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Numerical Modeling of Ejecta Dispersal from Transient Volcanic Explosions on Mars

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The dynamics of ejecta dispersal in transient volcanic eruptions on Mars are distinct from those on Earth and Venus because of the low atmospheric pressure and gravitational acceleration. Numerical modeling of the physical mechanisms of such activity, accounting for the different martian environmental conditions, can help constrain the style of emplacement of the eruptive products. The scenario envisaged is one of pressurized gas, contributed from either a magmatic or meteoric source, accumulating in the near-surface crust beneath a retaining medium. On failure of the confining material, the gas expands rapidly out of the vent, displacing both the "caprock" and a mass of atmospheric gas overlying the explosion site, in a discrete, transient event. Trajectories of large blocks of ejecta are computed subject to the complex aerodynamic interactions of atmospheric and volcanic gases which are set in motion by the initiation of the explosion.

Reservoirs of crustal and surface water and carbon dioxide may have increased the chances of occurrence of transient explosive events on Mars in two ways: by supplying a source of volatiles for vaporization by the magma and by acting to slow the ascent of the magma by chilling it, providing conditions favorable for gas accumulation.

Results of the modeling indicate that ejection velocities ranging up to ~580 m sec^{-1} were possible in martian H_2O-driven explosions, with CO_2-driven velocities typically a factor of ~1.5 smaller. Travel distances of large blocks of ejecta lie within the range of a few kilometers to the order of 100 km from the vent. The low martian atmospheric pressure and gravity would thus have conspired to produce more vigorous explosions and more widely dispersed deposits than are associated with analogous events on Earth or Venus.

Other phenomena likely to be associated with transient explosions include ashfall deposits from associated convecting clouds of fine material, pyroclastic flows, and ejecta impact crater fields. It is anticipated that the martian environment would have caused such features to be greater in size than would be the case in the terrestrial environment. Ash clouds associated with discrete explosions are expected to have risen to a maximum of ~25 km on Mars, producing deposits having similar widths. Another indication of a volcanic explosion site might be found in areas of high regolith ice content, such as fretted terrains, where ice removal and mass-wasting may have modified the vent's initial morphology.

The modeling results highlight the implications of the occurrence of transient explosive eruptions for the global crustal volatile distribution and provide some predictions of the likely manifestation of such activity for testing by upcoming spacecraft missions to Mars.

INTRODUCTION

The widespread presence of extensive lava flows indicates that volcanic activity on Mars was dominantly effusive, at least later in the planet's evolution. Evidence suggests that pyroclastic activity may have been more common earlier in Mars' history, as implied by the identification of pyroclastic origins for paterae edifices in ancient highland regions (e.g., Reimers and Komar 1979, Greeley and Spudis 1981, Greeley and Crown 1990, Crown and Greeley 1993, Robinson *et al*. 1993). Since the current martian environment would act to increase the likelihood of explosive eruptions (Wilson and Head 1983), the identification of innumerable young(er) lava flows has important implications for volatile loss from martian magmas (Fagents 1994, 1996, Fagents and Wilson 1995a): either volatiles are lost during magma storage or transport or the magma source region is depleted in volatiles, which has implications for planetary degassing processes.

However, even if magmas are volatile-poor, it has been

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suggested that transient vulcanian style explosive eruptions could occur as a result of gas accumulation in the shallow crust (Wilson 1980). There are two possible sources of these volatiles: juvenile or meteoric. In the former case, it is possible to initiate a transient explosion at low magmatic volatile contents, provided that sufficient time is allowed for the gas to accumulate. If a meteoric volatile is exploited, all that is required is a crustal store of the volatile and suitable conditions for heat from the magma to vaporize and pressurize the volatile. On Mars, the presence of crustal and surficial stores of H_2O and CO_2 ice over much of the planet, and the possibility of liquid water aquifers (Squyres *et al.* 1992), indicate a significant potential reservoir of gas for driving volcanic explosions. Indeed, it may be argued that the widespread crustal permafrost may have increased the likelihood of vulcanian eruptions on Mars.

In this paper we present the results of a model for transient volcanic eruptions on Mars, in which large blocks of ejecta are emplaced semi-ballistically from discrete explosions. By considering the effects of martian atmospheric characteristics and the acceleration due to gravity, together with data on analogous eruptions on Earth, we determine the eruption velocities and expected distribution of ejecta for plausible ranges of pre-eruption conditions in the vent. We also consider the implications of the martian environment for other phenomena associated with vulcanian eruptions (e.g., pyroclastic flows, ashfall, secondary crater fields), and provide some predictions for features that may identify the sites of vulcanian style explosions on Mars.

BACKGROUND

Transient Volcanic Explosions

Vulcanian eruptions are defined as discrete, intermittent explosions, separated by intervals of minutes to days, originating in the near surface parts of the volcano (Wilson 1980). This eruption style appears to be a consequence of dikes stalling as intrusions at shallow depths and enabling a gas-rich layer to accumulate at the top of the magma body (Self *et al.* 1979, Wilson 1980). If the pathways to the surface are efficiently sealed by a coherent "caprock," pressure may increase to the point where the caprock fails catastrophically such that the rock and gas are violently expelled in a transient eruption. Field evidence favors this mode of eruption since commonly less than 50% of the ejecta comprises fresh magma (Nairn and Self 1978), consistent with the expulsion of a pre-existing caprock. Gas to drive the eruption may be contributed from two sources: from the degassing magma body, and/or from the vaporization of meteoric water in porous country rocks in contact with the magma.

A number of workers have attempted to relate the observed distribution of large blocks ejected in discrete volcanic (vulcanian) explosions to the pre-eruptive parameters such as gas concentration and excess pressure by deriving the velocity with which they were ejected and relating this to the initial conditions (e.g., Minakami 1950, Gorshkov 1959, Decker and Hadikusumo 1961, Gorshkov and Bogoyavlenskaya 1965, Hédervári 1968, Fudali and Melson 1972, Steinberg and Steinberg 1975, Nairn 1976, Steinberg 1977, Steinberg and Babenko 1978, Self *et al.* 1980, Wilson 1980). A review of treatments of the mechanisms and dynamics of discrete volcanic explosions was given in Fagents and Wilson (1995b), in which it was pointed out that most efforts had suffered from inappropriate expressions describing the relationship between the pre-eruption gas pressure and the eruption velocity, and/or from inappropriate treatments of the influence of the atmosphere on the trajectories of ejected blocks.

A new model for transient volcanic eruptions has been developed and used to model terrestrial eruptions for which good field data on ejecta distribution were available (Fagents and Wilson 1993; see Fig. 1). Discrete explosions typically eject blocks having diameters ranging from a few tens of centimeters to several meters, to distances extending to several kilometers, from craters commonly several tens of meters wide (e.g., Fudali and Melson 1972, Nairn and Self 1978, Self *et al.* 1980). For the eruptions studied, it was demonstrated that pre-eruptive gas pressures lie in the range 0.2 to 20 MPa, entirely compatible with the expected range of tensile strengths of caprock materials. Ratios of the gas to caprock mass were found to be in the range 0.01 to 0.10 for andesitic eruptions, implying some concentration over typical magmatic gas values, consistent with the accumulation of pressurized gas under a retaining lid (Fagents and Wilson 1993). The success of this treatment over previous methods lies in the detailed treatment of the aerodynamic interactions between ejected blocks, driving gas and the atmosphere overlying the explosion source, which will be displaced *en masse* upon initiation of the explosion. Since the velocity of the accelerating clasts relative to the displaced gas is initially low, so will be the drag forces acting on the clasts, which is contrary to all previous treatments. Much lower pressures and gas concentrations are therefore required to explain the observed dispersal of ejecta.

The potential has also been explored for vulcanian eruptions to distribute pyroclastic material on Venus (Fagents and Wilson 1995b), where the extreme conditions of high atmospheric pressure and temperature (~10 MPa and ~740 K at the mean planetary radius (mpr), Kliore *et al.* 1985) require high volatile contents and eruption temperatures for sustained explosive activity to occur (Garvin *et al.* 1982, Head and Wilson 1986, Thornhill 1993). These different environmental conditions on Venus were shown to affect the ejection of material in transient explosions significantly. The model predictions suggest that coarse ejecta may travel a maximum distance of ~1 km from the vent (cf. ~11 km on Earth) in the extreme case, but will typically be distributed within a radius of a few tens to a few hundreds of meters from the vent. These blocky deposits would typically be below the limit of resolution in Magellan radar images. However, fallout of ash from the convecting eruption clouds associated with explosive events may result in deposits commonly having widths ranging up to ~5 km and much greater downwind dimensions (Fagents and Wilson 1995b). Thus explosive activity on a moderate scale on Venus may be relatively common (as born out by image data of possible ashfall deposits associated with small volcanic edifices (e.g., Head *et al.* 1991, Wenrich and Greeley 1992, Bulmer 1994).

Martian Volcanism

Mars displays evidence of extensive volcanic activity having occurred throughout its geologic history, with the most recent volcanism occurring on the order of 100 Ma ago (Strom *et al.* 1992, Tanaka *et al.* 1992). Older, pyroclastic activity appears confined to the southern highlands regions, as evidenced by the paterae edifices, whereas younger activity to the north appears to have been predominantly effusive, with Olympus Mons possessing some of the youngest surfaces (Strom *et al.* 1992, Tanaka *et al.* 1992).

Several independent lines of evidence indicate that compositions of martian igneous rocks are mafic to ultramafic (Greeley and Spudis 1981). Viking Lander analyses (Arvidson *et al.* 1989), spectroscopic data (Soderblom 1993), compositions derived from geophysical modeling (McGetchin and Smyth 1978), and the morphological similarity between volcanic edifices and lava flows on Mars and on Earth (e.g., Carr *et al.* 1977, Carr 1981) all suggest that martian volcanism is likely to have been basaltic in nature.

Evidence for more silicic rock compositions on Mars is inconclusive (Francis and Wood 1982). However, given that the environmental conditions conspire to encourage explosive volcanism on Mars, more evolved or more volatile-rich magmas are not deemed necessary for explosive eruptions to have occurred (Wilson and Head 1983; Mouginis-Mark *et al.* 1992).

The martian physical environment is anticipated to have influenced the style of volcanic activity strongly, increasing the likelihood of explosive eruptions (Wilson and Head 1983, Wilson 1984, Mouginis-Mark *et al.* 1992, Fagents 1994). Whereas on Venus the heavy atmosphere (~10 MPa at mpr) dominates control of the style of eruption and subsequent distribution of the eruptive products, on Mars both the low atmospheric pressure (~600 Pa at mpr, cf. ~10^5 Pa at sea level on Earth) and the acceleration due to gravity (~3.7 m sec^{-2}, cf. ~9.8 m sec^{-2} on Earth) will have strongly influenced the eruptive style. In terms of the explosion model, the greater degree of expansion undergone by pressurized gases from an initial high pressure down to the low ambient pressure means that there is a greater energy release per unit mass with which to drive the eruption. Hence expected eruption velocities are higher than those observed on Earth. The high velocities, together with the low gravity inhibiting clast settling, and the reduced aerodynamic drag offered by the low density atmosphere, imply larger clast ranges.

The high abundance of CO_2 (and lack of H_2O) in the martian atmosphere suggests that CO_2 may have been an important volatile species in the magma source region. However, it is well established that there are significant crustal reservoirs of H_2O in ice and possibly liquid phases (Carr 1979), some of which may be volcanic in origin. Surface deposits of H_2O and CO_2 ice have been detected visually (Cutts *et al.* 1976), thermally (Kieffer *et al.* 1976a,b), and spectroscopically (Herr and Pimentel 1969, Larsen and Fink 1972, Clark and McCord 1982), while several theoretical (Leighton and Murray 1966, Anderson *et al.* 1967, Smoluchowski 1968) and morphological studies (Allen 1979, Hodges and Moore 1979, Mouginis-Mark *et al.* 1984, Carr 1986, Mouginis-Mark 1987) have predicted the existence of subsurface H_2O ice. Recent calculations have shown subsurface ice to be thermodynamically stable over a much greater area of the martian surface than was previously thought (Paige 1992). At greater depths in the lithosphere the intersection of the martian geotherm with the melting point of H_2O ice implies the existence of a liquid water aquifer, and such a groundwater system has been proposed on a global scale (Carr 1979, Clifford 1986).

Despite the presence of H_2O and CO_2 in the martian crust and atmosphere, it is unclear whether significant stores of volatiles remain in the magma source region, since there appears to have been a general change from explosive to effusive volcanism throughout Mars' history (Mouginis-Mark *et al.* 1992). Even if martian magmas were low in volatiles, this may have been more than compensated for by the ability of magma to interact with crustal reservoirs of volatile compounds: the presence of these various crustal stores of volatiles would therefore ensure that plenty of accidental gas would be available to collect under the caprock overlying an intrusion.

Another important ramification of the existence of large volumes of ice and water within the martian crust is the potential for chilling and consequent retardation of the ascending magma. Since it is generally accepted that volcanic activity on Mars involved a predominance of mafic to ultramafic composition magmas, there may be a conceptual difficulty (despite the fact that upward buoyancy forces on the magma would be reduced as a result of the low martian gravity) in allowing such magmas, which typically would be of low viscosity (1 to 100 Pa sec, Basaltic Volcanism Study Project 1981) and hence very mobile, to halt their ascent and stall near to the surface so that a gas pocket may form. However, the presence of crustal liquid water or ice could act to slow the rise of the melt in a dike by causing significant convective cooling of its leading surface and inducing a much higher viscosity. Since it is typically the behavior of the outermost crust or "skin" of a magma that exerts the greatest control over its motion (Kilburn 1993), resistance to deformation of the chilled surface may be sufficient to delay the ascent of the magma. Convective cooling appears to have operated at Poas volcano, Costa Rica (Brown *et al.* 1987, 1991), and at Soufrière, St. Vincent (Shepherd and Sigurdsson 1978), where crater lakes provided temporary heat sinks for the rising magma, thus delaying the onset of the eruption. In a similar way a magma body on Mars may stall long enough for gas to establish itself as a discrete region above the magmatic liquid.

RESULTS OF MODELING

The two FORTRAN computer programs that comprise the explosion model (Fagents and Wilson 1993) were modified for the martian environment. Atmospheric temperature and pressure models were derived from Viking Lander data (Kieffer *et al.* 1976a,b, Seiff and Kirk 1977, Davies 1979), appropriate expressions were employed to describe the thermodynamic behavior of the predominantly CO_2 atmosphere (e.g., molecular weight $m_w = 43.486 \pm 0.066$ (Seiff and Kirk 1977) and specific gas constant $R = 191.17$ J kg^{-1} K^{-1}, corresponding to an atmosphere comprising 0.9555 mole fraction CO_2, 0.027 ± 0.003 mole fraction N_2, 0.016 ± 0.003 mole fraction Ar, and 0.0015 ± 0.0005 mole fraction O_2, as measured by the Viking Lander 2 mass spectrometer (Owen and Biemann 1976)), and suitable values were used for other environmental factors (e.g., gravitational acceleration).

Figure 1 illustrates the basis of the explosion model. Gas at pressure P_{gz} and density ρ_{gz}, having a mass $m_g = (1/3)\rho_g\Omega r_1^3$, accumulates beneath a caprock of density ρ_s and thickness r_{21} in a region of radius r_1 subtending a solid angle Ω at a point. r_2 represents the distance to the outer edge of the caprock. Upon failure of the caprock, the gas expands adiabatically out of the vent, ejecting the caprock (mass $m_s = (1/3)\rho_s\Omega(r_2^3 - r_1^3)$) and displacing the atmosphere (mass $m_a = (1/3)\rho_a\Omega[(r + r_{21})^3 - r_2^3]$) ahead of it. The mass ratio of pressurized gas to caprock material is defined as $n = m_g/m_s$. The equation of motion of the caprock and displaced atmospheric gas is given by

$$\left\{P_{gz}\left(\frac{r_1}{r}\right)^{3\gamma} - P_a\right\}\Omega r^2 = \frac{1}{3}\Omega\left(\frac{d^2r}{dt^2}\right)\{\rho_s(r_2^3 - r_1^3) + \rho_a((r + r_{21})^3 - r_2^3)\}, \tag{1}$$

in which γ is the ratio of the specific heats of the volcanic gas. This equation is integrated using a simple first-order scheme to obtain the maximum velocity u_0 after the initial gas expansion and the distance R_0 at which u_0 is attained. It is assumed that blocks of caprock of a given size and density are launched from this location with this velocity into the moving gas, the velocity of which now decays as a function of time, t, and distance from the vent, R, as (Fagents and Wilson 1993)

$$u = u_0\left(\frac{R_0}{R}\right)^2 e^{-t/\tau}, \tag{2}$$

in which τ is a time constant defined as the difference between the time taken for the maximum velocity, u_0, to be attained and the duration of the entire gas expansion phase (Fagents and Wilson 1995b).

A fourth-order Runge–Kutta scheme (Wilson 1972, Fagents and Wilson 1993, 1995b) is used to follow the trajectory of the clast until it intersects the ground, subject to the complex and varying drag forces imposed as a result of the motion induced in the atmosphere by the explosion.

The ranges of values of initial parameters crucial to the explosion model have been discussed previously (Fagents and Wilson 1993, Fagents and Wilson 1995b) and are held to apply here. The modeling was carried out exploring the full range of plausible values: P_{gz} between 0.01 and 20 MPa (in keeping with the likely range of caprock tensile strength); and gas/caprock mass ratio, n, between 0.01 and 0.10. The values used for n are based on plausible geometries of vent regions and magma contact with saturated country rock; the lower limit is suggested by typical mafic magma gas solubilities (Basaltic Volcanism Study Project, 1981), the upper limit by the efficiency of magma–country rock thermal interactions (Wilson 1980). The gas region radius, r_1 (see Fig. 1), is taken to be 50 m and the angle of ejection of each block is taken as 45°. In each of the cases modeled here the clast radius is taken as 1 m, since the ranges of clast sizes observed in terrestrial vulcanian eruptions have a mode near 1 m (e.g., Fudali and Melson 1972, Nairn and Self 1978, Self *et al.* 1980). This is likely to be a valid assumption for other planets since the fragmentation processes are dependent on rock properties, which are essentially planet-independent. The clast density is 2600 kg m^{-3}. Because of the uncertainty regarding which volatiles are likely to be in the magma, both H_2O and CO_2 are considered as possible driving volatiles, representing end members for the range of possible high- and low-molecular-weight volcanic gases. The explosions are modeled with starting gas temperatures of 1200 and 800 K. The former represents little cooling undergone by a gas from a basaltic magmatic source, whereas the latter case could represent either juvenile gas that has undergone some cooling, or a vaporized meteoric volatile which will not reach such a high temperature. Finally, explosions occurring at

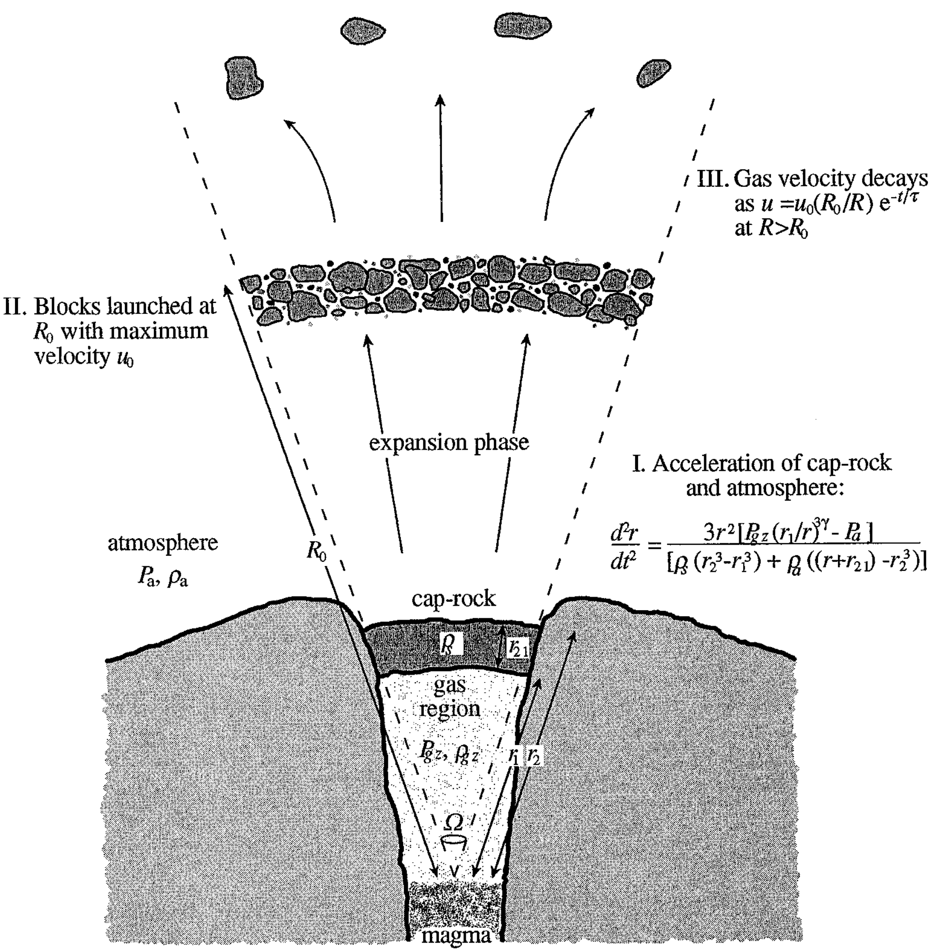

FIG. 1. Schematic representation of the model for vulcanian explosions (see text for detailed explanation).

varying elevations between the mean planetary radius (mpr = 3390 km) and the maximum planetary elevation (mpr +25 km) are modeled in order to investigate the influence of the varying atmospheric characteristics.

Figures 2 and 3 present the results of the modeling for Mars. Plots of ejection velocity against initial gas pressure are shown in Fig. 2. In each case it is clear that the velocity attained increases with pressure and gas content, as would be expected. In addition, the velocities for the cases when the vent elevation is taken as mpr +25 km (represented by the dashed lines) are significantly higher than for lower elevations. This is a result of the lower atmospheric pressure (which decreases by an order of magnitude over the 25-km rise in elevation) allowing greater gas expansion: hence more energy is available to drive the eruption. The velocity–pressure curves for mpr and mpr +25 km become more closely spaced at higher pressures. This reflects the nature of the martian atmosphere: at high gas pressures in the low pressure environment the gas involved in the explosion is starting to reach the limits of expansion, the ultimate limit being expansion into a vacuum; no matter how much the initial pressure is increased, the velocity reached at the end of expansion phase tends to a finite limit, which will be the same for any vent elevation.

The differences between a starting gas temperature of 1200 K (Figs. 2a and 2c) and one of 800 K (Figs. 2b and 2d) are also clear: significantly lower velocities are attained as a result of lesser expansion of cooler gas which liberates less energy and therefore the mass of caprock and atmosphere cannot be accelerated to such a great degree. Comparison of Figs. 2c and 2d with 2a and 2b also demonstrates the effect of advocating CO_2 as the gas driving the explosion: CO_2-driven velocities are lower than those for H_2O by a factor of ~1.5. This is a result of the greater molecular weight of CO_2 having the effect of increasing the gas density in the vent for any given gas pressure. The gas mass is much greater than for H_2O, which ensures a greater caprock mass for any chosen mass fraction, n. The consequent effect is a significant diminution of the accelerations of the caprock

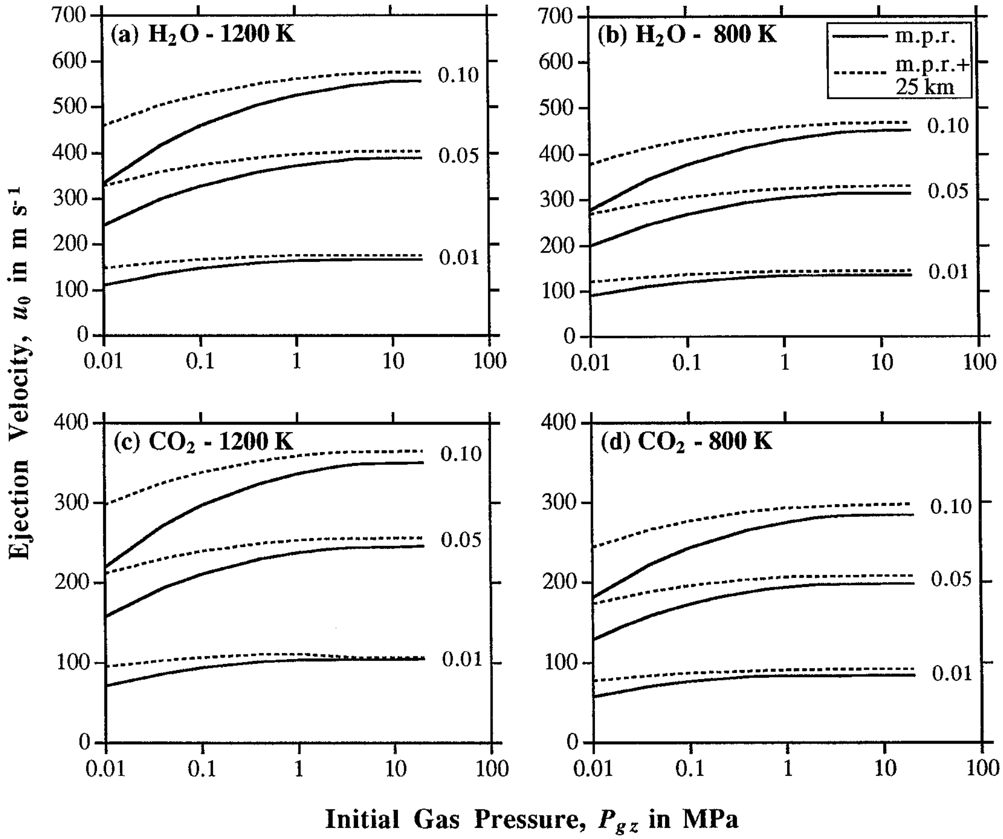

FIG. 2. Ejection velocity, u_0, as a function of excess gas pressure, P_{gz}, immediately prior to eruption. Curves are marked with the value of gas to caprock mass ratio, $n(=m_g/m_s)$; solid lines represent explosions taking place at the mean planetary radius (mpr), whereas dashed lines represent a vent altitude of mpr +25 km—equivalent to the summit region of Olympus Mons, the tallest edifice on Mars. (a) H_2O as the driving gas, initial gas temperature 1200 K; (b) H_2O initially at 800 K; (c) CO_2 at 1200 K; (d) CO_2 at 800 K.

and displaced atmosphere and hence much lower final velocities after the initial gas expansion phase.

The dependence of clast range on gas pressure and concentration (Fig. 3) follows essentially the same trends as for ejection velocity (Fig. 2). However, the tendency for mpr +25 km clast range values to be similar to mpr values at higher pressures is much reduced with respect to ejection velocity. This illustrates the fact that clast range is not simply a function of the ejection velocity, but is also strongly dependent on atmospheric density (and hence drag resistance offered by the atmosphere), which is lower at high altitudes.

A comparison of clast range as a function of excess gas pressure is shown for Venus, Earth, and Mars in Fig. 4. In these cases the gas/caprock mass ratio is taken as 0.1, the initial gas temperature is 1200 K, and the vent altitude is equal to the mpr on Venus and Mars and mean sea level on Earth. Model results are shown for both H_2O (solid lines) and CO_2 (dotted lines). The effect of the different planetary environments on the magnitude of the eruption can clearly be seen. The thin atmosphere and low gravity on Mars cause clast ranges to be greater by one to two orders of magnitude than those on Earth, as a result of both the greater ejection velocities attained for any chosen set of initial conditions and the longer clast ranges (due to low drag resistance and low gravity) attained for any eruption velocity. In contrast, the extreme atmospheric pressure on Venus has the effect of greatly reducing clast distances with respect to Earth (by an order of magnitude), by both suppressing gas expansion out of the vent and by offering much greater aerodynamic resistance to the ejected clasts (Fagents and Wilson 1995b).

Finally, while venusian clast ranges are more strongly dependent on the size of the gas region, velocities and ranges for Mars are generally so much larger that the contribution to the total range made by the initial expansion phase tends to be insignificant compared with the distance traveled by the clast in free flight (i.e., after the caprock has fragmented and the blocks are launched into the gas-stream). Thus a two order of magnitude increase in the initial gas region radius, r_1, produces very little percentage increase in clast range (and has been omitted from

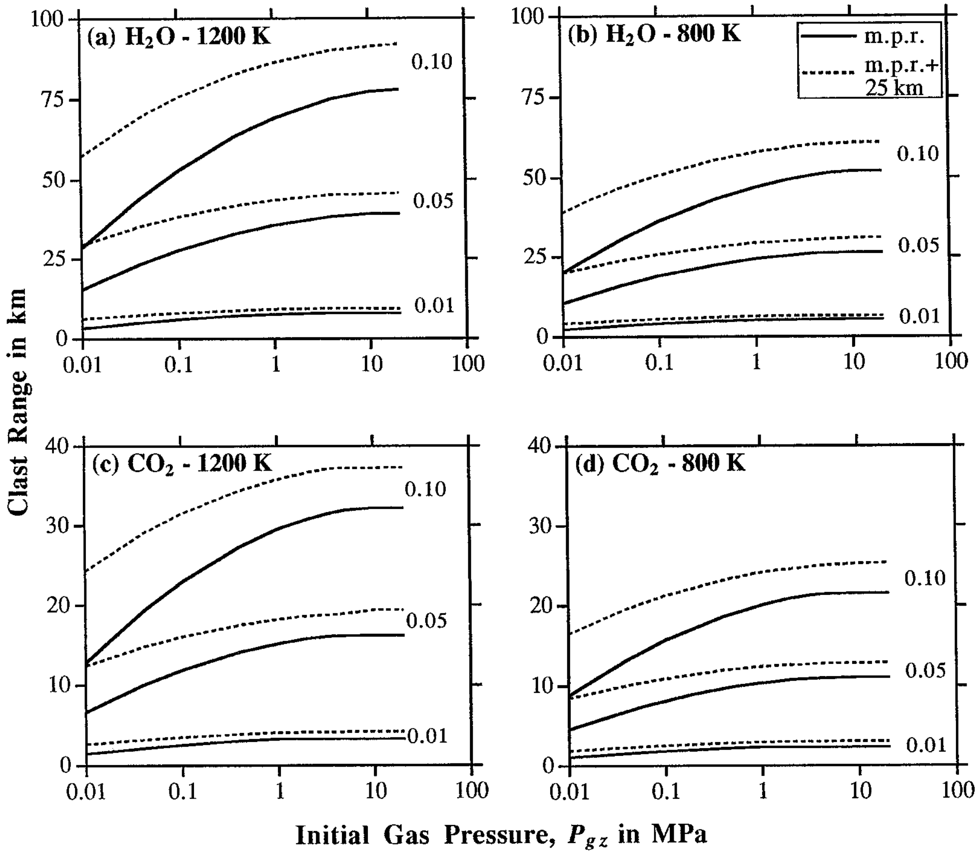

FIG. 3. Distance traveled by ejected clast as a function of initial gas excess pressure, P_{gz}, and gas/caprock mass ratio, n. Details as for Fig. 2.

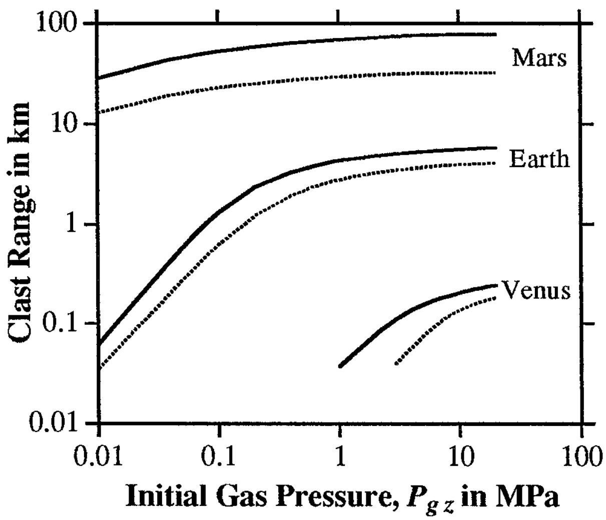

FIG. 4. Comparison of clast travel distance as a function of initial excess pressure, P_{gz}, for Earth, Venus, and Mars. The value of gas/caprock mass ratio is 0.1, initial gas temperature is 1200 K. Solid lines represent H_2O as the driving gas; dotted lines are for CO_2.

the figures for that reason), whereas on Venus it can significantly influence the distances achieved (Fagents and Wilson 1995b).

DISCUSSION

The low martian atmospheric pressure and acceleration due to gravity together would significantly influence the explosion process and the trajectories of blocks ejected in vulcanian style eruptions, such that the eruption velocities and ranges attained by the blocks are significantly greater than for Earth or Venus. The low atmospheric pressure allows greater expansion of gas out of the vent, so that higher velocities are attained with respect to those reached on Earth. The low gravity contributes to longer clast flights, whereas the thin atmosphere and hence reduced aerodynamic drag on the clasts ensures that clasts travel a significant proportion of their equivalent range in a vacuum. Deposits of coarse, blocky ejecta are therefore expected to be widely dispersed over radii ranging up to a maximum of ~100 km. In view of the limited volumes of material emitted in discrete explosions (generally less than 10^9 m^3

(Newhall and Self 1982)) thinly spread and laterally discontinuous deposits over areas up to $\sim 10^4$ km^2 are implied.

Comparison of clast ranges presented here with those of previous workers (Mouginis-Mark *et al.* 1992, Wilson and Head 1994) reveals an order of magnitude discrepancy. This discrepancy appears not to arise from inappropriate treatment of the atmosphere (as was the cause of most problems with early work on terrestrial explosions) since the martian aerodynamic influence on the clasts is sufficiently small that clast trajectories may be reasonably well approximated by the classical ballistics equations. Rather, the problem lies with the assertion that the ejection velocities attained on Mars would be approximately the same as for Earth. This is clearly not the case with the velocities calculated with our explosion model (Fig. 2): the numerator in the equation of motion (Fig. 1) expresses the way in which the lower atmospheric pressure of Mars would cause greater accelerations, even though the range of values for P_{gz} (based on rock strengths) will be essentially the same for both planets. The pressure–velocity expression used in the earlier work (from Self *et al.* 1979) was demonstrated to be inappropriate by Fagents and Wilson (1995b) since an assumption of the Self *et al.* (1979) model was that the ejected clasts were encountering atmospheric gas at rest. Our treatment now correctly accounts for the initial displacement of the atmosphere at the onset of the explosion.

Existing Viking Orbiter coverage of Mars is of adequately high resolution for features exceeding a few tens of meters in size to be discerned. However, given that the wide dispersal of material ejected from transient explosions on Mars would have created landforms of very low relief, even if repeated emissions were to occur, it is questionable whether volcanic centers exhibiting vulcanian behavior are detectable in Viking data. It may therefore be necessary to look for further indications of a transient explosive center.

It may be possible to identify crater fields associated with the impact of ejected blocks, since the considerable kinetic energy possessed by a martian projectile should be sufficient to create a substantial crater. For example, a clast 2 m in radius, ejected from an explosion involving an initial gas mass fraction of 0.10 and a vent pressure of 10 MPa, would land with a kinetic energy of $\sim 10^7$ J (calculated conservatively assuming that the vent site and clast landing point are both at the mpr). Following the treatment of Fudali and Melson (1972) this energy could excavate $\sim 10^6$ kg of the target material which corresponds to a crater diameter of 15 to 25 m, depending on the assumed values of the crater depth : diameter ratio and target material density. This is just at the limits of detection in the highest resolution Viking imagery, although the presence of an ejecta blanket and/or post-formation crater modification (e.g., mass-wasting) would increase the diameter of the feature. In addition, a concentrated zone of sub-pixel sized craters may display a significant textural anomaly (Wilson *et al.* 1984). There would, however, be an inevitable confusion of such craters with meteorite impact craters unless a significant and nonrandom pattern in the areal distribution of craters could be detected. Crater diameters should characteristically increase in size away from the vent as is observed for vulcanian eruptions on Earth (Gorshkov and Dubik 1970, Steinberg 1977, Self *et al.* 1980), since the ejected fragments are initially accelerated to the same velocity and, by virtue of their greater inertia, the largest blocks will travel the greatest distances.

In addition, identification of a central cone or volcanic crater may indicate that the origin of such a feature is vulcanian. Candidate areas for vulcanian explosions may include proposed composite volcanoes of greater relief (e.g., Elysium Mons (Malin 1977, Pike 1978) and Tyrrhena (Greeley and Crown 1990, Crown and Greeley 1993) and Apollinaris Paterae (Robinson *et al.* 1993)), which may have superimposed indications of vulcanian deposits.

The possible presence of pyroclastic flow deposits may increase the chances of detection of transient explosion deposits. Though such deposits commonly (but not always) accompany vulcanian eruptions on Earth (e.g., at Ngauruhoe in 1975 (Nairn and Self 1978); at Galeras (Calvache and Williams 1992)), they also accompany a range of other styles of volcanism, so careful interpretation would be required. On Earth, pyroclastic flows associated with discrete explosions tend to have shorter run-out distances than those associated with plinian activity. These flow lengths may be the result of the limited volume of material emitted in vulcanian events, or possibly be due to the flows being less fluidized as a result of the smaller proportion of juvenile material (Nairn and Self 1978).

On Mars, pyroclastic flows are more likely to have formed during sustained explosive activity than on Earth as a result of the low atmospheric density (Wilson *et al.* 1982). Simple energy conservation arguments imply flows up to 3 times longer on Mars since the distance traveled is proportional to the square of the vent velocity, which would be greater on Mars (Wilson *et al.* 1982). Crown and Greeley (1993) suggested that the lower martian gravity would further increase martian pyroclastic flow lengths with respect to the Earth, by reducing particle–particle frictional interactions. They calculated flow lengths in excess of 1000 km under favorable conditions (cf. <1 to >100 km on Earth (Cas and Wright 1988)). Several candidate examples of pyroclastic flow deposits have been identified (e.g., Greeley and Crown 1990, Zimbelman and Edgett 1992, Crown and Greeley 1993). Given that martian vulcanian eruption velocities are predicted to range up to several hundred m sec^{-1}, pyroclastic flows of considerable length may be expected.

Finally, another indication of a vulcanian feature might be the presence of an ashfall deposit resulting from the convective cloud of gas and fine material associated with such events. Again, these are associated with a wide variety of volcanic styles, but taken in context with other indications may lead to the positive identification of deposits associated with transient eruptive activity.

The rise height of a convecting eruption cloud produced from a transient source is related to the fourth root of the released heat energy, Q (Morton *et al.* 1956), which is in turn related to the total mass of ejecta, M (Settle 1978, Wilson *et al.* 1978),

$$H = k(c\,\Delta T\,F\,M)^{1/4}, \tag{3}$$

in which k is a constant relating to the ratio of the adiabatic to environmental temperature lapse rates, ΔT is the decrease in temperature undergone by the cloud particles between the vent and their final height, and F is a factor describing the efficiency of heat transfer between particle and gas phases.

Accounting for the differing atmospheric characteristics, cloud heights on Mars are found to be around 5 times greater than those calculated for similar eruptive conditions on Earth (Wilson *et al.* 1982). In a vulcanian eruption it is reasonable to assume that the efficiency factor would take a value of 0.25 (Fagents and Wilson 1995b), since typically only 50% of the ejecta is juvenile and hence able to supply heat to the eruption cloud, and of this proportion a further 50% is likely to comprise coarse material which does not remain in the column to act as a heat source. Thus a moderate-sized explosion ejecting 10^8 to 10^9 kg would produce a cloud rising to between 13 and 26 km. If it is assumed that the downwind width of the resulting ashfall deposit is roughly equal to the cloud height (Wilson 1978), a property which is essentially independent of planetary atmospheric structure (Head and Wilson 1986), deposits ranging up to 26 km in width should have been relatively easily produced by vulcanian events on Mars. The long axis of the deposit may greatly exceed the width, depending on the prevailing wind conditions (Mouginis-Mark *et al.* 1988). Greater rise heights and deposit dimensions are expected if repeated emissions take place over intervals less than the time required for the cloud from each pulse to dissipate, or if a greater value for F is assumed (corresponding to a greater proportion of, or a greater degree of fragmentation of, the juvenile component).

However, individual explosions ejecting even larger masses would, when deposited over such areas, produce an average deposit thickness of only a few millimeters at most. Coupled with the ease of erosion of ash-sized material in the martian environment, this may imply that such deposits would not be discernible in image data. Repeated explosions over relatively short time scales would be required to produce deposits capable of avoiding erosion.

An area of Mars to the west of the Tharsis volcanoes dubbed the "Stealth" region has recently been identified in groundbased synthetic aperture radar data (Muhleman *et al.* 1991, Butler 1994). The essentially zero radar reflectance that characterizes this area is interpreted to imply an extensive covering of fine, low density material several meters thick. The geographic relation to the Tharsis region is consistent with an accumulation of volcanic ash, indicating an origin in explosive eruptive activity from the Tharsis volcanoes, with the ash subsequently moved westward (either during the eruption or post-deposition) by prevailing local and regional winds (Muhleman *et al.* 1991, Zimbelman and Edgett 1994, Butler 1995). It is possible that transient outbursts on these otherwise effusive volcanoes may have made some contribution to the material in the Stealth region.

In summary, in order to be able to identify unequivocally the site of a vulcanian eruption on Mars, it may be necessary to employ more than one distinguishing criterion. These criteria could include the presence of a coarse, blocky deposit, a densely packed crater field with crater size increasing outward, pyroclastic flows, or ashfall deposits. A final possibility has been identified by Carruthers (1995), who interpreted features associated with fretted channels in southern Ismenius Lacus to be volcanic in origin. In particular, a closed depression is interpreted to be an explosion crater modified by removal of crustal ice as a result of heating due to igneous activity, with possible pyroclastic flow or surge deposits visible on the flanks. The hypothesis that igneous activity is a primary formative cause for fretted terrain (including fretted channels, Carruthers 1995), together with our postulated increased likelihood of transient explosions in areas of high crustal volatile content, implies that vulcanian eruptions and their resulting surface manifestations may be relatively common, although as yet little recognized, features in regions with significant crustal volatile stores. If so, then with the use of suitable planetary image data, the identification of sites of transient eruptions, and of their position in the martian stratigraphic record, may be used as a probe of global crustal volatile distribution, which would yield important information on Mars' history and evolution.

Although Viking imagery is suitable for looking at some aspects of transient explosions (e.g., pyroclastic flows, modified vent regions), there is insufficient coverage at adequately high resolution for other features to be distinguished (e.g., crater fields). The targeting of specific areas in order to identify possible vulcanian volcanic sites will be one objective for future missions to Mars that could provide evidence of magma–ground ice/water interactions. A global search for indications of transient explosive eruptions may help to establish the chances of detection and the frequency of occurrence of these events. The planned Mars Global Surveyor (MGS) Mars Orbital Camera (MOC) operates in three modes (Malin *et al.* 1992): the global and selective coverage at low (7.5 km/pixel) and moderate (280 m/pixel) resolutions would be useful for looking for larger scale explosion features; and selective very high resolution images (1.4 m/pixel) could resolve features a few meters in horizontal extent and hence detect individual craters or ejected blocks. Together with laser altimeter (MOLA) data on topography and surface reflectivity (Zuber *et al*. 1992), and thermal emission spectrometer (TES) information on surface ices (Christensen *et al*. 1992), significant advances could be made in the understanding of the relationship between igneous activity and modification of the martian surface.

CONCLUSIONS

Transient volcanic explosions on Mars may achieve clast ejection velocities ranging up to ~580 m sec^{-1} (H_2O as the driving gas at initial temperature 1200 K) or ~380 m sec^{-1} (CO_2 at 1200 K): velocities for CO_2-driven explosions are lower than those for H_2O by a factor of ~1.5. The effect of advocating a lower initial gas temperature (800 K) is to decrease eruption velocities by a factor of ~1.2.

Deposits of coarse, blocky ejecta are expected to lie within radii of a few km to ~100 km on Mars for the range of initial conditions explored (P_{gz} = 0.01 to 20 MPa, n = 0.01 to 0.1, vent elevation mpr to +25 km, initial gas temperature 800 to 1200 K, H_2O and CO_2 as driving gases). The deposit radii are typically one to two orders of magnitude greater than are predicted for Earth, and more than three orders of magnitude greater than for Venus. Such widely scattered deposits may not be readily recognized in Viking image data.

Ashfall deposits resulting from convection clouds associated with vulcanian eruptions may typically have widths ranging up to 25 km and much greater downwind lengths. It is interesting to note that the block ranges typically exceed fall deposit widths, in contrast to the venusian case (Fagents and Wilson 1995b). Since fine material is easily mobilized in martian atmospheric conditions, ash deposits may not remain *in situ* for long. Repeated emissions would be required for a significant thickness to accumulate.

Use of other identifying criteria, such as the presence of pyroclastic flows, dense ejecta impact crater fields, and modified explosion craters, may aid detection of sites of vulcanian eruptions. Regions of probable high crustal volatile content (such as the fretted terrain, or areas where rampart craters are common), are suitable candidate sites for vulcanian explosions, and may be considered important targets for imaging with MGS. The recognition of vulcanian explosion sites elsewhere on Mars would serve as a tool for determining global volatile distribution.

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