Transport and mixing dynamics from explosions in debris-filled volcanic conduits: Numerical results and implications for maar-diatreme volcanoes

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Abstract

Most volcanoes experience some degree of phreatomagmatism during their lifetime. However, the current understanding of such processes remains limited relative to their magmatic counterparts. Maar-diatremes are a common volcano type that form primarily from phreatomagmatic explosions and are an ideal candidate to further our knowledge of deposits and processes resulting from explosive magma-water interaction due to their abundance as well as their variable levels of field exposure, which allows for detailed mapping and componentry. Two conceptual models of maar-diatreme volcanoes explain the growth and evolution of the crater (maar) and subsurface vent (diatreme) through repeated explosions caused by the interaction of magma and groundwater. One model predicts progressively deepening explosions as water is used up by phreatomagmatic explosions while the other allows for explosions at any level in the diatreme, provided adequate hydrologic conditions are present. In the former, deep-seated lithics in the diatreme are directly ejected and their presence in tephra rings is often taken as a proxy for the depth at which that particular explosion occurred. In the latter, deep-seated lithics are incrementally transported toward the surface via upward directed debris jets. Here we present a novel application of multiphase numerical modeling to assess the controls on length scales of debris jets and their role in upward transport of intra-diatreme material to determine the validity of the two models. The volume of gas generated during a phreatomagmatic explosion is a first order control on the vertical distance a debris jet travels. Unless extremely large amounts of magma and water are involved, it is unlikely that most explosions deeper than ~250 m breach the surface. Other factors such as pressure and temperature have lesser effects on the length scales assuming

they are within realistic ranges. Redistribution of material within a diatreme is primarily driven by subsidence following debris jet passage. Natural debris jet deposits often contain wall rock material that originated above the explosion site; likely this material was first transported downward by subsidence associated with preceding explosions and debris jets. The length scales of debris jets decrease with increasing depth, all other factors staying equal. The characteristics of the domains formed from just one explosion along with the difficulty of erupting deep-seated material indicate that erupted material at maar-diatreme volcanoes has a complex history before it is erupted and is mostly sourced from the upper-most part of the diatreme.

Key words: maar, diatreme, phreatomagmatic, explosive volcanism, kimberlite

1. Introduction

Maar-diatremes are a type of volcano that primarily result from repeated, subsurface explosions caused by the interaction of magma and groundwater. These volcanoes are characterized by: (1) a maar crater, whose floor is below the pre-eruptive surface; (2) a low angle tephra ring that is rich in country rock derived lithics deposited from pyroclastic density currents, ballistic trajectories, and fallout; and (3) a subsurface vent structure (diatreme), which extends downward from the crater floor to as deep as ~2 km, narrowing with depth, until meeting a feeder dike (White and Ross, 2011). While maar-diatremes are the second most common sub-aerial volcanoes on Earth behind scoria cones, observations of actual eruptions are limited to a few cases, most notably the 1977 Ukinrek Maars in Alaska (Kienle et al. 1980; Self et al. 1980). As a result, models explaining maar-diatreme growth and eruption dynamics

are based on observations of deposits (e.g. Lorenz 1986; White 1991; Ross and White 2006; Valentine et al. 2011; Lefebvre et al. 2013; Jordan et al. 2013; van Otterloo et al. 2013; Delpit et al. 2014), experiments (e.g. Ross et al. 2008a,b; Valentine et al. 2012, 2015; Ross et al. 2013; Taddeucci et al. 2013; Andrews et al. 2014; Graettinger et al. 2014, 2015), and empirical relationships (Valentine et al. 2014). Even though maar-diatremes represent a unique form of volcanism that forms a negative landform, phreatomagmatic eruptions are common at many volcano types (likely including eruptions of diamond bearing kimberlite magma; e.g., Brown and Valentine 2013). As a result, improving our knowledge of the processes that occur at maar-diatreme volcanoes has broader volcanological and economic implications.

Two conceptual models have been proposed to explain the mechanisms by which phreatomagmatic maar-diatreme volcanoes form and evolve over the course of an entire eruptive episode. Lorenz (1986) explains the growth of diatremes by subsidence following successive molten fuel-coolant interactions (MFCI) between groundwater and magma. The usage of water in MFCI results in drawdown of the local water table, and as magma flows into the system the explosion locus progressively deepens and the diatreme widens by subsiding wall rock. Deep-seated lithics are directly ejected as this drawdown occurs and their presence in tephra rings is often taken as an indicator of the depth at which a particular explosion occurred. Valentine and White (2012) propose a model in which explosions can take place at any level within a diatreme provided adequate hydrologic conditions are present. Initial explosions probably occur near the surface and form an initial crater and “protodiatreme.” In this model, the newly formed diatreme grows more rapidly near the surface due to the increased effectiveness of explosions at shallower depths; deep-seated lithics are not

necessarily directly ejected by deep explosions, but rather are transported toward the surface during successive subsurface explosions via debris jets – upward moving streams of volcanoclastic debris, steam, magmatic gases, and in some cases liquid water that form above explosion sites (Ross and White 2006).

Although both models provide plausible explanations for maar-diatreme growth, little quantitative work has explored the subsurface dynamics within diatremes. Many explosions that occur within a diatreme probably do not actually erupt (discharge material into the air), but remain confined in the diatreme. However, these “non-erupting” explosions possibly have an important role in both the upward transport of material in diatremes and may contribute to the overall mixing of the vent-filling debris, though the extent to which they do remains unclear. While experiments (e.g. Ross et al. 2008ab; Andrews et al. 2014; Graettinger et al. 2014; Valentine et al. 2015) have validated that discrete subsurface blasts can form subvertical deposits similar to those found in nature and inferred to be formed from debris jets, several outstanding issues pertaining to debris jets remain, particularly those related to processes occurring at natural length and time scales. Key problems relate to the dynamics of vertical transport and mixing of material within the subsurface environment, which is the focus of this paper. Specifically, we address the following outstanding questions in this work: (1) How far do debris jets travel in diatremes and how do the initial conditions (pressure, volume, depth) control these length scales? (2) From what depth(s), if any, are debris jets more likely to erupt? And (3) how is material redistributed in diatremes following an explosion?

2. Debris jets

2.1. Debris jets in nature

McClintock and White (2006) introduced the notion of debris jets in a volcanological sense in a study on the Mawson Formation at Coombs Hills, Antarctica (Ferrar Province). They proposed that sub-vertical tuff breccia zones found within the vent complex were formed when jets of “debris + steam + water” passed through unconsolidated vent-filling debris. Based on simple spatial relationships, they found that the majority of the debris jets probably did not breach the surface and erupt, but rather remained in the vents, though they were unable to determine exactly what portion of debris jets remained confined to the vent. Furthermore, the composition of the sub-vertical deposits is indicative of both the lateral and vertical location where they originated (Ross and White 2006). They hypothesized that deposits filled with country rock-rich lapilli tuff and tuff breccia formed by explosions near the walls or floor of the diatreme, where significant fragmentation of country rock could occur. In contrast, the basalt-rich deposits were formed a distance away from the vent and floor where there was abundant basalt already present. Ross and White (2006) argued that once a debris jet has been formed above a site of explosive magma-water interaction, it entrains little to no material of the surrounding vent-filling debris as it propagates. Instead, incorporation of clasts takes place directly at or near the site of explosions.

Due to the natural structure of a maar-diatreme volcano (crater on top of diatreme), it is rare to be able to directly study a maar crater with its corresponding diatreme. In fact, we are aware of only one published example, in the Suona Crater, Japan, where a small maar and its diatreme are exposed in one coherent cross section (Geshi et al. 2011). Lefebvre et al. (2013) explored the relationship between an exposed unbedded diatreme remnant, Standing Rocks

West volcano (SRW) in the Hopi Buttes volcanic field (Arizona, USA) and complementary eruptive deposits preserved in a nearby tephra ring, providing the first semi-quantitative description of how debris jet activity distributes material within a diatreme. The SRW massif is exposed in a 150 m high vertical section, and contains a minimum of 35 individual pyroclastic columns, which generally extend either the entire height of the exposure or are rarely overlain by a different infill type (Figure 1a,b). Componentry data taken from within the massif elucidate a few key diatreme-forming processes. Wall rock abundance within the exposure is generally <10% and the wall rock that is present originated mainly from up section in the pre-volcanic stratigraphy (75 – 245 m below paleosurface), but span from the paleosurface down to 400 m. The presence of stratigraphically higher material is taken to mean that space generated at shallow depths encourages wall rock collapse as well as continual mixing of vent-filling debris during the eruption. Furthermore, the lack of deeper-seated wall rock (>700 m below the paleosurface in the case of SRW) suggests that at deep levels in the diatreme there was limited upward transport of disrupted material. In contrast, the complementary tephra ring deposit contains abundant wall rock clasts, primarily sourced from the upper 75 m of the pre-eruptive stratigraphy. To summarize, the diatreme preserves evidence of mixing of clasts from different levels and shifting explosion sites at depth, and the tephra ring indicates that most eruptive deposits were sourced from shallow explosions.

2.2 Experimental debris jets

Experiments at both the centimeter scale (e.g. Ross et al. 2008a,b; Andrews et al. 2014) and meter scale (e.g. Valentine et al. 2012, 2015; Ross et al. 2013; Taddeucci et al. 2013;

Andrews et al. 2014; Graettinger et al. 2014) have provided valuable information to our understanding of maar craters and diatreme structures. Discrete blasts in a homogeneous granular host (cm scale) can form columnar bodies similar to debris jet deposits with or without sub-aerial eruption (Ross et al. 2008a). Andrews et al. (2014) showed that a discrete blast within granular material produces a two-stage process of rapid gas expansion and subsequent granular fountaining. At the initial stage of a blast, a cavity is opened above the site where high-pressure gas is released. The cavity expands rapidly and an uprush of granular material from below the source follows. This two-stage process promotes segregation of source material and the host material, which results in the deposit being primarily composed of the source material that rushes up through the cavity. This result is consistent with fieldwork (e.g. Ross and White 2006), which has suggested that debris jet deposits are dominantly composed of material sourced near or at the explosion site.

Meter scale blast experiments in layered substrate have also had success in reproducing maar-diatreme structures (e.g. Graettinger et al. 2014 and references therein). The subsurface deposits in these experiments show evidence for progressive vertical mixing of host materials through successive explosions, which suggests that the presence of deeper seated material can be used as a natural length and time scale for upward mixing in a diatreme. Valentine et al. (2015) extended these experiments to laterally shifting explosion sites, and described how mixing occurs through upward displacement of materials directly above explosion sites, forming material domains with subvertical and irregular margins. A collar of material that experiences net subsidence surrounds the upward-displaced domains. Repeated explosions break down the domains and promote mixing. While analog and scaled experiments have yielded key insights

into maar-diatreme processes, numerical modeling is needed to explore the dynamics with realistic, natural length and energy scales. This paper is a first step in that direction.

3. Numerical methods

3.1. Numerical model and governing equations

Studying the dynamics of explosions in diatremes requires a numerical model that solves mass, momentum, and energy conservation for a system composed of compressible gas and particles, with imperfect coupling between these phases, in 2- and 3-dimensions. We adopt the Eulerian-Eulerian *Multiphase Flow with Interphase eXchanges* (MFiX; Syamlal et al. 1993)) code for this purpose. Modified versions of MFiX have been used to model various problems in volcanology such as pyroclastic density current dynamics (e.g. Darteville 2004a; Dufek and Bergantz 2007), Plinian eruption columns (e.g. Darteville 2004b), and transient processes during an eruption through a borehole in Iceland (Darteville and Valentine 2007). MFiX solves a generally accepted set of multifield partial differential equations for mass, momentum, and energy, and is closed by constitutive equations (see Agrawal et al. 2001). The approach is termed *multifield* because it treats the different phases (gas, solids) as fluid fields that are defined as volume fractions occupying the same space. In other words, particles are treated as a field whose behavior represents the mean behavior of particles within a control volume (e.g., Valentine 1994). In our simulations we assume no interphase mass transfer takes place (future applications will explore effects of condensation), resulting in the following governing equations (variables listed in Table 1; Supplementary Material):

Continuity equation for solids phases $m=1, M$:

$$\frac{\partial}{\partial t}(\epsilon_m \rho_m) + \frac{\partial}{\partial x_i}(\epsilon_m \rho_m U_{mi}) = 0$$

189 Continuity equation for gas phase g :

$$\frac{\partial}{\partial t}(\epsilon_g \rho_g) + \frac{\partial}{\partial x_i}(\epsilon_g \rho_g U_{gi}) = 0$$

190 Momentum equation for solids phases $m=1, M$:

$$\frac{\partial}{\partial t}(\epsilon_m \rho_m U_{mi}) + \frac{\partial}{\partial x_j}(\epsilon_m \rho_m U_{mj} U_{mi}) = -\epsilon_m \frac{\partial P_g}{\partial x_i} + \frac{\partial \tau_{mij}}{\partial x_j} + I_{gmi} - \sum_{k=1}^M I_{kmi} + \epsilon_m \rho_m g_i$$

191 Momentum equations for gas phase g :

$$\frac{\partial}{\partial t}(\epsilon_g \rho_g U_{gi}) + \frac{\partial}{\partial x_j}(\epsilon_g \rho_g U_{gj} U_{gi}) = -\epsilon_g \frac{\partial P_g}{\partial x_i} + \frac{\partial \tau_{gij}}{\partial x_j} - \sum_{m=1}^M I_{gmi} + f_{gi} + \epsilon_g \rho_g g_i$$

192 Granular energy balance:

$$\frac{3}{2} \rho_m \left[\frac{\partial \epsilon_m \Theta_m}{\partial t} + \frac{\partial \epsilon_m U_{mj} \Theta_m}{\partial x_j} \right] = \frac{\partial}{\partial x_i} \left(\kappa_m \frac{\partial \Theta_m}{\partial x_i} \right) + \tau_{mij} \frac{\partial U_{mi}}{\partial x_j} + \Pi_m - \epsilon_m \rho_m J_m$$

193 Energy balance equations for solids phases $m=1, M$:

$$\epsilon_m \rho_m C_{pm} \left[\frac{\partial T_m}{\partial t} + U_{mj} \frac{\partial T_m}{\partial x_j} \right] = -\frac{\partial q_{mi}}{\partial x_i} - \gamma_{gm} (T_m - T_g) + \gamma_{Rm} (T_{Rm}^4 - T_m^4)$$

194 Energy balance equation for gas phase g :

$$\epsilon_g \rho_g C_{pg} \left[\frac{\partial T_g}{\partial t} + U_{gj} \frac{\partial T_g}{\partial x_j} \right] = -\frac{\partial q_{gi}}{\partial x_i} - \sum_{m=1}^M \gamma_{gm} (T_m - T_g) + \gamma_{Rg} (T_{Rg}^4 - T_g^4)$$

195
196 The constitutive equations for the granular phase in MFiX are based on a modified Princeton
197 model (Agrawal et al. 2001) and close the momentum and energy equations by providing
198 explicit expressions for stress τ , heat exchange Π_m , and collisional dissipation J_m . These
199 equations along with the equations for the frictional stress model, interface momentum
200 transfer, drag correlation, interphase heat transfer, and conduction can be found in Appendix
201 A. The equations are solved in two dimensions on a 1 m by 1 m axisymmetric, structured mesh
202 using a semi-implicit, finite volume scheme.

203 The general multifield framework described by the above conservation equations has
204 been validated as adequately describing many physical processes associated with explosive
205 volcanic phenomena. This validation comprised comparison with experiments on particle-laden
206 overpressured (underexpanded) jets, and qualitative comparison with eruption of fragmented
207 basaltic magma and gas through a borehole in Iceland (see Darteville and Valentine 2007,
208 2008). These examples show that the model captures the combined effects of compressible
209 flow and imperfect momentum and energy coupling between carrier gas and particles.

210

211 *3.2. Initial and boundary conditions*

212 The simulations in this study are done with the purpose of elucidating intra-diatreme
213 transport and depositional processes, which are only one aspect of this complex style of
214 volcanic activity. As a result, our models do not capture the full physics of maar-diatreme
215 eruption dynamics. We focus on the *effects* of explosions on the subsurface and not on the
216 mechanisms of the explosions themselves, the dynamics of which are still poorly understood.

217 Furthermore, the infancy of a maar-diatreme eruption, whether through soft or hard substrate,
218 involves some level of fracturing and redistribution of country rock. We are not concerned with
219 this initial proto-diatreme phase and the models assume the diatreme has evolved to the point
220 where it is full of close packed debris. A list of our conceptual simplifications and their possible
221 effects on our results is given in Table 2 (Supplementary Material).

222 A volume of close packed particles is introduced into the lower portion of the
223 computational domain at various thicknesses depending on the run (h_p in Figure 1c). These
224 close packed particles are meant to represent the debris in a diatreme during an eruption. For
225 all simulations we assume this debris is of basaltic composition, with a density of 2800 kg/m^3
226 and specific heat of $1200 \text{ J/Kg}\cdot\text{K}$. All the particles are the same size (1 cm) and the same shape
227 (circular).

228 The domain has a simple cylindrical geometry, which defines the shape of the
229 boundaries, and the model is two-dimensional and axisymmetric. The boundary conditions are
230 the same for each run (Figure 1c). The bottom of the domain has a no-slip wall condition, along
231 with the right side of the domain adjacent to where particles are present, which is meant to
232 represent a boundary between a diatreme and surrounding hard country rock. The right wall,
233 above where particles are present, is given a constant pressure outflow boundary that allows
234 mass to flow out of the domain at a constant temperature and pressure (STP). The same
235 boundary condition is applied to the upper boundary. An axisymmetric coordinate system is
236 justified due to the natural cylindrical geometry of diatremes, which results in a reflection
237 boundary at the left wall of the domain.

The particle domain is stationary until a perturbation is introduced. A volume of high-pressure compressible gas, which is modeled as incondensable water vapor (from the MFiX thermo-chemical database), is then introduced into the domain (P_b in Figure 1c). The high-pressure gas overlaps a region of particles, so the initial “exploding” volume contains both the compressible gas and “juvenile” particles. This pressurized gas is analogous to the steam produced during explosive magma-water interaction. It is reasonable here to assume that steam production is instantaneous relative to the time scale of the dynamics of the resulting flow (debris jet) and the process is adiabatic with respect to the surrounding debris due to the speed of heat transfer during the thermohydraulic detonation phase in a natural molten fuel-coolant interaction (Büttner and Zimanowski 1998).

3.3. Energy budget for explosions at maar-diatreme volcanoes

To assign more precise initial conditions in our simulations, more assumptions must be made regarding the energy budget of the system. Different models have been proposed to estimate the energy budget of phreatomagmatic eruptions, based on two groups of proxies: (1) crater size, ballistics distribution, and ejecta volume; and (2) physical-thermodynamic properties of magmas, pyroclasts, and fluid phases (Büttner et al. 2006; Taddeucci et al. 2010). However, most estimates of maar forming energies are overestimated because the size of a maar crater is the result of many explosive events and not necessarily controlled by the largest explosion (Valentine et al. 2011). Energies vary considerably throughout an entire eruptive episode, so we consider a range of reasonable energy values: 10^9 - 10^{13} J, where the upper bound comes from the maximum explosion energy estimate by Self et al. (1980) for the Ukinrek

maar eruption and the lower bound is given by Valentine et al. (2014) from intra-diatreme
 intrusion volumes that provide the heat energy for phreatomagmatic explosions. The total
 thermal energy E_t released by a magma batch can be written as a product of its mass m ,
 specific heat C_p (1200 J/kg·K), and the temperature ΔT drop during MFCI (1000 K; Valentine et
 al. 2011), or, $E_t = mC_p\Delta T$. This assumes the magma is quenched rapidly during the interaction
 so that latent heat of crystallization is negligible. The conversion from thermal energy into
 usable mechanical energy via MFCI is not a conservative process, so there is then some
 efficiency factor e_c , which is in the range of 0.01 – 0.1 (Wohletz 1986; Büttner and Zimanowski
 1998; Büttner et al. 2005). So, the mechanical energy available for explosion can then be
 written as $E_m = e_cE_t$. From this equation, we can then calculate a mass or volume of magma
 for a given energy, and vice versa. For simplicity we assume that the volume ratio of initial
 liquid water to magma is unity (i.e., a mass ratio of about 1:3), although this is a subject of
 current research; thus the volume of calculated magma is equal to the volume of liquid water
 involved in the MFCI. Assuming an isochoric transformation between liquid water and vapor, a
 volume and mass of vapor produced from the MFCI is now attained, and pressure can be
 approximated using an ideal gas equation and a temperature must be assumed. During an
 MFCI, the water is heated so rapidly that it overshoots its boiling point by some amount
 (Wohletz et al. 2013). The exact temperature at which the phase transition from liquid water to
 vapor occurs is difficult to pinpoint exactly, but it is likely between the range of boiling (273 K)
 and the homogeneous nucleation temperature (~583 K), where all water spontaneously
 changes state. Here we assume an initial gas temperature of 500 K for all simulations.

4. Results

4.1. Phenomenology

While the dynamics of the debris jets formed in this work vary based on several factors such as depth of particles and initial gas pressure and volume, the overall phenomenology is consistent (Figure 2). At the instant the model begins, the high-pressure gas rapidly expands, pushing material out in a uniform way, and deforming the free surface. The amount the gas is able to expand (and thus the amount the surface is able to deform) depends largely on its initial depth. For very shallow depths, the gas expansion phase completely breaches the surface and erupts material above it, while at very large initial depths the gas is not able to expand at all, inhibiting effective debris jet formation. Following expansion, the gas bubble then begins to collapse, with adjacent particles rapidly filling the region and beginning to become vertically focused (analogous to the “granular uprush” observed by Andrews et al. 2014). The dynamics that follow vary from run to run, but generally the vertically focused jet remains coupled with some volume of gas that is rapidly collapsing, and sometimes expanding again behind the jet. Regardless of initial depth, the final deposits are similar in that they form subvertical to vertical domains dominated by the juvenile material and the material found directly adjacent to the initial high-pressure gas.

4.2. Length Scales

Some debris jets formed during the course of a maar-diatreme eruption do not breach the surface (Ross and White 2006; Valentine et al. 2014); in this section we focus on the vertical length scales of debris jets in order to determine exactly how far material is able to vertically

travel following the sudden release of a high-pressure gas in a deep environment. This is useful because it might allow for reasonable estimates on the source conditions and history of a deposit if the surrounding country rock strata are known (e.g., Lefebvre et al. 2013). Unless otherwise stated, in these simulations, all numerical parameters are fixed except the thickness of the particle bed (h_p in Figure 1c) above where the region of high-pressure gas (P_b) is assigned.

4.2.1. Vertical transport vs. initial pressure and depth

Two series (A and B; Table 3; Supplementary Material) of simulations were done to investigate the vertical transport scales of debris jets as a function of both the initial pressure of gas and the depth of release. In the first (A), the parameter P_b/P_a (pressure of high-pressure gas/pressure of ambient gas) is fixed at 200 and the second (B) it is fixed at 300. These values represent reasonable estimates of typical pressure differences between gas generated by MFCI and ambient gas pressure based on the assumptions in 3.2. The depth of the initial high-pressure gas with its juvenile particles varied from 100 m - 200 m (20 m increments), and from 200 m - 440 m (in 40 m increments). In all cases, we begin with a high-pressure gas volume of 100 m^2 , which is roughly equivalent to 3000 m^3 in three dimensions, thought to be a typical or slightly less than average amount generated in a natural MFCI (Valentine et al. 2014).

Simulations were run to completion (once the domain reaches a reasonable level of stability or the velocities become too small to resolve and the code becomes unstable) and the vertical extent of the “juvenile” material (solids species that occupied the same volume as the high-pressure gas) relative to the initial depth of particles was recorded (Table 3; Supplementary Material; Figure 3; a value of 1 means that the debris jet breached the surface). The variation in

l_d/h_p changes as a function of the initial depth as well. For large h_p , l_d remains relatively constant (Table 3; Supplementary Material). In each pressure-ratio case, l_d/h_p (height of debris jet deposit/depth of particles) decreases as the initial depth of particles (h_p) increases. In series A, the rate is steady from $h_p = 440$ m to 200 m where it increases slightly until $h_p = 160$ m, where it increases dramatically. In series B, the rate is steady from $h_p = 440$ m to 220 m where it increases slightly until $h_p = 180$ m where it increases dramatically. A 50% increase in pressure does not appear to have a large affect on the length scales.

4.2.2. Vertical transport vs. initial volume

The volumes of magma associated with phreatomagmatic explosions can be estimated reasonably well with some assumptions. Taking the energy range considered earlier, $10^9 - 10^{13}$ J and assuming an efficiency (e_c) of 0.05, the range of magma masses is 1.7×10^4 kg to 1.7×10^8 kg, which corresponds to volumes roughly (assuming $\rho = 2800$ kg/m³) of 6 m³ to 6×10^4 m³. Here, we run three series (A,C,D; Table 3; Supplementary Material) of simulations with initial high-pressure gas volumes of 3000 m³, 12500 m³, and 28000 m³. Simulations were run to completion, and again the vertical extent of the “juvenile” material (l_d) relative to the initial depth of particles (h_p) was recorded (Table 3; Supplementary Material; Figure 4). In series C, the debris jet breaches the surface from 240 m depth and in series D, the debris jet breaches the surface from 400 m depth. In general the vertical length of a debris jet deposit is larger for larger initial magma and water volumes, which in turn equate to larger explosion energies.

4.3. Width of domain

Both the lateral and vertical location of explosion sites changes throughout the course of a maar-diatreme eruption (e.g., Valentine and White 2012; Jordan et al. 2013). As a result, it is important to understand how both the width of the diatreme and the proximity of an explosion to the diatreme wall affect debris jet dynamics. Here we fix all numerical parameters but vary the width of the domain. We divide the domain into 5 identical layers, each 20 m thick with identical physical properties referred to as species A-E (A: 0 – 20 m, B: 20 – 40 m, C: 40 – 60 m, D: 60 – 80 m, E: 80 – 100m as measured from the bottom of the domain). Four simulations (W1, W2, W3, W4) were run with right boundary locations at 25, 50, 75, and 100 m, respectively, (which corresponds to actual values of 50, 100, 150, and 200 m due to reflection boundary). The jet formed during these runs weakly breached the surface, so it affected all the layers in the domain. The bulk densities (proportional to volume fractions) of each solid species are visualized in Figures 5-8. On the left of each pair of plots is the initial configuration at $t = 0$.

Several qualitative similarities are present between resulting deposit geometries in the W1 results (Figure 5). Each solid species deposit other than the juvenile species shows some level of subsidence away from the left boundary (central axis in plots), placing material below its initial starting position. The middle three species (B,C,D) all show a similar geometry where the highest concentrations are found at or below their original levels, and each is present in a thin vertically-oriented, central domain. The central juvenile deposit is narrow (<10 m) and vertical. The deposits found in the progressively wider W2 (Figure 6), W3 (Figure 7), and W4 (Figure 8) domains are considerably more complex than the narrow W1 deposits. The W2 deposits show some similar geometry to the W1 deposits – the highest concentrations of species B and C are found well below their initial positions and significant subsidence occurred

following the passage and deposition of the debris jet. In all four simulations, the juvenile species is found exclusively in a central, vertical deposit and at the surface.

Each of the simulations exhibits a central zone of upward-displaced material, which is analogous to natural debris jet deposits, surrounded by a collar of subsided materials. Pie charts show the composition of the deposits (average values from 10 m x 20 m areas; Left columns in Figure 9) in a manner similar to presentation of field data (Lefebvre et al. 2013). The 10 m is taken as an average debris jet deposit width inferred from both the simulations done here as well as the distribution of natural deposits described by Lefebvre et al. (2013). Fixing the control volume size, even though there is some variability in the simulated debris jet deposit widths, in consideration allows for an unbiased comparison between the different model runs. There are some similarities present across the four deposits: (1) there is a strong representation of both species A (green) and B (red) in all considered volumes and at no location do they represent less than 33% of the total lithologies present; (2) the concentration of the species A decreases as a function of height from the bottom of the domain; and (3) there is a nearly equivalent amount of the juvenile species found in all of the deposits.

Although there are similarities in the simulated deposits, there are also important differences. In the W1 deposit there are expressions of species D below, at, and above its original position (spanning 60 m). Furthermore, the distribution of concentrations of the W1 deposit are more uniform, with expressions of each species present in the upper 40 m of the domain and expressions of all but species E down to 60 m. In the W2 and W3 deposits, species D is present in the upper 20 m and in small amounts (less than 5% in each case), but there is also a small presence (<1%) of species D in the 60 – 80 m range in the W3 deposit. There is no

D or E present in the W4 deposit. In both the W3 and W4 deposits the concentration of species B increases as a function of height from the bottom of the domain. In addition, species C is found in the bottom 20 m in both deposits, though it is very scarce in the bottom of the W4 deposit. The results indicate that the width of a diatreme can play a role in defining the redistribution of material from individual explosions.

4.4. Vertical location of initial volume

Intra-diatreme explosions can occur in any location in a diatreme given suitable hydrologic conditions (Valentine and White 2012). As a result, it is necessary to determine how the vertical location of an explosion in the diatreme affects the mixing caused by debris jet passage. Here we discuss three simulations where the vertical location of the high-pressure gas volume is moved incrementally up through the bed of particles at locations of 10 – 20 m (V1), 20 – 30m (V2), and 30 – 40 m (V3; location levels relative to the bottom of the domain). The same initial stratification is used as above and the results are presented in the same manner (Figure 10).

Species A dominates all three deposits and decreases in concentration as a function of height from the bottom of the domain. This is the only feature that is shared by all three simulated deposits. In the debris jet deposit from V1 there is no expression of species D in any of the layers, whereas in the debris jet deposit from V2 there are expressions of species D in every layer, down to the bottom of the domain (up to 80 m below where it originated). In the deposit from V3 species D can be found down to the 20 – 40 m layer. Overall, there is a more uniform distribution of each species in the deposit from the V2 run.

5. Discussion

Our numerical modeling suggests that the vertical distance a debris jet travels depends strongly on the volume of gas generated during MFCI, with variations in the initial gas pressure having a smaller effect within the range studied. The absolute vertical transport distances tend to be relatively similar in deep runs, but increase drastically in shallower runs, indicating that the shallowest explosions in diatremes are the most efficient at transporting material. In addition, the volume of high-pressure gas dramatically affects the vertical distance the debris jets can travel within the close packed domain. Explosions at depths greater than ~250 m are unlikely to erupt unless anomalously large amounts of magma (and thus water) are present. Recent work by Valentine et al. (2014) combining relationships between scaled depth of explosions and energy constraints of phreatomagmatic explosions suggests that most eruptions in maar-diatreme systems are likely sourced by explosions in the uppermost ~200 m of the diatreme and shallower ones (<100 m) are likely to dominate deposition onto tephra rings; both are consistent with the numerical modeling results presented here.

Lorenz (1986) used the presence of deep seated lithics found in tephra rings as evidence for eruptions originating at great depths (many hundreds of meters to one or two kilometers) as well as for progressive deepening of explosions. Here we have shown that explosions deeper than 250 m are unlikely to actually vent any material originating at or near the explosion site. However, through multiple explosions at variable levels and step-wise mixing, it is possible to transport a deep seated lithic to the surface. Our modeling suggests the deeper explosions in the diatreme are not as efficient at transporting material toward the surface as shallow

explosions. It is not necessarily the case that there are any more or fewer explosions at any given depth in the diatreme, it is a matter of transport efficiency. This fact explains the general paucity of deep seated lithics in tephra rings as compared to lithics derived from shallow depths (e.g. Standing Rocks West; Lefebvre et al. 2013).

The simulated debris jet deposits commonly contain material that originated at higher levels in the initial stratigraphy. This phenomenon was observed at Standing Rocks West diatreme, where clasts in debris jet deposits were found up to 200 m below their stratigraphic position (Lefebvre et al. 2013), and was attributed to space generated at shallow depths encouraging wall rock collapse as well as evidence for continual mixing of vent filling debris during the eruption. The latter interpretation is favored here and is consistent with the simulation results. In a 100 m thick domain, the passage and sedimentation of a single debris jet was enough to redistribute material from three 20 m sections (A,B,C) throughout the entire deposit. This mixing within the debris jet deposit was driven by a combination of subsidence and entrainment. Specifically, the material that is sourced by the initial cavity opening and fountaining (e.g. species A) leaves a transient void at the bottom sides of the domain, which is progressively filled in by the material above. We note that this conclusion might be sensitive to the difference in material properties of the simulated diatreme materials (see further discussion below), which are homogeneous and may be able to respond in a more fluid-like manner than in a natural, “messy” diatreme environment.

Several simplifications in this work such as the coarse particle size (1 cm) and overall homogeneity of the particle domain were made in order to build a framework for future analyses using more complicated parameters, to build an understanding of debris jet

phenomenology, as well as to make comparisons to experimental and field work. The pressure field within the domain is affected by the pressure drop, which in turn, depends on the particle size and shape. In subaerial eruption jets, the grain size distribution controls the type of decompression pattern of the jet (e.g. Carcano et al. 2014) and large scale experiments using natural pyroclasts in a cylindrical conduit have found that coarse particles prevent efficient coupling of the gas with the particles (e.g. Dellino et al. 2007; Dellino et al. 2010; Dellino et al. 2014). In addition, the ability to form bubbles and cavities in fluidized particle beds is significantly affected by particle size (Gidaspow 1994). However, we note our current knowledge of both grain size and shape distributions in natural diatremes is very limited compared to well-fragmented ash in magmatic eruptions, and additionally that that little work has been done exploring explosion processes at many hundreds of meters depth as we have done here. Exposed diatremes in the southwestern United States (e.g. SRW; Figure 1a,b) preserve evidence of many grain sizes, shapes, and lithologies, and are often relatively coarse (tuff breccias). As a result, it is not feasible yet to attempt to simulate a specific diatreme. Furthermore, experiments using glass beads as a substrate (e.g. Ross et al. 2008; Andrews et al. 2014), modeling similar processes as we have here with smaller and uniform grain sizes, found little effect on the results when the particle size of the host glass beads was changed within the range used in their experiments (100-400 μm) and the overall phenomenology (i.e. cavitation and granular fountaining) was similar to what we observed in our simulations with coarser particles.

While we observe lithologies within the narrow debris jet deposit below their initial position, the debris jet deposit primarily records evidence of upward transport, as evidenced by

the strong presence of the lowest layers in our deposits. In contrast, the rest of the domain shows considerable subsidence. Broad downwarping in the domain generally characterize this subsidence, but more complex geometries are found in wider domains. Graettinger et al. (2014) and Valentine et al. (2015) found similar results using buried explosives, where a narrow domain recording the history of upward transport is surrounded by a subsidence collar. In a maar-diatreme system, which can experience hundreds or even thousands of explosions over an entire eruptive episode, these domains can be further destroyed and mixed resulting in a complex distribution of lithologies, making it difficult to relate the actual componentry of one debris jet deposit to a particular depth at which it formed. In a natural diatreme such as quantified by Lefebvre et al (2013), country rock clasts that originated at stratigraphic levels above a measured debris jet deposit probably had experienced subsidence from previous explosions, in order to bring the material to a depth where it could then be included into an upward-injecting jet that produced the deposit.

Several differences exist between the modeling results here and experimental results done at the meter scale (e.g. Graettinger et al. 2014) and centimeter scale (e.g. Andrews et al. 2014), particularly regarding transport of “deep” clasts. The results in Section 4.3. show that an explosion within the diatreme forms a debris jet that is able to cause upward transport of material that originated below the explosion site. Graettinger et al. (2014) did not observe eruption of material that was sourced below the explosion site, but did observe vertical transport of such material within the subsurface. This difference could possibly be attributed to the material properties of the substrate in those experiments versus the parameters used to model the particles in this work. In these models, the fluidity of the domain is exaggerated due

to the Eulerian-Eulerian framework, and the simple homogeneous single-particle size characteristics of the granular medium, which might allow for transport that would not be observed in nature. In addition, it is not clear how much the shape of the charge used in Graettinger et al. (2014) or the isometric volumes used here affect debris jet emplacement, so more work needs to be done modeling natural shaped intrusion volumes.

6. Conclusion

Numerical modeling of debris jet formation and deposition has elucidated several key intra-diatreme processes. The volume of gas generated during a MFCI is a first order control on the vertical distance a debris jet travels. Unless extremely large amounts of magma and water are involved, it is unlikely that most explosions deeper than ~250 m breach the surface. Other factors such as pressure and temperature have lesser effects on the length scales assuming they are within realistic ranges. Redistribution of material within a diatreme is primarily driven by subsidence following debris jet passage. Entrainment plays a noticeable, but small role in vertical transport of material. In every simulation done in this study, the debris jet deposit (defined as 10 m from z axis) contains >33% material sourced at or directly adjacent to the site of high-pressure release, though it was usually much more.

The results presented in this paper corroborate and advance previous field and experimental work studying debris jet activity. We have demonstrated that numerical modeling can be successfully used to investigate transport phenomena in diatremes, but there are still several open questions related to maar-diatreme volcanism. Many aspects of magma-water interaction in nature remain poorly understood, especially regarding natural length and

time scales of MFCI and the role of complex, heterogeneous properties of diatreme fill. In addition, eruptions of maar-diatreme volcanoes involve many explosions in variable locations within the diatreme over their lifetime so future modeling efforts will address how that would affect the mixing dynamics and final morphology of a diatreme. Valentine et al. (2015) noted that discrete subsurface explosions can produce structures and mixing that are similar to those which have been attributed to fluidization, particularly in the literature pertaining to kimberlite diatremes (pipes), and the modeling reported here further supports that conclusion. This, along with the general commonalities of the physical characteristics of kimberlite diatremes with those of more typical basalt compositions (Brown and Valentine 2013), suggests that phreatomagmatic processes are a viable alternative to mechanisms that rely on magmatic volatiles in the development of kimberlite pipes.

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Appendix A. Model equations (Benyahia et al. 2012)

A.1. Constitutive equations

Solids stresses:

$$\tau_{mij} = \left(-P_m + \eta\mu_b \frac{\partial U_{mi}}{\partial x_i} \right) \delta_{ij} + 2\mu_m S_{mij}$$

$$S_{mij} = \frac{1}{2} \left(\frac{\partial U_{mi}}{\partial x_j} + \frac{\partial U_{mj}}{\partial x_i} \right) - \frac{1}{3} \frac{\partial U_{mi}}{\partial x_i}$$

544 Solids pressure:

$$P_m = \epsilon_m \rho_m \theta_m \left[1 + 4\eta \sum_{n=1}^M (\epsilon_n g_{0,mn}) \right]$$

545 Solids viscosity:

$$\mu_m = \left(\frac{2 + \alpha}{3} \right) \left[\frac{\mu_m^*}{g_{0,mn} \eta (2 - \eta)} \left(1 + \frac{8}{5} \eta \sum_{n=1}^M (\epsilon_n g_{0,mn}) \right) * \left(1 + \frac{8}{5} \eta (3\eta - 2) \sum_{n=1}^M (\epsilon_n g_{0,mn}) \right) + \frac{3}{5} \eta \mu_b \right]$$

$$\mu_m^* = \frac{\rho_m \epsilon_m g_{0,mn} \theta_m \mu}{\rho_m \sum_{n=1}^M (\epsilon_n g_{0,mn}) \theta_m + \left(\frac{2\beta\mu}{\rho_m \epsilon_m} \right)}$$

$$\mu = \frac{5}{96} \rho_m d_p \sqrt{\pi \theta_m}$$

$$\mu_b = \frac{256}{5\pi} \mu \epsilon_m \sum_{n=1}^M (\epsilon_n g_{0,mn})$$

546 Solids conductivity:

$$\kappa_m = \left(\frac{\kappa_m^*}{g_{0,mn}} \right) \left[\left(1 + \frac{12}{5} \eta \sum_{n=1}^M (\epsilon_n g_{0,mn}) \right) \left(1 + \frac{12}{5} \eta^2 (4\eta - 3) \sum_{n=1}^M (\epsilon_n g_{0,mn}) \right) + \frac{64}{25\pi} (41 - 33\eta) \eta^2 \left(\sum_{n=1}^M (\epsilon_n g_{0,mn}) \right)^2 \right]$$

$$\kappa_m^* = \frac{\rho_m \epsilon_m g_{0,mn} \theta_m \kappa}{\rho_m \sum_{n=1}^M (\epsilon_n g_{0,mn}) \theta_m + \left(\frac{6\beta\kappa}{5\rho_m \epsilon_m} \right)}$$

$$\kappa = \frac{75\rho_m d_p \sqrt{\pi\theta_m}}{48\eta(41 - 33\eta)}$$

547 Collisional dissipation:

$$J_m = \frac{48}{\sqrt{\pi}} \eta (1 - \eta) \frac{\sum_{n=1}^M (\epsilon_n g_{0,mn})}{d_p} \theta_m^{3/2}$$

$$\eta = \frac{1 + e}{2}$$

548 Exchange terms:

$$\Pi_m = -3\beta\theta_s + \frac{81\epsilon_m \mu_g^2 |\mathbf{u}_g - \mathbf{u}_m|^2}{g_{0,mn} d_p^3 \rho_m \sqrt{\pi\theta_m}}$$

549 *A.2. Frictional stress model*

550 This model is used at the critical state when the solids volume fraction exceeds the maximum

551 packing limit.

$$P_c = \begin{cases} 10^{24} (\epsilon^* - \epsilon_g)^{10}, & \epsilon_g < \epsilon^* \\ 0, & \epsilon_g \geq \epsilon^* \end{cases}$$

$$\mu_f = \begin{cases} \min \left(\frac{P_c \sin(\phi)}{\sqrt{4I_{2D}}} \frac{\epsilon_s}{\sum_{m=1}^M \epsilon_m}, \mu_s^{\max} \right), & \epsilon_g < \epsilon^* \\ 0, & \epsilon_g \geq \epsilon^* \end{cases}$$

$$\mu_s^{\max} = 100$$

$$\mu_f^{\text{bulk}} = 0$$

$$I_{2D} = \frac{1}{6} \left[(D_{s,11} - D_{s,22})^2 + (D_{s,22} - D_{s,33})^2 + (D_{s,33} - D_{s,11})^2 \right] + D_{s,12}^2 + D_{s,23}^2 + D_{s,31}^2$$

$$D_{s,ij} = \frac{1}{2} \left(\frac{\partial u_{s,i}}{\partial x_j} + \frac{\partial u_{s,j}}{\partial x_i} \right)$$

552 *A.3. Interface momentum transfer*

553 Gas/solids momentum interface exchange:

$$I_{gmi} = \beta_{gm}(u_{gi} - u_{mi})$$

$$I_{kmi} = \beta_{km}(u_{ki} - u_{mi})$$

554 Wen-Yu drag correlation:

$$\beta_{gm} = \frac{3}{4} C_D \frac{\rho_g \epsilon_g \epsilon_m |\mathbf{u}_g - \mathbf{u}_m|}{d_{pm}} \epsilon_g^{-2.65}$$

$$C_D = \begin{cases} 24/\text{Re}(1 + 0.15\text{Re}^{0.687}), & \text{Re} < 1000 \\ 0.44, & \text{Re} \geq 1000 \end{cases}$$

$$\text{Re} = \frac{\rho_g \epsilon_g |\mathbf{u}_g - \mathbf{u}_m| d_{pm}}{\mu_g}$$

$$\beta_{km} = 3(1 + e) \left(\frac{\pi}{2} + c_{fkm} \frac{\pi^2}{8} \right) \frac{(d_{p,m} + d_{p,k})^2}{2\pi(\rho_m d_{p,m}^3 + \rho_k d_{p,k}^3)} \rho_m \rho_k g_{0,mn} |\mathbf{u}_m - \mathbf{u}_k| + s_{\text{coef}} P_c$$

555 *A.4. Gas momentum constitutive model*

556 Stresses:

$$\tau_{gij} = 2\mu_{gt} S_{gij}$$

$$S_{gij} = \frac{1}{2} \left(\frac{\partial U_{gi}}{\partial x_j} + \frac{\partial U_{gj}}{\partial x_i} \right) - \frac{1}{3} \frac{\partial U_{gi}}{\partial x_i} \delta_{ij}$$

$$\mu_{gt} = \min(\mu_{\max}, \mu_g + \mu_e)$$

$$\mu_e = 2l_s^2 \epsilon_g \rho_g \sqrt{I_{2Dg}}$$

$$I_{2Dg} = \frac{1}{6} \left[(D_{g,11} - D_{g,22})^2 + (D_{g,22} - D_{g,33})^2 + (D_{g,33} - D_{g,11})^2 \right] + D_{g,12}^2 + D_{g,23}^2 + D_{g,31}^2$$

557 Porous media model:

$$f_{gi} = -\frac{\mu_g}{c_1} U_{gi} - \frac{c_2}{2} \rho_g \left(\sqrt{U_{gj} U_{gj}} \right) U_{gi}$$

558 *A.5. Energy equation constitutive model*

559 Interphase heat transfer:

$$\gamma_{gm} = \frac{C_{pg} R_{gm}}{\left[\exp \left(\frac{C_{pg} R_{gm}}{\gamma_{gm}^0} \right) - 1 \right]}$$

$$\gamma_{gm}^0 = \frac{6\kappa_g \epsilon_m \text{Nu}_m}{d_m^2}$$

$$\text{Nu}_m = (7 - 10\epsilon_g + 5\epsilon_g^2)(1 + 0.7\text{Re}_m^{0.2}\text{Pr}^{1/3}) + (1.33 - 2.4\epsilon_g + 1.2\epsilon_g^2)\text{Re}_m^{0.7}\text{Pr}^{1/3}$$

Conduction:

$$q_{mi} = -\kappa_m \frac{\partial T_m}{\partial x_i}$$

$$q_{gi} = -\kappa_g \frac{\partial T_g}{\partial x_i}$$

References

- Agrawal K, Loezos PN, Syamlal M, Sundaresan S (2001) The role of meso-scale structures in rapid gas-solid flows. J Fluid Mech 445:151-185
- Andrews RG, White JDL, Dürig T, Zimanowski B (2014) Discrete blasts in granular material yield two-stage process of cavitation and granular fountaining. Geophys Res Lett 41:1-6
- Benyahia S, Syamlal M, O'Brien TJ (2012) Summary of MFIX Equations 2012-1. <https://mfix.netl.doe.gov/documentation/MFIFEQuations2012-1.pdf>
- Brown RJ, Valentine GA (2013) Physical characteristics of kimberlite and basaltic intraplate volcanism, and implications of a biased kimberlite record. Geol Soc Am Bull 125:1224-1238, doi:10.1130/B30749.1
- Büttner R, Zimanowski B (1998) Physics of thermohydraulic explosions. Phys Rev E 57:5726-5729

575 Büttner R, Zimanowski B, Mohrholz C-O, Kümmel R (2005) Analysis of thermohydraulic
 576 explosion energetics. J Appl Phys 98, 043524

577 Büttner R, Dellino P, Raue H, Sonder I, Zimanowski B (2006) Stress-induced brittle
 578 fragmentation of magmatic melts: Theory and experiments. J Geophys Res 111, B08204

579 Carcano S, Esposti Ongaro T, Bonaventura L, Neri A (2014) Influence of grain-size distribution on
 580 the dynamics of underexpanded volcanic jets. J Volcanol Geotherm Res 285:60-80

581 Darteville S (2004a) Numerical modeling of geophysical granular flows: 1. A comprehensive
 582 approach to granular rheologies and geophysical multiphase flows. Geochem Geophys
 583 Geosyst 5(8): Q08003

584 Darteville S (2004b) Numerical modeling of geophysical granular flows: 2. Computer
 585 simulations of plinian clouds and pyroclastic flows and surges. Geochem Geophys Geosyst
 586 5(8): Q08004

587 Darteville S, Valentine GA (2007) Transient multiphase processes during the explosive eruption
 588 of basalt through a geothermal borehole (Námafjall, Iceland, 1977) and implications for
 589 natural volcanic flows. Earth Planet Sci Lett 262:363-384

590 Darteville S, Valentine GA (2008) Multiphase magmatic flows at Yucca Mountain, Nevada. J
 591 Geophys Res 113, B12209

592 Dellino P, Zimanowski B, Büttner R, La Volpe L, Mele D, Sulpizio R (2007) Large-scale
 593 experiments on the mechanics of pyroclastic flows: design, engineering and first results. J
 594 Geophys Res 112, B04202 doi: 10.1029/2005jb003958

595 Dellino P, Dioguardi F, Zimanowski B, Büttner R, Mele D, La Volpe L, Sulpizio R, Doronzo DM,
 596 Sonder I, Bonasia R, Calvari S, Marotta E (2010) Conduit flow experiments help constraining
 597 the regime of explosive eruptions. J Geophys Res doi: 10.1029/2009JB006781
 598 Dellino P, Dioguardi F, Mele D, D'Addabbo M, Zimanowski B, Büttner R, Doronzo DM, Sonder I,
 599 Sulpizio R, Dürig T, La Volpe L (2014) Volcanic jets, plumes and collapsing fountains:
 600 evidence from large-scale experiments, with particular emphasis on the entrainment rate.
 601 Bull Volcanol 76, 834 doi: 10.1007/s00445-014-0834-6
 602 Delpit S, Ross P-S, Hearn BC (2014) Deep-bedded ultramafic diatremes in the Missouri River
 603 Breaks volcanic field, Montana, USA: 1 km of syn-eruptive subsidence. Bull Volcanol 76:832
 604 Dufek J, Bergantz GW (2007) Suspended load and bed-load transport of particle-laden gravity
 605 currents: the role of particle-bed interaction. Theor Comput Fluid Dyn 21:119-145
 606 Geshi N, Nemeth K, Oikawa T (2011) Growth of phreatomagmatic explosion craters: a model
 607 inferred from Suona crater in Miyakejima Volcano, Japan. J Volcanol Geotherm Res 201:30-
 608 38
 609 Gidaspow D (1994) Multiphase Flow and Fluidization: Continuum and Kinetic Theory
 610 Descriptions, Academic, San Diego, Calif.
 611 Graettinger AH, Valentine GA, Sonder I, Ross P-S, White JDL, Taddecucci J (2014) Maar-diatreme
 612 geometry and deposits: Subsurface blast experiments with variable explosion depth.
 613 Geochem Geophys Geosyst 15:740-764
 614 Graettinger AH, Valentine GA, Sonder I, Ross P-S, White JDL (2015) Facies distribution of ejecta
 615 in analog tephra rings from experiments with single and multiple subsurface explosions.
 616 Bull Volcanol, in press

617 Jordon SC, Cas RAF, Hayman PC (2013) The origin of a large (>3 km) maar volcano by
 618 coalescence of multiple shallow craters: Lake Parrumbete maar, southeastern Australia. J
 619 Volcanol Geotherm Res 286:331-347

620 Kienle J, Kyle PR, Self S, Motyka RJ, Lorenz V (1980) Ukinrek Maars, Alaska, I. April 1977
 621 eruption sequence, petrology and tectonic setting. J Volcanol Geotherm Res 7(1-2):11-37

622 Lefebvre NS, White JDL, Kjarsgaard BA (2013) Unbedded diatreme deposits reveal maar-
 623 diatreme-forming eruptive processes: Standing Rocks West, Hopi Buttes, Navajo Nation,
 624 USA. Bull Volcanol 75(8):739-756

625 Lorenz V (1986) On the growth of maars and diatremes and its relevance to the formation of
 626 tuff rings. Bull Volcanol 48(5):265-274

627 McClintock MK, White JDL (2006) Large-volume phreatomagmatic vent complex at Coombs
 628 Hills, Antarctica records wet, explosive initiation of flood basalt volcanism in the Ferrar LIP.
 629 Bull Volcanol 68:215-239

630 Ross P-S, White JDL (2006) Debris jets in continental phreatomagmatic volcanoes: a field stud of
 631 their subterranean deposits in the Coombs Hills vent complex, Antarctica. J Volcanol
 632 Geotherm Res 149(1-2):62-84

633 Ross P-S, White JDL, Zimanowski B, Büttner R (2008a) Rapid injection of particles and gas into
 634 non-fluidized granular material, and some volcanological implications. Bull Volcanol
 635 70(10):1151-1168

636 Ross P-S, White JDL, Zimanowski B, Büttner R (2008b) Multiphase flow above explosion sites in
 637 debris-filled volcanic vents: insights from analogue experiments. J Volcanol Geotherm Res
 638 178(1):104-112

639 Ross P-S, White JDL, Valentine GA, Taddeucci J, Sonder I, Andrews RG (2013) Experimental birth
 640 of a maar-diatreme volcano. *J Volcanol Geotherm Res.* 260:1-12
 641 Self S, Kienle J, Huot J-P (1980) Ukinrek Maars, Alaska, II. Deposits and formation of the 1977
 642 craters. *J Volcanol Geotherm Res* 7:39-65
 643 Syamlal M, Rogers W, O'Brien TJ (1993) MFIX Documentation Theory Guide. Technical Note,
 644 DOE/METC-94/1004, NTIS/DE94000087
 645 Taddeucci J, Sottili G, Palladino DM, Ventura G, Scarlato P (2010) A note on maar eruption
 646 energetics: current models and their application. *Bull Volcanol* 72:75-83
 647 Taddeucci J, Valentine GA, Sonder I, White JDL, Ross P-S, Scarlato P (2013) The effect of pre-
 648 existing craters on the initial development of explosive volcanic eruptions: an experimental
 649 investigation. *Geophys Res Lett* 40(3):507-510
 650 Valentine GA (1994) Multifield governing equations for magma dynamics. *Geophys & Astrophys*
 651 *Fluid Dyn* 78:1, 193-210
 652 Valentine GA, White JDL, (2012) Revised conceptual model for maar-diatremes: subsurface
 653 processes, energetics, and eruptive products. *Geology* 40:1111-1114
 654 Valentine GA, Shufelt NL, Hintz ARL (2011) Models of maar volcanoes, Lunar Crater (Nevada,
 655 USA). *Bull Volcanol* 73(6):753-765
 656 Valentine GA, White JDL, Ross P-S, Amin J, Taddeucci J, Sonder I, Johnson PJ (2012)
 657 Experimental craters formed by single and multiple buried explosions and implications for
 658 volcanic craters with emphasis on maars. *Geophys Res Lett* 39(2)
 659 Valentine GA, Graettinger AH, Sonder I (2014) Explosion depths for phreatomagmatic
 660 eruptions. *Geophys Res Lett* 41(9):3045-3051

661 Valentine GA, Graettinger AH, Macorps E, Ross P-S, White JDL, Döhring E, Sonder I (2015)
 662 Experiments with vertically and laterally migrating subsurface explosions with applications
 663 to the geology of phreatomagmatic and hydrothermal explosion craters and diatremes.
 664 Bull Volcanol 77:15
 665 van Otterloo J, Cas RAF, Sheard MJ (2013) Eruption processes and deposits characteristics at
 666 the monogenetic Mt. Gambier Volcanic Complex, SE Australia: implications for alternating
 667 magmatic and phreatomagmatic activity. Bull Volcanol 75:737
 668 White JDL (1991) Maar-diatreme phreatomagmatism at Hopi Buttes, Navajo Nation (Arizona),
 669 USA. Bull Volcanol 53:239-258
 670 White JDL, Ross P-S (2011) Maar-diatreme volcanoes: a review. J Volcanol Geotherm Res 201(1-
 671 4):1-29
 672 Wohletz KH (1986) Explosive magma-water interaction: thermodynamics, explosive
 673 mechanisms, and field studies. Bull Volcanol 48:248-264
 674 Wohletz KH, Zimanowski B, Büttner R (2013) Magma-water interactions. In Modeling Volcanic
 675 Processes: The Physics and Mathematics of Volcanism. Edited by Fagents SA, Gregg TKP,
 676 Rosaly MC. Cambridge University Press, U.K., 431 p.
 677

Figure captions

Figure 1. (a) Standing Rocks West diatreme. Entire vertical exposure is ~150 m, and its base is ~300 m below the ground surface at the time of its formation (eruption); (b) Close up view of Standing Rocks West diatreme with subvertical pyroclastic columns outlined in red-dashed lines (see detailed study by LeFebvre et al. 2013); (c) Computational domain compared to idealized maar-diatreme. The right and bottom walls of the domain are given a no slip boundary condition, while the left side is implicitly assigned a reflection boundary due to the axisymmetric coordinates. The free surface in the domain represents the boundary between the top of the diatreme (or floor of the maar crater) and the atmosphere. Mass is allowed to leave the system only above the free surface in both the vertical and horizontal directions. h_p – depth of particle domain, w_p – width of particle domain, l_d – vertical extent of debris jet, P_a – ambient gas pressure, P_b – pressure of initial gas bubble.

Figure 2. Idealized phenomenology of a “typical” debris jet. (a) initial layout of model. A volume of pressurized gas, which is co-located with particles, is introduced into the bottom of the domain. The particles that overlap with this pressurized gas are taken as an analog to juvenile particles; (b) the gas rapidly expands, pushing the juvenile particles outward uniformly in all directions; (c) the gas bubble begins to collapse, with the lower region collapsing faster as particles move in to fill in the area opened by the gas expansion; (d) the juvenile particles are forced upward through the gas, beginning to form a more vertically focused jet, while still entraining directly adjacent material; (e) the gas pockets are stretched and coupled to the jet as vertical motion begins to dominate – some minor entrainment occurs only at the base of the domain; (f) motion of the jet is purely vertical while the gas

begins to decouple from the particles resulting in (g) the final debris jet deposit. Tracer particles are illustrated showing how the initial layout of material relates to the final deposit. Figure 3. Results from series A ($P_b/P_a = 200$) and B ($P_b/P_a = 300$) simulations (refer to Table 3 for full list of parameters; Supplementary Material). Vertical extents of debris jets (l_d/h_p) are plotted as a function of the initial depth (h_p). If $l_d/h_p = 1$, that means that the juvenile material reached the surface (i.e. an eruption). Two regimes are clearly present in each series, one with a shallow slope at deep depths (>200 m) and one with a steep slope at shallower depths (>200 m) implying that debris jet transport is more efficient at shallower depths. A 50% increase in pressure has a negligible effect on the length scales. The lines between the data points are meant to help visualize the relationship between depth and transport and do not imply a fitted functional relationship.

Figure 4. Results from series A, C, and D simulations, which have initial pressurized gas volumes of 100 m^3 , 200 m^3 , and 300 m^3 , respectively (refer to Table 3 for full list of parameters; Supplementary Material). Vertical extents of debris jets (l_d/h_p) are plotted as a function of the initial depth (h_p). If $l_d/h_p = 1$, that means that the juvenile material reached the surface (i.e. an eruption). Series D (yellow) erupts from 400 m depth, series C erupts from 240 m depth, and series A erupts from 120 m depth. The lines between the data points are meant to help visualize the relationship between depth and transport and do not imply a fitted functional relationship.

Figure 5. Visualized results from W1. Initial (left panels) and final (right panels) bulk density of the material of interest, (density x volume fraction) is plotted side by side for each species

present in the domain. “Final” refers to the time at which the domain becomes stable or the velocities become too small to resolve and the code becomes unstable.

Figure 6. Visualized results from W2. Initial (left panels) and final (right panels) bulk density of the material of interest, (density x volume fraction) is plotted side by side for each species present in the domain. “Final” refers to the time at which the domain becomes stable or the velocities become too small to resolve and the code becomes unstable.

Figure 7. Visualized results from W3. Initial (left panels) and final (right panels) bulk density of the material of interest, (density x volume fraction) is plotted side by side for each species present in the domain. “Final” refers to the time at which the domain becomes stable or the velocities become too small to resolve and the code becomes unstable.

Figure 8. Visualized results from W4. Initial (left panels) and final (right panels) bulk density of the material of interest, (density x volume fraction) is plotted side by side for each species present in the domain. “Final” refers to the time at which the domain becomes stable or the velocities become too small to resolve and the code becomes unstable.

Figure 9. Componentry results for debris jet deposits in domains with variable widths. The initial volume distribution is on the far left. The colors represent individual layers in the domain. Pie charts in the left columns correspond to a 10 m by 20 m volume, while the pie charts in the right columns correspond to the adjacent 10 m by x m volume, where x is the width of the domain minus 10 m.

Figure 10. Componentry results for debris jet deposits where the initial location of the high-pressure volume was changed. The initial volume distribution is on the left. The colors

742 represent individual layers in the domain. Each pie chart corresponds to a 10 m by 20 m
743 volume. Note that the uppermost layer was not found in the final deposits.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10