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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

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

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

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

56

57 **Introduction**

58

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

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

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

- 92 i. Field deposits (e.g., Polacci et al. 2006a; Rust and Cashman 2007; 2011; Mattsson
93 2010);
- 94 ii. Petrological data (e.g., Larsen 2008; Shea et al. 2009; 2010b; Burgisser et al. 2010;
95 Bai et al. 2011);
- 96 iii. Chemical analyses (e.g., Piochi et al. 2005, 2008; Shimano and Nakada 2006;
97 Noguchi et al. 2006; Costantini et al. 2010; Schipper et al. 2010a; b; c; 2011; 2012;
98 2013; Balcone-Boissard et al. 2010, 2011, 2012; Shea et al. 2012; 2014)
- 99 iv. Geophysical measurements (e.g., Burton et al. 2007; Gurioli et al. 2008, 2013;
100 2014; Polacci et al. 2009b; Andronico et al. 2008; 2009a; 2009b; 2013a, 2013b;
101 Miwa et al. 2009; Miwa and Toramaru 2013; Colò et al. 2010; Landi et al. 2011;
102 Pistolesi et al. 2011; Leduc et al. 2015)

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

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

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

125 The core communal issues to be explored were agreed on:

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

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

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

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

148

149 **Sampling of pyroclasts and quantification of their textural features**

150

151 (i) Representative samples

152

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

157 This assumption has two requirements:

158 i. The textural signature that was quenched immediately at the fragmentation level can
159 be distinguished from the textural effects of post-fragmentation processes, including
160 microlite formation and bubble nucleation, expansion, collapse, coalescence and
161 Ostwald ripening that will change clast vesicularity or vesicle size and shapes once the
162 pyroclast has been formed (e.g., Thomas et al. 1994; Cashman et al. 1994; Herd and
163 Pinkerton 1997; Larsen and Gardner 2000; Gurioli et al. 2008; Costantini et al. 2010;
164 Stovall et al. 2011, 2012). The time window for post-fragmentation changes depends
165 on magma composition, viscosity and fragmentation depth.

166 ii. Because clast density is also a function of clast size (Houghton and Wilson 1989), only
167 clasts of similar sizes must be used in order to avoid non-uniform grain-size effects on
168 textural parameters.

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

- 171 i. Timing: This requires sampling of narrow stratigraphic intervals (Houghton and
172 Wilson 1989) in which juvenile clasts of similar dimensions can be assumed to
173 represent those parts of the magma fragmented at a particular time (n.b. conduit
174 processes can change over short timescales);
- 175 ii. Distribution: This requires selection of more than one outcrop for each event;
- 176 iii. Degree of fragmentation: This requires selection of a sampling methods that is
177 appropriate for the the full range of grain sizes in the deposit;
- 178 iv. Componentry: If the juvenile fraction is heterogeneous, then sampling should be done
179 based on preliminary componentry analysis of the clasts analyzed (e.g., Wright et al.
180 2011, Eychenne et al 2015)

181 In previous studies, only clast sizes of 16-32 mm, i.e. coarse lapilli (White and Houghton
182 2006) have been considered for textural purposes. Such clasts were considered to be large
183 enough to be easily sampled and studied, while being fully representative of the density
184 variation of the majority of erupted pyroclasts and unaffected by significant post-
185 fragmentation phenomena (Houghton and Wilson 1989). These requirements are not always
186 met. In basaltic magma, post-fragmentation effects can be a complication even for these sizes
187 (e.g., Cashman et al. 1994; Szramek et al. 2006; Costantini et al. 2010; Gurioli et al. 2008;
188 Pioli et al. 2014; Pistolesi et al. 2008; 2011; Stovall et al. 2011; 2012). In these cases, the
189 challenge is to identify, quantify and remove post-fragmentation effects in order to isolate
190 textures preserved across the fragmentation zone. For example, the original shapes of vesicles
191 may be reconstructed by de-coalescing large vesicles using the presence of broken, or
192 partially retracted, glassy septa.

193 However, if we study an ash-dominated or a bomb-dominated event, particle-texture
194 analyses must be performed on the fine or coarse juvenile fragments, respectively. Ash size
195 particles (<2 mm) have been investigated recently in terms of vesicle and crystal size
196 distributions (Taddeucci et al. 2002, 2004; Cioni et al. 2008; D’Oriano et al. 2011a, b; Miwa
197 et al. 2009; 2013; Miwa and Toramaro 2013; Proussevitch et al. 2011; Genareau et al. 2012;
198 2013; Colucci et al. 2013, Schipper et al. 2013), and an extensive work has been done in the
199 last 40 years in characterizing ash morphology and deposit componentry (Table 1). For the
200 ash fraction, post-fragmentation expansion can be excluded (e.g., Proussevitch et al. 2011;
201 Genareau et al. 2012; 2013; Colucci et al. 2013). Consequently, analyses allow comparison
202 between morphological and textural features of clasts sampled in proximal and distal areas.
203 Ash particles can record most of the information related to magma ascent dynamics (e.g.,
204 decompression-driven microlite crystallization) and fragmentation (Cioni et al. 2008;

205 D’Oriano et al. 2005; D’Oriano et al. 2011a, b; Proussevitch et al. 2011; Genareau et al. 2012;
206 2013; Colucci et al. 2013). Advantages of studying ash are that it can also be statistically more
207 representative of the variability of the magma properties and is less affected by density-driven
208 settling within the plume. However, ash fragments record only small-scale vesicularity. The
209 integration of observations made on the external shapes of clasts may give information about
210 the presence and importance of a coarser vesicularity which drives magma fragmentation
211 (e.g., Proussevitch et al. 2011; Genareau et al. 2012; 2013; Colucci et al. 2013). However,
212 they cannot provide complete information about the abundance and size of the full vesicle
213 population, if the magma included bubbles larger than the ash particles. Furthermore, ash
214 particles are not suitable for permeability studies, as they are often smaller than the bubbles
215 forming the permeability network. However, the presence of coalesced vesicles in a preferred
216 direction, and an abundance of ash clasts with an elongate shape, have been interpreted as an
217 indication of the development of a permeable bubble network (D’Oriano et al. 2011a).

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

224

225 (ii) Bulk measurements of particle characteristics

226

227 The fastest and most straightforward textural measurement of individual pyroclasts is density
228 (vesicularity), which provides basic information on processes related to gas exsolution and
229 escape (Houghton and Wilson 1989). The densities of lapilli and small bombs can be
230 determined by comparing their weights in water and air following the Archimedes principle.
231 Clasts can be made impermeable with silicone waterproofing spray, by immersion in cellulose
232 acetate, or by using ParafilmTM wax. This technique is fairly rapid and yields large arrays of
233 data with a reproducibility within 10-30 kg m⁻³ and accuracy within 30 kg m⁻³ (Barker et al.
234 2012). Quicker, more timely and precise, density measurements can now be performed using
235 a commercial envelope-density measurement device ([http://www.micromeritics.com/Product-
236 Showcase/GeoPyc-1360-Envelope-Density-Analyzer.aspx](http://www.micromeritics.com/Product-Showcase/GeoPyc-1360-Envelope-Density-Analyzer.aspx)). Following the same principles, a
237 battery-powered device has been used to vacuum-seal pumice or scoria in plastic bags in the
238 field (Kueppers et al. 2005).

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

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

251 Other bulk particle-texture measurements include vesicle connectivity, permeability (Klug
252 and Cashman 1996; Klug et al. 2002; Formenti and Druitt 2003; Rust and Cashman 2004; and
253 references in Table 1) and electrical conductivity (Le Pennec et al. 2001; Bernard et al. 2007;
254 Wright et al. 2009; Wright and Cashman 2014). The connectivity measurements are mostly
255 performed using gas-displacement helium pycnometers, and they deliver first-order
256 information on the outgassing capacity (i.e., potential for gas loss) of the magma near
257 fragmentation (Klug et al. 2002; Formenti and Druitt 2003; Giachetti et al. 2010; Shea et al.
258 2011; 2012). Permeability controls the rate at which magma outgases during decompression.
259 Several methods exist for permeability measurements in volcanology. Rust and Cashman
260 (2004) used a commercial permeameter to perform systematic steady-state gas-flow
261 experiments using porous samples, and the relationship between flow rate and pressure
262 gradient was determined. They also introduced the Forchheimer equation into volcanology,
263 which is a modified form of Darcy's law that includes the inertial effect of gas flow, and
264 specified the importance of this effect in volcanic degassing processes. Mueller et al. (2005)
265 used gas-pressure decay with time after sudden decompression in a fragmentation bomb for
266 the permeability measurements, without measuring gas-flow rate. A falling head permeameter
267 developed by Burbié and Zinszner (1985) has also been used to measure the permeability of
268 volcanic porous materials (Jouniaux et al. 2000; Bernard et al. 2007). Recently, a low-cost gas
269 permeameter was developed (Takeuchi and Nakashima, 2005) and improved (Takeuchi et al.,
270 2008), to measure permeability of natural samples and experimental products. Finally,
271 electrical conductivity measures how well a material transports electric charge. Rocks, in
272 general, are poor conductors, whereas ionic fluids are good conductors. Therefore a

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

277

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

279

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

294

295 *2D method*

296 Standard procedures for the 2D method have been recently published for vesicles (Shea et al.
297 2010a) and crystals (Higgins 2000; 2006). Two-D techniques can yield high-quality data and
298 account for both vesicle and crystal sizes in the sample and can be applied to particles ranging
299 in size from bombs (e.g., Gurioli et al. 2014, Leduc et al. 2015) to ash (Miwa et al. 2009;
300 2013; Miwa and Toramaru 2013). These measurements are best used when there is a broad
301 size distribution to be measured. The main limitation of the method is that is based on the
302 assumption of spherical shape of the textural objects, following Sahagian and Proussevitch
303 (1998). When this conversion is simply obtained by dividing the number of vesicles per unit
304 area by the median value of diameter of each size class (Cheng and Lemlich 1983), no shape
305 assumption is made. However, the 3D conversion is more precise when a shape is defined.
306 Empirical corrections are commonly used for crystal analyses (Higgins 2000 and 2006), but

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

309

310 *3D method*

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

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

340

341 (iv) Crystal size distribution

342

343 Crystal size distribution (CSD) is a well-established tool for interpreting the physical
344 processes and environmental variables that drive differentiation and crystallization in magma
345 chambers and conduits (e.g., Marsh 1988; Cashman and Marsh 1988; Cashman 1992;
346 Hammer et al. 1999; Cashman and McConnell 2005, Armienti 2008; also see references in
347 Table 1). CSD, coupled with vesicle distribution data, yield deeper insights into the physical
348 processes operating in the conduit (e.g., Gurioli et al. 2005; D’Oriano et al. 2005; Piochi et al.
349 2005; 2008; Noguchi et al. 2006; Giachetti et al. 2010; D’Oriano et al. 2011a; Vinkler et al.
350 2012). The CSD method has been well tested and widely applied (Table 1), so that it is now
351 quite straightforward to quantify CSD (Higgins 2000; 2006; and references in Table 1).

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

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

375 compositions, textures (crystal size, orientation, shape) and modes of minerals (e.g., Muir et
376 al. 2012; Leduc et al 2015). Another new technique uses an electron backscatter diffraction
377 detector (EBSD) attached to the SEM to obtain crystal orientations, which can provide
378 insights into shearing, accumulation and degassing processes (Prior 1999; Prior et al. 1999;
379 Hammer et al. 2010). Chemical mapping is now routinely and widely used (e.g., Leduc et al.
380 2015). In contrast, EBSD is more difficult to use and interpreting the data is harder than the
381 chemical maps. As described in the references cited, it produces a wealth of information on
382 various minerals, although the monocline structure of the feldspar can be problematic.

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

394

395 (v) Errors in particle-texture analyses

396

397 Uncertainties in textural analysis are due to several factors. Any textural parameter, such as
398 porosity or crystal size, has intrinsic measurement errors. These are linked to the apparatus
399 used and are generally easy to quantify using standards. A good practice, when a new method
400 is introduced, is to assess its intrinsic error with synthetic samples of well-known particles,
401 having textural components (e.g. crystals, vesicles/voids) with known size and distribution
402 (e.g., see review of Rust and Cashman 2004 for permeability, and Baker et al. 2011 for 3D
403 data from X-ray microtomography). Another type of uncertainty is linked to natural
404 variability, which is generally approached by using the concept of Representative Elementary
405 Volume (REV, Bear 1972). Parameters measured in small, neighboring, regions within a
406 sample have a large variability. As the analyzed regions become larger, this variability
407 decreases until a steady value is reached at the REV size. One complication is that the REV
408 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

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

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

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

466

467 **Quantification and sampling of pyroclastic deposits for the textural characterization of** 468 **their components**

469

470 (i) Preliminary field studies and sampling strategy

471

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

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

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

505

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

507

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

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

528

529 (iii) Selecting the outcrop

530

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

539 A second criterion for selecting the outcrop is that clear textural variations among the
540 juvenile clasts, in terms of color, general morphology, vesicularity, vesicle shape, and
541 crystallinity should be evaluated in the preliminary field survey, so that any lateral and
542 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 carefully 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 Cl, F, S, H₂O, CO₂) 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 (⁶Li, ⁷Li, H/D, ¹⁰B, ¹¹B) or radiogenic
632 isotopes (²¹⁰Pb-²²⁶Ra), 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:

- 642 - Experimental observations on phase equilibria (mineral-melt-vapor), crystallization
643 paths and liquid line of descent (e.g., Hammer and Rutherford 2002; Couch et al.
644 2003; Blundy et al. 2006; Hammer 2008);
- 645 - Calibration of decompression rates. While this has been carried out for rhyolitic
646 systems (e.g., Mourtada-Bonnefoi and Laporte 2002, 2004; Mangan and Sisson 2005;
647 Gardner 2007; Cichy et al. 2011; Cluzel et al. 2008) and phonolitic systems (e.g.,
648 Larsen 2008; Shea et al. 2010b), there are ongoing studies on basaltic systems (Bai et
649 al. 2008; Lesne et al. 2011; Pichavant et al. 2013);
- 650 - Diffusion coefficients of relevant chemical elements, including volatiles, to improve
651 kinetic modeling (Dohmen et al. 2007; Chakraborty 2008);
- 652 - Relationships between crystal morphologies, cooling rates and degree of undercooling
653 (e.g., growth of crystals with hopper and swallow tail shapes experiencing rapid late-
654 stage crystallization; Faure et al. 2003, 2007);
- 655 - Surface flux of volatiles (i.e. what leaves the system; see reviews by Fischer 2008;
656 Pyle and Mather 2009) compared with melt inclusion data (i.e. what is in the system
657 initially; e.g., Le Voyer et al. 2010; Rose-Koga et al. 2012; Schiavi et al. 2012).

658

659 2) Where geochemistry can help textural study

660

661 Measurements of volatile contents in quenched, phenocryst-hosted melt inclusions provide
662 estimates of initial (shallow crustal) values (e.g., Kent 2008). These are minimum estimates,
663 because H₂O can leak from melt inclusions during ascent by intracrystalline diffusion as the
664 far-field environment of the crystal evolves (Chen et al. 2011, 2013). Melt inclusion volatile
665 contents can be inverted to equivalent saturation pressures using multi-species (e.g., H₂O-
666 CO₂; H₂O-Cl) solubility laws (using, for example, VOLATILCALC, Newman and
667 Lowenstern 2002; MELTS, Ghiorso and Sack 1995; Asimow and Ghiorso 1998). These, in
668 turn, can be used to calculate total pressures (and hence depth) by assuming volatile
669 saturation, or minimum pressures if the sample is under-saturated in volatiles. Progressive
670 closure of melt inclusion networks in growing phenocrysts can result in zone-dependent melt
671 inclusion volatile contents that record the evolution of pressure conditions as magmas migrate
672 from depth (Blundy and Cashman 2008, and references therein). Combining major element
673 and volatile compositions of the melt with phenocryst contents allows calculation of initial
674 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., H₂O, CO₂, SO₂, 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 H₂O 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-hydration (Kyser and O'Neil 1984). Hydration can also be
699 assessed from the ratio of water species (molecular H₂O 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, H₂O 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, H₂O and CO₂ 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 H₂O on magma
720 viscosity, particularly at low H₂O 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 recycling? 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 pistonning 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 SO₂ 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 SO₂ fluxes relatively
820 close to the source; see Williams-Jones et al (2008) for full review. These approaches have
821 been recently supplemented by SO₂ camera systems, which allow 2D images of SO₂
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, SO₂ 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*

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

855 Seismic signals that accompany explosions are primarily short period (SP; high frequency
856 > 1 Hz) signals which are typically termed “explosion quakes”. These usually have high
857 amplitudes and mostly include frequencies up to a few hertz, with a possible higher frequency
858 acoustic phase (McNutt 1986, Mori et al. 1989, Braun and Ripepe, 1993). Below these
859 frequencies, short period (SP) signals are often hidden by very-long period (VLP) components
860 with much lower amplitudes (Neuberg et al. 1994; Kaneshima et al. 1996). In spite of an
861 enormous amount of work, it remains unclear as to how we can explain the VLP seismic
862 component, which itself is only one part of the seismic signal. It also remains unclear as to
863 whether, and/or how, SP and VLP components are related to the magnitude and intensity of an
864 explosion, although attempts have been made using tremor (Brodsky et al. 1999; Nishimura
865 and McNutt 2008; Prejean and Brodsky 2011). Clearly, better coupling with the physical
866 volcanology community could help narrow down much uncertainty and allow progress
867 towards better models to untangle the seismic signal associated with discrete explosive events.

868 Delay times in the arrival of seismic, infrasonic and thermal signals have been commonly
869 used to assess the depth at which various physical processes occur in explosive basaltic
870 systems (e.g., Ripepe and Braun 1994; Ripepe et al. 2001; 2002; Harris and Ripepe 2007).
871 However, the sound speed in the conduit needs to be assumed if, for example, the thermal-
872 infrasound delay is to be used to obtain the fragmentation depth. This will vary strongly with
873 conditions in the empty portion of the conduit, including mixture density, gas-to-particle ratio,
874 and temperature of the mixture through which the sound is propagating. Thus we need to
875 know these variables if we are to provide a realistic sound speed value and hence infer a
876 plausible depth. We thus need to constrain two fundamental parameters to strengthen
877 geophysical modelling of the shallow explosion mechanism and depth. First, the magma
878 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 $\frac{1}{4}$ 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 SO₂ sensor arrays are becoming increasingly common components of permanent
945 monitoring arrays at many persistently active sites (Harris 2013). However, such technology

946 will probably never be installed on every potentially active volcano, all of which give seismic
947 and pressure signals detectable by distant stations. From a practical point of view, it is more
948 realistic to push forward with operational correlations between seismic-infrasonic metrics and
949 deposit and particle textural deliverables to understand the ongoing progression of global
950 volcanic events. In doing so, we must remember that many geophysical signals tend to be
951 time-averages (e.g., tremor amplitude). We need to consider, however, geophysical
952 measurements that describe single, discrete explosions if we are to reasonably compare the
953 data with textural variations between many individual emission events, or emission phases,
954 that characterize the total eruption total energy is one prime example (e.g., Marchetti et al
955 2009).

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

968

969 **Questions, needs and recommendations**

970

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

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

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

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

999

1000 **DynVolc: an integrated database**

1001

1002 Inspired by this effort, a database – DynVolc (Dynamics of Volcanoes) – is now operative
1003 at <http://wwwobs.univ-bpclermont.fr/SO/televolc/dynvolc/index.php>. This database is part of
1004 an observation system within the services provided by Observatoire de Physique du Globe de
1005 Clermont-Ferrand (OPGC). It is an attempt to provide an integrated and accessible library for
1006 all multi-disciplinary data sets for explosive eruptive events. This database is an integrated
1007 collection of data from physical and geophysical observations of dynamic volcanic processes.

1008 DynVolc database spans the full range of explosive and effusive activity. Its intent is to
1009 provide a library of standards for eruptive styles, for each of which the data base provides:

- 1010 • field data (i.e., results of field mapping, outcrop and sample descriptions)
- 1011 • key deposit features (thickness, areal dispersion, sedimentary structure, grain size)
- 1012 • clast characterization (componentry, morphology, density, porosity, permeability)

- 1013 • clast texture (connectivity, vesicle and crystal size and size distributions)
- 1014 • chemical analyses of samples (bulk and glass chemistry)
- 1015 • associated geophysical measurements (e.g., fragmentation depth, ejection and ascent
- 1016 velocity, fragment and gas mass, seismic and acoustic energies)

1017 Integration of these data allows improved, better constrained, insights into the dynamics
1018 driving each eruptive style. It also allows improved definition of the rheological and
1019 degassing conditions associated with each activity style. At the same time it provides a library
1020 of key physical parameters that need to be assumed by geophysical data reduction methods, as
1021 well as during model-based enquiry.

1022 Central to this initiative will be the transformation of this database into a communal
1023 databank, involving a web-based GIS platform to allow huge amounts of cross-correlation and
1024 comparison between parameters relating to different processes and cross-correlation of
1025 different datasets obtained for the same eruption. It is intended as an open database into which
1026 anyone can input, and withdraw, citable cross-disciplinary information for scientific analysis.
1027 At the same time, through this library, we can provide cross-community time series, baseline
1028 and monitoring data for the full range volcanic activity.

1029

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1036

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2731 **Figure captions**

2732

2733 **Figure 1** Diagrammatic illustration showing a volcanic strombolian conduit (modified from
2734 Harris and Ripepe 2007), and the list of a few parameters that can be measured through the
2735 deposit (D), the texture of the pyroclasts (T), the geochemistry (G) and the geophysics (G)
2736 methods for small, magmatic explosions.

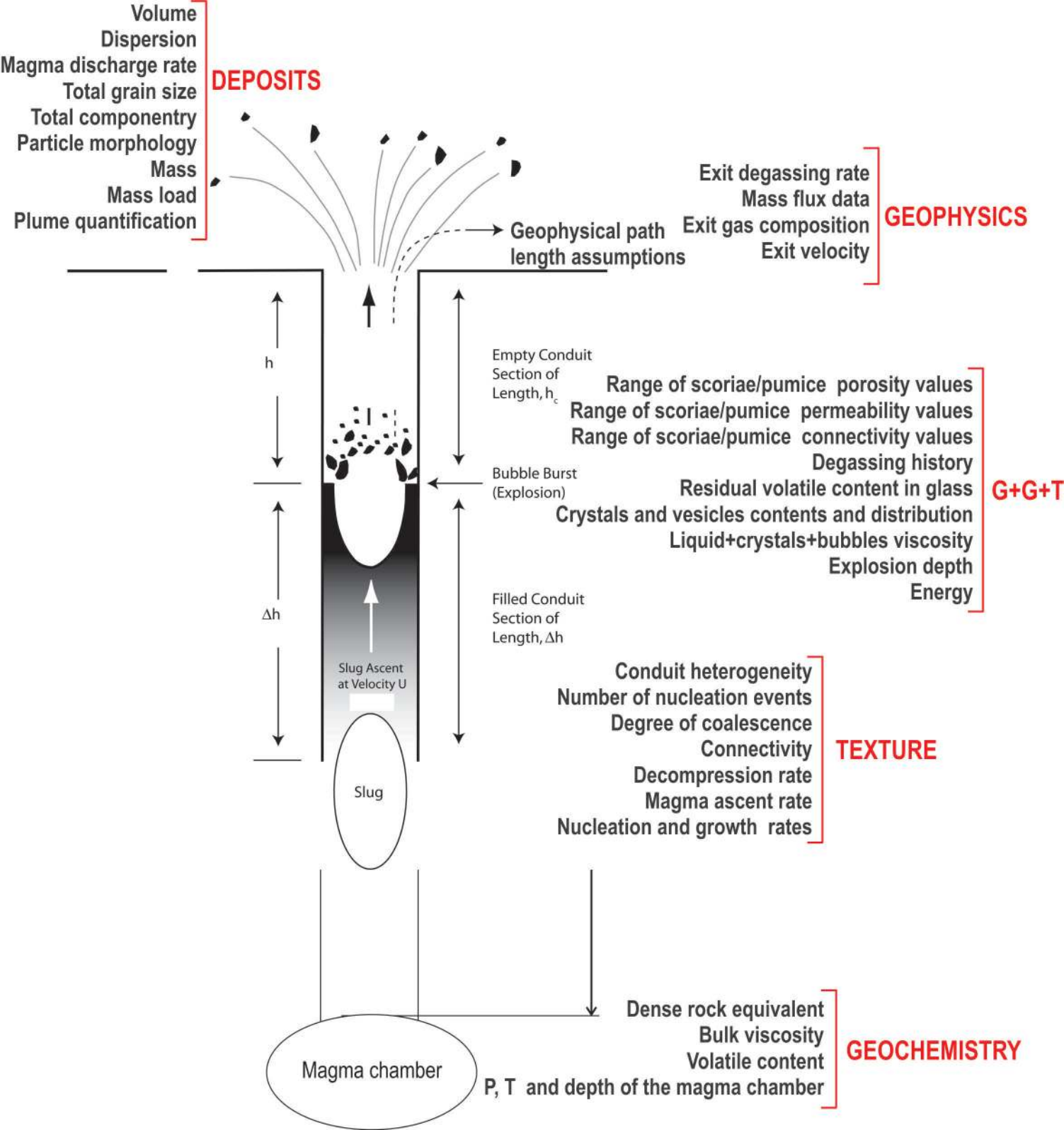


Table 1 Quantification of explosive dynamics from textural parameters of the pyroclast components

<i>Textural parameters</i>	<i>Quantification</i>		<i>References</i>
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 et al. 2010a, b, c; 2011; 2012; Rotella et al. 2013; 2014
Vesicle Number density (N _v)	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 Wörner, 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