Evidence for a massive phreatomagmatic eruption in the initial stages of formation of the Mangala Valles outflow channel, Mars

Lionel Wilson
Planetary Science Research Group, Lancaster University, Lancaster, UK

James W. Head III
Department of Geological Sciences, Brown University, Providence, Rhode Island, USA

Received 21 April 2004; revised 15 June 2004; accepted 18 June 2004; published 4 August 2004.

[1] New data reveal the presence of a system of parallel dune-like ridges extending in a band for up to 25–30 km around the eastern margin of the source trough for the Mangala Valles outflow channel system. We interpret these ridges to be formed during initial phreatomagmatic activity caused by dike emplacement, cryospheric cracking, magma-groundwater mixing, and explosive eruption to the surface. The erupted material consisted of fragmented magma, steam, and country rock which expanded from a choked state at the surface vent to form a near-ballistic, Io-like eruption plume. Outward flow of the plume at velocities of ~430 m/s from a ~30–60 km-long section of the graben is interpreted to have formed the system of dune-like ridges. Subsequent outpouring of groundwater formed the Mangala Valles outflow channel system. *INDEX TERMS:* 5400 Planetology: Solid Surface Planets; 5407 Planetology: Solid Surface Planets: Atmospheres—evolution; 5416 Planetology: Solid Surface Planets: Glaciation; 6225 Planetology: Solar System Objects: Mars. **Citation:** Wilson, L., and J. W. Head III (2004), Evidence for a massive phreatomagmatic eruption in the initial stages of formation of the Mangala Valles outflow channel, Mars, *Geophys. Res. Lett.*, *31*, L15701, doi:10.1029/2004GL020322.

1. Introduction

[2] The initial stages of some major outflow channel events on Mars appear to involve dike emplacement and breaching of a thick cryosphere to provide access to the surface for significant volumes of groundwater held under hydrostatic pressure. Here we analyze new high-resolution data that reveal circumferential ridges partly surrounding a linear graben that is the source for a major outflow channel system, Mangala Valles. We conclude that the most plausible explanation for the ridges is formation during a massive phreatomagmatic eruption as the initial emplacement of a dike cracked the cryosphere, allowing mixing of magma and groundwater, and explosive venting to the surface.

2. Description

[3] Memnonia Fossae is part of an extensive graben system radiating away from the Tharsis rise [*Scott and Tanaka*, 1986] and recently interpreted as the surface manifestation of radial dike systems [e.g., *Wilson and Head*, 2002]. This branch of Memnonia Fossae cuts across a broad, north-south oriented circum-Tharsis ridge which in turn is superposed on an 80 km diameter impact crater of Noachian age (Figure 1). *Tanaka and Chapman* [1990] proposed that outflow of water released by the dike-induced cracking of the cryosphere caused erosion through the notch in the northern part of the trough to form the Mangala outflow channel.

[4] Recent spacecraft have returned high-resolution data revealing the nature of the graben, trough and surrounding rim. THEMIS images have revealed the presence of an unusual pattern of ridges (Figure 2) that are developed adjacent and subparallel to the edge of the graben-trough system in its eastern part (Figure 3). The ridges occur along the southern rim of the trough over at least 60 km of its eastern extent and along at least 30 km on its northern rim. MOLA altimetry footprints are too close to the scale of the ridges to give reliable height measurements, but illumination geometry relationships suggest that the ridges are less than a few to tens of meters in height and are typically 500–1000 m in length. Measurements along five profiles normal to the strike of the ridges show that their spacing (Figure 2) ranges from ~235–400 m, with ridges proximal to the trough margin tending to be more closely spaced (typically 235–300 m) than distal ridges (typically 300–400 m). The ridges generally appear symmetrical at this scale and are widespread, but not evenly distributed. Ridges tend to be most well-developed on local highs and least well-developed in local lows where they appear less numerous or missing. Some ridges are disrupted by subsequent events such as impact craters. The underlying topography in the region of ridge development is quite varied, with a major scarp at the eastern end of the trough, and terraces and scarps on the southern rim. For example, in the region of ridges mapped in Figure 2, the scarp in the upper left is a few hundred meters high, while the scarp cutting NW-SE across the center of the image marks the edge of a terrace in which the area to the NE lies several hundred meters above the area to the SW (Figure 1). Despite these major changes in local topography and slope, most ridges do not appear significantly disrupted by or conform in detail to the local topography. Instead, Figure 2 shows that the ridges are generally parallel to subparallel to each other throughout this area almost regardless of local topography and that they broadly curve around the trough margins (Figure 3).

[5] THEMIS data show that the pattern of ridges described in detail in Figure 2 continues around the margins of the trough in a broad band up to 20–30 km wide (Figure 3). The detailed continuity of ridges cannot be

0094-8276/04/2004GL020322$05.00

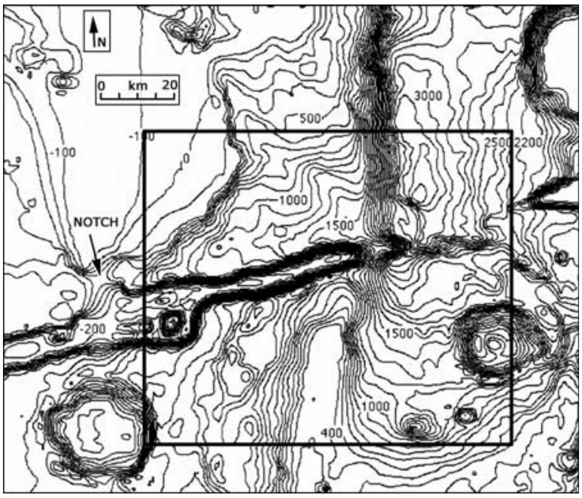

Figure 1. Topographic map of the Mangala Valles-Memnonia Fossae region where the system of parallel ridges is developed; bounded by −17, −19 latitude, 210.2, 212.6 longitude. Square shows area of Figure 3.

conclusively established in the areas of missing data, or in areas that have been modified by subsequent events such as the regions indicated as lobes in Figure 3 [*Head et al.*, 2004]. Even so, sufficient data exist to reveal the presence of a circumferential pattern of ridges around the trough (Figure 3) covering an area estimated at about 5000 km^2.

3. Discussion and Interpretation

[6] We test the hypothesis that these ridges could form from interaction of the two major components of such systems, magma and water [e.g., *Head et al.*, 2003]. Volcanic eruptions involving water can result in large phreatomagmatic plumes that collapse and create flow outward from the vent, resulting in erosion and deposition of material in the surrounding regions, often producing dune-like forms [e.g., *Dellino et al.*, 2004]. Could these ridges have resulted from an early stage of explosive eruption associated with the dike-induced graben?

[7] During dike emplacement, heat is conducted out of the magma into the dike walls. In the aquifer system it boils water that is then incorporated into the magma as vapor bubbles. This occurs because growth of the bubbles into the magma at the dike wall requires less work than forcing the water in the aquifer away from the interaction site [*Bonafede and Mazzanti*, 1997]. In the overlying cryosphere it has to melt ice to water before the water vaporizes and is incorporated. The most conservative estimate of the rate at which heat penetrates laterally is found by assuming that no convection occurs; heat transfer is then limited by the thermal diffusivity of the dike wall material. For the rock component alone this is $\sim 7 \times 10^{-7}$ m^2 s^{-1}. The value for cryosphere ice (mean temperature −35°C, density 920 kg m^{-3}, specific heat 1860 J kg^{-1} K^{-1}, thermal conductivity 2.65 W m^{-1} K^{-1}) is 15×10^{-7} m^2 s^{-1}, and for water just above the freezing point (density 1000 kg m^{-3}, specific heat 4200 J kg^{-1} K^{-1}, thermal conductivity 6.3 W m^{-1} K^{-1}) is also $\sim 15 \times 10^{-7}$ m^2 s^{-1}. Weighting in proportion to component masses, and assuming that ice and water fill pore space that varies from ∼15% near the surface to ∼5% near the base of the cryosphere [*Hanna and Phillips*, 2003], the bulk diffusivity κ_c varies from $\sim 8.2 \times 10^{-7}$ m^2 s^{-1} near the surface to $\sim 7.4 \times 10^{-7}$ m^2 s^{-1} at depth. Thus over any time interval τ, the depth λ into the country rock to which heat penetrates is given by $\lambda = (\kappa_c\ \tau)^{1/2}$.

[8] The relevant value for τ is the time taken by a given batch of magma to rise through the zone occupied by water and ice. Let the cryosphere thickness be C and let the aquifer system extend for a vertical distance A below the

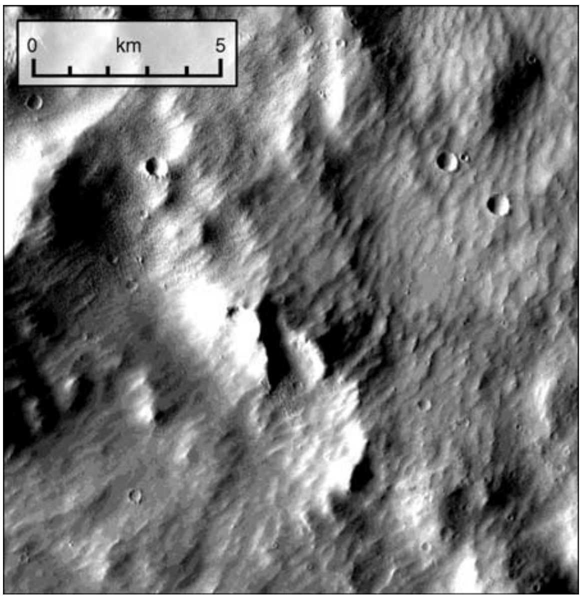

Figure 2. Characteristics and relationships of the ridges. Portion of THEMIS image V05848003. Location shown by box in Figure 3.

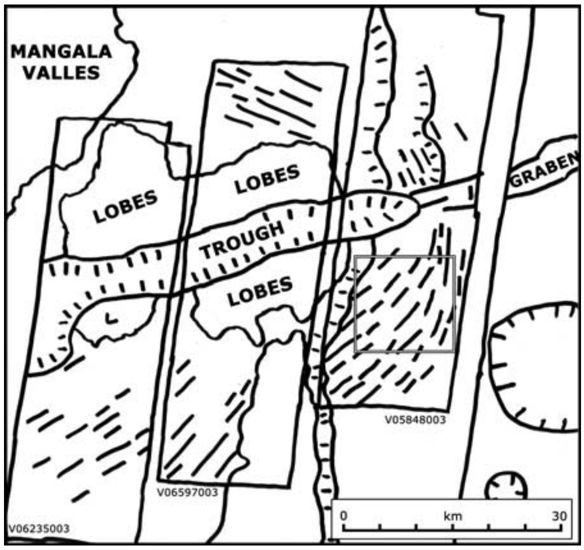

Figure 3. Sketch map of the distribution of ridges around the eastern end of the trough eroded from the Memnonia Fossae graben. General trends of the ridge orientations as seen in the THEMIS images (marked by parallelograms) are shown. Square is location of Figure 2 which shows details of ridges.

base of the cryosphere. With a value for the geothermal gradient on Mars of $\sim$20 K km^{-1} and an annual average mean surface temperature near the equator of 210 K [*Clifford*, 1993], *C* is expected to be about (273 K–210 K)/(20 K km^{-1}) = $\sim$3 km. Estimates of the variation of porosity with depth [*Hanna and Phillips*, 2003] imply that *A* may be close to 5 km. The vertical rise speed of the magma as the dike is emplaced can be obtained from an estimate of the horizontal propagation speed of a giant dike of the dimensions relevant here [*Wilson and Head*, 2002] under the typical stress fields driving them [*Fialko and Rubin*, 1999; *Wilson and Head*, 2002], which yields a value in the range 10–30 m s^{-1}. As the tip advances laterally, the upward growth speed is dictated by the shape of the dike tip. Given the aspect ratio of a dike reaching the proximal part of the Memnonia Fossa, about 1800 km from the likely source under Arsia Mons, the upward speed will be $\sim$1% of the horizontal speed, at this distance $\sim$20 m s^{-1}, leading to a vertical magma rise speed of $\sim$0.2 m s^{-1}. The time taken to rise a total of ($C + A$ =) 9 km would then be τ = $\sim$45000 s, and the thermal penetration depth would be λ = $\sim$187 mm at 9 km depth, zero at the surface (assuming no delay between the dike tip reaching shallow depth and the start of explosive venting), and proportional to depth$^{1/2}$ in between.

[9] To estimate the volume of magma emplaced into the upper 9 km of the dike we assume that the dike tip has a triangular section. The width at the upper tip would of course be zero and, given the half-height, the width at 9 km depth would be $\sim$30 m. The volume of magma per meter along strike would then be (0.5 $\times$ 30 $\times$ 9000 =) 1.35 $\times$ 10^5 m^3 m^{-1}. If the magma contained 0.65 wt% CO_2 and 0.3 wt% H_2O as it left the mantle, values typical of the magma feeding the Hawaiian volcanoes on Earth [*Gerlach*, 1986] which, as a primary mantle melt, we assume to be a plausible candidate mantle-derived magma for Mars, the magma bulk density can be calculated by assuming that the pressure at the moving dike tip is buffered at the saturation pressure of the water, $\sim$4.3 MPa, and that below this CO_2 exsolves to be in equilibrium with the ambient magma pressure, which is close to the ambient lithostatic pressure. This yields bulk densities that increase from $\sim$1200 kg m^{-3} near the dike tip to 2400 kg m^{-3} at the base of the aquifer system at 9 km depth. The masses of liquid magma and CO_2 in any given cross-sectional area of the dike can be evaluated as a function of depth using the solubility as a function of pressure given by *Harris* [1981] and *Dixon* [1997]. The total magma mass is 3.1 $\times$ 10^8 kg per meter along strike of the dike and the mass of exsolved CO_2 is 1.9 $\times$ 10^6 kg m^{-1}. Using water and ice densities of 1000 and 917 kg m^{-3}, respectively, the mass of ice and water contained in the pore space of the 9 km high, on average $\sim$95 mm thick layer along the two dike walls would be $\sim$6.6 $\times$ 10^5 kg m^{-1} and the mass of wall rock in the layer would be $\sim$3.9 $\times$ 10^6 kg m^{-1}.

[10] We now assume that the magma in the dike mixed with the wall layer and that the evaporation of the ice and water produced pressurized water vapor bubbles in the magma that led to an explosive eruption which eventually decompressed the mixture to Martian atmospheric pressure, P_a = $\sim$500 Pa. This event would have involved the propagation of an expansion wave downward into the intrusion, at least as far as the maximum depth of vapor addition, i.e., the base of the aquifer system. The typical speed of an expansion wave is about half of the speed of sound in the medium it is decompressing, and the speed of sound in the vesicular dike with an exsolved CO_2 mass fraction of $\sim$0.6% is found from formulae given by *Kieffer* [1981] to be $\sim$100 m s^{-1}. Thus the duration of the explosion, the time taken for the expansion wave to travel 9 km at $\sim$50 m s^{-1}, would have been $\sim$3 minutes. Assuming that essentially all of the magmatic H_2O also exsolved during the magmatic decompression, its mass was 0.94 $\times$ 10^6 kg m^{-1} along strike. Adding the crustal H_2O to the magmatic H_2O and CO_2 and dividing by the total of the volatiles, the magmatic liquid and the crustal rock, the total volatile mass fraction of the exploding mixture was about n = 0.011 with the average volatile molecular weight (the mass-weighted mean of the two volatile components) being m = 33.7. The thermal energy required to melt the ice and evaporate all of the crustal water would have decreased the magma temperature from its initial $\sim$1450 K to T = $\sim$904 K at the dike tip and $\sim$912 K at 9 km depth. Good thermal mixing between all of the ejected components is likely because the mean grainsize of phreato-magmatic explosion products on Earth is 0.1–0.3 mm [*Wohletz*, 1983] and should be the same on Mars since the magma-water interaction process is not dependent on gravity. The specific energy E released due to gas expansion from an initial magma pressure P_i to Martian atmospheric pressure P_a would then have been

$$E = [(n\ Q\ T)/m] \ln(P_i/P_a) + [(1-n)(P_i - P_a)]/\rho_s - g\ D \quad (1)$$

where ρ_s is the average density of the magmatic liquid and wall rock and D is the depth from which a given batch of ejecta rises. E varied from 2.2 $\times$ 10^4 J kg^{-1} of the total mixture in the upper part of the dike to 2.0 $\times$ 10^4 J kg^{-1} in the lower part, sufficient to produce a maximum ejection speed U given by E = 0.5 U^2 that varied from U = 209 m s^{-1} as the upper part of the dike was emptied to 201 m s^{-1} as the lower part erupted. These speeds would have led to maximum ejecta ranges of R = (U^2/g) = $\sim$12 to $\sim$11 km using a ballistic approximation, justified by calculations [*Fagents and Wilson*, 1996] implying that in short-lived explosive eruptions on Mars an initial shock wave pushes the ambient atmosphere ahead of the ejecta which are then dispersed near-ballistically in a manner more similar to eruptions on Io than on Earth.

[11] These ranges are only about 40% of the maximum radial extent of the observed ridges, suggesting that more of the wall rock and its volatiles were incorporated into the magma than assumed so far, and that a fuel-coolant interaction (FCI) occurred. Such events involve rapid dynamic mixing of equal volumes of magma and wall material, as suggested by *Wilson and Mouginis-Mark* [2003] for other explosive eruptions on Mars at Hrad Vallis. The equivalent of the above calculations can be carried out with a volume of wall rock, with its attendant ice or water, as appropriate, equal to the magma volume being substituted for the thin conductive layer. In this case a much larger volume of wall material was mixed with the magma and it is much more accurate to model the subsequent explosion as the expansion of a pseudo-gas, as described by *Kieffer* [1981] and developed for the present geometry in equations (14)–(19) of

Wilson and Mouginis-Mark [2003]. Allowance is made in the calculation for the fact that gas expansion would have been choked in the vent, and using the method of *Wilson and Mouginis-Mark* [2003], typical vent conditions are found to be a pressure of 5.6–5.8 MPa and an exit speed of $\sim$120–125 m s^{-1}. Above the vent, gas and pyroclasts would have completed their expansion to the local atmospheric pressure, reaching speeds that varied from U = 338 m s^{-1} as the upper part of the dike was emptied to 247 m s^{-1} as the lower part erupted. Comparison of clast sizes with the mean free path of gas molecules in the expanding cloud behind the interaction shock with the atmosphere shows that all clasts larger than $\sim$30 microns avoided entering the Knudsen molecular-flow regime and thus experienced the full force of the expanding gas; even clasts as small as 1 micron reached at least 80% of the above gas speeds. Trans-sonic to supersonic fire-fountains on Earth eject pyroclasts mainly at angles up to $\sim$20° from the vertical, and computer simulations of explosive eruptions on Mars show that choking of the gas expansion at the surface vent leads to similar conditions [*Mitchell*, 2001]. Thus final speeds of 247–338 m s^{-1} would imply ballistic ranges between 10 and 20 km, 50–60% of the maximum observed extent of the ridges.

[12] This is consistent with our proposed model for the final stage of emplacement of the ejected solids. The mass flux leaving the vent during the $\sim$3 minute duration of the explosion would have increased steadily (given the assumed triangular dike cross section) up to a maximum of $\sim 6 \times 10^6$ kg s^{-1} per meter along strike of the fissure vent. This represents the product of the expansion wave speed ($\sim$50 m s^{-1}), the width of the zone of excavation (double the dike width in the case of the FCI interaction, i.e., up to 60 m), and the average density of the material being excavated (dike density $\sim$2000 kg m^{-3}, host rock bulk density $\sim$2350 kg m^{-3}). The mean volatile content of the erupting mixture was given above as 0.01 mass fraction. For this mass fraction of volatiles, the models of *Wilson and Walker* [1987] and *Wilson and Head* [1994] show that fissure eruptions on Mars will form collapsing eruption columns, rather than convection plumes, for mass fluxes greater than $\sim 1 \times 10^6$ kg s^{-1} m^{-1}. Thus we expect the bulk of the eruption products to consist of materials deposited from a short-lived (minutes) pyroclastic surge produced by the collapse process. Conservation of energy dictates that the initial horizontal speed of the surge as material accumulated at distances out to 10 km from the vent would have been a large fraction of the eruption speed in the vent, at least $\sim$200 m s^{-1}. This speed would have allowed the surge to flow over topographic obstacles up to 5 km high on Mars, and so the $\sim$300–500 m topography in the distal parts of the ridge system would have been surmounted even after the flow speed had decayed to $\sim$60 m s^{-1}. Models of pyroclastic flow run-out on Mars [*Crown and Greeley*, 1993; *Wilson and Head*, 1994] show that travel distances of tens of km would easily be achieved by surges with these initial kinetic energy levels even on very shallow slopes. Unfortunately, despite recent advances in modelling processes inside volcanic surge clouds [e.g., *Dellino et al.*, 2004], it is not possible to extract quantitative information from the width and spacing of the ridges formed by the deposits.

[13] In summary, we interpret these ridges to be formed during initial phreatomagmatic activity caused by dike emplacement, cryospheric cracking, magma-groundwater mixing, and explosive eruption to the surface. The resulting eruption of fragmented magma and steam formed a near-ballistic eruption plume akin to those seen on Io, albeit on a smaller scale. Pyroclast accumulation under the control of the complex gas expansion pattern is interpreted to have created the system of dune-like ridges.

References

Clifford, S. M. (1993), A model for the hydrological and climatic behavior of water on Mars, *J. Geophys. Res.*, *98*, 10,973–11,016.

Dellino, P., R. Isaia, L. La Volpe, and G. Orsi (2004), Interaction between particles transported by fallout and surge in the deposits of the Agnano-Monte Spina eruption (Campi Flegrei, southern Italy), *J. Volcanol. Geotherm. Res.*, *133*, 193–210.

Dixon, J. E. (1997), Degassing of alkali basalts, *Am. Mineral.*, *82*, 368–378.

Fagents, S. A., and L. Wilson (1996), Numerical modelling of ejecta dispersal from transient volcanic explosions on Mars, *Icarus*, *123*, 284–295.

Fialko, Y. A., and A. M. Rubin (1999), Thermal and mechanical aspects of magma emplacement in giant dike swarms, *J. Geophys. Res.*, *104*, 23,033–23,049.

Gerlach, T. M. (1986), Exsolution of H_2O, CO_2, and S during eruptive episodes at Kilauea Volcano, Hawaii, *J. Geophys Res.*, *91*, 12,177–12,185.

Hanna, J. C., and R. J. Phillips (2003), A new model for the hydrologic properties of the Martian crust and implications for the formation of valley networks and outflow channels, *Proc. Lunar Planet. Sci. Conf.* [CD-ROM], *XXXIV*, Abstract 2027.

Harris, D. M. (1981), The concentration of CO_2 in submarine tholeiitic basalts, *J. Geol.*, *89*, 689–701.

Head, J. W., L. Wilson, and K. L. Mitchell (2003), Generation of recent massive water floods at Cerberus Fossae, Mars by dike emplacement, cryospheric cracking, and confined aquifer groundwater release, *Geophys. Res. Lett.*, *30*(11), 1577, doi:10.1029/2003GL017135.

Head, J. W., D. R. Marchant, and G. J. Ghatan (2004), Glacial deposits on the rim of an Hesperian-Amazonian outflow channel source trough: Mangala Valles, Mars, *Geophys. Res. Lett.*, *31*, L10701, doi:10.1029/2004GL020294.

Kieffer, S. W. (1981), Blast dynamics at Mount St. Helens on 18 May 1980, *Nature*, *291*, 568–570.

Mitchell, K. L. (2001), Explosive volcanic eruptions on Mars: An improved numerical model, *Proc. Lunar Planet. Sci. Conf.* [CD-ROM], *XXXII*, Abstract 1189.

Scott, D. H., and K. Tanaka (1986), Geologic map of the western region of Mars, *U.S. Geol. Surv. Misc. Invest. Map*, *I-1802-A*, 1:15,000,000.

Tanaka, K. L., and M. G. Chapman (1990), The relation of catastrophic flooding of Mangala Valles to faulting of Memnonia Fossae and Tharsis volcanism, *J. Geophys. Res.*, *95*, 14,315–14,323.

Wilson, L., and J. W. Head (1994), Mars: Review and analysis of volcanic eruption theory and relationships to observed landforms, *Rev. Geophys.*, *32*, 221–263.

Wilson, L., and J. W. Head (2002), Tharsis-radial graben systems as the surface manifestation of plume-related dike intrusion complexes: Model and implications, *J. Geophys. Res.*, *107*(E8), 5057, doi:10.1029/2001JE001593.

Wilson, L., and P. J. Mouginis-Mark (2003), Phreatomagmatic explosive activity at Hrad Vallis, Mars, *J. Geophys. Res.*, *108*(E8), 5082, doi:10.1029/2002JE001927.

Wilson, L., and G. P. L. Walker (1987), Explosive volcanic eruptions: VI. Ejecta dispersal in plinian eruptions: The control of eruption conditions and atmospheric properties, *Geophys. J. R. Astron. Soc.*, *89*, 657–679.

Wohletz, K. H. (1983), Mechanisms of hydrovolcanic pyroclast formation: Grain-size, scanning electron microscopy, and experimental studies, in *Explosive Volcanism*, edited by M. F. Sheridan and F. Barberi, pp. 31–63, Elsevier Sci., New York.

J. W. Head III, Department of Geological Sciences, Brown University, Box 1846, Providence, RI 02912, USA. (james_head_III@brown.edu)

L. Wilson, Planetary Science Research Group, Lancaster University, Lancaster, UK.