrestrial atmospheres have differing properties [Wilson and Head, 1981a] and it can be shown that discrete eruption clouds will rise about 3 times higher on Mars than on earth.

To reach a height of 70 km on Mars, a discrete eruption cloud must consist of a mass equal to about \( 5 \times 10^{10} \) kg. The emplacement of the total deposit mass of \( 7 \times 10^{10} \) kg would then require about 1400 discrete explosions. We cannot completely exclude this eruption style, but no terrestrial equivalent exists.

The plinian eruption style is commonly associated with silicic magmas on earth, and it is possible that such magmas may exist on Mars [Rutherford and Hess, 1981]. However, the only general prerequisite for the plinian eruption style is a sufficiently large magma volatile content to ensure thorough disruption of the magma in or below the vent. Calculations [Wilson and Head, 1981a] show that a basaltic eruption like the 1886 eruption of Tarawera, New Zealand (G. P. L. Walker, personal communication, 1981), could have produced a Hecates-type mantle deposit had it occurred on Mars. Thus if the magma had both a sufficiently low viscosity and a sufficiently low rise velocity through the crust, then gas bubble coalescence could be so significant that it would lead to a more intermittent explosive style producing a coarse range of pyroclast grain sizes and a less widely dispersed strombolian or vulcanian deposit. For the wide range of magma viscosities and volatile contents that are typical for both lunar and terrestrial basalts, it is found that the magma rise velocity must exceed a critical value of about 0.5 to 1 m/s if intermittent explosions are to be avoided [Wilson and Head, 1981b]. We now demonstrate that if the Hecates Tholus magma were of basaltic to intermediate composition, its rise velocity would have been high enough to exceed this critical value.

For all except the highest magma rise speeds in the ensuing calculations, the motion of the erupting magma will be laminar as it rises through the crust; only within a few hundred meters of the surface (where gas exsolution causes a rapid velocity increase) will this approximation be generally invalid. Wilson and Head [1981b] show that, under laminar conditions and below the level of appreciable gas exsolution, the rise velocity \( u \) of a magma with density \( \rho \) and viscosity \( \mu \) driven by a density difference \( \Delta \) between the magma and the surrounding crust through a fissure of width \( W \)

\[
u = \frac{gW^2\Delta}{12\mu}
\]

(3)

where \( g \) is the acceleration due to gravity. The mass eruption rate \( M \) through the fissure of width \( W \) and length \( L \) (where both \( W \) and \( L \) are measured in the horizontal plane) is then

\[
M = \rho W L
\]

(4)

We already have a value for \( M \) in the Hecates eruption and can put constraints on the ratio of \( L \) and \( W \): the values of \( L/W \) expected for crustal fissures lie in the range \( 10^2 \) to \( 10^3 \) [Fedotov, 1978], though much smaller values (close to unity) are found locally where eruptive activity concentrates into part of a fissure [Crowe et al., 1981]. If we write \( Z = L/W \) to eliminate \( L \) from (4) and use (3) for \( u \), (4) becomes

\[
M = \frac{Z W^4 \rho g \Delta}{12 \mu}
\]

(5)

Solving for \( W \),

\[
W = \left( \frac{12 \mu M}{Z \rho g \Delta} \right)^{0.25}
\]

(6)

so that (3) can be written

\[
u = \left( \frac{Mg\Delta}{12Z \rho \mu} \right)^{0.5}
\]

(7)

\( L \) is just \( Z W \), by definition, so using (6),

\[
L = \left( \frac{12 \mu Z M^3}{\rho g \Delta} \right)^{0.25}
\]

(8)

Thus for any chosen value of \( Z \) (which we state above should lie in the range of 1 to 1000), we can use the measured value of \( M = 3 \times 10^7 \) kg/s, the standard value of \( g = 3.74 \) m/s\(^2\) and assumed values of \( \rho \), \( \Delta \), and \( \mu \) to find \( W \), \( u \), and \( L \). We choose \( \rho = 2800 \) kg/m\(^3\) as a plausible average magma density and \( \Delta = 300 \) kg/m\(^3\). The latter value is a compromise between typical density differences between silicate melts and solids, on the other hand, and somewhat larger apparent
density differences exist in eruptions where magma is accelerated by an excess pressure in the magma chamber [Wilson and Head, 198Ib]. A useful value to choose for is 10^3 Pa s. For viscosities of this order or larger, it can be shown [Wilson and Head, 1981b] that the rise speeds of the largest bubbles which can form in the magma as it approaches the surface (bubble diameters of a few millimeters [Sparks, 1978]) are less than 0.1% of the minimum rise speed of the magma (of the order of 10 mm/s, the value being set by the requirement that the magma must not cool too much during its ascent) and bubble coalescence is of negligible importance under all conditions. Also, since is inversely proportional to and equation (7)), lower values of than 10^2 Pa s will certainly lead to higher rise velocities than those calculated with this value. Thus if the velocities found using is 10^2 Pa s are greater than the critical velocity of 0.5 to 1 m/s, then it can be guaranteed that bubble coalescence could not have been important in the Hecates Tholus eruption.

Table 1 shows the results of the calculations of , , and for for Z = 1, 10, 100, 1000, and 10,000. The Reynolds number, Re = Wp/µ, is also included in each case to confirm the assertion that magma motion is laminar (Re < 500) in all cases except Z = 1. Clearly, greater than the critical value of 0.5 to 1 m/s as long as is less than 1000, which is exactly the result we are seeking. The implied active fissure (and, hence, surface vent) length could lie anywhere in the range from a few tens of meters to about 3 km, again consistent with the range of sizes of topographic structures visible in and around the caldera. Even if were as large as 10,000, the eruption would only be slightly unsteady with . However, if is 10^4, then and a linear vent structure of this dimension would easily be visible in the Viking images, since the vent would exceed the 10-km diameter of the summit caldera complex (see Figure 2). Thus we conclude that, at a discharge rate of 3 x 10^7 kg/s, the Hecates eruption would have been of a steady plinian type whatever the magma composition.

A final factor in the eruption deserves attention: the source of the volatile component in the erupting magma. We have demonstrated earlier that a stable convecting eruption would have demonstrated earlier that a stable convecting eruption could only have existed if the erupting magma contained at least 1 wt % H_2O or 2 wt % CO_2 and have so far assumed that the volatile phase was present in the magma at depth in the crust. There are two possible mechanisms of introducing volatiles into the magma during the eruption: (1) incorporation of volatile-bearing country rock into the erupting magma by attrition and assimilation of conduit walls and (2) entrainment by the magma of liquid water from an aquifer intersecting the conduit system.

In order to introduce 1 wt % water into the total erupted deposit volume of 23 km^3 (dense rock equivalent) we could reasonably postulate that the erupting magma incorporated 10% by weight, i.e., 2.3 km^3, of country rock which contained 10% by weight H_2O as ice. This proportion of country rock is quite common in plinian deposits. The latent heat required to melt and evaporate the ice could be taken from the magma with negligible effect on the temperature of this low volatile concentration. Permafrost may exist on Mars down to depths of several kilometers [Carr and Schaber, 1977] and if we assume a depth of 5 km, the required volume of country rock could be supplied by the excavation of a region having the shape of an inverted circular cone if the cone had a diameter of 1.3 km at the surface. This diameter happens to be close to (and, as is necessary, less than) the 1.6-km-diameter surface vent size calculated above as being needed to allow the emerging mixture of clasts and gas to decompress to the local atmospheric pressure. It is likely that a vent structure with the steep-sided geometry implied by this calculation would be unstable after the eruption ceased. The collapse of the deeper parts of the vacated region would enlarge the surface expression of the vent. If the collapse continued until the slopes were everywhere at a typical angle of rest of 30°, the resulting structure would be a depression just over 3 km in diameter and 900 m deep. Since the greatest depth of the caldera currently appears to be no more than 450 m, the diameter of the depression should in fact be at least 4.4 km, a value somewhat larger than the size of the largest of the four discrete craters visible within the caldera complex. It would clearly be necessary to postulate that at least two of these craters formed as a direct consequence of the country rock excavation process. The above calculation would not be changed significantly if CO_2 ice were used as the volatile; given the current uncertainties in estimates of the permafrost content of the Martian near-surface layers, it seems that we should not exclude the possibility that a significant contribution to the magma volatiles was made by such near-surface layers.

The main problem associated with supplying water to the magma from an aquifer is the need to maintain the temperature of a large enough volume of subsurface layers within the appropriate range (0°C to a few hundred degrees Centigrade, the exact upper limit depending on the pressure and, hence, depth of the base of the aquifer). Given the availability of water, there seems no reason to think that the magma cannot assimilate it [Stewart and Hulston, 1976]. A constraint on the required geometry can be set by considering an aquifer intersecting the fissure feeding the vent. Let the horizontal flow speed of water through the contact area be v and let the contact area be of vertical extent h along both sides of the fissure of length . The mass rate at which water must be supplied to the magma is 0.01 M, since we require 1 wt % of water to be mixed with the magma erupting at mass rate M = 3 x 10^7 kg/s. Hence we have

\[0.01 M = 0.3(2L)\rho_w u h \tag{9}\]

where \(\rho_w\) is the water density and the factor 0.3 is a deliberately generous estimate of the pore space fraction in the aquifer. For any chosen value of L (see Table 1) we can find pairs of values of v and h which satisfy the equation. If we choose L = 3300 m, the largest allowed value for Z less than 1000, we shall obtain the most conservative values for v and h. Information on likely values of v comes from data on hydraulic conductivities of terrestrial aquifers [De Weist, 1965]: in coarse gravels, v may be as large as 0.1 m/s; more typical values for sandstones cover the range 10^-2 to 10^-6 m/s. Table 2 shows the values of h implied by assuming values

<table>
<thead>
<tr>
<th>Z</th>
<th>W, m</th>
<th>u, m/s</th>
<th>L, m</th>
<th>Re</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>18.3</td>
<td>32</td>
<td>18.3</td>
<td>1.6 x 10^4</td>
</tr>
<tr>
<td>10</td>
<td>10.3</td>
<td>10</td>
<td>103</td>
<td>288</td>
</tr>
<tr>
<td>10^2</td>
<td>5.8</td>
<td>3.2</td>
<td>580</td>
<td>52</td>
</tr>
<tr>
<td>10^3</td>
<td>3.3</td>
<td>1.0</td>
<td>3300</td>
<td>9</td>
</tr>
<tr>
<td>10^4</td>
<td>1.8</td>
<td>0.3</td>
<td>18000</td>
<td>1.5</td>
</tr>
</tbody>
</table>
of the diameter, $X$, of the (circular) region around the vent which must be capable of supplying water to make up the total volume required for the eruption. Clearly, solutions with $h$ much greater than about 3 km are forbidden by the inferences that martian permafrost extends only down to depths of this order [Carr and Schaber, 1977; Cintala and Mouginis-Mark, 1980].

The problem is now resolved into the need to supply magmatic heat to a region of extent $X$ in sufficient quantities to melt an ice thickness about 0.3 km. For $h = 3$ km, $X = 550$ m we would need to assume dike-like magma bodies around the main fissure while for $h = 30$ m, $X = 5.5$ km the required structure is a sill. A simple heat sharing calculation shows that the volume of magma required is about 80% of the volume of ice to be melted, so in the case of $h = 30$ m, $X = 5.5$ km, the sill would need to be 8 m thick. The time required to melt the ice would be of the order of a year, being dictated by the low thermal conductivity of the magma. Similar time scales and geometries could be contrived for the case $h = 3$ km, $X = 550$ m, by a suitable set of assumptions about dikes radiating from the main fissure. The volume of magma needed to melt the ice would be about 0.2 km$^3$, nearly 1% of the total volume of erupted magma. Again it seems possible for a significant quantity of near-surface volatiles to be supplied to the erupting magma; however, numerous favorable circumstances must combine to allow this. There must be an episode of near-surface sill or dike emplacement which takes place at just the right interval of time before the main eruption to ensure that liquid water is available; the water must be formed in strata which have a large hydraulic conductivity (typical of that of clean sand or gravel on earth); the strata probably have to dip toward the center of the volcano to ensure a continued water supply from a large enough area (although outward dipping strata are more likely to be encountered on volcanoes). It is difficult to evaluate the likelihood of all these conditions being met simultaneously, but it is probably low.

**Implications and Conclusions**

The preceding analysis of both the morphology and eruption characteristics of Hecates Tholus mainly considers the processes by which the surficial deposits on the present shield were produced and modified. At a second level of investigation, however, certain inferences can also be drawn concerning the general characteristics of volcanic activity on Hecates and the physical properties of the parent magma chamber. With varying degrees of confidence, estimates of the age of the air fall deposit, the repose period, and the duration of intermittent activity can all be made.

Of particular relevance to resolving the history of Hecates is the determination of the ages of the air fall deposit and the surrounding flanks of the volcano. At the 40-m per pixel resolution of the best available Viking images, the lack of any fresh craters within the mantled region is striking. Although no absolute age can be assigned to this deposit, the existing crater flux models for Mars [Soderblom et al., 1974; Neukum and Hiller, 1981] would indicate a very young age.

In particular, it appears to be significant that this area of Hecates has a lower crater density than the youngest part of the Olympus Mons caldera, which Neukum and Hiller [1981] assign an age of less than 300 m.y. Contrasting with this very young age for the Hecates air fall deposit, Plescia and Saunders [1979] and Neukum and Hiller [1981] observed that variations in crater number density over the entire Hecates shield clearly show that earlier resurfacing events have also occurred, indicating that the volcano was active for a protracted period of time, probably in excess of hundreds of millions of years.

Recognition of such an extended period of volcanic activity, together with the constraints imposed on magma chamber depth by the solubility of magma volatiles, in turn provides some restriction on the average repose period between successive eruptions on Hecates. Based on cooling calculations for terrestrial magma bodies [Lachenbruch et al., 1976], a magma body of any given size can only remain molten for a certain period of time at any given depth (unless it is replenished by fresh, hot magma from below). For example, a magma chamber equivalent in volume to the Hecates air fall would have to be buried to a depth in excess of 100 km in order for it to remain molten for more than about 300 m.y. Although the lithospheric structure is poorly constrained for the region of Mars beneath Hecates Tholus (due to the absence of circumferential graben around the volcano, poor topographic information, and the lack of gravity data; cf. Comer et al. [1980]), such a depth of origin for the Hecates magma would most likely place the source region within the mantle. Solidification times for the same magma body at shallower depths would be shorter than 300 m.y., due to the lower temperature of the surrounding rocks.

In the case of CO$_2$ being the driving volatile for activity on Hecates, it appears, therefore, that one major eruption every 300 m.y. might be possible if the parental magma came directly from mantle depths. The eruption of mantle-derived magmas seems to characterize certain high effusion rate basaltic eruptions on earth [Parke et al., 1968; Swanson et al., 1975] and also appears to be theoretically possible on Mars [Wilson and Head, 1981a]. For the Hecates event, a 100-km depth of origin (and the long repose period) is also possible for H$_2$O, but it is not essential. As discussed above, the solubility of H$_2$O in a magma dictates that the source region lies at a depth of at least 0.2-4 km. If this minimum depth were the actual value, however, the magma solidification would take place in about 0.1 m.y., indicating that each new cycle of magma chamber replenishment would have to occur within this time interval. Therefore if the explosive activity is to be attributed to the magma adsorbing groundwater, it is possible to constrain the average repose period of Hecates to be significantly less than a million years. On the other hand, if the volatile was juvenile water, eruptions might only have occurred once every few hundred million years.

We believe that the salient features of the hypothesized Hecates eruption, and the implications which it holds for the long-term evolution of the volcano, can be summarized as follows:
1. The lack of superposed craters indicates that the deposit (and hence the eruption) is very young, probably less than 300 m.y. B.P.

2. From the spatial variation of the number of small impact craters on the volcano, the air fall deposit appears to mantle an elongate region at least 50 to 70 km in extent to a depth of up to 100 m, implying a dense rock volume of more than 20 km$^3$ and a corresponding mass of about 7 × 10$^{15}$ kg. On the basis of both theoretical and observational studies of the shapes of eruption clouds, the deposit width implies a best estimate of 70 km for the eruption cloud height and an implied mass eruption rate of 3 × 10$^7$ kg/s. The eruption duration must have been at least 20 to 30 days.

3. Because a stable eruption column was able to form (rather than a low fountain feeding pyroclastic flows), the implied magma volatile content must have been at least about 1 wt % if the volatile were H$_2$O and greater than about 2 wt % if the volatile were CO$_2$.

4. The implied magma volatile contents in turn require that the magma originated at depths in excess of 50 to 100 km if the volatile were CO$_2$ and 150 m to 4 km if it were H$_2$O. These depths essentially restrict the magma sources to the region of the Martian mantle if CO$_2$ were the sole volatile but impose no restrictions on the depth of origin if H$_2$O were the main component. These conditions for H$_2$O also include the possibility of the absorption of near-surface groundwater as an important volatile source as a consequence of these volatile contents. The associated velocities of the gas and small pyroclasts within the vent would have been 140 to 240 m/s for CO$_2$ or 220 to 370 m/s for H$_2$O, indicating that the largest pyroclastic blocks should be 10 to 30 mm in diameter (irrespective of volatile species).

5. The cratering flux models derived by Soderblom et al. [1974] and Neukum and Hiller [1981], together with the variations in impact crater number density reported here (Figure 7), show that earlier resurfacing events have probably taken place on Herats. A more detailed analysis of the volcano's crater distribution may therefore allow the products of these earlier eruptions to be recognized.

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