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Magmatic versus phreatomagmatic fragmentation: Absence of evidence is not evidence of absence

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ABSTRACT

Fragmentation processes in eruptions are commonly contrasted as phreatomagmatic or magmatic; the latter requires only fragmentation of magma without external intervention, but often carries the connotation of disruption by bubbles of magmatic gas. Phreatomagmatic fragmentation involves vaporization and expansion of water as steam with rapid cooling and/or quenching of the magma. It is common to assess whether a pyroclast formed by magmatic or phreatomagmatic fragmentation using particle vesicularity, shape of particles, and degree of quenching. It is widely known that none of these criteria is entirely diagnostic, so deposit features are also considered; welding and/or agglomeration, particle aggregation, lithic fragment abundance, and proportion of fines. Magmatic fragmentation yields from rhyolite pumice to obsidian to basaltic achneliths or carbonatitic globules, making direct argument for magmatic fragmentation difficult, so many have taken an alternative approach. They have tested for phreatomagmatism using the fingerprints listed above, and if the fingerprint is lacking, magmatic fragmentation is considered proven. We argue that this approach is invalid, and that the criteria used are typically incorrect or incorrectly applied. Instead, we must consider the balance of probabilities based on positive evidence only, and accept that for many deposits it may not be possible with present knowledge to make a conclusive determination.

INTRODUCTION

One of the long-standing challenges in volcanology is to determine the fragmentation processes, which define the transition from a bubbly magma to a gas with dispersed fragments, during explosive eruptions (e.g., Gonnermann, 2015). A strong distinction has long been drawn between phreatomagmatic (also termed hydrovolcanic or hydromagmatic) and magmatic fragmentation in eruptions. Strictly, the latter simply means fragmentation of magma without involvement of external water, such as by disruption by bubbles of magmatic gas or by brittle failure of magma due to high strain rates (such as might happen during rapid decompression; Dürig and Zimanowski, 2012). Phreatomagmatic fragmentation implies that external water, with resulting rapid melt quenching and water vaporization, played a role in fragmenting the magma (Table 1).

Magma is necessarily involved in phreatomagmatic fragmentation, and a common approach to assessing whether a particular deposit was formed by magmatic or/and phreatomagmatic fragmentation is to make a stepwise assessment. This varies in detail, but commonly is based on particle shape, degree of quenching, and whether there is a glassy fluidal exterior film, and for a bulk sample, proportions of differently classified clasts. It is widely known that no single one of these clast-specific criteria is entirely diagnostic and other criteria at the local deposit level are often considered, such as welding, particle aggregation, and lithic fragment abundance. For whole-deposit grain populations the proportion of fines is often considered, and is sometimes applied even at local deposit level.

The range of products produced by magmatic fragmentation is broad and well known, and includes dense to highly vesicular fragments with many different shapes. This hinders establishment of a magmatic "fingerprint," and many have instead chosen to demonstrate a magmatic origin by "ruling out" a phreatomagmatic one. A test is run against criteria considered proof of phreatomagmatic origin, and if the test fails a magmatic origin is considered proven. In other words, the absence of evidence for phreatomagmatic fragmentation is taken as evidence of its absence. Such "argument from ignorance" is logically fallacious at its core, and further weakened by problems with the criteria applied (Table 2; Fig. 1). Positive arguments are needed, and an acceptance that fragmentation process(es) of an eruption cannot always be determined from its deposits.

VESICULARITY

Vesicles are among the most obvious and varied features of particles in primary volcaniclastic rocks (White and Houghton, 2006). Dense, glassy, almost vesicle-free volcanic particles from deep-sea eruptions are one possible magma-water end member, opposite basaltic pumice or reticulite from Hawaiian fountains as the magmatic end member (Table 2). This simple distinction fails for rhyolitic compositions, where both pumice and obsidian are formed from dry magmatic eruptions.

A more rigorous way of using vesicles to distinguish between magmatic and phreatomagmatic (used in this paper broadly for all magmatism involving external water) particles was clearly expressed in an empirical study by Houghton

TABLE 1. EXAMPLES OF SOME HISTORICAL ERUPTIONS WITH RELATIVELY CLEAR INFERENCES OF MAGMATIC OR PHREATOMAGMATIC MECHANISMS

Eruption observed* (A.D.)	Phreatomagmatic or magmatic	Basis	Deposit features (not all underpin overall interpretation)
Surtsey 1963–1967[1,2,3]	phreatomagmatic then magmatic	PM: eruption emerged from ocean, abundant water vapor, salt on pyroclasts M: eruption shifted to lava fountaining after vent isolated, formed lava delta	PM: accretionary lapilli, vesicular tuff, sideromelane pyroclasts, moderate vesicularity M: lava, spatter
Taal 1965[4]	magmatic, then phreatomagmatic	M: incandescent fountaining of scoria cone PM: explosions, crater filled with water, low-temperature surges, incandescence not described	M: spatter PM: surge dunes, ash plastering, high abundance of country-rock lithic fragments
Littoral eruptions Kilauea (various years)[5]	phreatomagmatic from magmatic	M: lava flow to coastline PM: explosions at ocean, observed response to change in lava entry processes at ocean, abundant steam	M: feeding lava flow PM: spatter, Pele's hair, Pele's tears, and limu (Pele's seaweed; Hon et al., 1988), lithic blocks, delicate vesicular fluidal sideromelane
Myojinsho 2011[6,7]	phreatomagmatic	underwater explosions, steamy explosions breaching surface, tephra jets	pumice, dense juvenile clasts
Pu'u 'Ō'ō Kilauea 1983–1986[8]	magmatic	lava fountains, lava lake	spatter plus scoria cone, lava
Kilauea 2008[9,10]	magmatic	in-conduit observations, circulating lava, but discrete explosions	vesicular ash, tiny volumes
Stromboli, various years[11,12]	magmatic	open vent, ponded lava, incandescent bursts	bombs, spatter, scoria
Parícutin 1943–1952[13,14]	magmatic	incandescence, dry tephra fall, lava flow	lava, scoria and ash falls widespread
Montserrat dome 1995–1998[15]	magmatic	extrusive growth, incandescent block and ash flows, incandescent fountains	block and ash deposits, pumiceous pyroclastic flow deposits

Note: M—magmatic, PM—phreatomagmatic.
*References: [1]Thorarinsson et al. (1964); [2]Thorarinsson (1967); [3]Lorenz (1974); [4]Moore et al. (1976); [5]Mattox and Mangan (1997); [6]Morimoto and Ossaka (1955); [7]Fiske et al. (1998); [8]Heliker and Mattox (2003); [9]Houghton and Gonnermann (2008); [10]Carey et al. (2012); [11]Blackburn et al. (1976); [12]Lautze and Houghton (2007); [13]Foshag and Gonzalez (1956); [14]Pioli et al. (2008); [15]Watts et al. (2002).

TABLE 2. EXAMPLES OF CRITERIA FOR MAGMATIC VERSUS PHREATOMAGMATIC FRAGMENTATION, AND COUNTEREXAMPLES

Indicative deposit feature	General interpretation	Counterexample	References
Fluidal pyroclast	magmatic, dry	littoral cones, submarine spatter	1, 2, 3, 4, 5, 6, 7
Blocky pyroclast	phreatomagmatic	vulcanian, blast-surge pyroclasts	8, 9, 10
Abundant fines	phreatomagmatic	White Island, Eyjafjallajökull	11, 12, 13
High lithic content	phreatomagmatic	Tarawera proximal fallout, caldera-collapse breccias, lithic lag breccias and lithic breccias of pyroclastic flows, blast surges	11, 13, 14, 15,
High vesicularity	magmatic	submarine pumice, Loihi, Surtseyan	16, 17, 18, 19
High bubble number density	intense magmatic	Ilchulbong, Black Point	7, 20, 21
Low vesicularity	phreatomagmatic	vulcanian, Stromboli, block and ash deposits	8, 9, 22, 23, 24, 25, 26, 27, 28

Note: 1—Fisher (1968); 2—Jurado-Chichay et al. (1996); 3—Mattox and Mangan (1997); 4—Simpson and McPhie (2001); 5—Schipper and White (2010); 6—Kaneko et al. (2005); 7—Jutzeler et al. (2016); 8—Miwa et al. (2009); 9—Burgisser et al. (2010); 10—Belousov (1996); 10—Houghton and Nairn (1991); 11—Carey and Houghton (2010); 12— Dellino et al. (2012); 13—Lipman (1976); 14—Walker (1985); 15—Macías et al. (1998); 16—Fiske et al. (2001); 17—Kano and Yoshikawa (2005); 18—Schipper et al. (2010a, 2010b); 19—Schipper and White (2016); 20—Murtagh et al. (2011); 21—Murtagh and White (2013); 22—Hoblitt and Harmon (1993); 23—Kennedy et al. (2005); 24—Polacci et al. (2006); 25—Lautze and Houghton (2007); 26—Shea et al. (2012); 27—Schwarzkopf et al. (2005); 28—Belousov et al. (2002).

and Wilson (1989), in which eruptions inferred to have involved magma-water interaction produced particle populations with a greater range of vesicularity, and a lower median value, than did eruptions or eruptive phases without water. The explanation for the broader vesicularity range of phreatomagmatic particles is that those with lower vesicularity were formed and quenched either before their volatiles could exsolve or before small bubbles could grow under decompression to reach high volume proportions. In contrast, higher vesicularity particles encountered water higher in the conduit, or not at all, allowing further exsolution of volatiles and growth of bubbles during depressurization. This distinction has been complicated by recent studies of deposits from observed Hawaiian and Strombolian eruptions, which were clearly driven by magmatic volatiles but contain clasts with a wide range of vesicularity, even within individual clasts, due to entrainment of degassed melt in conduits by erupting more gas-rich melt (e.g., Stovall et al., 2012; Gurioli et al., 2014).

Vesicle-population analysis has become an important tool for assessing conditions and rates of magma ascent (e.g., Shea et al., 2010), and yields information not available from total-vesicularity (particle density) studies or simple descriptive analysis of vesicle shapes. Despite the detail available from such

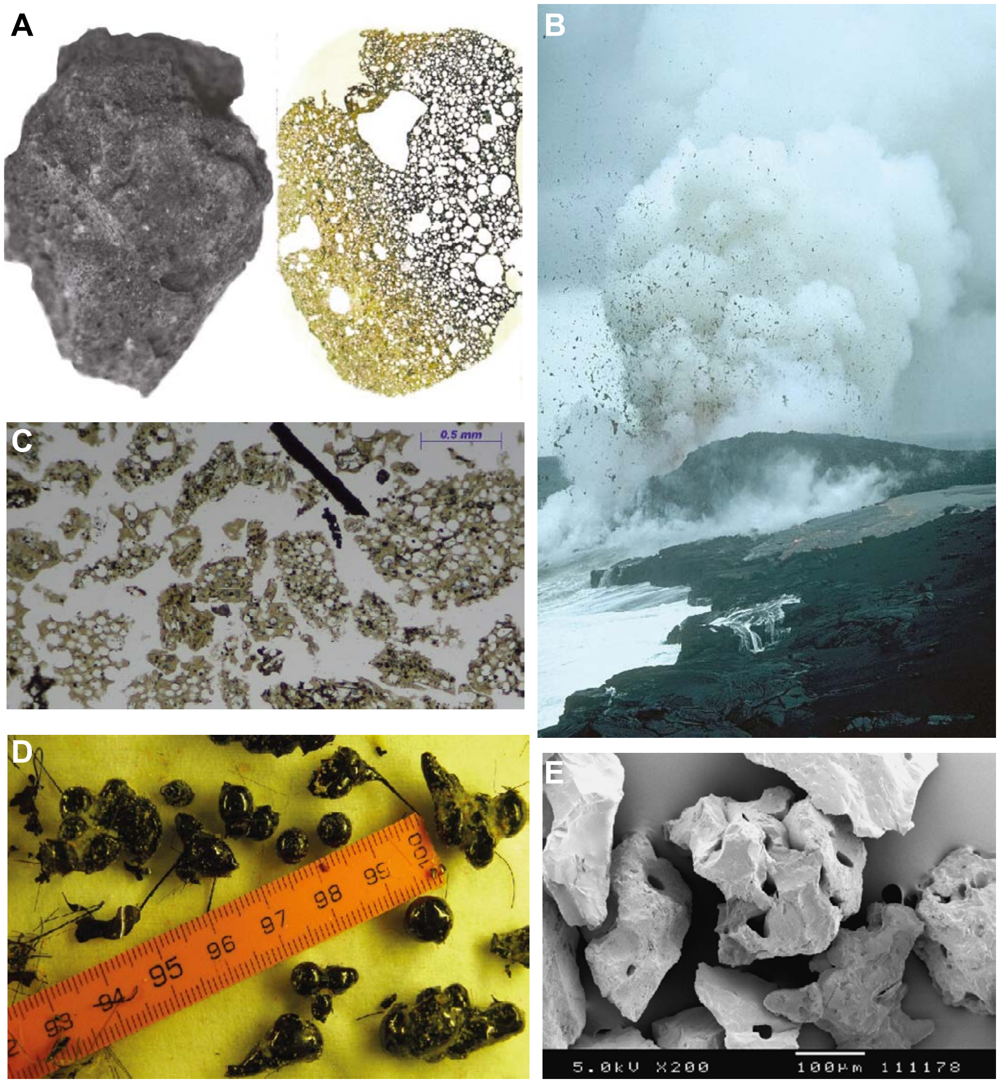

Figure 1. Particles from different eruptions that would be misdiagnosed using standard magmatic versus phreatomagmatic criteria. (A) Eruptions at Loihi submarine volcano, Hawai'i, have produced vesicular (to 60%) pyroclasts that are neither dense nor blocky, despite eruption in 1.2 km of water. (B) Littoral eruptions driven by water vaporization produce littoral cones dominated by fluidal spatter clasts that weld and/or agglomerate upon landing. Similar deposits are produced by magmatic eruptions, so the spatter is not diagnostic of either. (Photo of littoral activity on Kilauea, Hawai'i, by L. Keszthelyi, U.S. Geological Survey.) (C) Highly vesicular Surtsey (Iceland) pyroclasts sampled at 120 m below sea level by drilling. Increments on scale are millimeters. (D) Basaltic glass spheres, some with tails, some welded together, from pouring basalt melt into a wet slurry with 60% sediment (Schipper et al., 2011b). (E) Blocky grains with low vesicularity from a 2003 Vulcanian eruption of Karymsky volcano. Photo courtesy of A. Belousov.

studies, no unique fingerprint of magma-water interaction has emerged (e.g., Schipper et al., 2010a, 2011a; Murtagh and White, 2013). Analyzing both vesicle populations and the volatile abundances in glass (adjusted to reflect the crystal population), coupled with information on original volatile abundance from glass inclusions in phenocrysts, enables construction of an eruption's volatile budget and tracking of volatile transfer from melt to vesicles through time, but does not yield a direct magma-water signature (e.g., Schipper et al., 2010b).

PARTICLE MORPHOLOGY

Walker and Croasdale (1971) recognized a relationship that is now presented in textbooks and many papers; in the basaltic deposits they studied inferred phreatomagmatic particles were typically equant and fracture bound, while many from inferred dry magmatic eruptions were achnelithic, having aerodynamic forms (Table 2). Heiken (1972), followed by many others, termed the equant fracture-bound particles blocky. Heiken and Wohletz (1985) presented a comprehensive atlas and analysis of pyroclastic particles, including comparisons with those produced in experimental explosions by interaction of water with a magma analogue (Wohletz, 1983; but see Büttner et al., 1999, and the following discussion of molten fuel-coolant interactions).

A key concept applicable to both magmatic and phreatomagmatic fragments is that not all particles hold information about fragmentation dynamics that relates specifically to the explosion mechanism. Magma surrounding an explosion or expansion site can be broken into fragments without contributing to that explosion or expansion; the internal textures of those fragments carry no information about the explosion mechanism. Fluid magmas, for example, can be torn apart to form a shower of fluidal achnelithic fragments by bursts of air in experiments, by magmatic gas in hornitos, lava fountains, and Strombolian bursts, or by vaporization of external water trapped below or in the magma (Fisher, 1968; Zimanowski et al., 1997a; Mattox and Mangan, 1997). An additional problem in small-scale eruptions such as those of Stromboli (Capponi et al., 2016), Surtsey (Schipper and White, 2016) or of maar-diatreme volcanoes (White, 1996) is that recycling of material within the vent means that many, sometimes most, particles erupted in a single explosion are just passive participants rather than active in causing the explosion. For particles formed in more highly explosive magmatic eruptions there can be a similar distinction; where outgassed magma is shattered by explosive decompression of an underlying magma body, the particle population reflects the history of the shattered magma. Blocky dense particles are formed by magmatic eruptions of this sort (Heiken and Wohletz, 1985), including blast (e.g., Belousov, 1996) and Vulcanian ones (e.g., Miwa et al., 2009), and to complicate the matter further, some of these particles show stepped fracture surfaces somewhat reminiscent of those formed by molten-fuel coolant interactions (MFCI) (Miwa et al., 2009).

For particles formed by experimental phreatomagmatic explosions, only a narrow size range (grains <130 µm), comprising ~30 wt% of the total particles, carries the signature of direct magma-water interaction (Büttner et al., 2002). No equivalent fingerprint for magmatic eruptions yet exists, not least because such eruptions form fragments in many ways, but experimental work continues (Küppers et al., 2006; Dürig and Zimanowski, 2012), as do efforts to ascertain fingerprints empirically (Liu et al., 2015).

QUENCHING VERSUS GLASSY SKIN

Basalt must solidify quickly to form glass (sideromelane), with slower cooling rates yielding glass charged with a cryptocrystalline mass of nanolites or/and microlites (tachylite). Basalt magma quenched by water commonly forms sideromelane granules (e.g., black sand where Kilauea lava is granulated at the shoreline, or in laboratory experiments), and ash fragments of many phreatomagmatic eruptions are dominated by sideromelane (e.g., Heiken, 1974). Such quenching is rare for dry basaltic eruptions, but is sometimes achieved in vigorous lava fountains (e.g., Stovall et al., 2012). Silicic magma, in contrast, can form glass even when cooling is relatively slow, such as in obsidian domes (Fink and Manley, 1987), so silicic glass is of little help in inferring cooling rates.

PARTICLE WELDING OR AGGLUTINATION

To stick primary volcanic particles back together after they are formed, heat must be retained or applied, along with some force. Basaltic bombs or coarse lapilli formed as fluidal, (spatter) clasts can become agglutinated upon landing or with subsequent loading when their interiors remain hot and ductile during emplacement and their accumulation rate is sufficiently high (Head and Wilson, 1989). Retaining this characteristic requires that the clasts not be fully quenched, but this is not a difficult threshold to meet because loss of heat from the interior of a clast is limited to diffusive rates. Agglutinate is common in hornitos (small lava-flow–fed spatter cones), around lava fountains, and in layers within scoria cones. It is also known from submarine eruptions (Kaneko et al., 2005), and is a common feature of littoral cones formed by magma-water explosions at the coastline (Fisher, 1968; Mattox and Mangan, 1997) and rootless cones where lava flows flood swamps and bogs (Hamilton et al., 2010), i.e., in situations clearly involving external water. Agglutination of original pyroclasts also takes place within vents, and perhaps especially in phreatomagmatic ones, to form composite bombs (e.g., Rosseel et al., 2006; White and Houghton, 2006; Lefebvre et al., 2013; Valentine and van Wyk de Vries, 2014; Schipper and White, 2016) but also in magmatic eruptions of Strombolian or Hawaiian fountains, for example, where spatter accretes around a nucleus as it rolls down steep inner crater walls back toward the vent (accretionary bombs; Heiken, 1978; Valentine and Cortés, 2013).

PARTICLE AGGREGATION

Particle aggregates, such as accretionary and armored lapilli, commonly form by adhesion of damp ash to a nucleus of ice or rock. They are a well-known product of phreatomagmatic eruptions (Moore et al., 1966; Lorenz, 1974; Self, 1983), and occasionally form when meteoric moisture wets ash in eruption plumes (Folch et al., 2010). Different types can be distinguished, and a range of inferences drawn (Moore and Peck, 1962; Schumacher and Schmincke, 1995; Brown et al., 2010; Van Eaton and Wilson, 2013). They can provide compelling evidence of water droplets in the eruption column, plume, or density currents, but do not indicate the nature of fragmentation that produced the ash being aggregated.

LITHIC FRAGMENT ABUNDANCE

Fragments of country rock are important constituents of maar-forming eruptions, which are defined by their crater cut below the preeruptive surface that remains open at the end of the eruption (e.g., Ollier, 1967; White, 1991). It is widely accepted that basaltic maars result from phreatomagmatic eruptions

(White and Ross, 2011, and references therein), and lithic-rich horizons in tuff cones and rings, and in scoria cones, are commonly attributed to episodes of phreatomagmatic activity (e.g., Womer et al., 1980; Houghton et al., 1996; Ross et al., 2011). In other contexts, such as eruptions produced during caldera subsidence, increased abundance of lithics need not involve water or phreatomagmatic processes (Druitt and Bacon, 1986; Bear et al., 2009), and even in basaltic eruptions high abundances of lithic clasts in deposits can result from other processes (Table 2; Houghton and Nairn, 1991; Carey and Houghton, 2010).

PROPORTION OF FINE PARTICLES

An important feature often used to diagnose effects of water on an eruption is the proportion of fine particles. This is embedded in the well-known Walker eruption classification that plots fall-deposit dispersal against grain size (Walker, 1973), and finds support also in thermodynamic analyses of magma-water fuel-coolant interactions (Sheridan and Wohletz, 1981; Wohletz and McQueen, 1984; Wohletz, 1986). Strong fragmentation of magma requires considerable energy (Zimanowski et al., 1997a; Alatorre-Ibargüengoitia et al., 2010; Dürig et al., 2012), so it makes sense that intense magma-water explosions add to the proportion of fine fragments produced.

Diagnosing phreatomagmatic interaction on the basis of high proportions of fines in an examined deposit remains, despite this sensible background, problematic (Table 2). First, only a whole-deposit grain-size distribution can be considered representative. It is a common misperception that all juvenile particles erupted by a phreatomagmatic explosion should have the signature of magma-water interaction. Zimanowski et al. (1997a) showed that even in strong MFCI in laboratory experiments, only ~33% of the melt mass interacts with water and fragments to fine ash, while the other ~66% of particles are produced as melt around the explosion site is torn apart and ejected, (hydrodynamic fragmentation) and form fluidal clasts. Environmental factors can introduce a bias in grain size; for example, local water flushing (Talbot et al., 1994) or deposition of aggregate particles (Lorenz, 1974; Rosi, 1992; Van Eaton and Wilson, 2013) are known to increase the proportion of fine particles deposited in a given site. Specific to the evolution of maar-diatreme eruptions, there is also good evidence that the tephra ring deposits, for which one might obtain a whole-deposit grain-size inventory, do not represent well the overall fragment population produced by the volcano because of systematic retention of coarser fragments within the maar-diatreme subsurface structure (e.g., Valentine and White, 2012; Lefebvre et al., 2013; Graettinger et al., 2015). The abundant fines in many inferred phreatomagmatic deposits have conspicuously high proportions of country rock or recycled fragments that were not formed by fragmentation during the explosion that deposited them (Houghton and Nairn, 1991; White, 1991). Fine-grained lithic debris can be produced by repeated explosions before being finally ejected onto a tephra ring, for example, but also depends upon the original characteristics of the host material. A mudstone or shale, or poorly indurated fine sands and silts, will eject as fine-grained material that is not reflective of any particular degree of volcanic fragmentation.

Magmatic eruptions can also produce very high proportions of juvenile fines in some instances (Palladino and Taddeucci, 1998; Dellino et al., 2012). In contrast, there is no published evidence that submarine or other subaqueous eruptions generate particularly large proportions of fines. This may be because fines in such settings are very efficiently separated by hydraulic sorting, but until some whole-deposit grain-size data become available, this must be considered no more than a reasonable hypothesis. Surtsey is the classic example for near-vent enrichment in fine particles produced from a shallow-submarine phreatomagmatic vent, but even there separating the signature of water in the jets and plumes, which drive near-source deposition and leave their imprint on depositional features, from a fragmentation signature is difficult. It is compounded by strong evidence for particle recycling through the vent (Thorarinsson et al., 1964; Kokelaar, 1983; Moore, 1985; Schipper and White, 2016). In deposits of both Surtseyan eruptions and deeper submarine ones, interactive particles have been identified (Schipper et al., 2010b, Murtagh and White, 2013). Such particles are known from MFCI experiments, and imply phreatomagmatic fragmentation (Zimanowski et al., 1997a), but the subaqueous edifices formed by these eruptions are not rich in fines, whether from a paucity of fines produced or efficient hydraulic sorting.

OTHER CLAST FEATURES

There are other simple features that are less commonly available but that can be helpful. Normal jointing on large particle surfaces is well known from water-chilled fragments (e.g., Yamagishi, 1987; McPhie et al., 1993; Allen and McPhie, 2000). Cauliflower bombs (Lorenz, 1974) have deeply quenched rough surfaces; these differ in surface-fracture geometry and internal vesicularity from the breadcrust bombs common in some dry eruptions (e.g., Fisher and Schmincke, 1984; Wright et al., 2007), although bombs with some similarities to cauliflower bombs are also known from dry eruptions (Miyabuchi et al., 2006; Alvarado et al., 2011). A range of fine ash textures, including mossy fragments and fragments with surface hydration, cracking (Büttner et al., 1999), and pitting have been described from phreatomagmatic deposits (Heiken and Wohletz, 1985), but may not be distinguishable from diagenetically formed textures in less than pristine older deposits.

MAGMA:WATER RATIOS—RELEVANCE, USE, AND ABUSE

Sheridan and Wohletz (1981) introduced a diagram summarizing their analysis of the effects of different amounts of water relative to magma on the energy released when magma acts as a fuel in interaction with water; a fuel-coolant interaction (FCI or MFCI). These proportions, described by a ratio of fuel:coolant mass, were argued to release maximum kinetic energy when the fuel:coolant mass ratio is ~3:1, for basalt and water close to a 1:1 volume ratio. MFCI sensu stricto in relatively low viscosity magmas requires fluid-fluid mixing of melt and water and development of an insulating vapor film, which catastrophically fails in response to an external or internal trigger. Vapor-film collapse brings liquid melt and water into contact with a large surface area for interaction; rapid heat transfer fractures the melt, allowing further penetration of water and further melt quenching and water expansion. In experiments, low bubble content in the melt favors explosive MFCI as the compressible gas within bubbles dampens the melt fracturing process (Zimanowski et al., 1995). Austin-Erickson et al. (2008) demonstrated a type of MFCI involving high-viscosity rhyolite melt, wherein fluid-fluid premixing does not occur (due to the high viscosity contrast of the melt compared to water), but fracturing due to deformation and quenching at the melt-water interface produces a similar detonation-like process.

Wohletz and Sheridan (1983) explicitly linked different volcanic edifices to different water:magma ratios, from cinder cones with little or no water, to pillow lavas with deep water; maximum explosivity forms tuff rings with groundwater. Sheridan and Wohletz (1983) linked water:magma ratios to a range of bedding structures, eruptive phenomena, and dominant products. Welded spatter, together with lava fountains and Strombolian ballistic and ash fall, is at the minimum end of hydrovolcanic activity, while pillow breccias, lahars, and pillow lava occupy the excessive end of the spectrum. Phreatoplinian and Surtseyan eruptions are at optimum ratios, although phreatoplinian cases, which assume interaction of water in a vent with high sustained magma discharge rate, arguably cannot be governed by the same processes associated with MFCI (see also Palladino et al., 2015).

The equating of large-scale features and phenomena with water:magma ratio has been widely embraced, even at the textbook level (e.g., Schmincke, 2004), but has critical flaws that have been addressed in some studies (e.g., White, 1996; Wohletz, 2002, 2003) and that need to be accepted broadly. The key feature of this ratio is that it applies to domains of interaction that are much smaller than an entire volcano or even an entire single fragmentation domain. In natural settings, the efficiency of total fragmentation is probably lower than in experiments, although the domain size mingled prior to explosion will be larger. Only fragments generated in the interacting domain show the characteristic features of magma-water interaction, and these particles are largely confined, in both experiments and natural deposits, to a narrow size range of small grains that reflect the fragmentation process. Particles larger than this range are produced by other processes, including simple tearing of fluid melt that produces spatter indistinguishable from spatter produced by any other explosion. This is why littoral spatter is identical to some magmatic spatter, and shows a wide range of vesicularity depending only on the state of vesicularity of the lava entering the sea at the time littoral explosions are generated.

In evaluating the importance of water:magma ratios in the ocean, Wohletz (2003), in a discussion that deserves widespread attention, explained that the likelihood of an MFCI is dependent on domain size with respect to vapor-film thickness and heat-transfer parameters. Wohletz (2003, p. 38) concluded that for a wavelike curling layer of magma effectively enclosing a 1.5-m-diameter domain of water of similar volume (Fig. 2), there is an "…effective R in the range of 0.01 to 3.8; the lower extent of this range is compatible with R values thought to be typical of terrestrial explosive hydrovolcanism." In other words,

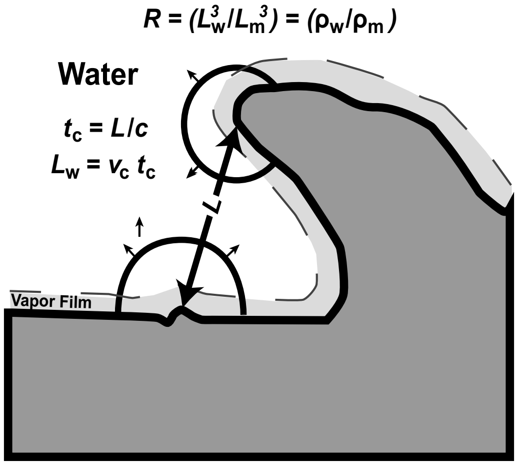

Figure 2. The key feature of the water:magma ratio is that it applies to domains of interaction that are much smaller than an entire volcano or even an entire single fragmentation domain. Here for an interaction in an ocean, distance *L* is 1.5 m and defines the domain size to give an effective water:magma ratio (*R*) between 0.1 and 3.8 (diagram redrawn after Wohletz, 2003). L_w is characteristic length for water, L_m is characteristic length for magma, ρ_w is density of water, ρ_m is density of magma, t_c is characteristic time, and v_c is growth speed of the supercritical fluid layer.

even though perhaps a few cubic meters of magma is under consideration within the volume of the ocean, the effective water:magma ratio may be as low as ~1 part water to 100 parts magma (R ~0.01). White (1996) discussed implications of domain size for interaction of liquefied water-saturated sediment with magma and concluded that sediment grains within the interaction domains would further reduce the water:magma ratio, but this effect is tiny compared to the vast difference between effective domain ratios versus whole-environment ratios.

The significance of the effective water:magma mass ratio cannot be overstated, but is commonly ignored: the effective water:magma mass ratio is the only thermodynamically relevant one, and may differ by orders of magnitude from the ratio of water to magma in general in a volcanic setting. Intense explosions can result in settings with the wrong ratio (too much water, or too little) if the right local mingling and trigger conditions are met; conversely, interaction of ideal amounts of magma with water can yield lavas or spatter if the interaction domains are small, or if explosive interaction is not triggered. Note that triggering events and their required energies have not been established fully even for well-characterized laboratory systems, and there are no quantitative or repeated observations from any natural eruption that allow certain prediction of whether an explosive interaction will take place. From littoral observations and thermal granulation experiments it is known that magma can nonexplosively interact with water volumes ranging from a film on wet sand to the ocean. In addition, from rootless cones in swampy or littoral settings, it is known that explosions can be generated by changing confinement, e.g., by having interactions in lava tubes (Fisher, 1968; Mattox and Mangan, 1997) or with water entrapped from below (Hamilton et al., 2010), even in submarine lavas (Tribble, 1991). It is uncertain whether either of these involve full-fledged MFCI, but they produce explosions and powerful jets driven by transfer of heat from magma to water.

Additional issues may call into question the robustness of the optimal water:magma ratio as a universal value, even accounting for the effective ratio. For example, in experiments by Zimanowski et al. (1997b), optimal explosion energies in experiments involving basaltic melts (compared to thermite; Wohletz and McQueen, 1984) were for water:magma mass ratios of 0.03–0.04 (~25–35 times as much magma as water), which is an order of magnitude different from the Wohletz and Sheridan (1983) water:magma optimal mass ratio of ~0.3. This difference has largely been unnoticed by other researchers who continue to refer to the value of ~0.3 that was argued by Wohletz and Sheridan (1983). This difference points to the possibility that optimal conditions might depend upon many factors, including length scale, system geometry, imposed dynamics (e.g., magma flowing in a dike compared to a relatively static experimental container), and fluxes of melt and water. We suggest that our state of knowledge on magma-water interaction is still in its infancy, and it is premature to assume simple relationships between magma:water ratios, explosion efficiencies, and fragmentation characteristics.

MAAR-DIATREME VENT STRUCTURES AND ERUPTIVE FACIES

One of the clear characteristics of magma-water interaction within the context of MFCI is that it can result in discrete energetic explosions. When these occur in the subsurface (interaction with groundwater), the explosions brecciate host rocks, and cause both upward mixing of subsurface host rock and juvenile materials (debris jets; Ross and White, 2006; Lefebvre et al., 2013) and downward subsidence (Lefebvre et al., 2013; Delpit et al., 2014), both of which have been amply documented by field studies, laboratory- and field-scale experiments, and multiphase numerical modeling (Ross and White, 2006; Lefebvre et al., 2013; Delpit et al., 2014; Ross et al., 2008a, 2008b; Andrews et al., 2016; Graettinger et al., 2014; Valentine et al., 2015; Sweeney and Valentine, 2015). The resulting funnel-shaped structures (diatremes) are typically several hundred meters in diameter (to a few kilometers) and can extend hundreds of meters, even >1 km, into the preeruptive subsurface. On the surface, the explosions produce tephra rings around broad craters. Deposits in the tephra rings reflect the repeated occurrence of discrete explosions in the form of numerous, often lithic-rich, thin packages of deposits from pyroclastic density currents, many of which preserve dune bedforms (pyroclastic surge deposits; Fisher and Waters, 1970) and associated fallout. In these maar-diatreme systems, the diatreme and tephra facies relationships are the most direct evidence for dominant phreatomagmatic activity.

Volcanoes dominated by purely magmatic volatile-driven processes can also have discrete explosions, but produce different conduit structures and deposits. In low magma-flux mafic systems such as scoria cones, the nonphreatomagmatic cousins of many maar-diatremes, discrete magmatic explosions are usually manifested as Strombolian explosions that discharge mostly gas with small quantities of melt over short time scales of seconds (e.g., Taddeucci et al., 2012). These explosions are much less damaging to host rocks beneath the vent, and a growing number of field studies show that vent structures (e.g., flared zones above a feeder dike) are generally limited to the upper 100 m, more typically the upper tens of meters, of the feeding system (Keating et al., 2008; Valentine, 2012; Friese et al., 2013; Geshi and Oikawa, 2014; Geshi and Neri, 2014; Harp and Valentine, 2015). Eruptions of more viscous magmas can also produce discrete explosions in the form of Vulcanian bursts (for simplicity we include Mount St. Helens–like lateral blasts in this category). Such explosions often initiate at or near the top of a magma column, or in an active lava dome, due to sudden decompression either by fracturing of a solidified plug or collapse of a confining slope. Although we are not aware of any direct geologic observations of conduits that are demonstrably related to Vulcanian eruptions, the fact that the explosions are driven by downward-propagating expansion waves, lasting tens of seconds, rather than causing radially directed shocks from a localized MFCI source, also suggests that the effects on host rocks are minimal compared to phreatomagmatic explosions and that the vent structures have relatively limited lateral extents. For example, Vulcanian explosions at Soufrière Hills (Montserrat) excavated craters on the active lava dome (Druitt et al., 2002), and apparently these did not extend downward into preeruption host rocks. Conduit diameters for those explosions are thought to have been ~30 m and relatively constant between the shallow crater and feeding reservoir (Melnik and Sparks, 2002). In both of these magmatic explosion processes (Strombolian and Vulcanian), eruptive products are relatively poor in country-rock lithics (e.g., Valentine, 2012; Druitt et al., 2002), compared to those of maar-diatremes.

Thus within the context of MFCI as an explosion mechanism, vent structures and ejecta deposits may be considered as the strongest evidence of phreatomagmatic activity, perhaps the best evidence we can get with our current knowledge, even if individual clasts yield ambiguous evidence. It is ironic that diatremes with juvenile material of kimberlitic composition (kimberlite pipes) have essentially the same structures as normal maar-diatremes (e.g., White and Ross, 2011), and the general large-scale characteristics of kimberlite fields are the same as other small-volume intraplate volcanoes (McClintock et al., 2009; Brown and Valentine, 2013), yet many rule out phreatomagmatic activity because of perceived absence of evidence at the scale of individual clasts.

CONCLUSIONS

Wohletz et al. (2012, p. 252) concluded that a "…well-developed catalog exists of diagnostics that are useful to characterize magma–water interaction from analysis of hydrovolcanic (phreatomagmatic) products, including vent/construct morphology, deposit dispersal, tephra bedding sequences, and ash analyses." We argue that this catalog is not so clear-cut as the statement implies, but by combining a number of the lines of evidence addressed here, with the understanding that each line carries its own caveats and uncertainties, one can make a strong case that an eruption's fundamental behavior was probably significantly shaped by interaction with water. It is rare that a strong case can be made on the basis of only one or two features, particularly if representation of the overall deposit is poor because of outcrop limitations or alteration. The fine ash that has been demonstrably linked with MFCI is among the first diagenetic casualties, unfortunately, and also is less likely to be deposited in proximal-medial areas that are the focus of most field studies. Whole-deposit grain size is difficult to obtain, even for fallout deposits, which probably represent the most straightforward deposits for which to make such an estimate, let alone pyroclastic surge deposits.

We return to our title, i.e., "absence of evidence is not evidence of absence." The criteria that have been traditionally used to rule out a phreatomagmatic component focus mainly on clast characteristics, especially blocky, poorly vesicular juvenile ash (or with a wide range of vesicularity) and accretionary lapilli. We have reviewed how these are problematic criteria, and shown that features commonly linked to some types of purely magmatic fragmentation can also form by purely phreatomagmatic fragmentation, or by fragmentation that combines those processes. We argue that the long-accepted optimal ratio of water mass to magma mass is not so clear, has not held up under subsequent experimental work, and has been applied in ways that do not account for the dynamics of MFCI. Textures within pyroclasts, such as vesicle populations and microlites, are contingent on the magma's history up to and often beyond the point of fragmentation; they do not reflect the fragmentation process. Phreatomagmatic processes, involving interaction of liquid water with complex bubble- and crystal-bearing magmas, in complex geometric and dynamical situations that characterize natural volcanic systems are not well-understood phenomena, but rather represent a major grand challenge in volcanological research.

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