Origin of small pits in martian impact craters

Joseph M. Boyce a,*, Lionel Wilson a, b, Peter J. Mouginis-Mark a, Christopher W. Hamilton c, Livio L. Tornabene d

a Hawaii Institute for Geophysics and Planetology, University of Hawaii, Honolulu, HI 96822, USA
b Lancaster Environment Centre, Lancaster University, Lancaster LA1 4YQ, UK
c NASA Goddard Space Flight Center, Greenbelt, MD 20771, USA
d Center for Planetary Science and Exploration, University of Western Ontario, 1151 Richmond Street, London, ON, Canada N6A 5B7

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A B S T R A C T

We propose a numerical model for the formation of the closely-spaced pits found in the thin, ejecta-related deposits superposed on the floors, interior terrace blocks, and near-rim ejecta blankets of well-preserved martian impact craters. Our model predicts the explosive degassing of water from this pitted material, which is assumed to originally be water-bearing, impact melt-rich breccia at the time of deposition. This process is analogous to what occurred in the fall-out suevite deposits at the Ries impact structure in Germany. At Ries, impact heating of water-bearing target material resulted in the rapid degassing of its water and other volatiles. The martian environment plays an important role in enhancing the effects of this degassing by increasing the flow-speed of the escaping gas. The high flow-rate of gas through particulate materials, such as suevite, tends to quickly form segregation channels or vent pipes, similar to those found in the Ries deposits. These pipes act as conduits for the efficient high-speed escape of the gas and small clasts that it entrains. Escaping gas and entrained clasts abraded and eroded the conduit walls, flaring them to form pits above a network of pipes.

1. Introduction

Impact craters provide considerable information about the composition and structure of target materials, physical properties of the impactors, and impact cratering processes. For example, on Mars, impact structures show evidence of the effects of substantial water in the target material (e.g., Carr et al., 1977; Mouginis-Mark, 1979). Here, we focus on a recently-identified impact ejecta and crater-fill facies first discovered by Mouginis-Mark et al. (2003) in Mars Orbiter Camera (MOC) images. These deposits were later mapped globally by Tornabene et al. (2012). This facies generally occurs as thin, heavily pitted deposits in relatively well-preserved martian impact craters (Fig. 1). The closely-spaced pits found in these deposits are morphologically distinct from the larger, single central pit features studied by Wood et al. (1978), Hale (1982), and Barlow (2010a). In the parent craters that contain both these thin ejecta facies and large single central pits, the thin ejecta facies appears to overlie the large pits. This thin ejecta facies is thought to be composed of overlay-melt rich breccia impactite similar to suevite at the Ries impact structure on Earth, and may have initially contained substantial water (McEwen et al., 2007; Tornabene et al., 2007, 2012; Mouginis-Mark and Garbeil, 2007; Morris et al., 2010). The closely-spaced pits have been attributed to sublimation (Hartmann et al., 2010), or to undermining and collapse (McEwen et al., 2007; Tornabene et al., 2007; Mouginis-Mark and Garbeil, 2007) caused by removal of water that had seeped or defused into these deposits following their deposition. This material is found in impact craters that are widely distributed over the surface of Mars. Consequently, if the pits require a substantial amount of water to form, then their occurrence has important implications for the history of volatiles on Mars (Tornabene et al., 2012).

Here we present a numerical model of pit formation in this ejecta and crater-fill facies (herein called martian pitted material) that involves explosive degassing of the water initially contained in the target material. This model is summarized in Fig. 2, which is intended to serve as a guide to the reader in the presentation of our model. The model is inspired by evidence that hot materials in the Ries fall-out suevite degassed immediately after deposition (Pohl et al., 1977; Newsom et al., 1986), resulting in fluidization of the deposits, elutriation or separation of fine particles and formation of vent pipes that carried gas from the deposit.

2. Background

The martian pitted material was first recognized in MOC images by Mouginis-Mark et al. (2003) at a resolution of 3–6 m/pixel. As
High Resolution Science Imaging Experiment (HiRISE) images (25 cm/pixel) became available, workers such as McEwen et al. (2007), Tornabene et al. (2007, 2012), Mouginis-Mark and Garbeil (2007), and Morris et al. (2010) studied this material, and concluded that its morphology and stratigraphic position indicated that it is a facies of impact ejecta, probably impact melt-rich breccia (similar to suevite at Ries crater, Germany). They proposed that the pits resulted from collapse of voids left in the pitted material by escape of water from pockets, or of ice from lenses, at a time well after deposition of this material. In contrast, Hartmann et al. (2010) suggested that the pits are produced by sublimation of ice distributed more evenly throughout these deposits. However, in this paper we show that all of these models are inconsistent with many characteristics of the pits and pitted material.

McEwen et al. (2007) and Tornabene et al. (2012) listed three reasons why the pitted material is impact melt-rich breccia, similar to the suevite at the Ries impact structure. These reasons include (1) the apparent fluid flow of the material unit during emplacement of the host deposit, (2) the presence of blocks embedded in pits walls, and (3) the pitted material’s uppermost stratigraphic position relative to other impactite deposits. Tornabene et al. (2012) suggested that the likelihood that the pitted material is impact melt-rich breccia implies the presence of water in the target materials. They based this suggestion on experimental, theoretical

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**Fig. 1.** Examples of pitted material at Tooting crater, a 27 km diameter fresh martian impact crater. (a) Location image for other sub-scenes in this figure. CTX image P01_001538_2035. (b) Pits on the floor, segment of HiRISE image PSP_002158_2035. (c) Pits on the exterior rim, segment of HiRISE image PSP_002580_2035. (d) Pits on a terrace block, segment of HiRISE PSP_003569_2035.
and terrestrial field studies (Kieffer and Simonds, 1980; Sheridan and Wohletz, 1983; Stewart et al., 2003; Pope et al., 2004; Stewart and Ahrens, 2005; Osinski, 2006; Osinski et al., 2008) implying that craters formed in volatile-rich targets, like the Earth or Mars, should produce an abundance of this type of material.

Furthermore, while the pitted material is best exposed in the freshest martian single- and multiple-layer impact craters (it is rare in double-layer ejecta craters), it is also identified in some older, but still well-preserved, craters of the same type. This, combined with their observation that craters containing pitted material occur everywhere outside the polar regions, led Tornabene et al. (2012) to suggest that abundant H₂O (probably in the form of ice) was present in the subsurface of Mars, at least within the excavation depth of the craters, for much of the planet's history. For this reason, it is important to determine the origin and timing of these pits, and specifically, whether they formed by processes that require substantial water in the target materials at the time of impact.

2.1. Morphology of the pits

The martian pits (Fig. 1) are bowl-shaped, depressions that range in size from a few meters to more than \( \frac{24}{3} \) km in diameter, but most pits are \( \approx 200 \text{ m} \) in diameter (see Tornabene et al. (2012) for detailed geomorphic descriptions of the pits). These pits occur on the floors, terraces, and exterior ejecta deposits of fresh craters (Tornabene et al., 2012). The pits commonly occur in closely-spaced groups whose individual pits are so close together that their rims share straight segments situated between individual pits. The tendency for the pits in these closely-spaced groups to

![Fig. 2. Conceptual model for the formation of pitted materials within impact structures on Mars.](image-url)
be nearly circular and to share straight rim segments produces a geometry that resembles a 2-D cross-section though soap-froth, or a honeycomb (Fig. 3). There are isolated pits that do not exhibit shared rims (Fig. 4), but in general pits occur within closely-packed groups with rims that do not appear to be raised above the surrounding terrain. In contrast, isolated pits commonly have relatively low, raised rims (Fig. 3).

The martian pits commonly exhibit evidence of mantling (flat floors and subdued topography), even in the freshest of craters. However, in some large pits, even those that appear to be mantled, meter-scale, light-toned blocks are embedded in their walls (Fig. 5). McEwen et al. (2007) and Tornabene et al. (2007, 2012) have proposed that these blocks are large clasts within an impact-melt-bearing breccia deposit.

Tornabene et al. (2012) found that the size of the largest pits in a given crater generally scales with the diameter of the host crater. They also found that the largest pits in a particular crater are typically located near the center of the pitted materials within that crater. This relationship of pit diameter to crater size and location may reflect a relationship between pit size and deposit thickness.

2.2. The pitted material

In this section, we discuss the aspects of the pitted material most relevant to our model for pit formation, including the general morphology of the pitted material, and its physical properties based on knowledge gained from Mars observations, theoretical studies, and terrestrial analog studies (n.b., for more extensive discussion of the morphology and possible origins see Tornabene et al., 2012). We also estimate the volatile content of the pitted material based on models of the martian crust and indirect observational data. This information provides a basis for constructing and testing our model against observed pit characteristics.

The pitted material mainly occurs as flat, thin deposits superposed on other impact crater material on the floors of fresh Martian craters (Fig. 1). These deposits can be several hundred meters thick in the largest craters containing these materials (e.g., Hale crater). Pitted material also is found within topographic low areas on interior terrace blocks and on ejecta blankets near crater rims. Considering the general relief of hummocks in some of the ejecta blankets, the deposits found there may only be several meters thick. This material also commonly occurs on interior slopes of craters as thin sheets. These sheets appear to have drained from high places in the crater (e.g., on terrace blocks) to lower elevations (e.g., crater floors) (Fig. 6). They often contain...
curvilinear ridges and grooves that appear to be flow lines, suggesting that this material was initially relatively fluid, but solidified before it could completely drain into the crater floor. In contrast to the pitted deposits on the floor and in topographic lows, pits are typically rare on these drained sheets.

In some fresh craters, solidification and stabilization of the surface of the pitted material appears to have occurred in some places before others. For example, fully-formed pits located around the edges of the floors of some relatively large craters appear to be partially buried by pitted material that flowed from interior walls (Fig. 7). This overlapping relationship suggests that the surface of the pitted material on the floor of these craters had stabilized enough to allow pit formation, while the pitted material elsewhere was still a fluid. The length of time between formation of the floor pits and their partial flooding is not clear, but considering cooling rates and deposit thicknesses it could be quite short, possibly days (Mouginis-Mark and Boyce, 2012). These observations suggest that some process rapidly indurated the pitted material (e.g., cementation, sintering, or solidification of the melt phase), which is consistent with the observation that the pits appear to have resisted the subsequent flow of lobate materials (Morris et al., 2010).

In this study, we use the pitted material in the northern portion of the floor of fresh martian impact crater Tooting (~28.8 km diameter) to test our model because it contains well-developed pitted deposits and we can reasonably estimate the average thickness of the deposit at this location. We also use the suevite of Ries impact structure as a terrestrial analog to the pitted material in Tooting crater as a way to estimate physical properties of the pitted material (see e.g., Pohl et al., 1977; Stöffler, 1977; Hörz et al., 1983; Newsom et al., 1986; Osinski, 2003, 2008). We contend that this is reasonable because the two craters are approximately the same size (the gravity-scaled diameter of Ries is ~32 km; see Stewart and Valiante, 2006), relatively uneroded, and produced in similar target rock (i.e., water-rich silicate rock). In addition, similarities in ejecta physical properties are suggested by the calculations of Pope et al. (2006, Tables 1 and 2) that indicate that, even considering the effects of differences in targets composition (i.e., basalt versus granite) (Pierazzo et al., 1997, 2005), as well as impact velocities (Bjorkman and Holsapple, 1987; O'Keefe and Ahrens, 1994), and gravity (Pope et al., 2000) these two craters should have produced a similar volumes of impact melt (i.e., ~11 km$^3$). Furthermore, because the shock fragmentation process follows a power function that is influenced by maximum clast or block size which is related to crater size (Moore, 1971, 1972; Bart and Melosh, 2007), stress history, and the degree of pre-impact fracturing, it should fragment the targets at both craters in a similar way (see Melosh, 1989; Fujiwara et al., 1989; Melosh et al., 1992). Hence, the clasts in their impact melt-rich breccia should have similar size distributions. However, the clast-size distribution for Ries suevite shows some variations, Bringemeier (1994) reported a median particle diameter of ~4.5 mm (i.e., 22 mm, whereas Engelhardt (1972) found a median clast size of ~2.75 mm (i.e., 6.75 mm). This difference may be the result of the natural inhomogeneity of impact-melt-rich breccia deposits. We consider the effects of these differences in grain size later.

Heat is a critical element of our model because it is the driving force behind water loss. A considerable amount of heat is generated by conversion of the kinetic energy of the impact to produce hot ejecta in the form of impact melt and impact melt-rich breccia (Ahrens and O'Keefe, 1972; Melosh, 1989; Pope et al., 2006). These heated materials progressively lose heat from the time they are excavated through transport and deposition. In this study, only the heat remaining in the pitted material after its deposition is considered and relevant to our model (Fig. 2d). The temperature of shock melted silicates should be ~2000 °C (e.g., see Engelhardt, 1968). However, on planets where this melt is mixed with cooler clasts to produce impact melt-rich breccia rapid cooling occurs, quickly reducing the average temperature (Grieve and Cintala, 1992; Cintala and Grieve, 1994). For example, Engelhardt (1990) and Engelhardt et al. (1995) estimate that the temperature of the Ries suevite was an average of ~750 °C (i.e., 1023 K) at the time of its deposition. We use this value as the initial temperature of the pitted material to test our model.

The water content of the martian target material is also a critical factor because it dictates the water content of the pitted material at the time of deposition. Although the water content of martian target material cannot be measured directly, it can be estimated from indirect observations (Barlow, 2010b) and theoretical calculations (see Clifford, 1993; Carr, 1996). For example, based on layered ejecta morphology and terrain softening Barlow (2010b) found that ~20–25% ice is likely to reside within the regolith in non-polar regions. This is consistent with calculations based on the approach employed by Clifford (1993) that used the lunar impact fracturing model of Binder and Lange (1980) to estimate pore space in the martian surface target material (i.e., the megaregolith). This calculation provides an estimate of the volume of possible storage space for water/ice in the martian target
materials. Using this approach, we calculate that the maximum volume percent of pore water in the target rocks of Tooting crater ranged from ~10% to ~25% (assuming the upper 150–200 m of the target rock at Tooting crater was dry due to pre-impact diffusion). If this volume were filled with water, then it would imply a water content of ~4 wt.% to ~10 wt.% (assuming an average density of 2400 kg m$^{-3}$ for the target rock). In addition to this source, water bound in minerals may add another ~4–7 wt.% (e.g., Ming et al., 2008; Boynton et al., 2008) making a total of ~12 ± 5 wt.%.

Target material may not be the only source of water that could invade the pitted material immediately after deposition. Sources such as water-rich basement rock, groundwater emanating from the intensely fractured wall rock (Morris et al., 2010), and/or local rainfall from water to vapor released by the impact, may all be possible, but we consider them unlikely. For example, the floors of large craters, such as Hale crater (~136 km diameter), excavate below the theoretical water/ice-rich layer into dry rock (e.g., Clifford, 1993; Carr, 1996) eliminating them as a source of water. Also, impact craters are commonly so hot, especially on their interiors, that liquid water is not stable on their surfaces immediately after deposition. This would prevent the presence of such water for some time after crater formation, eliminating its infiltration as a source.

The steam production rate from the clasts and impact melt in impact melt-rich breccia is also a critical element behind water loss. Such loss should be initially quite high because the gas diffusion rate is temperature dependent, and that these deposits contain a relative high abundance of small clasts and hot impact melt particles for which the diffusion pathways are short (Sparks et al., 1999). For this study, like with average temperature and median grain-size, we use the steam production rate inferred for the Ries suevite to test our model (Newsom et al., 1986).

Because our model is developed using the pitted material in the northern floor of Tooting crater, a knowledge of the thickness of this material in this area is required. We estimate that the minimum thickness is at least that of the depth of largest pits in this area, which is ~30 m (i.e., based on digital elevation model data shown in Fig. S.10 of Mouginis-Mark and Boyce (2012)). A maximum thickness is suggested by the observations of Mouginis-Mark and Boyce (2012) who noted a ~190 m elevation difference between the higher, heavily-pitted northern crater floor and the sparsely-pitted southern crater floor. They speculated that this difference may be due to the presence of pitted deposits only on the northern floor where it lay above a relatively flat, impact-melt similar to that in complex lunar craters (Howard and Wilshire, 1975; Bray et al., 2010; Denevi et al., 2012a,b). However, Mouginis-Mark and Boyce (2012) also point out that some of this elevation difference may be due to buried slide deposits originating from collapse of the northern crater wall as suggested by extensive terrace blocks development and comparatively lower rim elevation (~290 m) of the northern side. However, the presence of such buried slide material beneath the pitted material cannot be verified with available data. Consequently, for the purposes of testing our model, we assume 100 m as a reasonable value for the maximum thickness of pitted material in the northern floor of Tooting crater.

Although conduction is the dominant mechanism for heat loss from the solid clasts in impact melt-rich breccia, water is lost mainly by diffusion through and between grains, and fluid flow through vent pipes, like those found at Stac Fada, Scotland (Amor et al., 2008; Branney and Brown, 2011) and in the Ries fall-out suevite (Newsom et al., 1986; Engelhardt et al., 1995; Osinski, 2004). As with volcanic vents and vent pipes in pyroclastic deposits (e.g., Engelhardt, 1972; Pohl et al., 1977; Newsom et al., 1986), such conduits in impact melt-rich breccia are the most efficient means of releasing large volumes of water vapor from a particulate deposit. For example, at Ries, the fall-out suevite is dissected by vent pipes that are typically 2–4 cm wide. In places, the pipes are so closely spaced that they occupy up to 10% of the volume of the deposit (Newsom et al., 1986). We note that such pipes have yet to be found in many of the impact melt-rich deposits elsewhere on Earth (Osinski et al., 2008). This may be due to mineralization in the Ries and Stac Fada vent pipes that makes them easier to observe. Also, it is not clear that a search for vent pipe features within other impact structures has yet been specifically conducted. We will discuss the implications for vent pipe formation later.

3. Degassing of water vapor from the pitted material

The approach used here for developing a numerical model for the origin of the pits through the loss of heat and water from the martian pitted material borrows from previous work on degassing behavior of pyroclastic materials during their transport, deposition, and cooling. This approach was also used to model degassing of the Ries Crater fall-out suevite by Newsom et al. (1986), and has provided valuable insight. We suggest that, to a first order, the degassing behavior of all of these materials (i.e., martian pitted materials, known terrestrial suevite, and pyroclastic materials) should be similar because they are all composed of silicate melts and water-bearing clasts of a wide range of sizes. The rate and duration of water loss from these hot clastic deposits is mainly a function of their initial water content, and their physical properties such as temperature and grain-size distribution. Together these factors control the production rate of gas within the deposit and its migration through and from the deposit. Below, we outline a numerical-based approach to how these factors control water loss in impact melt-rich breccia on Mars and the probable outcome of this water loss (Fig. 2f–h).

Furthermore, we test this model to determine if it produces reasonable results consistent with observations and morphometric measurements, using the pitted material in the northern portion of the floor of Tooting crater because of its similarities to Ries (discussed in detail above). It should also be noted that even though these two craters have many similarities, martian environmental factors (in particular the comparatively thinner atmosphere and smaller acceleration due to gravity) may have dramatic effects on the escape rate of water vapor from the impact melt-rich breccia of Tooting crater compared with that at Ries crater. Consequently, it is important to pay special attention to these environmental factors.

3.1. Flow rate and volume loss

The rate at which water vapor or steam flows through and from freshly deposited impact melt-rich breccia is a major factor controlling how those deposits behave during degassing. The rate at which steam escapes is dependent on a number of factors such as average temperature of the deposit, the rate at which the steam is generated, the physical properties of the clasts contained in the deposit, and the type of flows within and from the deposit. Additionally, the amount of steam that escapes and how long it takes to escape is dependent on the inventory of water in the materials of these deposits. Appendix A presents the detailed numerical basis for the discussion that follows.

Like fluids in other clastic deposits, the ease with which the steam moves diffusively through particulate materials is strongly dependent on the dimensions of pore space and permeability of the material, which depends on the shape and size of the grains. As water vapor is exsolved and diffuses from the impact melts and hot clasts, it is released into the pore space between the grains in the deposit (Fig. 2e and f) where it flows through the connected pore spaces (here approximated as narrow tubes whose size is determined by the mode of the clast-size distribution). Consequently, the steam flow-rate is dependent on the average size
and number per unit area of these idealized small tubes (estimated by considering slices through a cube exactly containing a void space approximated by a sphere, shown in Fig. 8). This geometry allows estimation of the average fractional area of a slice through the void space.

Considering this model for the pitted material in Tooting crater, the gas pressure required to support the static weight of this material must be at least equal to the 0.74 MPa equivalent weight of the 100 m deep bed. Using this value for the gas pressure and the corresponding void space $v = 0.125$ found above, the steam viscosity $\eta_s = 6.0 \times 10^{-8} \text{ Pa s}$ at $T_s = 1023 \text{ K}$ (from Engelhardt et al. (1995)), an average martian atmospheric pressure of $P_0 = 700 \text{ Pa}$ (n.b. the escaping steam should be rapidly carried away by thermal convection, diffusion, and cross winds, and as a result have little local effect on $P_s$), assuming a median particle diameter of $-4.5 \Omega$ (i.e., 22.6 mm, so that the median radius of the particles is $r = 11.3 \text{ mm}$) in the deposit from Bringemeier (1994), suggests that the appropriate solution involves turbulent gas flow, with the minimum value of $F_1$ required being $\sim 0.101 \text{ kg m}^{-3} \text{s}^{-1}$, so that $F_m$ is $\sim 4.8 \times 10^{-5} \text{ kg kg}^{-1} \text{s}^{-1}$. The resulting variations of pressure and gas flow speed with distance upward from the base of the 100 m-deep suevite layer in the floor of Tooting crater are shown in Fig. 9a and b. The steep pressure gradient at shallow depth results in an enormous gas flow speed close to the surface of nearly 300 m s$^{-1}$.

Clearly this state could not be maintained for an appreciable time because at these near-surface gas flow speeds all fine material would rapidly be elutriated from the shallow part of the deposit, destroying the assumption of a uniform grain size throughout the deposit and requiring a more elaborate model. We have calculated the size of the largest spherical clast that can resist being elutriated with mid-level gas flow speeds from Fig. 9b of $\sim 50 \text{ m s}^{-1}$ and pressures from Fig. 9a of $\sim 0.7 \text{ MPa}$, giving a steam density at $T_s = 1023 \text{ K}$ of $\sim 1.5 \text{ kg m}^{-3}$; the implied value of $r_{\text{max}}$ is of order 0.16 m. This is so much larger than the $\sim 11 \text{ mm}$ median clast radius in the deposit that if these conditions existed the deposit would immediately be explosively disrupted. However, there is no evidence that this happened, suggesting that some other factor also was at play to prevent explosive disruption.

This apparent conundrum can be resolved in several ways. One of these is that small clasts can be vigorously elutriated from much of the deposit without any requirement that the whole deposit be fluidized (Fig. 2f). This underlines an inevitable limitation of experiments to simulate the behavior of particulate deposits like suevite or ignimbrites using fluidized beds to fluidize the whole deposit layer by passing gas in through the base of the equipment (e.g., Wilson, 1980). We suggest that, in nature, impact-generated impact-melt-rich deposits (and also pyroclastic density current deposits) produce gas from clasts within the flow, not just at its base. So the gas flux, and hence the gas flow speed, must increase upward in the way found here, and not require that the entire deposit be fluidized. For example, if all particles smaller than 100 $\mu\text{m}$ radius are elutriate from a deposit, then (see Appendix B, Eq. (1)) the typical gas flow speed must be $\sim 1.3 \text{ m s}^{-1}$. However, for median size clasts of $r = 11.3 \text{ mm}$ (from Bringemeier (1994)), the solution that leads to the required typical speed involves laminar gas flow with a basal gas pressure of $\sim 2000 \text{ Pa}$. This is only about three times the atmospheric pressure, and results in a gas release rate from clasts of $F_s = 2.5 \times 10^{-5} \text{ kg m}^{-3} \text{s}^{-1}$ (i.e. $F_m = \sim 1.2 \times 10^{-8} \text{ kg kg}^{-1} \text{s}^{-1}$).
If the pitted deposit in the floor of Tooting crater contains 10\% by mass of water (i.e. 0.1 kg kg$^{-1}$) this gas release rate would need to exceed 8.3 $\times$ 10$^5$ kg m$^{-3}$ s$^{-1}$ (i.e., $F_0 = 3.0 \times 10^8$ kg kg s$^{-1}$), and a 10\% by mass water content would imply $\sim3.3 \times 10^5$ s ($\sim38$ days) to release all of the water vapor. But, the actual duration of such release may be much shorter. Engelhardt et al. (1995) and Oskin (2003) found that the impact glass at Ries contains about half of the water content of the target material (6 wt.% water), suggesting that only half of the initial water may have been lost as a result of the impact. If such loss also occurred in the pitted material of Tooting crater, then degassing may have occurred over a factor of 2 less time than calculated above.

In addition, the important effect of median clast diameter on these calculations can be illustrated by changing the median clast diameter from $\sim$4.5 $\phi$ to $\sim$2.75 $\phi$ (r = 3.365 mm), i.e., the average median clast size determined for Ries suevite by Engelhardt (1972). The gas motion is again laminar, but the required basal gas pressure is now close to 10,000 Pa. Hence, the required gas release rate is $F_0 = 6.2 \times 10^5$ kg m$^{-3}$ s$^{-1}$ (i.e., $F_0 = 3.0 \times 10^8$ kg kg s$^{-1}$), and a 10\% by mass water content would imply $\sim39$ days to release all of the water vapor with a 4\% by mass water content releasing its water vapor over $\sim16$ days.

Using the values assumed in this analysis, it is clearly possible to remove a significant fraction of the small particles from the pitted material in the floor of Tooting crater on a timescale that is in the order of a few days to a few tens of days (Fig. 2h). Fig. 9b shows that the gas flow-speed near to the surface of the deposit can readily exceed 300 m s$^{-1}$. This is considerably higher than the $\sim1.3$ m s$^{-1}$ terminal velocity of particles of $\sim100$ μm in size deeper in the deposit, and suggests that particles of this size will be lofted above the surface with essentially all of the gas speed. This gas flow speed assumes that the gas stays in close contact with the bulk of the suevite layer and thus stays at the same average temperature as the clasts in the layer. We have not taken into account the energy release due to the decompression of the gas, but this would increase its flow speed even more than the existing large increase due to the pressure gradient. Just how much the speed would increase would depend on the mass loading of the fine particles being carried with the gas.

3.2. Effects of vent pipe formation

To a large degree, the conundrum of disruption of the deposit by high gas-flow speeds can be solved by another mechanism, i.e., the formation of numerous preferred pathways, or vent pipes that provide an efficient means for escape of the gas (Fig. 2f and g). Vent pipes are common in some terrestrial suevite and pyroclastic deposits and have been shown to provide an efficient means for gas escape from those deposits (see Newsom et al., 1986). We suggest that vent pipes are likely to have formed in the martian pitted material for the same reason they formed in these other particulate deposits, and that they not only played a major role in preventing the explosive disruption of the pitted materials, but also in the formation of the pits.

Vent pipes are produced as a step in a systematic sequence of behavior in particulate deposits that results from increasingly higher gas flow-rates through that material (e.g., Kuni and Levenspiel, 1969; Rhodes, 1998; Wilson, 1980, 1984). At low flow-rates, gas merely percolates through void spaces in the material to escape. At increasing flow-rates, fluidization occurs (i.e., when the drag force from the ascending gas equals the buoyant weight of the bed) followed by bubble formation. In such cases, bubbles tend to coalesce into non-bubbling channels (here interchangeably called segregation, degassing, or vent pipes) where small, easily-transported particles are carried to the surface by escaping gases to form a vigorously bubbling layer. Newsom et al. (1986) suggest that there is evidence of such elutriation of fine material at the top of the Ries fall-out suevite as a result of this process. At high gas flow-rates where the drag forces are large enough to propel particles, such phenomena as spouting, fountaining and explosive-like eruptions may occur at the top of the pipes, producing pit-like features.

It also should be noted that experimental studies have shown that in deposits of particulate materials that have high content of fine-grain particles, high temperatures, and/or high shear stresses, conditions favor homogeneous fluidization, instead of vent pipe formation (Gravina et al., 2004; Druitt et al., 2004, 2007). However, vent pipes may readily form in the pitted deposits because (1) elutriation by the high-speed escaping gas should rapidly decrease the fines content, (2) shear stresses are likely to be low in these relatively flat and even static deposits during pit formation, and (3) although high temperature can eliminate moisture-derived interparticle cohesion needed for pipe formation, the water vapor-rich environment expected in the pitted materials during its degassing may counter the drying effect of high temperature.

As at Ries, vent pipes in the pitted material probably developed at nearly the same time that fine particle elutriation in the deposit began (Kuni and Levenspiel, 1969; Wilson, 1980, 1984; Newsom et al., 1986; Druitt et al., 2004, 2007). However, because these pipes provided a much more efficient means for gas to escape compared with uniform diffusion, they quickly dominated as the mechanism for gas escape from the deposit (Kuni and Levenspiel, 1969; Wilson, 1980; 1984; Newsom et al., 1986). Furthermore, we suggest that this mechanism should be nearly the same as the mechanism that controls behavior of escaping gas and entrained clasts through volcanic conduits during explosive eruptions because both involve high-speed escape of such materials driven by gas expansion through pipe-like conduits that vent at the surface. For this reason, we have adopted the numerical approach of Wilson and Head (2007) for calculation of the escape speed of products generated by explosive pyroclastic eruptions on Mars and Earth, and applied it to the calculation of the escape speed of gas and entrained clasts from the vent pipes that underlie the martian pits. The escape speed of these materials is critical to erosion of the deposit surrounding the vent pipes, and is especially important to erosion that results in flaring of the vent pipes near the surface.

In theory, expansion of gas through the vent pipes in the pitted material can be accommodated by increasing the speed of the gas/clast mixture, and flaring of the conduits toward the surface (Fig. 2g). This is similar to what happens in terrestrial explosive volcanic conduits where expansion of gas is accommodated in the same way (Wilson and Head, 1981, 2007; Glaze and Baloga, 2002; Mitchell, 2005). Wilson and Head (2007, see Table 1) calculated eruption speeds in vents for martian and Earth pyroclastic material that originate from basaltic magma with water contents comparable to those expected for the martian pitted material. Based on their calculations, the eruption speeds range from $\sim360$ to $\sim410$ m s$^{-1}$. We suggest that these speeds are likely to be similar to eruption speeds from the vent pipes in pitted material containing similar water contents because both situations involve the expansion of gas through a conduit that is accommodated by increasing the speed of the gas and the clasts it entrains.

Considering the gas eruption velocities in the vents calculated by Wilson and Head (2007), and the size of the largest spherical clast in a deposit (Eq. (14)), these flow speeds would easily be enough to move spherical clasts on the order of $\sim1$–4 cm in diameter from the vents. It is clear that such high-speed escape of gas and clasts, no matter whether they originate from volcanic or impact
melt-rich breccia sources, not only can eject particles great distances, but can substantially erode the walls of the conduits that carry them, as well as the vents at the surface.

3.3. Pit formation

Insight into the morphologic effects of the high-speed escape of steam from the vent pipes in the pitted materials is offered by the modeling of Mitchell (2005). Mitchell (2005) considered the coupling between high-speed gas flow in an explosive volcanic conduit and the shape of the vent. He found that abrasive erosion by escaping high-speed gas and clasts results in conduit flaring near the surface to produce a pit. His model was based on the work of Macedonio et al. (1994) who derived the abrashon-rate of wall rock as a function of the mechanical properties of the wall material, and the vertical flow-speed of solid particles within the conduit. Mitchell (2005) calculated the probable effects of abrasive erosion on conduit shape, and how these changes in shape feedback to affect the flow parameters. He started with parallel-sided circular geometry, setting the choking point at the surface, allowing a unique first solution for a given mass flux, and proceeded in a step-wise manner, recalculating the conduit shape at every iteration. Fig. 10 shows the generalized results based on his analysis of the evolution of conduit radius and wall stress as a function of depth for a terrestrial basaltic eruption (also see Fig. 5d in Mitchell (2005)).

In comparison with Mitchell’s (2005) finding, the higher water content of the martian impact-melt-rich breccia deposits, combined with the lower acceleration due to gravity and atmospheric pressure of Mars, should increase the velocities of the escaping steam/clasts mixture from this material, and duration of flow to enhance conduit wall erosion and pit growth. These environmental differences have the effect of increasing the rate of pit growth on Mars compared with Earth given the same escape velocities, but differences have the effect of increasing the rate of pit growth on Mars compared with Earth given the same escape velocities, but do not change the general form of pit growth. However, many other parameters may cause small secondary effects on the quantitative calculations based on Mitchell’s (2005) model. Consequently, Mitchell’s erosion model’s main value to our work is to show that substantial flaring of the pipes near the surface is expected to result from erosion of the vent pipe walls in the martian pitted material by the high-speed escape of the steam and its entrained clasts.

4. Discussion

We have developed a model for the formation of the closely-spaced pits found in thin deposits in and around well-preserved martian impact craters (Fig. 2). The material in which these pits have formed is likely to be composed of impact melt-rich breccia derived from water-rich target materials (Fig. 2a–d). In contrast to previous models of formation of these pits by sublimation or collapse, our model predicts that, as in the Ries suevite, impact-induced heating of the water-bearing components in the pitted material caused them to rapidly release their water as steam (Fig. 2e–h). Our model predicts that degassing of the martian pitted material should be more explosive and have greater morphologic effects on the surface compared with degassing of terrestrial suevite deposits because of the effects of the lower atmospheric pressure and lower acceleration due to gravity of Mars. These differences act to increase the escape velocity of the steam from these deposits. In addition, water-rich, impact melt-rich breccia produced by martian cratering of a broad range in sizes should degas similarly even if the average grain-size (Moore, 1971, 1972; Bart and Melosh, 2007) and proportions of impact melt to clasts (Pope et al., 2006) in these particulate deposits vary somewhat with crater size. This is because the physics that controls the degassing behavior of such particulate deposits as the impact melt-rich breccia on Mars should produce generally similar results (Kuni and Levenspiel, 1969; Wilson, 1980, 1984; Druitt et al., 2004, 2007). This behavior is in contrast to that in impact generated hydrothermal systems where water vapor slowly escapes through fractures produced by impacts that radiate outward into water-bearing rock (Abramov and King, 2005).

Escape of the steam released by these hot, water-rich materials should predominantly be through a network of vent pipes, a natural consequence of the high-flow rate of gas through particulate materials. The high-speed flow of steam and the clasts it entrains through these pipes should result in abrasive erosion of the pipe walls, and flaring of these walls near the surface to form pits. Using the pitted material on the floor of Tooting crater to test our model, we calculate that the gas-flow speeds through the vent pipes could readily exceed 300 m s\(^{-1}\) at the vent, and may be as high as 410 m s\(^{-1}\). At these gas-flow speeds the rate of gas loss from the deposit would have expended the gas supply in only a few days to a few weeks, depending on the water content of the pitted material. Even though the eruption of gas and entrained clasts may have lasted only a short time, the gas-flow speeds are easily high enough to cause substantial abrasion and erosion of the vent pipe walls to produce the observed pits. These speeds are high enough to eject particles as large as a few centimeters diameter to very great distances, leaving little trace of these materials around the pit rims. Fine-grain particles ejected from these vents are likely to be carried upward by the escaping hot steam to form a turbulent, convecting cloud above the deposit. This cloud would interact with the atmosphere at its outer edges, sharing all the properties of the fine ash plumes that form over pyroclastic density currents, such as those seen at Mount St. Helens and Montserrat, and produce so-called co-ignimbrite fall deposits. Depending on local conditions, the fine particles in these plumes could be blown sideways to significant distances by the wind; alternately, under low-wind conditions they could settle into the parent crater to form a mantle of fine materials on what otherwise would be the coarse upper part of the deposit from which the fines were stripped.

The location, spacing, and size of the pits are reflections of conditions in the pitted material during eruption (e.g., porosity, permeability, and deposit thickness). Vent pipes probably form initially in the pitted deposits as a closely-spaced network of nearly vertical tubes like those at Ries (Kuni and Levenspiel, 1969; Rhodes, 1998; Wilson, 1980, 1984; Newsom et al., 1986). Following their formation, and as steam begins to rapidly escape through them, some of these pipes should begin to grow at the expense of others. This is a result of a low-pressure zone that should
develop in the material surrounding the pipes, with the best-developed of these pipes tending to draw the greatest volumes of steam to them. This produces an even lower pressure zone around these pipes that acts to draw even more steam toward them (see Kokelaar, 1983). These pipes are likely to produce the largest and deepest pits because they tend to carry the most steam. This also implies that the largest pits should develop in the thickest pitted deposits because these thick deposits are likely to harbor greater volumes of steam, and have more vertical space for the pits to grow deeper before the escaping steam shuts off. This supposition is consistent with the observation of Tornabene et al. (2012) that the largest pits are in the largest craters, and the largest pits in a particular crater are in places that would be expected to have the thickest deposits (i.e., the floors).

The model presented here also provides an explanation for why isolated pits commonly have low, raised rims, whereas the pits in closely-spaced groups typically lack rims that rise above the surrounding terrain. It is clear that the presence of raised rims is an important morphologic feature suggesting involvement of constructional processes, which is not expected for pits of collapse or sublimation origin. However, even though solitary martian pits have low, raised rims, they typically lack evidence of substantial ejecta deposits. This lack of thick ejecta deposits, together with the presence of low rims, gives these pits a morphologic appearance resembling hydroeruption pits at Mount St. Helens (Fig. 11). These hydroeruption pits were produced by steam-driven explosive eruptions caused when hot pyroclastic flows covered a wet substrate. The rims and ejecta of these hydro-eruption pits are so subdued that they are difficult to identify even in aerial photographs (Moyer and Swanson, 1987). Consequently, it is not surprising that ejecta deposits around martian pits produced by similar explosive degassing processes would be difficult to identify in orbital images.

Furthermore, we suggest the reason that the closely-spaced groups of martian pits lack raised rims between individual pits is a result of their close-packing combined with their simultaneous formation. The pits in these closely spaced groups are commonly packed so close together that their rims butt against each other, resulting in shared straight rim segments between individual pits (Fig. 2). Their close packing probably is a reflection of the spacing of vent pipes beneath. This close spacing has produced a geometry pattern that resembles a 2-D cross-section through soap froth that is the result of the balance of pressure between adjacent bubbles (as see Glazier and Weaire, 1992; Li et al., 2007; Vasconcelos et al., 2003). We suggest that the geometry of the pits in these closely-spaced groups is likely to be a result of the simultaneous degassing of nearly all pits in a group with individual pits growing so large that their rims intersect the rims of adjacent pits. With continued degassing and pit growth, an equilibrium point would be reached where outward growth of an individual pit is counterbalanced by growth of the adjacent pits. After this equilibrium is reached, and as degassing continues, all but the largest size-fraction of particles in the ejecta are likely to be removed from around the pits by the high-speed eruption of material from adjacent pits. This process prevents substantial raised rims from forming between pits in the closely-spaced groups and leaves only relatively low rim ridges at the equilibrium distance between the pits. These rim ridges should be composed dominantly of coarse-grain clasts such as those shown in Fig. 5.

The morphological expression of simultaneously formed pits in martian impact breccia is analogous to the formation of mud pots found in some locations on Earth, such as those at Crystal Geyser, Utah (Fig. 12). The geometry of these closely-spaced mud pots results from simultaneous venting of pulses of CO2-rich water from a deep source. These pulses travel through multiple near-surface fractures that act as vent pipes that propagate into the mud deposit above (Shipton et al., 2005). This allows all mud pots to form simultaneously, which influences the formation and geometry of adjacent mud pots. This reinforces our conjecture that the pits of the martian pitted material are likely to have formed nearly simultaneously.

Although rare, the morphology of these pits and closely-spaced groups of pits is not unique to Mars. For example, recently pits and closely-spaced groups of pits that appear to be morphologically indistinguishable to the martian pits have been discovered in and
around impact craters on the asteroid Vesta (Denevi et al., 2012a,b). Denevi et al. (2012a,b) speculate that these pits may also be caused by degassing of volatile-rich impactite in the low-gravity, low-atmospheric pressure environment of the asteroid.

We note that there are other types of closely-spaced groups of pits found on Mars and other planets that only superficially resemble the pits and pit groups of the martian pitted material. These other types of pit groups owe their origin to a variety of different processes including volcanic (Frey and Jarosewich, 1982; Greeley and Fagents, 2001; Keszthelyi et al., 2010; Hamilton et al., 2011; Ghent et al., 2012), sedimentary (Greeley and Iversen, 1987; Burr and Fagents, 2001; Keszthelyi et al., 2010; Hamilton et al., 2011; Engelhardt and Graup, 1984) we suggest the pitted material at Tooting crater may have released a substantial amount of water vapor into the atmosphere in a relatively short time (i.e., hours to a few tens of days). A first-order estimate of the amount of water vapor can be calculated by multiplying the volume of pitted material at Tooting crater by the average water content of the pitted material (~10 wt.%), assuming all water was released. The volume can be estimated by multiplying the area of pitted material (i.e., ~177 km²) mapped by Mouginis-Mark and Boyce (2012) by the average thickness of the pitted material. The average thickness can be approximated by averaging the maximum thickness of ~100 m (estimated in Section 3) with the minimum thickness based on the likely thickness of the numerous small patches of pitted material pooled in topographic lows on the rim and terraces (judged from the typical topographic relief in those areas), and the thin sheets of this material found on the rim. Employing this approach results in an average thickness values of ~55 m, a volume of ~9.7 km³, and a maximum amount of water released of ~9.2 × 10¹¹ kg. Alternatively, the volume can be estimated using theoretical calculations. Stöffler et al. (2002) and Pope et al. (2006) calculated that a volume of ~11 km³ of impact melt would be produced by formation of martian impact craters the size of Tooting crater and Ries crater on Earth. Using the approximate ratio of impact melt to clasts in the Ries suevite of approximately 1–1 (Engelhardt and Graup, 1984) we suggest the pitted material at Tooting crater may have a total volume of ~22 km³. As a result, it initially may have contained ~22 × 10¹¹ kg of the water. However, considering the evidence that only about half of the original water content of the target material is found in Ries suevite samples (Engelhardt et al., 1995; Osinski, 2004) these estimates may be a factor of 2 greater than the actual value. These values may provide useful input parameters to models of the effects of such rapid water vapor release on the local martian atmosphere (Segura et al., 2002; Toon et al., 2010; Kite et al., 2011). This is in contrast to the slow escape of water vapor from impact-generated hydrothermal systems that may develop in water-bearing fractured rock surrounding impact craters (Abramov and King, 2005).

5. Conclusion

We propose that the martian pits formed by rapid degassing of water from the pitted material in a way that is similar to the degassing of the fall-out suevite of the Ries impact structure. We calculate that the escaping high-speed gas and entrained clasts substantially eroded walls of the numerous vent pipes in the pitted material to produce flaring of the vents at the surface, thereby forming pits. Furthermore, the honeycomb-like geometry of the closely-spaced pit groups is likely to have been produced by eruptive degassing across the deposits that caused simultaneous growth of pits until they eventually reached an equilibrium point where their outward growth was counterbalanced by growth of the adjacent pits.

If this model is correct, it provides a new method for investigating the distribution of subsurface volatiles at the time of crater formation. Previous studies of martian impact craters have concentrated on the analysis of the layered ejecta materials (Carr et al., 1977; Mouginis-Mark, 1979; Barlow, 2010b). Future analysis of the morphology and distribution of pitted material as a function of target material and location on the planet appears to be warranted to further understand the role of volatiles in the impact cratering process on Mars.

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Appendix A. Flow rate and volume loss

In our treatment of steam release from the freshly-deposited melt-rich deposits, we let \( F_v \) be the steam generation rate, expressed as the gas mass released from the clasts per unit volume of the deposit per unit time (in units of kg m⁻³ s⁻¹). \( F_v \) can be related to the gas mass release rate per unit mass of silicate material per unit time, \( F_m \), via the void space fraction, \( \nu \), and the typical silicate clast density, \( \rho_c \):

\[
F_m = \frac{F_v}{\rho_c (1-\nu)}.
\]  (1)

Like fluids in other clastic deposits, the ease with which the steam moves diffusively though particulate materials is strongly dependent on the pore space and permeability of the material, which depends on the shape and size of the grains. As a basis for calculating the flow rate of steam through the pitted deposits (Fig. 2f), we consider that the total mass flux of gas passing through

![Fig. 12. Oblique view of mud pots during dry period at Crystal Geyser in Utah. Similar to the martian pits, these pits share rims.](image-url)
a surface of area $A$ at height $z$ above the base of the layer is the sum of all the gas generated at heights less than $z$, $(F(Az))$. This gas flows through the connected pore spaces, approximated as narrow tubes of equivalent mean radius $r$, where $r$ represents the typical radius of gaps between the most common clast-size in the layer, determined by the mode of the clast-size distribution. To calculate the rate of gas flowing through these connected pore spaces, we estimate the number per unit area of the idealized small tubes through which the gas flows by noting that if there are $q$ tubes cutting vertically through the area $A$, the fractional area they represent, $[(q\pi r^2)/A]$, is equal to the average cross-sectional area corresponding to the void space fraction, $\nu$. By considering slices through a cube exactly containing a void space approximated by a sphere (Fig. 8), this average fractional area of void space is found to be exactly equal to $\nu$ as long as the void space fraction does not become so large that any random cube contains parts of more than one sphere; hence $[(q\pi r^2)/A] = \nu$, and the number of tubes per unit area is $q/[A = q/(\pi r^2)]$. This geometry allows estimation of the average fractional area of a slice through the void space idealized as a spherical space of radius $R$, at the center of a cube of silicate material of side length $(2L)$. The fractional area of void space in a slice of thickness $dx$ is zero between $x = 0$ and $x = (L - R)$, and is $[(\pi r^2)/4L^3]$ between $x = (L - R)$ and $x = L$. The average fractional area, $A$, between $x = 0$ and $x = L$ (by symmetry the same as the average fractional area between $x = 0$ and $x = 2L$) is therefore given by

$$A_s = \frac{1}{L} \left[ \int_{x=L-R}^{x=L} \pi r^2 \, dx + \int_{x=L-R}^{x=L} \pi r^2 \, dx \right].$$

(2)

Fig. 8 shows that $R^2 = r^2 + (L - x)^2$, i.e.,

$$r^2 = R^2 - L^2 + 2Lx - x^2,$$

(3)

and so

$$A_s = \frac{1}{L} \left[ \int_{x=L-R}^{x=L} \pi r^2 \, dx = \frac{\pi}{4L^3} \int_{x=L-R}^{x=L} (R^2 - L^2 + 2Lx - x^2) \, dx \right] = \frac{\pi}{6L^3}.$$

But note that the void space fraction, $\nu$, is by definition equal to $[(q/4^3)\pi R^3]/(8L^3)$, which reduces to $(\pi r^2)/(6L^3)$, and so we have the surprisingly simple result that $A_s$ is exactly equal to $\nu$.

Collectively, the tubes carry the total mass flux $(F(Az))$ passing through the area $A$ at height $z$ above the base of the layer and so the mass per unit time passing through any one tube, $F_t$, must be equal to $(F(Az))/q$, i.e.,

$$F_t = \frac{F(Az)}{q}.$$

(5)

The mass flux calculated in Eq. (5) can be related to the pressure gradient, $-dp/dz$ (pressure decreases upward as $z$ increases) driving the gas through the tube. If the gas motion is laminar, the required relationship is the Poiseuille flow law, in terms of mass flux:

$$F_t = \frac{\eta gQ_s}{8r^4} \left( \frac{dp}{dz} \right).$$

(6a)

where $\eta g$ is the viscosity of the gas, $Q$ is the universal gas constant, $8314.7$ J kmol$^{-1}$ K$^{-1}$, $m$ is the molecular mass of H$_2$O, $18.0153$ kg kmol$^{-1}$, and it is assumed that the perfect gas law applies, so that the density of the gas, $p_0$, is equal to $[\eta g(Q)t]/(4r^4)]$. If the gas motion is turbulent, the relationship is

$$F_t = \frac{\pi r^5}{4} \left[ \frac{\eta gQ_s}{8r^4} \right]^{1/2} \left[ \frac{\eta gQ_s}{8r^4} \right]^{1/2},$$

(6b)

where $f$ is the friction factor for gas flow past particles from the wall of the tube. By expressing the equation for $F_t$ given by Eq. (5) with those given by Eqs. (6a) and (6b) we can find $dp/dz$ as a function of $z$, and hence $P$ as a function of $z$ and the flow speed of the steam through any one pathway, $U_g$ as a function of $z$, for each gas flow regime. The decision as to which flow regime is relevant hinges on the gas flow speeds found for the two assumptions, and whichever is the smaller is the relevant one.

In the case of laminar flow, using Eqs. (5) and (6a), we find

$$-\frac{dp}{dz} = \frac{8\pi r^4 \eta g Q_s}{4m \eta g Q_s} \frac{dz}{m \eta g Q_s}.$$

(7)

Performing the integration, with the boundary conditions that the pressure is equal to $P_0$, the Mars atmospheric pressure at the surface, and some prescribed value $P_0$ at the base of the suevite layer at depth $Z$ below the surface, we have:

$$P_0 = \frac{P_0^2}{z} + \frac{8\pi r^4 \eta g Q_s}{4m \eta g Q_s} Z,$$

(8)

which enables the value of $F_t$ required to produce a given value of $P_0$ to be found. The pressure $P$ as a function of $z$ is obtained by integrating Eq. (7) between $P_0$ and $P$:

$$P = \frac{P_0^2}{z} + \frac{8\pi r^4 \eta g Q_s}{4m \eta g Q_s} Z.$$

(9)

Using the Poiseuille flow law: $U_g = (r^2 dp/dz)/(8\pi g)$, $U_g$ can be found at any height $z$ above the base of the layer. Using Eq. (6) to express $dp/dz$ as a function of $P$ and $z$ gives $U_g$ as a function of $z$:

$$U_g = \left[ \frac{F_t Q_s}{P_m} \right] z.$$

(10)

In the case of turbulent flow, using Eq. (5) and (6b), we find

$$-\frac{dp}{dz} = \frac{2\pi r^5 Q_s}{3m \eta g Q_s} Z^3,$$

(11)

Performing the integration with the $P_0$, $P_m$, and $Z$ boundary conditions,

$$P_0 = \frac{P_0^2}{Z} + \frac{2\pi r^5 Q_s}{3m \eta g Q_s} Z^3,$$

(12)

which again enables the value of $F_t$, required to produce a given value of $P_0$ to be found. The general expression for the pressure $P$ as a function of $z$ is now

$$P = \frac{P_0^2}{Z} + \frac{2\pi r^5 Q_s}{3m \eta g Q_s} Z^3,$$

(13)

and differentiating this to find $dp/dz$ as a function of $P$ and $z$, the turbulent gas flow speed, $U_g = [(rQ_s dp/dz)/(4m \eta g)]^{1/2}$, is found (entirely predictably since it only involves the gas density, void space fraction, and gas volume production rate) to be given by the same expression as that found for laminar flow and given by Eq. (10).

Appendix B. Radius of the largest spherical clast that can resist being elutriated

The radius, $r_{max}$ of the largest spherical clast that can resist being elutriated by gas flowing at speed $U_g$ can be found by equating the downward weight of the clast to the upward drag force exerted on the clast by the gas flow. When the gas flow around the clast is turbulent, the drag force is given by $(1/2)\rho_0 C_d \pi r_{max}^2 U_g^2$, where $C_d$ is a drag coefficient, dependent on clast shape but of order unity. Equating this force to the clast weight, $(4/3)\pi r_{max}^3 \rho_s g$, we find

$$r_{max} = \left[ \frac{3\rho_0 C_d U_g^2}{8 \rho_s g} \right]^{1/3}.$$