MeMoVolc consensual document: a review of cross-disciplinary approaches to characterizing small explosive magmatic eruptions

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Abstract A workshop entitled "*Tracking and understanding volcanic emissions through cross-disciplinary integration: A textural working group*." was held at the Université Blaise Pascal (Clermont-Ferrand, France) on the 6-7 November 2012. This workshop was supported by the European Science Foundation (ESF). The main objective of the workshop was to

establish an initial advisory group to begin to define measurements, methods, formats and standards to be applied in the integration of geophysical, physical and textural data collected during volcanic eruptions. This would homogenize procedures to be applied and integrated during both past and ongoing events. The workshop comprised a total of 35 scientists from six countries (France, Italy, Great Britain, Germany, Switzerland and Iceland). The four main aims were to discuss and define:

- Standards, precision and measurement protocols for textural analysis
- Identification of textural, field deposit, chemistry and geophysical parameters that can best be measured and combined
- The best delivery formats so that data can be shared between, and easily used by different groups;
- Multi-disciplinary sampling and measurement routines currently used, and measurement standards applied, by each community

The group agreed that community-wide, cross-disciplinary integration, centered on defining those measurements and formats that can be best combined, is an attainable and key global focus. Consequently, we prepared this paper to present our initial conclusions and recommendations, along with a review of the current state of the art in this field that supported our discussions.

Introduction

A major goal of modern volcanology is to relate conditions of magma ascent to the resulting eruption style using information preserved in volcanic deposits. Because it is impossible to directly observe magma ascent, vesiculation and fragmentation, one way to obtain quantitative information on magma-ascent dynamics is through textural quantification of the sampled particles. Textural quantification involves full description of the vesicle and crystal properties of erupted products (e.g., Sparks 1978; Sparks and Brazier 1982; Whitham and Sparks 1986; Houghton and Wilson 1989; Marsh 1988, 1998; Cashman and Marsh 1988, Toramaru 1989, 1990; Cashman and Mangan 1994; Higgins 2000; 2006; Blower et al. 2002; Burgisser and Gardner 2005; Shea et al. 2010a; Rust and Cashman 2011; Baker et al. 2012 and references therein). Magma viscosity, ascent rate, vesiculation processes, fragmentation style and explosion dynamics all imprint characteristic and measurable properties on the textures of volcanic particles, as shown by theoretical and experimental studies (e.g., Rust and

Cashman 2011; Gonnerman and Houghton 2012; Degruyter et al. 2012; Nguyen et al. 2013 and references therein). The main assumption is that most of the pyroclast properties are acquired during ascent in the conduit, with few changes occurring after fragmentation or in the atmosphere, if the pyroclasts are lapilli size or smaller (e.g., Houghton and Wilson 1989; Nguyen et al. 2013). Specifically, the textural parameters of the pyroclastic components can yield insights into the dynamics of explosive eruptions, as reviewed in Table 1.

However, the physical characteristics of individual pyroclasts must not be considered in isolation from detailed studies of (i) the deposits from which they were collected, (ii) their chemical properties, (iii) geophysical signatures of the related explosive event, or (iv) petrological and/or analogue experiments. Indeed, attempts to understand eruption dynamics have been increasingly coupled to traditional fieldwork and geophysical measurements made synchronously with sample collection. In 2004, a special issue of the *Journal of Volcanology and Geothermal Research* (Volume 137) focused on multidisciplinary approaches, proposing "*simultaneous collection of multiple geophysical data sets, such as seismic, infrasonic, thermal and deformation data, as well as sampling of ejecta and detailed mapping*". The argument was that "*complete constraint of a volcanic system is not possible using one data set, so that an integrated multiparametric approach involving simultaneous collection of multiple geophysical and petrological data sets will increase our ability to reach tightly constrained and confident conclusions regarding the mechanics and dynamics of volcanic systems and eruptions*" (Harris et al. 2004). Since 2004, numerous studies have borne these predictions out, combining textural data with:

i. Field deposits (e.g., Polacci et al. 2006a; Rust and Cashman 2007; 2011; Mattsson 2010);
ii. Petrological data (e.g., Larsen 2008; Shea et al. 2009; 2010b; Burgisser et al. 2010; Bai et al. 2011);
iii. Chemical analyses (e.g., Piochi et al. 2005, 2008; Shimano and Nakada 2006; Noguchi et al. 2006; Costantini et al. 2010; Schipper et al. 2010a; b; c; 2011; 2012; 2013; Balcone-Boissard et al. 2010, 2011, 2012; Shea et al. 2012; 2014)
iv. Geophysical measurements (e.g., Burton et al. 2007; Gurioli et al. 2008, 2013; 2014; Polacci et al. 2009b; Andronico et al. 2008; 2009a; 2009b; 2013a, 2013b; Miwa et al. 2009; Miwa and Toramaru 2013; Colò et al. 2010; Landi et al. 2011; Pistolesi et al. 2011; Leduc et al. 2015)

Together, these studies have delivered complete pictures of explosive eruptions and their dynamics (Fig. 1).

Despite this progress, we remain far from establishing the best protocols for sampling pyroclasts and for correlating and comparing the many parameters that can be measured using individual clasts and field deposits. Only a few papers address some of these issues (e.g., Bonadonna et al 2013; Engwell et al. 2013; Klawon et al. 2014). In addition, no study has yet attempted to correlate all derivable textural parameters with the full range of multidisciplinary data available. To partially resolve these issues, a working group funded by the European Science Foundation, through the MeMoVolc program (http://www.memovolc.fr/), was set up. The group was composed of experts actively working on integration of textural, deposit and geophysical data, equally balanced between four theme areas: (i) particle-texture studies, (ii) deposit analysis, (iii) chemistry and (iv) geophysics. The priorities of the meeting were discussion and definition of:

- Improved standards, precision and measurement protocols needed by the particle-texture studies.
- Best practices for particle-texture studies in order to have comparable datasets from different types of eruptions.
- Parameters obtained from particle-texture, deposit, geochemical and geophysical data that need to be measured, and the best delivery format if each discipline's output is to be of use to all workers.
- Multi-disciplinary sampling and measurement routines, as well as measurement standards.

The core communal issues to be explored were agreed on:

1. What are the best sampling and measurement strategies for the quantification of pyroclast textural features, and what are their precision and uncertainty?
2. What are the best sampling and measurement strategies for pyroclastic deposits to allow textural characterization of their particles?
3. How can we link chemistry and particle-texture properties?
4. How can we link geophysical data and the particle-texture quantification?
5. What is the best multi-disciplinary strategy for combining output from each field in a meaningful way?

The paper reviews these topics in the light of a workshop consensus. Because of the time constraints and the complexity of the arguments, the paper focuses only on the study of explosive subaerial magmatic eruptions that generate sustained columns or fountains, and the associated fallout deposits (Fig. 1). Further workshop or working groups should be organized

to synthesise and integrate all work in progress, and already completed, in the areas of phreatic, phreatomagmatic and submarine explosions, as well as pyroclastic density current and lava flow deposits (Table 1). .

The final objective of this paper is to ensure that data collected in the field and laboratory can be shared effectively and ingested in a multi-disciplinary sense into experiments, modeling and monitoring. In the longer term, the objective is to publish and update standards, as well as to propose, support and organize field meetings to test integrated collection methodologies. The ultimate aim is to increase the number of open-access data-bases of standard and community-accepted quality, thereby increasing resources available for cross-disciplinary correlations.

Sampling of pyroclasts and quantification of their textural features

(i) Representative samples

Pyroclasts reflect degassing of the parent magma, from the conduit to the plume. Part of the textural signature is assumed to reflect the fragmentation (or explosion) zone. Consequently, texture can be used as an indicator of magma properties (composition, porosity, connectivity, permeability, vesicle and crystal content, size, shape and distribution) at that time (Table 1). This assumption has two requirements:

i. The textural signature that was quenched immediately at the fragmentation level can be distinguished from the textural effects of post-fragmentation processes, including microlite formation and bubble nucleation, expansion, collapse, coalescence and Ostwald ripening that will change clast vesicularity or vesicle size and shapes once the pyroclast has been formed (e.g., Thomas et al. 1994; Cashman et al. 1994; Herd and Pinkerton 1997; Larsen and Gardner 2000; Gurioli et al. 2008; Costantini et al. 2010; Stovall et al. 2011, 2012). The time window for post-fragmentation changes depends on magma composition, viscosity and fragmentation depth.
ii. Because clast density is also a function of clast size (Houghton and Wilson 1989), only clasts of similar sizes must be used in order to avoid non-uniform grain-size effects on textural parameters.

We thus recommend choosing samples that are representative of the studied explosion, or unit, in terms of:

i. Timing: This requires sampling of narrow stratigraphic intervals (Houghton and Wilson 1989) in which juvenile clasts of similar dimensions can be assumed to represent those parts of the magma fragmented at a particular time (n.b. conduit processes can change over short timescales);
ii. Distribution: This requires selection of more than one outcrop for each event;
iii. Degree of fragmentation: This requires selection of a sampling methods that is appropriate for the the full range of grain sizes in the deposit;
iv. Componentry: If the juvenile fraction is heterogeneous, then sampling should be done based on preliminary componentry analysis of the clasts analyzed (e.g., Wright et al. 2011, Eychenne et al 2015)

In previous studies, only clast sizes of 16-32 mm, i.e. coarse lapilli (White and Houghton 2006) have been considered for textural purposes. Such clasts were considered to be large enough to be easily sampled and studied, while being fully representative of the density variation of the majority of erupted pyroclasts and unaffected by significant post-fragmentation phenomena (Houghton and Wilson 1989). These requirements are not always met. In basaltic magma, post-fragmentation effects can be a complication even for these sizes (e.g., Cashman et al. 1994; Szramek et al. 2006; Costantini et al. 2010; Gurioli et al. 2008; Pioli et al. 2014; Pistolesi et al. 2008; 2011; Stovall et al. 2011; 2012). In these cases, the challenge is to identify, quantify and remove post-fragmentation effects in order to isolate textures preserved across the fragmentation zone. For example, the original shapes of vesicles may be reconstructed by de-coalescencing large vesicles using the presence of broken, or partially retracted, glassy septa.

However, if we study an ash-dominated or a bomb-dominated event, particle-texture analyses must be performed on the fine or coarse juvenile fragments, respectively. Ash size particles (<2 mm) have been investigated recently in terms of vesicle and crystal size distributions (Taddeucci et al. 2002, 2004; Cioni et al. 2008; D'Oriano et al. 2011a, b; Miwa et al. 2009; 2013; Miwa and Toramaro 2013; Proussevitch et al. 2011; Genareau et al. 2012; 2013; Colucci et al. 2013, Schipper et al. 2013), and an extensive work has been done in the last 40 years in characterizing ash morphology and deposit componentry (Table 1). For the ash fraction, post-fragmentation expansion can be excluded (e.g., Proussevitch et al. 2011; Genareau et al. 2012; 2013; Colucci et al. 2013). Consequently, analyses allow comparison between morphological and textural features of clasts sampled in proximal and distal areas. Ash particles can record most of the information related to magma ascent dynamics (e.g., decompression-driven microlite crystallization) and fragmentation (Cioni et al. 2008;

D'Oriano et al. 2005; D'Oriano et al. 2011a, b; Proussevitch et al. 2011; Genareau et al. 2012; 2013; Colucci et al. 2013). Advantages of studying ash are that it can also be statistically more representative of the variability of the magma properties and is less affected by density-driven settling within the plume. However, ash fragments record only small-scale vesicularity. The integration of observations made on the external shapes of clasts may give information about the presence and importance of a coarser vesicularity which drives magma fragmentation (e.g., Proussevitch et al. 2011; Genareau et al. 2012; 2013; Colucci et al. 2013). However, they cannot provide complete information about the abundance and size of the full vesicle population, if the magma included bubbles larger than the ash particles. Furthermore, ash particles are not suitable for permeability studies, as they are often smaller than the bubbles forming the permeability network. However, the presence of coalesced vesicles in a preferred direction, and an abundance of ash clasts with an elongate shape, have been interpreted as an indication of the development of a permeable bubble network (D'Oriano et al. 2011a).

Bombs may provide a plethora of information regarding pre-eruptive degassing and ascent rate (e.g. Hoblitt and Harmon 1993; Wright et al. 2007), timing and degree of thermal interaction of magma with wall-rock material prior to ejection (Rosseel et al. 2006; Sottili et al. 2009; 2010), post-fragmentation changes due to bubble growth, coalescence or shape changes (e.g. Herd and Pinkerton 1997, Shin et al. 2005) and mingling between stagnant and fresh magma (Gurioli et al. 2014; Leduc et al. 2015).

(ii) Bulk measurements of particle characteristics

The fastest and most straightforward textural measurement of individual pyroclasts is density (vesicularity), which provides basic information on processes related to gas exsolution and escape (Houghton and Wilson 1989). The densities of lapilli and small bombs can be determined by comparing their weights in water and air following the Archimedes principle. Clasts can be made impermeable with silicone waterproofing spray, by immersion in cellulose acetate, or by using Parafilm^{TM} wax. This technique is fairly rapid and yields large arrays of data with a reproducibility within 10-30 kg m^{-3} and accuracy within 30 kg m^{-3} (Barker et al. 2012). Quicker, more timely and precise, density measurements can now be performed using a commercial envelope-density measurement device (http://www.micromeritics.com/Product-Showcase/GeoPyc-1360-Envelope-Density-Analyzer.aspx). Following the same principles, a battery-powered device has been used to vacuum-seal pumice or scoria in plastic bags in the field (Kueppers et al. 2005).

For pyroclasts characterized by fine vesiculation (with largest vesicles smaller than 2-3 mm), the density can be measured with the glass-bead method (Nakamura et al. 2008) that allows the calculation of the density as well as the volume of an object of irregular size. For large bombs (from 15 to 40 cm in diameter), a "natural waterproofing" effect was exploited (Gurioli et al. 2013). Extensive tests showed that decimetric size bombs collected at Stromboli acquired a "natural waterproofing" from their quenched margins and thus could be weighed in water without waterproofing. This represents an easy, precise and fast strategy for large bombs.

The derived density distributions, coupled with external morphology variation, can be used as filters to select a few clasts, representative of the low, modal and high density values, from each subpopulation observed (e.g., Shea et al. 2010a). Selected clasts are then used for textural quantification.

Other bulk particle-texture measurements include vesicle connectivity, permeability (Klug and Cashman 1996; Klug et al. 2002; Formenti and Druitt 2003; Rust and Cashman 2004; and references in Table 1) and electrical conductivity (Le Pennec et al. 2001; Bernard et al. 2007; Wright et al. 2009; Wright and Cashman 2014). The connectivity measurements are mostly performed using gas-displacement helium pycnometers, and they deliver first-order information on the outgassing capacity (i.e., potential for gas loss) of the magma near fragmentation (Klug et al. 2002; Formenti and Druitt 2003; Giachetti et al. 2010; Shea et al. 2011; 2012). Permeability controls the rate at which magma outgases during decompression. Several methods exist for permeability measurements in volcanology. Rust and Cashman (2004) used a commercial permeameter to perform systematic steady-state gas-flow experiments using porous samples, and the relationship between flow rate and pressure gradient was determined. They also introduced the Forchheimer equation into volcanology, which is a modified form of Darcy's law that includes the inertial effect of gas flow, and specified the importance of this effect in volcanic degassing processes. Mueller et al. (2005) used gas-pressure decay with time after sudden decompression in a fragmentation bomb for the permeability measurements, without measuring gas-flow rate. A falling head permeameter developed by Burbié and Zinszner (1985) has also been used to measure the permeability of volcanic porous materials (Jouniaux et al. 2000; Bernard et al. 2007). Recently, a low-cost gas permeameter was developed (Takeuchi and Nakashima, 2005) and improved (Takeuchi et al., 2008), to measure permeability of natural samples and experimental products. Finally, electrical conductivity measures how well a material transports electric charge. Rocks, in general, are poor conductors, whereas ionic fluids are good conductors. Therefore a

measurement of conduction through fluid-saturated rocks provides information about the connected pore pathway through the sample. Although the influence of pathway tortuosity and pore shape on permeability is useful for numerical simulations on gas percolation, it has been the object of only a few studies (Table 1).

(iii) Comparison between 2D and 3D particle-texture measurements

Two different methods are currently available for extracting vesicle and crystal sizes, shapes and distributions in pyroclasts. The first is by conversion of 2D data from a planar surface (such as a thin section or photograph) to 3D data through stereology. The second method derives 3D data directly from X-ray tomographic reconstructions and visualization of clast textures without the need of stereological conversions (Song et al. 2001; Shin et al. 2005; Polacci et al. 2006b; 2008; 2009a, b; 2010; Degruyter et al. 2010b; Gualda et al. 2010; Giachetti et al. 2011; Baker et al. 2012), using computer software especially developed for geo-textural purposes (e.g. Ketcham and Carlson 2001; Ketcham 2005; Friese et al. 2013). Other 3D methods include serial sectioning (e.g. Bryon et al. 1995), serial focusing with optical microscope (Manga 1998), serial grinding (e.g., Marschallinger 1998a, b, c; Mock and Jerram 2005), and constructing digital elevation models of individual ash grains to calculate vesicle volume (Proussevitch et al. 2011). Two-D and 3D observations have different limitations and potential, and the two methods are becoming complementary, not competitive (e.g., Giachetti et al. 2011; Baker et al. 2011).

2D method

Standard procedures for the 2D method have been recently published for vesicles (Shea et al. 2010a) and crystals (Higgins 2000; 2006). Two-D techniques can yield high-quality data and account for both vesicle and crystal sizes in the sample and can be applied to particles ranging in size from bombs (e.g., Gurioli et al. 2014, Leduc et al. 2015) to ash (Miwa et al. 2009; 2013; Miwa and Toramaru 2013). These measurements are best used when there is a broad size distribution to be measured. The main limitation of the method is that is based on the assumption of spherical shape of the textural objects, following Sahagian and Proussevitch (1998). When this conversion is simply obtained by dividing the number of vesicles per unit area by the median value of diameter of each size class (Cheng and Lemlich 1983), no shape assumption is made. However, the 3D conversion is more precise when a shape is defined. Empirical corrections are commonly used for crystal analyses (Higgins 2000 and 2006), but

for vesicles, whose shapes are less uniform, they risk introducing systematic, uncontrolled errors in the data (Sahagian and Proussevitch 1998; Proussevitch et al. 2007a; 2007b).

3D method

X-ray-computed microtomography is the only available high-resolution, non-invasive 3D technique that allows reconstruction, visualization and processing of samples. Data acquisition is generally relatively straightforward, and several scales can be examined and combined, ranging from centimeters to <1 micron, depending on the resolution (Giachetti et al. 2011). In addition, the so-called 'local area' tomography technique (e.g., Lak et al. 2008) enables high resolutions to be attained, even with samples larger than the field of view of the camera. However, 3D quantification of textures can also be labor intensive, depending on the size of the volume that needs to be analyzed and on the textural parameters required. The results show the internal structures of samples, highlighting how objects and apertures are linked together. This information provides an excellent suite of data for studies of vesicle size, shape and distribution, collapse, deformation, coalescence, permeability, and tortuosity, as well as for determining crystal volume, size and distribution and visualizing crystal aggregates in 3D (Polacci et al. 2009a, b; 2012; Bai et al. 2010, 2011; Degruyter et al. 2010a, b; Zandomeneghi et al. 2010; Gualda 2010a, b; Baker et al. 2012; Castro et al. 2012; Okumura et al. 2013). Vesicles with complex shapes are easily identified, while in a 2D section they might be interpreted as two or more vesicles, thus biasing vesicle size distribution (VSD) and vesicle number density (N_v) (e.g. Giachetti et al. 2011). The 3D method is particularly effective for determining N_v if the study is focused on a specific size range; vesicle number densities over a wide range of sizes is achieved with nested studies in which a series of scans are done at different sizes and resolutions (e.g. Giachetti et al. 2011; Pardo et al. 2014b). However, the resolution of the reconstruction is still critical. Klug et al. (2002) showed that vesicle walls may be as thin as 0.1μm. To achieve this sort of spatial resolution using tomography requires very small samples. When the attained resolution is 5-15 μm, thin vesicle walls are not resolved.

There is currently no unique protocol for 3D measurements of different types of pyroclastic (or lava) samples; however the SYRMEP group of the Elettra Synchrotron Light Source (Trieste, Italy), together with researchers at McGill University of Montreal and INGV Pisa (M. Polacci), are developing protocols for volcanic samples of different vesicularities and crystallinities.

(iv) Crystal size distribution

Crystal size distribution (CSD) is a well-established tool for interpreting the physical processes and environmental variables that drive differentiation and crystallization in magma chambers and conduits (e.g., Marsh 1988; Cashman and Marsh 1988; Cashman 1992; Hammer et al. 1999; Cashman and McConnell 2005, Armienti 2008; also see references in Table 1). CSD, coupled with vesicle distribution data, yield deeper insights into the physical processes operating in the conduit (e.g., Gurioli et al. 2005; D'Oriano et al. 2005; Piochi et al. 2005; 2008; Noguchi et al. 2006; Giachetti et al. 2010; D'Oriano et al. 2011a; Vinkler et al. 2012). The CSD method has been well tested and widely applied (Table 1), so that it is now quite straightforward to quantify CSD (Higgins 2000; 2006; and references in Table 1).

However, we must keep in mind that crystals are commonly anisotropic, and therefore shape cannot be ignored. Most studies use the Higgins technique to account for shape. However, the Higgins method assumes that all crystals are the same shape. This is clearly not true, as small crystals are often more anisotropic than large crystals. Treating all crystals in the same way can introduce artifacts (see Castro et al. 2003). In addition, there are still resolution issues for microlites, as well as problems in both back-scattered electron (BSE) and cathode ray tube (CRT) analyses when the crystals have a density (Z number) near that of the glass. Several methods can be used to facilitate the extraction and quantification of crystals. CSDs of larger crystals (phenocrysts, antecrysts, etc.) can be measured from transmitted light microscopy images of thin sections and analyzed with digital image analysis to automate and hence speed up the quantification process (e.g., Armienti et al. 1994; Launeau et al. 1994; Lumbreras and Serrat 1996; Goodchild and Fueten 1998; Launeau and Cruden 1998; De Keyser 1999; Heilbronner 2000; Armienti and Tarquini 2002; Boorman et al. 2004). Tarquini and Favalli (2010) used a slide scanner to acquire input imagery in transmitted light from thin sections and GIS software to analyze the data.

Crystals can also be identified using a scanner and a polarizing filter placed at different angles (Pioli et al. 2014). Three pictures are then combined, and their correlation allows the individual grains to be classified by their characteristic orientation. To measure smaller crystals (microphenocrysts and microlites), a scanning electron microscope is commonly used in backscattered electron (BSE) mode (Cashman 1992; Hammer et al. 1999; Cashman and McConnell 2005; Nakamura 2006; Ishibashi and Sato 2007; Salisbury et al. 2008; Blundy and Cashman 2008; Wright et al. 2012). Development of rapid x-ray mapping techniques now allows CSD analysis of x-ray element maps, which provide information on crystal

compositions, textures (crystal size, orientation, shape) and modes of minerals (e.g., Muir et al. 2012; Leduc et al 2015). Another new technique uses an electron backscatter diffraction detector (EBSD) attached to the SEM to obtain crystal orientations, which can provide insights into shearing, accumulation and degassing processes (Prior 1999; Prior et al. 1999; Hammer et al. 2010). Chemical mapping is now routinely and widely used (e.g., Leduc et al. 2015). In contrast, EBSD is more difficult to use and interpreting the data is harder than the chemical maps. As described in the references cited, it produces a wealth of information on various minerals, although the monocline structure of the feldspar can be problematic.

Crystal size distribution can also be obtained directly in 3D via X-ray computed microtomography. Using this approach, it is possible to obtain the total crystal volume, as well as the crystal volume of each mineral phase present: crystallinity, crystal size and crystal shape (e.g., Zandomeneghi et al. 2010; Voltolini et al. 2011). Again, resolution can be a problem. First, crystals may span a large size range, which requires imaging at several different resolutions (e.g., Pamukcu et al. 2010; 2012). Additionally, as in BSE analysis, the compositional similarity between some crystal phases, such as alkali feldspars, and silicic matrix glass can make automated analysis challenging (e.g., Baker et al. 2012). However, excellent results can be obtained by working in phase-contrast tomographic mode (Polacci et al. 2010), and applying a procedure known as phase retrieval to the reconstructed sample volumes (Arzilli et al. 2013).

(v) Errors in particle-texture analyses

Uncertainties in textural analysis are due to several factors. Any textural parameter, such as porosity or crystal size, has intrinsic measurement errors. These are linked to the apparatus used and are generally easy to quantify using standards. A good practice, when a new method is introduced, is to assess its intrinsic error with synthetic samples of well-known particles, having textural components (e.g. crystals, vesicles/voids) with known size and distribution (e.g., see review of Rust and Cashman 2004 for permeability, and Baker et al. 2011 for 3D data from X-ray microtomography). Another type of uncertainty is linked to natural variability, which is generally approached by using the concept of Representative Elementary Volume (REV, Bear 1972). Parameters measured in small, neighboring, regions within a sample have a large variability. As the analyzed regions become larger, this variability decreases until a steady value is reached at the REV size. One complication is that the REV should be significantly smaller than the sample (not guaranteed for ash, or even lapilli,

409 particles), and that some parameters have an REV at the deposit scale, which means that many
410 clasts have to be analyzed. If the sample location is such that eruptive parameters were steady
411 during deposition, application of REV at the deposit scale allows analysis of magma at the
412 point of fragmentation in the conduit. Taking porosity as an example, one 2D SEM image will
413 yield one porosity measurement with a typically small (~1 %) intrinsic error due to
414 thresholding of the grayscale values that represent vesicles. Several 2D images of the same
415 sample taken at different locations and/or different resolutions (larger than the REV) typically
416 yield larger (~10 %) uncertainties that are caused by small-scale spatial heterogeneity. Finally,
417 if we assume that – or if we have a – very well sorted deposit, then the density distribution of
418 all clasts at that location indicates the variability of porosity at the conduit scale, which can be
419 quite large (e.g., Houghton and Wilson 1998). The situation is more complex with poorly
420 sorted deposits in which particles range from bombs through lapilli to ash.
421 Raw data in terms of size (area, long axis, short axis, perimeter) and orientation of crystals
422 and vesicles yield negligible intrinsic errors, because they are computed with programs on 2D
423 binary images with high resolution (>10^6 pixels). In this phase, the uncertainty is due to the
424 image clean-up process, which is generally unquantified (because it takes too long to have
425 four people complete the task independently and then take the average for every image).
426 The greatest source of intrinsic error here is thresholding, which is set by the operator
427 (Baker et al. 2011). When converting 2D data to a 3D projection, however, the error depends
428 on the stereological model used (i.e. particle shapes have to be assumed, Cashman 1988) and
429 is thus harder to estimate.
430 Most 2D textural parameters have well-established techniques and protocols to quantify
431 intrinsic errors, including:
432 • VSD (Toramaru 1990; Mangan et al. 1993; Klug and Cashman 1994; 1996; Klug
433 et al. 2002; Adams et al. 2006b; Shea et al. 2010a),
434 • CSD (Higgins 2006), fabric indicators (Launeau et al. 1990),
435 • vesicle shape (Moitra et al. 2013),
436 • clast shape (Marshall 1987; Capaccioni and Sarocchi 1996; Dellino and Liotino
437 2002; Riley et al. 2003; Ersoy et al. 2006).
438 However, conversion from 2D to 3D distributions introduces errors linked to stereological
439 assumptions. The Cheng and Lemlich (1983) method does not involve assumptions of object
440 shape, but it does not take into account a truncation effect (e.g. Pickering et al. 1995).
441 Truncation is related to the sensitivity of the measurement process; smaller objects are

increasingly difficult to detect. On the other hand, large-scale truncation occurs under several circumstances related to sample-size (volume and area) limitations. The Sahagian and Proussevitch (1998) conversion assumes spherical shapes and corrects for the cut effect (this being the effect induced by rarely cutting a spherical object through its exact center). Giachetti et al. (2011) found that N_v obtained by 2D and 3D methods for the same lapilli agreed within 15 %, and that VSD were also very similar. They recommended the Cheng and Lemlich (1983) method for 2D vesicle analysis, as the Sahagian and Proussevitch (1998) method may generate negative values for some size classes.

In terms of parameters that we can derive from textural analyses, decompression rate is probably one of the most important to quantify due to its implications for eruption dynamics. To achieve this, microlite shape, N_v and size distribution have been used in combination with experimental data for low-mass flux and effusive eruptions (Couch et al. 2003; Cashman and McConnell 2005; Szramek et al. 2006; Clarke et al. 2007; Martel 2012; Wright et al. 2012). Martel et al. (2006) consider this approach to be highly reliable, because different generations of microlites (nucleated pre-eruptively in the reservoir or syn-eruptively in the conduit) can be distinguished on the basis of chemical composition. Decompression rates deduced from N_v (e.g., Toramaru 2006), however, tend to be maximum estimates, because there could be more nucleation events during ascent that add to the signature left by decompression. Maximum decompression rates associated with the final, rapid, stages of ascent could be calculated directly from the smallest bubbles formed during the final fragmentation event (Giachetti et al. 2010; Shea et al. 2011; 2012). Another developing method is to use chemical gradients of volatiles in melt inclusions in crystal embayments to infer rise rates (Ferguson et al. 2013).

However, the relationships between bubble shape, nucleation, coalescence, deformation and/or fragmentation are not well established yet.

Quantification and sampling of pyroclastic deposits for the textural characterization of their components

(i) Preliminary field studies and sampling strategy

Field-based studies of pyroclastic deposits aim to relate both the whole-deposit characteristics (thickness and grain size) and the physical properties of the constituent particles to the eruption conditions. Particle-texture studies are time consuming, especially when they provide complete size distributions of the vesicle and crystal population. For these measurements, the

choice of a limited number of "representative" clasts selected for the analysis is critical, particularly when using these data to model eruption processes and their variability in time and space. Obtaining such clasts requires a cautious sampling strategy with well-defined scientific goals during field work. These studies are best performed only on well-documented deposits, supported by a robust stratigraphic reconstruction and correlation, as well as an accurate compositional stratigraphic framework. When not familiar with the deposit, a preliminary survey at different locations is useful to evaluate the significance of the outcrops used for detailed analysis. Well-defined sublayers (or units) should be identified in the deposit on the basis of clear, unequivocal lithologic and sedimentologic features and cross-correlated over the whole dispersal area of the deposit. Stratigraphic data are critical for placing each studied layer within an appropriate temporal framework within the stratigraphic sequence.

Pyroclasts can be collected after the eruption, from fall deposits of old (unobserved) or recent (observed) eruptions, for which sampling is done preferably within hours to days of the event (e.g., Gurioli et al. 2008; 2013). Sampling may also take place during eruptive activity, with samples collected using sampling device placed inside the fallout field. Three simple collection methods that can be applied to active fallout, as currently used, are: (1) the hand collection method involves collecting (and quenching) bombs or lapilli as they fall out of the plume by people standing in the active fallout field (e.g., Lautze and Houghton, 2007, 2008; Gurioli et al. 2014); (2) the "clean surface" strategy, whereby plastic sheets are laid out close to the vent, or a preexisting surface is cleaned before the eruption. In both cases the pyroclasts falling in a known area are collected (e.g., Rose et al. 2008; Andronico et al. 2009a; 2013a and b; Eychenne et al. 2012; Houghton et al. 2013, Harris et al. 2013b, Schipper et al. 2013); (3) the bucket strategy, in which a large number of buckets are distributed across a discrete area of fallout for a certain period of time (e.g., Yoshimoto et al. 2005; Swanson et al. 2009; Bustillos and Mothes 2010). When possible, the aims are to collect a sufficient number of samples to estimate the magnitude of the event through the mass load per unit area; and to obtain a sufficient number of clasts for chemical and textural characterization. Other promising methods are just coming on-line, such as automatic ash sampling collectors (e.g., Bernard 2013; Shimano et al. 2013, Marchetti et al. 2013).

(ii) Definition of essential, basic physical properties of the deposit to the study

Most particle and deposit texture studies aim at characterizing magma heterogeneity and ascent dynamics, and at understanding the fragmentation process, beginning with the size,

morphology and componentry of the particles (Table 1). Clasts selected for particle-texture analysisare usually sampled in a deposit at a single location (reference section). Lateral variability across the deposit is filtered by transport and sedimentation processes, which primarily depend on eruption intensity, along with related plume dynamics and other dynamic effects such as wind direction and velocity and rainfall. Therefore, clast properties can differ both in time (from the base to the top of a vertical sequence) and in space (from the main axis of dispersal to lateral outcrops at the edge of the fallout zone across the cloud, and from proximal to distal sites). Volcanic plumes (and clouds) are thus complex systems, the properties of which do not vary linearly with the main eruption parameters. They are also affected by external variables, such as wind direction and velocity. The external variables add additional complexity to the clast-type distribution. For this reason, the deposit should be preliminarily characterized at least in terms of stratigraphy, dispersal, thickness variation and volume before more detailed study is initiated (e.g., Fisher and Schmincke 1984; Cas and Wright 1987; Thordarson et al. 2009; Cioni et al. 2011). Estimation of plume height, eruption duration, volume and magma eruption rate can then also be derived for past eruptions from such analyses (e.g., Carey and Sparks 1986; Pyle 1989; Fierstein and Nathenson 1992; Sparks et al. 1997; Bonadonna et al. 1998; Freundt and Rosi 1998; Bonadonna and Costa 2012; Fagents et al. 2013).

(iii) Selecting the outcrop

There are three basic criteria for sample outcrop selection. First: minimize the effect of wind dispersal. Outcrops located along the main dispersal axis are preferred to lateral exposures, unless the effect of wind is the target of study. If wind direction or eruption intensity changes during different phases of the same eruption, it is more appropriate to sample each tephra layer at different ‘equivalent’ locations rather than to collect all samples at a single type outcrop. If sampling is restricted to a single location, the inferred dispersal pattern and distance from the main dispersal axis of each layer should be noted and taken into account when analyzing clast variability among different layers.

A second criterion for selecting the outcrop is that clear textural variations among the juvenile clasts, in terms of color, general morphology, vesicularity, vesicle shape, and crystallinity should be evaluated in the preliminary field survey, so that any lateral and vertical variability within the deposit is already defined following field reconnaissance. This

543 ensures that, when clast types are chosen in the laboratory, the main textural types are easily
544 identified and separated.
545 The third criterion for outcrop selection, if one of the goals of the study is quantification
546 of the proportion of distinct textural clast types, is to remember that sedimentation from the
547 volcanic plume is affected by clast density, shape and size (Bonadonna et al. 1998; Pfeiffer et
548 al. 2005; Barsotti et al. 2008; Eychenne et al. 2013, and references therein). This is especially
549 relevant when a single explosion produces a juvenile population with a wide range of textural
550 and other physical features: their relative proportions within the deposit can vary laterally in
551 the deposit as well as with distance from the vent. Thus, at any single site, the sample is not
552 necessarily representative of the abundance within the eruption mixture. This is especially
553 true in the case of small plumes and mid-intensity eruptions (e.g., Rose et al. 2008; Cioni et
554 al. 2008; 2011; D'Oriano et al. 2011a; Andronico et al. 2013a and references therein). While
555 the textural features of the different clast types can be studied at a single outcrop, the relative
556 proportions between clast types need to be determined across the whole deposit by integrating
557 componentry data on samples collected at outcrops at differing azimuths and distances from
558 the source.
559
560 (iv) Sampling
561
562 After identification of the outcrops where the deposit shows the best and most complete
563 exposure, a suitable approach is random collection of a statistically relevant number of clasts
564 from a single layer. Several techniques can be used, ranging from sieving in the field to find
565 the dominant clast size (for coarse clasts), or sampling the bulk deposit for later clast selection
566 in the laboratory (for small clasts). In the case of fine-grained deposits, it can be useful to
567 apply sampling techniques that preserve structural and textural characteristics of the whole
568 deposit. Samples can be retrieved using tubes or boxes manually pressed into the deposits, or
569 carfully carved out and surround-wrapped deposit blocks. In situ and/or laboratory
570 impregnation techniques of deposits exist for a broad range of grain sizes and compositions
571 (Bouma 1969), some of which are applicable to fragile or loose volcanic deposits (e.g. Fiske
572 et al 2009). The applicability of such techniques to fine-to-medium grained volcanic deposits
573 should be tested, since they would allow both 2D (e.g., X-ray radiography and thin section
574 analysis) and 3D analysis (X-ray tomography and anisotropy of magnetic susceptibility) to be
575 applied to deposits, rather than single pyroclasts; these, techniques are frequently used for
576 hard rocks (e.g., Lanza and Meloni 2006).

577 The number of samples should be defined depending on the purpose of the study.
578 Fixing the number of samples per stratigraphic layer based on the layer characteristics (e.g.,
579 extent of zoning/fluctuations in grain-size, componentry, etc.) for characterizing eruption
580 dynamics, or focusing on the layer thickness for conduit dynamic characterizations, are two
581 examples of such pre-selection decisions. Before selecting clasts, basic grain-size studies
582 (when the bulk deposit is collected) on each sampled layer (median and sorting of grain-size
583 distribution) and componentry analysis should be carried out to ensure effective sub-sampling
584 for textural studies. Following White and Houghton (2006) componentry analysis is the
585 subdivision of the sample into three broad components: juvenile, non-juvenile particles and
586 composite clasts. The juvenile components are vesicular or dense fragments, as well as
587 crystals, that represent the primary magma involved in the eruption; non-juvenile material
588 includes accessory and accidental fragments, as well as crystals, that predate the eruption
589 from which they are deposited. Finally, the composite clasts are mechanical mixtures of
590 juvenile and non-juvenile (and/or recycled juvenile) clasts. More-detailed componentry can
591 subdivide the juvenile and non-juvenile materials into subpopulations that have important
592 dynamic meanings (e.g., Eychenne et al. 2015).
593 Finally, after choosing the size intervals of the clasts for physical and textural
594 measurements (i.e., bulk and solid density, vesicularity, microtextures, permeability), it is
595 useful to compare the grain-size distribution of each interval with the total grain-size
596 distribution of the sampled layers, especially when the grain-size distribution is highly
597 variable within the sampled stratigraphy. This strategy allows checking of sample
598 representativeness. For example, sampling may be from (i) bimodal or complex multimodal
599 distributions, or (ii) anomalous, poorly sorted deposits. In the second case, sampling should
600 avoid features that can be indicative of contamination from other sources, such as ballistic
601 components, elutriated ash from pyroclastic density currents or from reworking (e.g., Fierstein
602 et al. 1997, Eychenne et al. 2012). It is useful, whenever possible, to show variance, or
603 invariance, of the textural features by comparing data collected in the selected size class with
604 textural data for different size classes. This should, at least, be carried out for a few selected
605 samples.
606
607 **How to link petrological, geochemistry and textural quantifications**
608
609 (1) Initial parameters and conduit processes
610

611 Geochemical and petrological analysis of pyroclastic products can constrain the initial
612 conditions in the shallow crustal holding chamber through to the surface via the conduit
613 system (Fig. 1). In transit through this system, the textural features are imprinted on the
614 pyroclasts quenched upon eruption. The geochemical and petrologic analysis can help:

615 • Define pre-eruptive P-T storage conditions from mineral-melt equilibria or
616 disequilibria (e.g., Rutherford et al. 1985; Scaillet and Evans 1999; Pichavant et al.
617 2002; Blundy and Cashman 2008; Schipper et al. 2010b);
618 • Assess initial viscosity, temperature, melt composition and volatile budget, including
619 input of gases from deeper sources (e.g., Wallace 2001; Blundy and Cashman 2008;
620 Métrich et al. 2010);
621 • Define the evolution of volatile contents (specifically CI, F, S, H_2O, CO_2) using
622 electron probe, ion probe (SIMS), Raman and FTIR in melt inclusions and host
623 minerals, while combining results with vesiculation studies and gas release
624 measurements (e.g., Wallace 2005; Métrich and Wallace 2008, Schipper et al. 2010c).
625 In such a way we can determine whether the magma was saturated, over-saturated or
626 under-saturated at a certain depth, and how these conditions affect vesiculation in the
627 conduit (e.g., Anderson 1991; Hurwitz and Navon 1994; Dixon 1997; Roggensack et
628 al. 1997; Schipper et al. 2012);
629 • Measure residual volatiles in glasses and bulk-rock samples to reveal how degassed
630 the magma is (Newman et al. 1988; Villemant and Boudon 1998; Shea et al. 2014);
631 • Provide variable diffusion of stable elements (^{6}Li, ^{7}Li, H/D, ^{10}B, ^{11}B) or radiogenic
632 isotopes (^{210}Pb-^{226}Ra), which are used as tracers for melt degassing and interaction
633 with hydrothermal fluids (e.g., Berlo et al. 2004; Kent et al. 2007; Humphreys et al.
634 2008b; Schiavi et al. 2010; Berlo and Turner 2010; Vlastélic et al. 2011);
635 • Measure mineral diffusion profiles and derive pre-eruptive residence times, ascent
636 rates and cooling rates (e.g., Kahl et al. 2011);
637 • Provide crystal shapes, zoning schemes, and dissolution stages, while determining
638 which magmatic process and physical parameters control crystal shape/zoning (e.g.,
639 Hammer and Rutherford 2002; Rutherford and Devine 2003; Blundy et al. 2006; Costa
640 et al. 2008; Streck 2008);

641 In addition, laboratory petrological investigations can provide:

- Experimental observations on phase equilibria (mineral-melt-vapor), crystallization paths and liquid line of descent (e.g., Hammer and Rutherford 2002; Couch et al. 2003; Blundy et al. 2006; Hammer 2008);
- Calibration of decompression rates. While this has been carried out for rhyolitic systems (e.g., Mourtada-Bonnefoi and Laporte 2002, 2004; Mangan and Sisson 2005; Gardner 2007; Cichy et al. 2011; Cluzel et al. 2008) and phonolitic systems (e.g., Larsen 2008; Shea et al. 2010b), there are ongoing studies on basaltic systems (Bai et al. 2008; Lesne et al. 2011; Pichavant et al. 2013);
- Diffusion coefficients of relevant chemical elements, including volatiles, to improve kinetic modeling (Dohmen et al. 2007; Chakraborty 2008);
- Relationships between crystal morphologies, cooling rates and degree of undercooling (e.g., growth of crystals with hopper and swallow tail shapes experiencing rapid late-stage crystallization; Faure et al. 2003, 2007);
- Surface flux of volatiles (i.e. what leaves the system; see reviews by Fischer 2008; Pyle and Mather 2009) compared with melt inclusion data (i.e. what is in the system initially; e.g., Le Voyer et al. 2010; Rose-Koga et al. 2012; Schiavi et al. 2012).

2) Where geochemistry can help textural study

Measurements of volatile contents in quenched, phenocryst-hosted melt inclusions provide estimates of initial (shallow crustal) values (e.g., Kent 2008). These are minimum estimates, because H_2O can leak from melt inclusions during ascent by intracrystalline diffusion as the far-field environment of the crystal evolves (Chen et al. 2011, 2013). Melt inclusion volatile contents can be inverted to equivalent saturation pressures using multi-species (e.g., H_2O-CO_2; H_2O-Cl) solubility laws (using, for example, VOLATILCALC, Newman and Lowenstern 2002; MELTS, Ghiorso and Sack 1995; Asimow and Ghiorso 1998). These, in turn, can be used to calculate total pressures (and hence depth) by assuming volatile saturation, or minimum pressures if the sample is under-saturated in volatiles. Progressive closure of melt inclusion networks in growing phenocrysts can result in zone-dependent melt inclusion volatile contents that record the evolution of pressure conditions as magmas migrate from depth (Blundy and Cashman 2008, and references therein). Combining major element and volatile compositions of the melt with phenocryst contents allows calculation of initial magma physical properties (viscosity, density, surface tension, and others). Derivations of

675 such parameters are necessary for modeling magma ascent, vesiculation and groundmass
676 crystallization.
677 Pre-ascent storage conditions can also be inferred from phase-equilibria studies of natural
678 compositions. Comparison of natural and experimental phase abundances and compositions,
679 combined with constraints of volatile content (from melt inclusions) and temperature (from
680 e.g., Fe-Ti oxides) allows estimation of total pressure if the degree of volatile saturation is
681 established through use of mixed-volatile experiments (Pichavant et al. 2007; Cadoux et al.
682 2014).
683 Residual volatile content (e.g., H_2O, CO_2, SO_2, Cl, F) measured in the glass or directly
684 from gases emitted at the vent can be correlated with textures (e.g., Piochi et al. 2005; 2008;
685 Schipper et al. 2010a; Balcone-Boissard et al. 2011, 2012; Shea et al. 2012, 2014; Burton et
686 al. 2007; Polacci et al. 2009b; Miwa and Toramaru 2013). The residual volatile contents can
687 also be compared with pre-eruptive volatile contents obtained from melt inclusion to evaluate
688 both the extent and efficiency of syn-eruptive degassing (e.g., Shimano and Nakada 2006;
689 Noguchi et al. 2006, Métrich et al. 2001; 2010). Residual water content or Cl content (when
690 Cl partitions into a H_2O vapor phase, so that it can thus be used as an indicator of degassing
691 processes; Balcone-Boissard et al. 2010) is typically plotted against V_g/V_l, where V_g is the
692 volume of vesicles corrected for phenocrysts and V_l is the volume of melt and microlites
693 (Villemant and Boudon 1998, Balcone-Boissard et al. 2011, 2012). An important issue is to
694 assess the extent of post-eruption hydration. Recently, thermal gravimetric studies have
695 proved to be quite effective in allowing this correction based on oxygen or hydrogen isotopic
696 compositions (e.g., Giachetti and Gonnerman 2013; Shea et al. 2014). Studies of hydrogen
697 isotopes, correlated with SEM glass textures, permit distinction of magmatic water from
698 meteoric water generated by re-hydratation (Kyser and O'Neil 1984). Hydration can also be
699 assessed from the ratio of water species (molecular H_2O vs. OH) in residual glass, as
700 determined by FTIR data or Raman analyses (Hammer et al. 1999; Le Losq et al. 2012).
701 Ascent and decompression in the conduit can result in chemical changes that can be
702 quantified by a range of microbeam analytical techniques (e.g., EPMA, LA-ICPMS, FTIR, μ-
703 Raman). As the pressure drops, H_2O will migrate out of melt inclusions and crystals (Le
704 Voyer et al 2010; Hamada et al 2010), and light elements (Li, B) will try to re-establish
705 equilibrium between crystals, host melt and any vapor or brine phase present (Berlo et al.
706 2004). At the same time, H_2O and CO_2 migrating out of melt inclusions will become apparent
707 as re-entrant tubes at the edges of crystals (Liu et al. 2007; Humphreys et al. 2008a). Each of
708 these processes will establish diffusive gradients frozen into the pyroclast that can be

709 measured and modelled using experimentally determined kinetic laws to infer decompression
710 rates during ascent (e.g. Gonnermann and Manga 2013). These decompression rates can then
711 be compared with values derived from other approaches, including those based on analyses of
712 microlite sizes and shapes, vesicle number densities, and hornblende-breakdown reactions
713 (e.g., Martel 2012; Cluzel et al. 2008; Giachetti et al 2010; Shea et al. 2011).
714
715 3) Contentious points
716
717 Care needs to be taken when converting decompression rate to magma ascent rate, and
718 especially when comparing decompression rates obtained using different methods. Pressure
719 gradients in conduits are highly nonlinear due to the strong effect of dissolved H_2O on magma
720 viscosity, particularly at low H_2O contents (Gonnermann and Manga 2013). Moreover,
721 different processes will likely record different decompression rates, according to the time
722 available for the process to take place. For example, microlite growth is relatively slow, so
723 that microlite size and shape distributions are likely to record an average decompression rate
724 during ascent (Martel 2012). Bubble nucleation and growth, on the other hand, can occur very
725 rapidly, so that N_v may record just the peak decompression rate immediately beneath the
726 fragmentation zone (Cluzel et al. 2008; Giachetti et al 2010). Comparison of rate calculations
727 from different methods therefore requires caution. However, integration of decompression
728 rates as obtained from different textural and chemical characterizations, when combined with
729 mass eruption rate estimation from deposit analysis or direct observations, can provide
730 quantitative insights into the processes involved in magma ascent from the deep source to the
731 surface.
732 Another outstanding issue is the role of dense clasts. That is, did they originate (i) from
733 magma quenched at depth prior vesiculation, (ii) by vesicle collapse in an originally vesicular
734 clast, (iii) from volatile-poor magma or from reciclying? It is important to provide a correct
735 interpretation, because the three conclusions relate to very different mechanisms. In several
736 eruptions it has been found that the densest clasts were depleted in water through syneruptive
737 bubble collapse and coalescence (Rust and Cashman 2007; Piochi et al. 2008, Shea et al.
738 2014). In Plinian eruptions at Vesuvius (Pompeii and Avellino), the densest clasts have been
739 interpreted as magma that lost water during transition from closed-to open-system degassing
740 (Balcone-Boissard et al. 2011; 2012). Water depletion can also result from syn-eruptive
741 processes, such as clast recycling at magmatic temperature (Gurioli et al. 2014) and intrinsic
742 magmatic redox conditions, as shown by the experiments of D'Oriano et al. (2012). No study

743 has yet have demonstrated that dense clasts retain all of their original gas and were quenched
744 at great pressure.
745 Another key question is whether the measured compositions (including volatile content)
746 of bulk rock, glass, or minerals represent equilibrium or disequilibrium processes, and if
747 equilibrium or disequilibrium conditions pertain to local subsystems or to the whole magmatic
748 body under investigation (see for example Pichavant et al. 2007). Chemical species with
749 different diffusivities, for example, record equilibrium or non-equilibrium conditions in the
750 same sample (e.g. De Campos et al. 2008; Schipper et al. 2012). Equilibrium kinetics is also
751 composition-dependent, because it is dictated in part by melt viscosity which is itself related
752 to viscosity. This issue will generally affect silicic to intermediate magmas more than basaltic
753 magmas. However, we note that even for basaltic systems crystal-fluid-bubble magma
754 mixtures can achieve apparent viscosities that range over six orders of magnitude, up to 10^6
755 Pa s (e.g., Gurioli et al. 2014), depending on the degree of cooling, degassing and
756 crystallization. Such rheological variation even within a single composition, and its effect on
757 eruption mechanisms, deserves increased attention.
758
759 **How to link the geophysical data with pyroclast textural quantification**
760
761 A wide array of remote sensing and geophysical approaches can be used to parameterize an
762 explosive event, both within and outside the conduit (Fig. 1). Geophysical signals are
763 generated by fluid and gas flow in the magma-filled part of the conduit and during
764 fragmentation. Magma-gas ascent dynamics and conduit conditions extracted from
765 geophysical data for this part of the system are particularly difficult to validate because the
766 system cannot be directly observed. They are thus effectively "invisible" to direct observation.
767 Measurements outside the conduit can be made of the emitted mixture of gas and particles as
768 it (i) exits the vent, (ii) ascends above the vent as a plume, and then (iii) drifts away from the
769 vent as the cloud. Models and dynamic parameters extracted for geophysical and remote
770 sensing data outside the conduit are a little easier to validate because they can be directly
771 observed.
772 The invisible part of the system is the realm of studies using seismic, pressure
773 (infrasonic), and deformation data. All three data sets have long been shown capable of
774 detecting the geophysical signature of explosive events spanning weakly explosive Hawaiian-
775 to-Strombolian through Plinian events. Seismic data sets are available, for example, for gas-
776 pistoning events, puffing, fountains, and strombolian eruptions at mafic systems (e.g.,

777 Goldstein and Chouet 1994; Ripepe et al. 1996; Sciotto et al. 2011; Ripepe and Braun 1994),
778 as well as for events that generate somewhat larger plumes during silicic eruptions, as at
779 Santiaguito, Soufriere Hills, and Redoubt. Associated pressure impulses (typically recorded
780 by infrasound and barometers) have long been recorded for such energetic events, famous
781 examples including the pressure response to the 1883 eruption of Krakatoa and the 1967
782 caldera-forming eruption of Fernandina (Simkin and Howard 1970). Magma-gas ascent has
783 also been shown to generate rapid, but recordable, deformation signals detected by tiltmeters
784 (Aoyama and Oshima 2008; Genco and Ripepe 2010; Iguchi et al. 2008; Zobin et al 2007).
785 Velocities, masses and size distributions of particles leaving the vent have typically been
786 measured by visible and thermal video (e.g., Chouet et al. 1974; Ripepe et al. 1993; Harris et
787 al. 2012; Delle Donne and Ripepe 2012; Taddeucci et al. 2012; Bombrun et al. 2014; Gaudin
788 et al., 2014a, b) and Doppler radar (e.g., Dubosclard et al 1999; Hort and Seyfried 1998; Vöge
789 et al 2005; Gouhier and Donnadieu 2008; 2011; Gerst et al 2013). Infrasonic array methods
790 are also available to locate the emission in x,y space (Ripepe and Marchetti 2002). Plume
791 front velocities, density and entrainment rates have also been successfully tracked using
792 visible and thermal cameras, as well as radiometers, for a few stronger, ash-rich, buoyant
793 plumes at Stromboli, Santiaguito and Eyjafjallajökull (Patrick 2007; Sahetapy-Engel and
794 Harris 2009; Bjornsson et al. 2013; Valade et al. 2014); See Chapter 9 of Harris (2013) for
795 review.
796 Satellite remote sensing has long been used to track and measure properties of the
797 eruption cloud as it drifts and disperses. These data are available for all cloud sizes, from
798 those associated with small Strombolian and fountaining events (e.g., Heiken and Pitts 1975;
799 Dehn et al. 2000; 2002) to sub-Plinian and Plinian events (e.g., Holasek and Self 1995;
800 Koyaguchi and Tokuno 1993; Holasek et al. 1996). Cloud dispersion dynamics are especially
801 well revealed by geostationary satellite data with nominal imaging of one image every 15
802 minutes and higher. Basic cloud properties that can be measured by satellite data include
803 cloud dimensions, drift velocity and height (e.g., Robock and Matson 1982; Denniss et al.
804 1998; Aloisi et al. 2002; Zakšek et al 2013). Prata (1989) and Wen and Rose (1994)
805 introduced a method to potentially extract particle size distribution and mass from “split
806 window” (10-12 μm) thermal data. While specially modified ground-based thermal cameras
807 were adapted to extract ash particle size and plume mass (Prata and Bernardo 2009), newly
808 available technology such as LiDAR and PLUDIX were shown of value in detecting, tracking
809 and measuring fine particles in the Eyjafjallajökull cloud (e.g., Bonadonna et al. 2011).
810 Disdrometers and ash collectors, however, currently show greater potential for measuring

811 particle size and terminal velocity (Marchetti et al. 2013; Shimano et al. 2013) than PLUDIX,
812 which was designed more for meteorological applications (Caracciolo et al. 2006; Prodi et al.
813 2011).

814 For the gas content of the cloud, satellite-based sensors such as TOMS, AIRS, OMI,
815 MODIS, GOME and IASI have been used to obtain the SO_2 content in the far field, once the
816 gas cloud has decoupled from the ash cloud (e.g., Krueger et al. 1990; Carn et al. 2003; 2005;
817 Watson et al. 2004; Yang et al. 2007; Thomas et al. 2011; Rix et al. 2012; Walker et al. 2012).
818 Ground-based sensors, such as COSPEC, FLYSPEC and DOAS (e.g., Caltabiano et al. 1994;
819 Horton et al. 2005; Oppenheimer et al. 2011), have been used to measure SO_2 fluxes relatively
820 close to the source; see Williams-Jones et al (2008) for full review. These approaches have
821 been recently supplemented by SO_2 camera systems, which allow 2D images of SO_2
822 concentrations to be collected at ~1 Hz rates (Mori and Burton 2006). Such studies have,
823 though, tended to focus on passive degassing and gas-puffing systems, because the presence
824 of ash interferes with UV-light transmission on which the technique relies, making
825 measurements problematic. Recently, SO_2 cameras have been used to measure the gas masses
826 and fluxes involved in discrete explosive events (Mori and Burton 2009; Holland et al 2011;
827 Barnie et al. 2014).

828 However, none of these remote sensing techniques directly collects or makes contact with
829 the magma or particles they measure. Thus the need exists for quality ground-truth data to
830 validate particle velocities and sizes extracted from what is, basically, an electronic response,
831 as well as to test the assumptions and models used to convert received "power" to a more
832 meaningful and useful parameter (such as mass). At the same time, any single data set can be
833 inverted to support a conduit or plume dynamic model; but results need to fall within
834 constraints provided by ground-truth data. In this case, ground truth is provided by analyses of
835 the magma and particles themselves to extract parameters such as magma temperature,
836 chemistry, density, crystallinity and vesicle content, as well as vesicle shape and size and
837 particle density, size, shape and roughness. Magma ascent, explosion-source and
838 fragmentation models based on geophysical data likewise need to be consistent with
839 independent measurements made for physical volcanology for the same processes if they are
840 to be valid. We explore below these needs, mostly focusing on weakly explosive, basaltic
841 cases, the usual targets because they provide a reliable and easy-to-measure source for testing
842 new technology, methods and algorithms for ground-based geophysical enquiry.
843

844 *The basic need: Realistic assumptions and validation*

The basic response of a remote sensing instrument is a voltage which, through calibration, can be converted at higher level physical value, such as spectral radiant intensity or power. The conversion of this value to higher level and more volcanologically useful parameters (such as particle-size distribution, mass flux or plume density) requires an increasingly complex system of assumption stacking. Thus, to adequately reduce geophysical data, a number of input parameters are required and many assumptions need to be made, all of which can be provided by the physical volcanological community. Data sets from this community, especially if provided simultaneously with geophysical data collection during an active event, or provided as a library typical of that event, can also be used to "ground truth" or check the precision and reality of the geophysically applied input or generated output.

Seismic signals that accompany explosions are primarily short period (SP; high frequency > 1 Hz) signals which are typically termed "explosion quakes". These usually have high amplitudes and mostly include frequencies up to a few hertz, with a possible higher frequency acoustic phase (McNutt 1986, Mori et al. 1989, Braun and Ripepe, 1993). Below these frequencies, short period (SP) signals are often hidden by very-long period (VLP) components with much lower amplitudes (Neuberg et al. 1994; Kaneshima et al. 1996). In-spite of an enormous amount of work, it remains unclear as to how we can explain the VLP seismic component, which itself is only one part of the seismic signal. It also remains unclear as to whether, and/or how, SP and VLP components are related to the magnitude and intensity of an explosion, although attempts have been made using tremor (Brodsky et al. 1999; Nishimura and McNutt 2008; Prejean and Brodsky 2011). Clearly, better coupling with the physical volcanology community could help narrow down much uncertainty and allow progress towards better models to untangle the seismic signal associated with discrete explosive events.

Delay times in the arrival of seismic, infrasonic and thermal signals have been commonly used to assess the depth at which various physical processes occur in explosive basaltic systems (e.g., Ripepe and Braun 1994; Ripepe et al. 2001; 2002; Harris and Ripepe 2007). However, the sound speed in the conduit needs to be assumed if, for example, the thermal-infrasound delay is to be used to obtain the fragmentation depth. This will vary strongly with conditions in the empty portion of the conduit, including mixture density, gas-to-particle ratio, and temperature of the mixture through which the sound is propagating. Thus we need to know these variables if we are to provide a realistic sound speed value and hence infer a plausible depth. We thus need to constrain two fundamental parameters to strengthen geophysical modelling of the shallow explosion mechanism and depth. First, the magma crystal and bubble content (as well as size, shape and distribution), plus fluid chemistry and

879 temperature, are needed to define magma rheology properties and bubble ascent dynamics.
880 Second, the exact proportions and character of the mixture of gas and particles that ascends
881 the final section of the conduit to exit the vent and feed the emission must be known.

882 Velocities, mass fluxes and particle size distributions (PSDs) for lapilli through bomb-size
883 particles have been derived from high spatial and temporal resolution video data obtained
884 using both near-infrared and thermal cameras (Chouet et al. 1974; Ripepe et al. 1993; Harris
885 et al. 2012; Delle Donne and Ripepe 2012; Bombrun et al. 2014).

886 Generally, these studies have focused on Stromboli. For such camera data, the lower limit
887 of a particle size that can be extracted is limited by pixel size. This is typically about one
888 centimeter in dimension, depending on the detector's instantaneous field of view and distance
889 to the target (Harris 2013). A pixel-mixture model can be applied to obtain the size of a sub-
890 pixel particle, but it needs to assume a temperature for the particle and then uses the pixel-
891 integrated temperature to solve for the pixel portion occupied by that particle (Harris et al.
892 2013a). Symmetry then needs to be assumed to convert from particle area to particle volume,
893 and a density needs to be assumed to derive particle mass (Bombrun et al. 2014). For ash-rich
894 plumes, methods have been applied to extract total plume mass and air entrainment properties
895 from ascent dynamics of buoyant thermals (Wilson and Self 1980; Patrick 2007; Valade et al.
896 2014). However, all methods need particle shape, particle density, plume density and/or size
897 distribution data to: (i) determine whether the input assumptions are valid; and (ii) ground
898 truth the remote-sensing-data-derived size and mass data (Harris et al. 2013a and b). The
899 advantage is, if a validated method can be developed, particle size distribution, mass and mass
900 flux data for the plume leaving the vent can potentially be provided multiple times per second
901 using camera data (e.g., Taddeucci et al. 2012; Bombrun et al. 2014).

902 Deducing the erupted mass from Doppler radar data requires the assumption of a particle
903 size distribution for the eruption. Because this distribution is unknown, an average particle
904 size can be constrained from the Doppler radar measurement, typically using the eruption
905 velocities themselves deduced from either terminal fall velocities (Hort et al 2003) or by
906 discriminating between lapilli (larger than a few millimeters or 1 cm, depending on the radar
907 wavelength) and fine ash particles (<1 mm) using their temporal velocity evolution (Valade
908 and Donnadieu 2011). Both methods can be used to obtain an estimate for the erupted mass of
909 ballistics. We thus need to know whether the constrained average particle size can be used for
910 mass calculation, whether the assumption is a good approximation, and what the difference
911 between the derived value and true value is.

912 The radar is able to measure particles of all sizes, provided enough particles are available
913 to return a signal. The relationship between particle size and number of particles required for
914 a signal that exceeds the noise level is not linear, however. It also depends on the radar
915 wavelength and the distance between the radar and target. The smaller the radar wavelength,
916 and/or the smaller the distance between the radar and target, the smaller the number of fine
917 particles needed for a return signal. For particles <1 mm, halving the particle size increases
918 the number of required particles by a factor of 64. Doubling the size of particles to >1 cm
919 means that only ¼ of the number of particles is needed to return the same signal amplitude. In
920 addition, radar can measure at points (gates) across the entire plume thickness. Currently,
921 radar's best role is to provide radial velocity measurements, with well-stated limits as to the
922 particle size to which these data relate, through the entire plume thickness.
923
924 *Questions, points and issues*
925 The main question from the geophysical community to the textural community is: "*What*
926 *does the magma look like at the point of fragmentation?*" Geophysical analysts need to know
927 everything possible about the fragments physically in order to reduce and model the data
928 correctly. To help with this, we concluded that:
929 - Measurements of basic geophysical parameters (such as seismic energy, acoustic
930 energy, energy partitioning, spectral radiance, and radar power) are the most
931 straightforward to consider for correlations with parameters derived from physical
932 volcanology.
933 - Multi-disciplinary correlations lead to improved understanding of explosion dynamics,
934 and only a complete set of measurements can enable a complete and well-constrained
935 understanding of the system (e.g., Gurioli et al. 2013; 2014; Leduc et al 2015).
936 - A wealth of textural and geophysical data exist for Strombolian events, and some data
937 for larger events. They have been used to define the characteristic geophysical and
938 textural signatures that allow distinguishing each event type (e.g. Patrick et al. 2007;
939 Leduc et al 2015). Focus on such relatively low energy events is appropriate, because
940 they are frequent and approachable (Harris and Ripepe 2007).
941 - There is an unfortunate, but understandable, lack of multi-disciplinary data for larger
942 (Vulcanian-to-Plinian) events; because they are rare. With multi-disciplinary
943 approaches becoming more routine, this situation is improving.
944 Thermal and SO_2 sensor arrays are becoming increasingly common components of permanent
945 monitoring arrays at many persistently active sites (Harris 2013). However, such technology

will probably never be installed on every potentially active volcano, all of which give seismic and pressure signals detectable by distant stations. From a practical point of view, it is more realistic to push forward with operational correlations between seismic-infrasonic metrics and deposit and particle textural deliverables to understand the ongoing progression of global volcanic events. In doing so, we must remember that many geophysical signals tend to be time-averages (e.g., tremor amplitude). We need to consider, however, geophysical measurements that describe single, discrete explosions if we are to reasonably compare the data with textural variations between many individual emission events, or emission phases, that characterize the total eruption total energy is one prime example (e.g., Marchetti et al 2009).

We are at an exciting point in our ability to track and understand explosive volcanic emissions through true cross-disciplinary integration of deposit, geochemical, textural and geophysical data. Studies are increasingly bringing together multiple approaches in the field (e.g., Rosi et al. 2006), in the laboratory (Clarke et al. 2009), at large-scale experiments (Sonder et al. 2013) and during field deployments (Harris et al. 2013b). As a community, we appear to be converging on the correct, multi-disciplinary approach. We are at the beginning of a new age, one which links particle texture to seismology (Miwa et al. 2009; Miwa and Toramaru 2013; Gurioli et al. 2014; Leduc et al. 2015) and infrasound (Colò et al. 2010, Landi et al. 2011); as well as petrology to geophysics (Saunders et al. 2012; Martí et al. 2013). Continuation of this trajectory can be aided by further support for pan-disciplinary workshops, meetings and working groups, the objectives of which are to totally understand the system and to constrain measurements with the least uncertainty.

Questions, needs and recommendations

Tables 2–4 summarize the main results from the previous discussions. Table 2 is the summary of major conclusions to date from cross-disciplinary approaches. Table 3 suggests improvements to methods to facilitate cross-disciplinary approaches. Finally, Table 4 groups outstanding questions that might be addressed if the recommended methods are used.

The list of key issues and questions defined allows us to distil the following community-wide points and initiatives as priorities:

1. We need to define, and adhere to, standard sampling, data collection, experimental and methodological procedures to allow full integration of the four disciplines;

2. In doing this, we need to understand each other's needs, and then follow each other's well-recognized sampling etiquette in order to work together as a truly integrated team;
3. We should aim to collate all data and measurements that can be provided by each discipline at some central host site and evaluate whether we need more from each field;
4. Quantification and statement of the precision of the measurements must always be made, and a set of standards must be produced to allow data quality control has to be followed;
5. The community needs to explore and discuss the best means to improve the quality of the measurements and the amount of data available;
6. Guidelines should be agreed on regarding essential key parameters that need to be extracted, versus those that are less important. Common standards need to be established that allow these key parameters to be shared by all groups;
7. Central to this is creation of an open access data bank to support essential geophysical, deposit, textural and geochemical data integration and sharing. This means creation of a repository of data grouped by eruptive style and/or geographic location into which members can make deposits and withdrawals;
8. All of this should ideally be integrated into a GIS platform to allow for easy cross-correlation and comparison of different types of parameters.

DynVolc: an integrated database

Inspired by this effort, a database – DynVolc (Dynamics of Volcanoes) – is now operative at http://wwwobs.univ-bpclermont.fr/SO/televolc/dynvolc/index.php. This database is part of an observation system within the services provided by Observatoire de Physique du Globe de Clermont-Ferrand (OPGC). It is an attempt to provide an integrated and accessible library for all multi-disciplinary data sets for explosive eruptive events. This database is an integrated collection of data from physical and geophysical observations of dynamic volcanic processes.

DynVolc database spans the full range of explosive and effusive activity. Its intent is to provide a library of standards for eruptive styles, for each of which the data base provides:

- field data (i.e., results of field mapping, outcrop and sample descriptions)
- key deposit features (thickness, areal dispersion, sedimentary structure, grain size)
- clast characterization (componentry, morphology, density, porosity, permeability)

- clast texture (connectivity, vesicle and crystal size and size distributions)
- chemical analyses of samples (bulk and glass chemistry
- associated geophysical measurements (e.g., fragmentation depth, ejection and ascent velocity, fragment and gas mass, seismic and acoustic energies)

Integration of these data allows improved, better constrained, insights into the dynamics driving each eruptive style. It also allows improved definition of the rheological and degassing conditions associated with each activity style. At the same time it provides a library of key physical parameters that need to be assumed by geophysical data reduction methods, as well as during model-based enquiry.

Central to this initiative will be the transformation of this database into a communal databank, involving a web-based GIS platform to allow huge amounts of cross-correlationand comparison between parameters relating to different processes and cross-correlation of different datasets obtained for the same eruption. It is intended as an open database into which anyone can input, and withdraw, citable cross-disciplinary information for scientific analysis. At the same time, through this library, we can provide cross-community time series, baseline and monitoring data for the full range volcanic activity.

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Figure captions

Figure 1 Diagrammatic illustration showing a volcanic strombolian conduit (modified from Harris and Ripepe 2007), and the list of a few parameters that can be measuremed trought the deposit (D), the texture of the pyroclasts (T), the geochemistry (G) and the geophysics (G) methods for small, magmatic explosions.

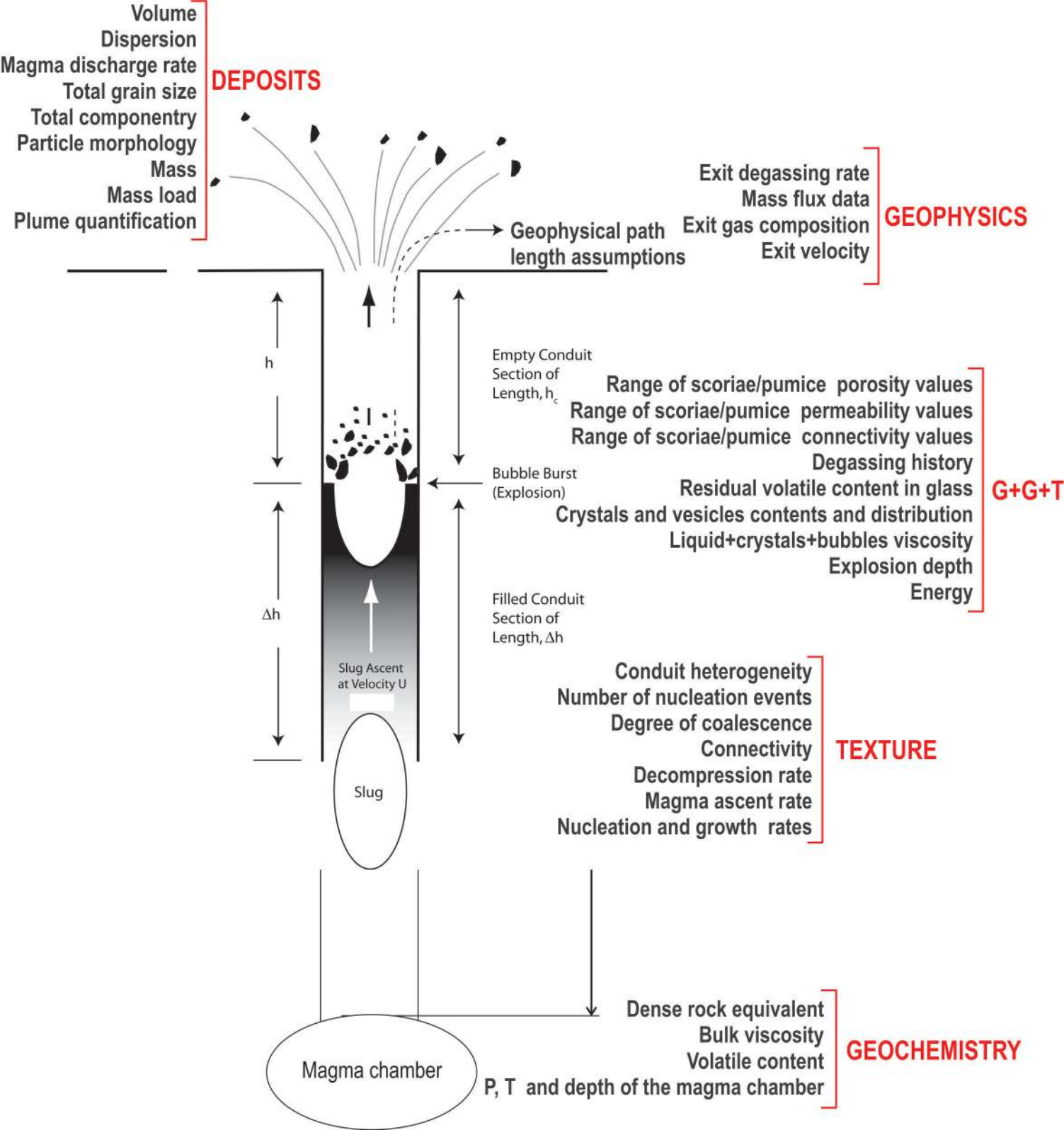
Volume
Dispersion
Magma discharge rate
Total grain size
Total componentry
Particle morphology
Mass
Mass load
Plume quantification
DEPOSITS
Geophysical path length assumptions
Exit degassing rate
Mass flux data
Exit gas composition
Exit velocity
GEOPHYSICS
h
Empty Conduit Section of Length, h_c
Range of scoriae/pumice porosity values
Range of scoriae/pumice permeability values
Range of scoriae/pumice connectivity values
Degassing history
Residual volatile content in glass
Crystals and vesicles contents and distribution
Liquid+crystals+bubbles viscosity
Explosion depth
Energy
G+G+T
Bubble Burst (Explosion)
Δh
Filled Conduit Section of Length, Δh
Slug Ascent at Velocity U
Slug
Conduit heterogeneity
Number of nucleation events
Degree of coalescence
Connectivity
Decompression rate
Magma ascent rate
Nucleation and growth rates
TEXTURE
Magma chamber
Dense rock equivalent
Bulk viscosity
Volatile content
P, T and depth of the magma chamber
GEOCHEMISTRY

Table 1 Quantification of explosive dynamics from textural parameters of the pyroclast components

Textural parameters	*Quantification*		*References*
Clast shape, morphology and size	Discriminate between different fragmentation mechanisms		Wohletz 1983; 1986; 1987; Heiken and Wohletz 1985; Sheridan and Marshall 1983; 1987; Dellino and La Volpe 1996a; b; Palladino and Taddeucci 1998; Büttner et al. 1999; 2002; Kueppers et al. 2006; Dellino et al. 2001; 2012; Zimanowski et al. 2003 Németh 2010 Pardo et al. 2014a
	Conduit stratigraphy and processes		Taddeucci et al. 2002; D'Oriano et al. 2005; Cioni et al. 2011; Perugini et al. 2011; Andronico et al. 2009b; 2013a; Lautze et al. 2012; 2013; Perugini et al. 2002;2007
	State of the magma at the fragmentation		Carey et al. 2000; Dellino and Liotino 2002; Maria and Carey 2002; 2007; D'Oriano et al. 2011; Ruth and Calder 2014
	Link between vesicularity and particle morphology, and particle morphology with cloud dispersal and sedimentation		Wilson and Huang 1979; Dellino et al. 2005; Mattsson 2010 Alfano et al. 2011; Mele et al. 2011
Clast density and vesicularity	Lateral variability of magma within the conduit		Houghton and Wilson 1989; Kennedy et al. 2005; Kueppers et al. 2005; Mueller et al. 2011; Barker et al. 2012
	Dense juvenile	Presence of outgassed magma	Sable et al. 2006; Lautze and Houghton 2005; 2007; 2008; Polacci et al. 2008; 2009a; b; 2012; Gurioli et al. 2005; 2014; Shea et al. 2011; 2012; 2014; Cimarelli et al. 2010
		Presence of a plug	Hoblitt and Harmon 1993; D'Oriano et al. 2005; Sable et al. 2009; Adams et al. 2006a; 2006b; Giachetti et al. 2010; Barker et al. 2012; Lavallée et al. 2012
Clast permeability and connectivity	Degassing history experienced by the magma		Eichelberger et al. 1986; Klug and Cashman 1996; Saar and Manga 1999; Jouniaux et al. 2000; Blower 2001a, b; Klug et al. 2002; Melnik and Sparks 2002; Rust and Cashman 2004; 2011; Mueller et al. 2005; 2008; Wright et al. 2006; 2007; 2009; Plats et al. 2007; Bernard et al. 2007; Takeuchi et al. 2008; Nakamura et al. 2008; Bouvet de Maisonneuve et al. 2009; Yokoyama and Takeuchi 2009; Bai et al. 2010; 2011; Degruyter et al. 2010a; 2010b; 2012

		Vinkler et al. 2012; Polacci et al. 2012; 2014; Nguyen et al. 2014; Pioli et al. 2008; Formenti and Druitt 2003; Giachetti et al. 2010; Shea et al. 2011; 2012
Clast conductivity	Input parameters for numerical percolation simulations	Le Pennec et al; 2001; Bernard et al. 2007; Wright et al. 2009; Wright et Cashman 2014
Vesicle shape and size	Bubble coalescence, ripening or collapse signatures	Klug and Cashman 1996; Mangan and Cashmann 1996; Gurioli et al. 2005; Shin et al. 2005; Sable et al. 2006; Polacci et al. 2008; Castro et al. 2012
	Shear conditions in the conduit Convection in the conduit	Marti et al. 1999; Polacci et al. 2001; 2003; Rust et al. 2003; Okumura et al. 2006; 2008; Bouvet de Maisonneuve et al. 2009; Wright and Weinberg 2009; Laumonier et al. 2011; Shea et al. 2011; 2012 Carey et al. 2013
	Eruptive style	Moitra et al. 2013
Vesicle size distributions (VSDs)	Vesicle nucleation processes and growth in magmas	Klug and Cashman 1994; Shea et al 2010a; LaRue et al. 2013, and references therein
	Total number of nucleation, coalescence or ripening events	Gaonac'h et al. 1996a; b Klug and Cashman 1996; Herd and Pinkerton 1997; Blower et al. 2001; 2002; Gaonac'h et al. 2003; 2005; Lovejoy et al. 2004, Yamada et al. 2008; Bai et al. 2008; Costantini et al. 2010
	Post-fragmentation evolution as indicator of : i) fountaining mechanisms ii) transportation and dispersal of the pyroclasts in submarine environment	Polacci et al; 2006a; Gurioli et al; 2008; Stovall et al. 2011; 2012; Schipper at al. 2010a, b, c; 2011; 2012; Rotella et al. 2013; 2014
Vesicle Number density (Nv)	Link with magma mass eruption rate (MER), link with column height	Polacci et al. 2006b; Toramaru 2006; Gurioli et al. 2008; Carey et al. 2009; Houghton et al. 2010; Rust and Cashman 2011; Alfano et al. 2012
	Magma decompression rate	Mangan and Sisson 2000; Suzuki and Nakada 2001; 2002; Toramaru 2006; Cluzel et al. 2008; Shea et al. 2010b; 2011; 2012; Wright et al. 2012
	Phreatomagmatic fragmentation	Tsukui and Suzuki 1995; Suzuki, Nakada 2001, 2002; Shimano and Nakada 2006; Mattsson 2010; Murtagh et al. 2011; Murtagh and White 2013
	Link vesicularity with external trigger mechanisms (crystallinity, pressure changes)	Belien et al. 2010; Carey et al. 2012; Gurioli et al. 2014

Crystal size distribution (CSD)	Crystal size (mean, modal, and maximum crystal size), crystallization kinetics (nucleation and growth rates), annealing, crystal accumulation, and fractionation	Cashman and Marsh, 1988; Marsh 1988; 1998; 2007; Cashman 1993; Armienti et al. 1994; Higgins 2000; 2002a;b; 2006; 2011; Wilhelm and Worner, 1996; Bindeman 2003; Gualda 2006; Gualda and Rivers 2006; Mock et al. 2003; Simakin and Bindeman, 2008; Spillar and Dolejs 2013
	Magma ascent rate	Cashman 1992; Rutherford and Hill 1993; Rutherford and Devine 2003; Noguchi et al. 2008; D'Oriano et al. 2011
	Pre-eruptive decompression paths	Hammer et al. 1999; Szramek et al. 2006; Clarke et al. 2007; Innocenti et al. 2013
	Magma storage conditions prior to eruption and residence times	Mangan 1990; O'Driscoll et al. 2007; Cigolini et al. 2008 Simakin and Bindeman 2008; Magee et al. 2010; Shea et al. 2009
	Water exsolution rate meter	Toramaru et al. 2008
	Magma mixing	Morgan et al. 2007; Jerram et al. 2003
Crystal+Vesicle size and percentage	Three phases magma rheology, fluid mechanical behavior of magma	Gurioli et al. 2014; Noguchi et al. 2006