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DOI:
10.1038/s41561-019-0468-6

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Fabio Arzilli[1], Giuseppe La Spina[1], Mike R. Burton[1], Margherita Polacci[1], Nolwenn Le Gall[2], Margaret E. Hartley[1], Danilo Di Genova[3], Biao Cai[4], Nghia T. Vo[5], Emily C. Bamber[1], Sara Nonni[5], Robert Atwood[5], Ed W. Llewellin[6], Richard A. Brooker[7], Heidy Mader[7] and Peter D. Lee[2]

[1]School of Earth and Environmental Sciences, University of Manchester, Manchester M139PL, UK

[2]Department of Mechanical Engineering, University College London, London, UK

[3]Institute of Non-Metallic Materials, Clausthal University of Technology, Zehntner Str. 2a, 38678 Clausthal-Zellerfeld, Germany

[4] School of Metallurgy and Materials, University of Birmingham, Birmingham B15 2TT, UK

[5]Diamond Light Source, Harwell Science and Innovation Campus, Didcot OX11 0DE, UK

[6]Department of Earth Sciences, Durham University, Durham DH1 3LE, UK

[7]School of Earth Sciences, University of Bristol, Bristol BS8 1RJ, UK

[*]Corresponding author: Dr. Fabio Arzilli

Corresponding author present affiliation: School of Earth and Environmental Sciences, The University of Manchester, Oxford Road, Manchester, M13 9PL, UK

E-mail address: fabio.arzilli@manchester.ac.uk

Phone: +393298429732; +447904104670

Basaltic eruptions are the most common form of volcanism on Earth and planetary bodies. The low viscosity of basaltic magmas inhibits fragmentation, favouring effusive and lava-fountaining activity, yet highly explosive, hazardous basaltic eruptions do occur. The processes that promote fragmentation of basaltic magma remain unclear, and are subject to debate. Here, we use a numerical conduit model to show that rapid ascent of magma during explosive eruption produces large undercooling. Novel *in situ* experiments reveal that undercooling drives exceptionally rapid (~minutes) crystallisation, inducing a step-change in viscosity that triggers magma fragmentation. Experimentally-produced textures are consistent with products of basaltic Plinian eruptions. We apply the numerical model to investigate basaltic magma fragmentation over a wide parameter space and find that all basaltic volcanoes have the potential to produce highly explosive eruptions. The critical requirements are initial magma temperatures lower than 1100 °C, in order to reach a syn-eruptive crystal content of > 30 vol.%, and thus a magma viscosity ≥ 10^5 Pa s, which our results suggest is the minimum viscosity required for the fragmentation of fast ascending basaltic magmas. Our study provides both a demonstration and explanation of the processes that drive basaltic Plinian eruptions, revealing how typically effusive basaltic volcanoes can produce unexpected highly explosive, and hazardous, eruptions.

Explosive eruptions are undoubtedly the most violent and catastrophic expression of volcanic activity[1,2]. Highly explosive Plinian eruptions can inject > 0.1 km^3 of pyroclastic material producing volcanic plumes that can reach up to 40 km in height, impacting both regional and global climate, and producing a significant threat to proximal populations[3,4]. Basaltic volcanoes are usually characterised by effusive and mild-explosive lava fountaining activity[5], however, basaltic Plinian eruptions can occur[6-11]. The main process characterising highly explosive activity is magma fragmentation, which is the transition from ductile to brittle behaviour of molten magma[1,2]. For silicic magmas, fragmentation is attributed to high strain rates associated with acceleration of the bubbly mixture[1,12-14] or from the bubble overpressure associated with restricted bubble expansion[14,15-18]. The low viscosity of basaltic magmas, however, makes high strain rates, or the bubble overpressure necessary to trigger fragmentation difficult to achieve[1,19].

Crystallisation during magma ascent may significantly increase magma viscosity, leading to fragmentation[20]. However, crystallisation in basaltic magmas has previously been thought to occur on timescales significantly longer than the time required for magma to ascend from a crustal storage chamber to the vent[21], particularly in the case of highly explosive eruptions.

In volcanic conduits, the crystallisation kinetics of an ascending magma are driven by degassing and cooling[22-23]. Plagioclase and pyroxene crystallisation are sensitive indicators of magma dynamics in volcanic conduits[21,24-27] and their kinetically controlled abundance can rapidly change magma rheology[28-29]. Our understanding of crystallisation kinetics in magmas is underpinned by *ex situ* crystallisation quench experiments. Here we perform novel *in situ* experiments in order to visualize and quantify the evolution of rapid crystallisation in basaltic magmas, as with this approach it is possible to ascertain in real time when crystals start to form and how quickly magma crystallinity evolves.

A frequently used model to describe crystallisation as function of time is given by an exponential law[21,23], where the rate of crystallisation is controlled by the characteristic time $\tau^{(c)}$. The characteristic time is a measure of how fast a process will approach the equilibrium[21,30]. The smaller

$\tau^{(c)}$, the faster crystals reach their equilibrium abundance (see Methods section). La Spina et al. (ref. 21) demonstrated that the time required to reach the equilibrium crystal fraction in basalts is ~4.6 times $\tau^{(c)}$. Furthermore, during mild lava-fountaining basaltic activity[21], $\tau^{(c)}$ is in the order of thousands seconds, resulting in a crystallisation time of ~2 hours. Crystallisation rate increases with magma ascent rate, as cooling and decompression rates increase[30-34]. Therefore, $\tau^{(c)}$ is smaller and the equilibrium crystal fraction will be reached faster. However, the characteristic times of crystal growth during fast magma ascent have not been quantified.

In order to quantify the characteristic times of plagioclase and pyroxene crystallisation during rapid ascent of basaltic magma, here we present the first *in situ* 4D (3D plus time) crystallisation kinetics experiments under fast cooling rates, using fast synchrotron X-ray microtomography. Our experiments provide the first estimation of the characteristic time for plagioclase and pyroxene crystallisation in trachybasaltic magmas during a rapid and continuous increase of undercooling, where ΔT is defined as the difference between the highest temperature at which plagioclase and pyroxene is expected to crystallise and the temperature of the magma[24-27].

Crystallisation experiments at high undercooling

Crystallisation experiments were performed *in situ* at beamline I12-JEEP, Diamond Light Source, Harwell, UK, using a trachybasaltic glass (Supplementary Table 1) from the 2001 Etna eruption as the starting material (see Methods). We combined a bespoke high-temperature environmental cell[35] with fast synchrotron X-ray microtomography to image the evolution of crystallisation in real time[36] in two experiments. In the first part of the experiment, crystallisation was induced by decreasing temperature from 1250 °C to either 1170 °C or 1150 °C at 0.4 °C s^{-1} at ambient pressure with a dwell time at the final temperature of 4 h[36] (Supplementary Fig. 1). During these 4 h, blocky and prismatic clinopyroxene and oxide crystals were able to grow[36] (Fig. 1 and Supplementary Figs. 2 and 3), and the residual melt evolved to a basaltic trachyandesitic composition (Supplementary Table 1; Supplementary Figs. 2 and 3). After 4 h at sub-liquidus

conditions (1170 °C and 1150 °C), the system was perturbed through a rapid cooling rate of 0.4 °C s^{-1}, inducing a sudden and continuous increase of undercooling (ΔT). This produced a rapid crystallisation event from the residual melt (Fig. 1; Supplementary Figs. 2 and 3). Our results show that skeletal plagioclase crystals grew during this final stage of rapid cooling, specifically between 1112 and 1073 °C after a dwell time of 4 hours at 1150 °C (Fig. 1) and between 1131 and 1053 °C after a dwell time of 4 hours at 1170 °C (Supplementary Fig. 3a,b,c). Plagioclase crystals grew to equilibrium abundance in ~100 seconds (Fig.1), i.e. between two 3D scans. Following this initial burst of rapid plagioclase growth, dendritic clinopyroxene crystals began to nucleate heterogeneously on plagioclase and grew to their final size in the following 180 seconds (Fig. 1b,c; Supplementary Figs. 2b,c, 3d,e, 4 and 5).

The large ΔT reached in a relatively short time during our *in situ* 4D crystallisation experiments generated distinctive skeletal plagioclase crystals with swallow-tail morphology and dendritic pyroxene (Fig. 2a-c), similar to the skeletal plagioclase and dendritic pyroxene crystals observed in the products of explosive basaltic Plinian eruptions[7-11,37,38], such as Etna 122 B.C.[7,8,37,38] (Fig. 2d). Heterogeneous nucleation of dendritic pyroxene on skeletal plagioclase (Fig. 1b) is observed to occur in ~180 s during our 4D experiments. These distinctive textures are also reported in products of the Fontana Lapilli (Nicaragua)[11,37,38] and 1886 Tarawera (New Zealand) eruptions[7,9,10]. Therefore, all the studied examples of basaltic Plinian deposits show features that are consistent with the textures produced in our experiments. Furthermore, the signature skeletal and dendritic pyroxene is also observed in sub-Plinian eruption (Yufune 2) products of Mt. Fuji (Japan)[39].

Skeletal plagioclase crystallisation occurred at ΔT between 60 and 140 °C with a growth rate of $3x10^{-5}$ cm s^{-1} (mm h^{-1}) (Supplementary Table 2), whilst dendritic pyroxene crystallised at ΔT between 60 and 175 °C with a growth rate of $2x10^{-5}$ cm s^{-1} (mm h^{-1}) (Supplementary Table 2). This indicates that a rapid increase of ΔT (>60 °C) induces fast crystallisation. As the equilibrium pyroxene crystal content is achieved within ~180 s, we can infer that the pyroxene characteristic time under large ΔT is < 40 s. For plagioclase, where the equilibrium crystal content is achieved

within 90 s the characteristic time is < 20 s. These are about two orders of magnitude less than the characteristic time found by La Spina et al. (ref. 21) for effusive and mild lava fountaining activity at Etna (Italy), Stromboli (Italy) and Kilauea (Hawaii), which involved much smaller ΔT (30-60 °C)[21] and magma ascent rates of ~3 m s^{-1}.

Ex situ experiments were also performed using a TZM cold seal pressure vessel apparatus, in order to investigate whether fast crystallisation also occurs in a hydrous trachybasaltic melt during rapid and continuous cooling and decompression (see Methods), simulating fast magma ascent in the conduit. The experimental approach consisted of holding the sample for 30 minutes at 75 MPa and 1070 °C before the sample was decompressed and the temperature decreased. After 30 minutes pressure and temperature were decreased continuously for 300 s at 0.2 MPa s^{-1} and 0.2 °Cs^{-1} respectively, reaching 15 MPa and 1010 °C (Supplementary Table 3; Supplementary Fig. 6). The rapid decompression and cooling enabled large ΔTs to be the achieved (>100 °C; see Supplementary Fig. 6) in 300 s, favouring predominantly clinopyroxene crystallisation, with minor plagioclase and oxide (Supplementary Fig. 7). This indicates that $\tau^{(c)}$ is < 60 s, in agreement with the order of magnitude estimated from *in situ* experiments.

Large undercooling can produce significant syn-eruptive microlite crystallisation during rapid magma ascent[24-27]. This increase in crystallinity dramatically increases the viscosity of the magma[19]. This process has been proposed to explain explosive basaltic Plinian eruptions, supported by evidence of high microlite contents[7-11,37,38]. However, a mechanism for this exceptionally fast crystallisation during rapid magma ascent in basaltic Plinian eruptions, favouring fragmentation, has not been demonstrated so far.

Numerical simulation of the 122 B.C. Etna basaltic Plinian eruption

Magma fragmentation in basaltic Plinian eruptions has been investigated with conduit models, where crystallisation has been assumed either to be constant[20] or at equilibrium[40]. Recent results demonstrate that disequilibrium crystallisation plays a fundamental role in magma dynamics within

the conduit[21], but syn-eruptive disequilibrium crystallisation has not yet been considered for basaltic explosive volcanism[20,40].

We used the conduit model described by La Spina et al. (ref. 21, 23) to investigate the effect of the new experimentally constrained characteristic times for crystallisation with large undercooling derived here on the ductile-brittle transition of basaltic magma during highly explosive eruptions. Since we are focussing our attention on explosive eruptions with high mass eruption rates (>10^6 kg/s), it is reasonable to assume that outgassing is negligible for this kind of activity. Indeed, outgassing will be inhibited by fast magma ascent and fragmentation will be achieved faster than outgassing[41]. As a test case, we consider the 122 B.C. Etna basaltic Plinian eruption[6-8]. To model fragmentation we adopt the strain-rate criterion introduced by Papale (ref. 1):

$$\dot{\gamma} = k \frac{G_\infty}{\mu} \quad (1)$$

where $\dot{\gamma}$ is the elongational strain rate, k=0.01 is a constant, μ is the magmatic viscosity and G_∞ is the elastic modulus at infinite frequency. The other constitutive equations are reported in the Methods section. In Figure 3 we report the calculated plagioclase undercooling, crystal content and viscosity as function of depth for $\tau^{(c)}$ = 10 and 1000 s. Large undercooling is produced by adiabatic expansion of exsolved volatiles, and mitigated by the latent heat of crystallisation for $\tau^{(c)}$ = 10 s (Fig. 3a). In this case, numerical results show a rapid increase in crystallinity and in viscosity at depths below 2 km (Fig. 3c), leading to fragmentation and explosive Plinian eruption. Importantly, numerical results indicate that strain-induced fragmentation is favoured by a combination of rapid viscosity increase (about one order of magnitude in 5 seconds approaching the fragmentation depth) and fast decompression/ascent rates (~0.2 MPa s^{-1} and ~15 m s^{-1}) that generate high strain rates. For $\tau^{(c)}$ = 1000 s, instead, the crystallisation rate is slow, and viscosity stays within the fragmentation threshold all along the conduit.

Sensitivity analyses on basaltic magma fragmentation driven by rapid crystallisation

171 Having established that fast crystallisation plays an important role on basaltic Plinian eruptions,
172 we performed a sensitivity study with our numerical model to investigate the parameter space
173 whereby basaltic fragmentation driven by rapid crystallisation may occur. We use the Etna 122 B.C.
174 eruption as a test case. We focus on the behaviour of ΔT as a function of characteristic time of
175 crystallisation, pressure, temperature and magmatic H_2O content at the conduit inlet, conduit radius,
176 and initial phenocryst content. A detailed description of the initial condition of the sensitivity study
177 can be found in the Methods section. We performed several sensitivity studies assuming a
178 characteristic time of $\tau^{(c)}$ = 10 s, as observed in our experiments, and $\tau^{(c)}$ = 1000 s, as observed for
179 Etna 2001 in mild explosive activity. We also examined $\tau^{(c)}$ = 1 and 100 s for completeness.
180 Numerical results show that undercooling is principally controlled by $\tau^{(c)}$ (Fig. 4a), because the
181 release of latent heat during rapid crystallisation (i.e. small $\tau^{(c)}$) affects the temperature of the
182 system, and, consequently, the undercooling. Therefore, undercooling and characteristic time are
183 linked, affecting one each other. However, our numerical results indicate that, within the parameters
184 space investigated, ΔT is always maintained between 60 and 190 °C (Fig. 4b), which is enough to
185 enable rapid crystallisation in any case. Furthermore, the sensitivity analyses indicate that, after the
186 characteristic time, initial temperature and water content (exsolved + dissolved) play an important
187 role on controlling undercooling (Fig. 4a). ΔT at the point of fragmentation increases as initial
188 temperature decreases (Fig. 4b), meaning that cooler magma in the chamber is more likely to
189 produce microlite crystallisation during ascent, as ΔT will already be large under the pre-eruptive
190 condition. An increase in the initial total H_2O (exsolved + dissolved) also produces an increase of
191 ΔT (Fig. 4c), caused by greater cooling of the system due to enhanced adiabatic gas expansion.
192 Furthermore, higher water content favours rapid crystallisation of microlites, due to the increase in
193 diffusivity within the melt.
194 Our sensitivity studies show that a lower initial temperature and a higher pre-eruptive crystal
195 content results in a greater likelihood of explosive eruptions (Fig. 4d,e). This is consistent with
196 estimates of pre-eruptive temperatures obtained for the Etna and Fontana Plinian eruptions, which

range between 1000 and 1060 °C[37]. Regarding the pre-eruptive crystal content, however, products erupted from basaltic Plinian eruptions are characterized by a small phenocryst content (<10 vol.%)[7,8,10,11,37,38,40]. Therefore, in this context of low initial temperature and low phenocryst content, our results highlight that the characteristic time of crystallisation has a fundamental role on the likelihood of producing an explosive eruption. Indeed, a small $\tau^{(c)}$ is the primary cause for a significant increase of the syn-eruptive crystal fraction over short timescales during magma ascent, which consequently increases magma viscosity and, thus, the probability of magma fragmentation (Fig. 4f). Moreover, our sensitivity analysis demonstrates that when the syn-eruptive crystal content exceeds 30 vol.%, all numerical solutions reach the fragmentation threshold (Fig. 4f). Indeed, in a fast ascending magma (from our sensitivity analyses we have an average ascent rate between ~5-50 m/s), this increase in syn-eruptive crystal content produces a rapid and dramatic increase of viscosity (about one order of magnitude in less than 10 s close to the fragmentation depth), increasing the likelihood of magma fragmentation. Numerical results also show that, as soon as viscosity exceeds 10^5 Pa s, all the numerical solutions reach the fragmentation threshold, generating an explosive eruption (Fig 4g). Indeed, for a fast ascending magma where outgassing can be neglected and strain-rates are high, the key parameter controlling the triggering of fragmentation is viscosity. Therefore, 10^5 Pa s represents a minimum viscosity determining a drastic change in eruptive style for activities characterised by high mass eruption rates. This minimum viscosity is one order of magnitude lower than previously reported for low-viscosity magmas[1,42].

Experimental and natural observations combined with a numerical model allow us to conclude that pre-eruptive temperatures <1100 °C favour the formation of highly explosive basaltic eruptions, such as Plinian volcanism, driven by fast syn-eruptive crystal growth under high undercooling. This implies that all basaltic systems on Earth have the potential to produce powerful explosive eruptions.

METHODS

Starting material. The starting material, used for our crystallisation experiments, is a trachybasalt from the lower vents of the 2001 Mt. Etna eruption[36,43]. The anhydrous, glassy starting material was obtained by melting crushed rock samples in a Pt crucible. Melting was performed in a Nabertherm® $MoSi_2$ box furnace at 1400 °C and at atmospheric pressure. The melt was left in the furnace for four hours to allow the melt to fully degas and to dissolve the crystals present. The melt was then quenched in air to glass. This procedure was repeated two times to homogenise the melt. Finally, anhydrous glassy cylinders 3 mm in diameter and 4 mm in length were drilled from the synthesized glass for *ex situ* and *in situ* 4D crystallisation experiments.

The chemical composition of the anhydrous glassy starting material has been analysed with a Jeol JXA 8530 F microprobe at the Photon Science Institute, University of Manchester, Manchester, UK, and are reported in Supplementary Table 1. Analyses were performed using a 15 kV accelerating voltage, 10 nA beam current and beam size of 10 μm. Standards used for calibration were albite for Na, periclase for Mg, corundum for Al, fayalite for Fe, tephroite for Mn, apatite for P, sanidine for K, wollastonite for Ca and Si and rutile for Ti. Sodium and potassium were measured first to minimize loss owing to volatilisation.

Hydrous trachybasaltic glass with ~3 wt. % of H_2O was obtained melting the starting material and homogenizing H_2O in a Pt capsule at 100 MPa and 1200 °C. The hydrous starting glasses were produced using a TZM cold seal pressure vessel apparatus at the School of Earth Sciences, University of Bristol, Bristol, UK. Before to perform crystallisation experiments Raman spectra were collected from the hydrous starting glass, using a Thermo ScientificTM DXRTMxi Raman Imaging Microscope at the School of Earth Sciences, University of Bristol (Bristol, UK) in order to check that the amount of H_2O was ~3 wt. %.

***Ex situ* decompression and cooling experiments.**

Ex situ experiments were performed in order to investigate the process of fast crystallisation in a hydrous trachybasaltic melt during a continuous decompression (0.2 MPa s^{-1}) and cooling (0.2 °Cs^{-1}), simulating magma ascent in the conduit. Two experiments were performed using a TZM cold seal pressure vessel apparatus at the School of Earth Sciences, University of Bristol, Bristol, UK. The redox condition of the apparatus was NNO, adding ~1 vol.% of hydrogen to the pressuring argon. Capsules (Au_{80}-Pd_{20}) were loaded with hydrous glass with a cylindrical shape. The experimental approach consisted of holding the sample for 30 minutes at 75 MPa and 1070 °C before being decompressed and cooled. The first experiment was quenched after 30 minutes at 75 MPa and 1070 °C, in order to texturally characterise the initial conditions before that fast decompression and cooling were applied. In the second experiment, after 30 minutes pressure and temperature were released continuously in 300 s at 0.2 MPa s^{-1} and 0.2 °Cs^{-1} respectively, reaching 15 MPa and 1010 °C (Supplementary Table 3; Supplementary Fig. 6). The sample was rapidly quenched at the final conditions. Final pressure was chosen as approximate fragmentation pressures calculated by modeling. Decompression and cooling rates were predicted from the conduit model for magma ascent during basaltic Plinian eruptions.

***In situ* synchrotron X-ray microtomography experiments.** The experiments were performed at the beamline I12-JEEP[44], Diamond Light Source, Harwell, UK. For these *in situ* crystallisation experiments, we used the high-temperature resistance Alice furnace[45], which allows us to control cooling at 0.05 °Cs^{-1} to 0.4 °Cs^{-1}, and the P2R in situ rig[35] for high speed rotation. Temperature was measured with a R type thermocouple, which was close to the sample, positioned in the middle of the hot spot. The dimension of the Alice furnace's hot spot with homogeneous temperature is ~5x5x5 mm; our samples were positioned within this area. The R type thermocouple provided us the temperature of the sample with an uncertainty of ±5 °C. The sample holder was an alumina crucible (Supplementary Fig. 2), which is suitable for the temperature range investigated and it has low X-ray attenuation coefficient. The glassy cylinders were heated *in situ* in the Alice furnace up

to 1250 °C for 30 minutes (Supplementary Figure 1). After the initial annealing period, crystallisation was induced by decreasing temperature from 1250 °C to 1170 °C or 1150 °C at ambient pressure, holding at the final temperature for 4 h[36] (Supplementary Figure 1). After this step, the system was perturbed by a rapid cooling at rate of 0.4 °Cs^{-1} in order to investigate the rapid crystallisation in real time (Supplementary Figure 1), reaching high undercooling (up to ~180 °C) in a short time.

The experiments were performed in phase-contrast mode, setting the sample-to-detector distance at 2300 mm in order to work in the edge-detection regime[46] (Supplementary Table 4). The projections were acquired using a monochromatic X-ray beam with energy of 53 keV. In each scan, 1800 tomographic projections were acquired by the detector with equiangular steps over a full rotation angle of 180° (Supplementary Table 4). The exposure time for the acquisition of each projection was 0.05 s (Supplementary Table 4), therefore, the temporal resolution of each scan was of 90 seconds. The isotropic pixel size is 3.2 μm. The detector was a high-resolution imaging PCO edge camera with optical module 3, corresponding to a field of view of 8.0 mm × 7.0 mm. Scan acquisition started before the end of the annealing, covering the cooling period between 1250 °C and the dwell temperatures (1170 and 1150 °C), the entire duration of the dwell time and the final rapid cooling rate of 0.4 °Cs^{-1}.

Image reconstruction and processing. Tomographic projections were reconstructed into 32-bit slices by using Diamond I12 in-house python codes, using the *gridrec* algorithm[47,48] (http://confluence.diamond.ac.uk/display/I12Tech/Reconstruction+scripts+for+time+series+tomography)[49,50]. The pre-processing pipeline includes centre of rotation calculation[49], zinger removal, blob removal[50], and regularisation-based ring removal[51].

The reconstructed slices were converted to 8-bit raw format and stacked using ImageJ software[52] to obtain volumes in which the isotropic voxel has an edge size of 3.2 μm. Reconstructed volumes of experiments ET1150 and ET1170 were then cropped using Avizo® software v.8.0 (FEI

298 Visualization Sciences Group) in order to select the volume of interest (VOI) (Supplementary Table
299 4). In the experiment ET1150 plagioclase and pyroxene crystals nucleated and grew in a relatively
300 large pocket of melt (Figs 1 and 2). Therefore, the VOI selected consists of a volume of melt where
301 the rapid crystallisation of plagioclase and pyroxene occurred during the final rapid cooling rate of
302 0.4 $°Cs^{-1}$. In the experiment ET1170 plagioclase and pyroxene crystals formed in narrow layers of
303 melt (Supplementary Fig. 3), during rapid continuous cooling at a rate of 0.4 $°Cs^{-1}$.
304 Three-dimensional visualization (volume rendering) of the reconstructed volumes was obtained
305 using the commercial software VGStudio 3.0 (Volume Graphics), which allowed us to make 3D
306 textural observations of the plagioclase and pyroxene crystal morphologies (Fig. 2). Therefore, the
307 reconstructed volume of each scan allowed us to quantify when and at which range of temperature
308 plagioclase and pyroxene crystals were able to form.
309
310 **Image segmentation and analysis of plagioclase**. Segmentation is the process that allows
311 separation of objects from the background to obtain binary volumes containing only the feature of
312 interest. Segmentation of plagioclase crystals from the glassy matrix was performed using the semi-
313 automatic volume segmentation[53,54] in Avizo software v. 8.0 (Supplementary Table 4). This
314 segmentation requires manual drawing of the outlines of crystals on the 2D slices. This is repeated
315 every 5–10 slices, depending on the size of the crystal and the complexity of their shape, along the
316 crystal length. The crystal shape is reconstructed automatically by the software through an
317 interpolation procedure. The advantage of this technique is that the operator can obtain the real
318 morphology of the object of interest by visual inspection[53,54].
319 The reconstructed 3D images were processed and analysed with the Pore3D software library,
320 custom-developed at Elettra[55]. The Pore3D software allowed us to quantify the number of
321 plagioclase crystals, the volume and the maximum length of each crystal, operating directly in the
322 3D domain[55,56]. As we were able to obtain the 3D shape of plagioclase crystals and the real
323 maximum axis length (L_{3D}) we could calculate the growth rate (Y_{L3D}) of plagioclase crystals

(Supplementary Table 2), using the experimental duration of growth (experimental duration). The growth rate was estimated using the following equation[53]:

$$Y_{L3D} = (L_{3D}*0.5)/t_{growth}$$

where t_{growth} is the time required for crystal growth. The microtomography images give us the opportunity to measure the volume of crystals. The volumetric growth rate (Y_V) was calculated (Supplementary Table 2) using the following relationship[53]:

$$Y_V = (V*0.5)/t_{exp}$$

where V is the volume of the crystal.

Image analysis of pyroxene growth kinetics. Back-scattered electron (BSE) images were collected, using a JEOL JSM-6390LA FE-SEM at the School of Earth and Environmental Sciences, University of Manchester, Manchester, UK, in order to analyse the pyroxene morphologies and kinetics. We used an acceleration voltage of 15 kV and beam current of 10 nA. The sizes of dendritic pyroxene crystals were measured in the 2D domain, using BSE images and ImageJ software[52], as pyroxene morphologies formed during continuous cooling are difficult to resolve and analyse in the 3D domain. The pyroxene growth rate is calculated by dividing the entire length of the dendritic crystal over the duration of the pyroxene growth (Supplementary Table 2), as dendritic crystals grow in one direction.

Electron microprobe analysis. Samples obtained with *ex situ* and *in situ* crystallisation experiments were analysed with a JEOL JXA-8530F field emission electron microprobe at the Photon Science Institute, University of Manchester, Manchester, UK. For both instruments, the operating conditions were as follows: 15 kV accelerating voltage, 10 nA beam current, and beam diameter of 10 or 5 μm (the latter for microlites). Sodium and potassium were measured first to minimise loss by volatilisation. Calibration standards were albite for Na, periclase for Mg, corundum for Al, fayalite for Fe, tephroite for Mn, apatite for P, sanidine for K, wollastonite for Ca

and Si and rutile for Ti. Several compositional maps, with a nominal resolution of 1 μm2, have been also collected using the EPMA (EPMA map, are reported in Supplementary Figs. 4, 5 and 7).

Constitutive equations for the conduit model. In this work we use the 1D steady-state model for magma ascent described by ref. (21, 23, 57). The governing equations used in this work are reported in ref. (57). The application to a specific volcano is achieved by providing constitutive equations to describe the specific rheological, solubility, crystallisation, outgassing, and fragmentation behaviour of the system.

Following ref. (58), the viscosity of the liquid phase is modelled as:

$$\mu_l = \mu_{melt} \cdot \theta(x_c^l),$$

where μ_{melt} is the viscosity of the bubble-free, crystal-free liquid phase and θ is a factor which increases viscosity attributed to the presence of crystals[59].

We use an empirical relationship to estimate μ_{melt} as a function of water concentration and temperature, as in ref. (60) (based on the Vogel-Fulcher-Tammann equation):

$$\log(\mu_{melt}) = \mathrm{A} + \frac{\mathrm{B}(\mathrm{y}, x_{d_{H_2O}}^{md})}{T - \mathrm{C}(\mathrm{y}, x_{d_{H_2O}}^{md})},$$

where the viscosity μ_{melt} is in Pa s and *T* is the temperature in Kelvin. The parameter A is the logarithmic value of the viscosity at infinite temperature and it is assumed to be constant for all melts. The parameters B and C, instead, are functions of the melt composition *y* and of the dissolved water content $x_{d_{H_2O}}^{md}$. In this work, we used the composition of the average melt inclusion composition (Etna 122 B.C.) from ref. (61). Furthermore, as crystallisation proceeds, viscosity is increased according to the empirical model described in ref. (62):

$$\theta = \frac{1 + \varphi^{\delta}}{[1 - F(\varphi, \xi, \gamma)]^{B\phi^*}},$$

where

373 $$F = (1-\xi)erf\left[\frac{\sqrt{\pi}}{2(1-\xi)}\varphi(1+\varphi^{\gamma})\right], \quad \varphi = \frac{\left(\sum_{j=1}^{n_c} x_{c_j}^{l}\right)}{\phi^{*}}.$$

374 The fitting parameters B, δ, ξ, γ and ϕ^{*} chosen for this work are the same used in ref. (63).

375 The model proposed in this work takes into account two different gas components: water and
376 carbon dioxide. The equilibrium profile of the dissolved gas content $x_{d_i}^{md,eq}$ of component i follows
377 the Henry's Law, i.e.

378 $$x_{d_i}^{md,eq} = \sigma_i \left(\frac{P_{g,i}}{\bar{P}}\right)^{\varepsilon_i},$$

379 where $P_{g,i} = \alpha_{g_i} P_g / \alpha_g$ is the partial pressure of the i-th gas component expressed in Pa, $\bar{P} = 1$
380 Pa is used to make the expression in the brackets adimensional, σ_i is the solubility coefficient and ε_i
381 is the solubility exponent. We assume that the solubility parameter σ_i and ε_i are constant during
382 ascent. For this work we adopted the following parameters $\sigma_{H_2O} = 1.8911 \times 10^{-6}$; $\varepsilon_{H_2O} = 0.5257$;
383 $\sigma_{CO_2} = 2.2154 \times 10^{-12}$; $\varepsilon_{CO_2} = 1.075$. In this work, we assume also equilibrium exsolution, which
384 means that the dissolved volatile contents always follow the equilibrium profile.

385 The crystallisation model adopted here has been proposed in ref. (23). We consider the three
386 different major crystal components erupted by Etna volcano: plagioclase, pyroxene and olivine. We
387 assume that crystals stay coupled with the melt (i.e. no fractional crystallisation). For a better
388 modelling of crystal nucleation and growth, we also assume that the equilibrium crystal contents are
389 functions of temperature, pressure and dissolved water content. With these assumptions, the
390 equilibrium mass fraction $x_{c_j}^{l,eq}$ of crystal phase j is computed using the polynomial function

391 $$x_{c_j}^{l,eq}(P^{*}, T^{*}, x_d^{*}) = \zeta_{j,1}(P^{*})^2 + \zeta_{j,2}(T^{*})^2 + \zeta_{j,3}(x_d^{*})^2 + \zeta_{j,4}(P^{*})(T^{*}) +$$
$$+\zeta_{j,5}(T^{*})(x_d^{*}) + \zeta_{j,6}(x_d^{*})(P^{*}) + \zeta_{j,7}(P^{*}) + \zeta_{j,8}(T^{*}) + \zeta_{j,9}(x_d^{*}) + \zeta_{j,10},$$

392 where P^{*} is the liquid pressure expressed in bars, T^{*} is the temperature expressed in Celsius
393 degrees and x_d^{*} is the dissolved water concentration in weight percent. From $x_{c_j}^{l,eq}$, the equilibrium
394 crystal volume fraction β_j^{eq} can be computed using the relation

$$\beta_j^{eq} = \frac{\rho_l x_{c_j}^{l,eq}}{\rho_{c_j}}.$$

The parameters $\zeta_{j,i}$ are calculated fitting the polynomial function over a large range of data obtained at different pressures, temperatures and water contents with alphaMELTS[64], a command line version of MELTS[65]. As previously, we used the average melt inclusion composition (Etna 122 B.C.) from ref. (61).

Disequilibrium crystallisation is considered in our model and characteristic times controls the time needed to reach equilibrium crystal content. The differential equation in the conduit model which describes the crystal volume fraction within the conduit is the following[21]:

$$\frac{\partial \alpha_l \rho_c \beta u_l}{\partial z} = -\frac{1}{\tau^{(c)}} \alpha_l \rho_c (\beta - \beta^{eq}).$$

Here, α_l is the volume fraction of the liquid phase, ρ_c is the density of the crystals, u_l is the velocity of the liquid phase, z is the vertical axis, β is the actual total crystal volume fraction, whereas β^{eq} is the equilibrium total crystal volume fraction. If we multiply all the terms by the characteristic time $\tau^{(c)}$, we notice that the smaller is $\tau^{(c)}$, the smaller has to be $(\beta - \beta^{eq})$. In other terms, the smaller is the characteristic time, the faster the crystals will reach the equilibrium value.

Formally, as described in La Spina et al. (ref. 21), the characteristic time reflects the time required to reduce the difference between the actual and the equilibrium value to e^{-1} (~37%) of the initial difference. This means that, if β_0 is the initial value of a physical parameter β (which, in our case, is the crystal volume fraction), and β_{eq} is the equilibrium value in response to a perturbation of the system, at the characteristic time τ, we have

$$\beta(\tau) = \beta^{eq} + e^{-1}(\beta_0 - \beta^{eq}).$$

The experimental phase diagram for Etna basalt erupted during the 122 B.C. eruption[37] provides the plagioclase and pyroxene liquidus at different pressures and temperatures, whilst the conduit model is able to track temperature evolution within the conduit. Combining both of these data, we can estimate ΔT with respect to the plagioclase and pyroxene liquidus during magma ascent.

For this work, as we are interested in the highly explosive activity, we assumed no relative velocity between gas and melt. Furthermore, as we indicated in the main text, we used as fragmentation model the strain-rate criterion introduced by ref. (1).

Initial condition for the sensitivity analysis.

The range of input parameters adopted for the sensitivity analysis are the following: 140–160 MPa for the inlet pressure at 6000 m depth, 1050–1100 °C for the magma inlet temperature, 5–30 m for the radius of the conduit, 2.0–4.0 wt.% for the total water content, 0.1–2.0 wt.% for the total CO_2 content, 0–20 vol.% for the initial phenocrysts, and 1–1000 s for the characteristic time of crystallisation. As we do not know the probability distribution of the uncertain input parameters, we have assumed a uniform distribution within the aforementioned ranges. The sensitivity analysis was performed on using the DAKOTA toolkit (Design Analysis Kit for Optimization and Terascale Applications)[66], an open-source software developed at Sandia National Laboratories that provides a flexible and extensible interface between analysis codes and iterative systems analysis methods such as uncertainty quantification, sensitivity analysis, optimization, and parameter estimation.

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Supplementary information is available in the online version of the paper.

Acknowledgements

The research leading to these results has received funding from the RCUK NERC DisEqm project (NE/N018575/1) and (NE/M013561/1). The beamtime on I12 was provided by Diamond Light Source (EE16188-1) and laboratory space by the Research Complex at Harwell.

Author Contributions

M.P., F.A., M.R.B., and P.D.L. conceived the research project. F.A., M.P., G.L.S., N.L.G., B.C., M.E.H., D.D.G., N.T.V., S.N., R.C.A., E.W.L., P.D.L. and M.R.B. contributed to the beamline experiments. F.A. collected the volcanic rocks for the starting material. D.D.G., H.M and R.A.B. prepared the starting material. F.A., M.P. and G.L.S performed image reconstruction. F.A. and M.P. performed image processing. F.A. performed image segmentation and analysis. G.L.S. performed simulations using the conduit model. R.A.B. and F.A. performed ex situ decompression experiments. E.C.B., F.A. and G.L.S. collected samples of the Etna 122 BC Plinian eruption. E.C.B. and F.A. acquired and analysed back-scattered electron images of Etna Plinian eruption's samples. F.A., G.L.S., M.R.B., M.P. and E.C.B. wrote the manuscript, with contributions from all other authors.

Author Information

Reprints and permissions information is available at www.nature.com/reprints. The authors declare no competing financial interests. Readers are welcome to comment on the online version of the paper. Correspondence and requests for materials should be addressed to F.A. (fabio.arzilli@manchester.ac.uk)

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636

637

638 **FIGURES**

639 Figure 1. Crystallisation through time. Reconstructed axial slices during continuous cooling at
640 0.4 °Cs^{-1}: (a) frame after 24 s from the onset of the cooling, in which the temperature ranges
641 between 1144 and 1112 °C (average 1128 °C); (b) frame after 208 s, in which the temperature
642 ranges between 1073 and 1034 °C (average 1054 °C); (c) frame after 392 s, in which the
643 temperature ranges between 997 and 959 °C (average 978 °C). m = melt; plg = plagioclase; px =
644 pyroxene.

645

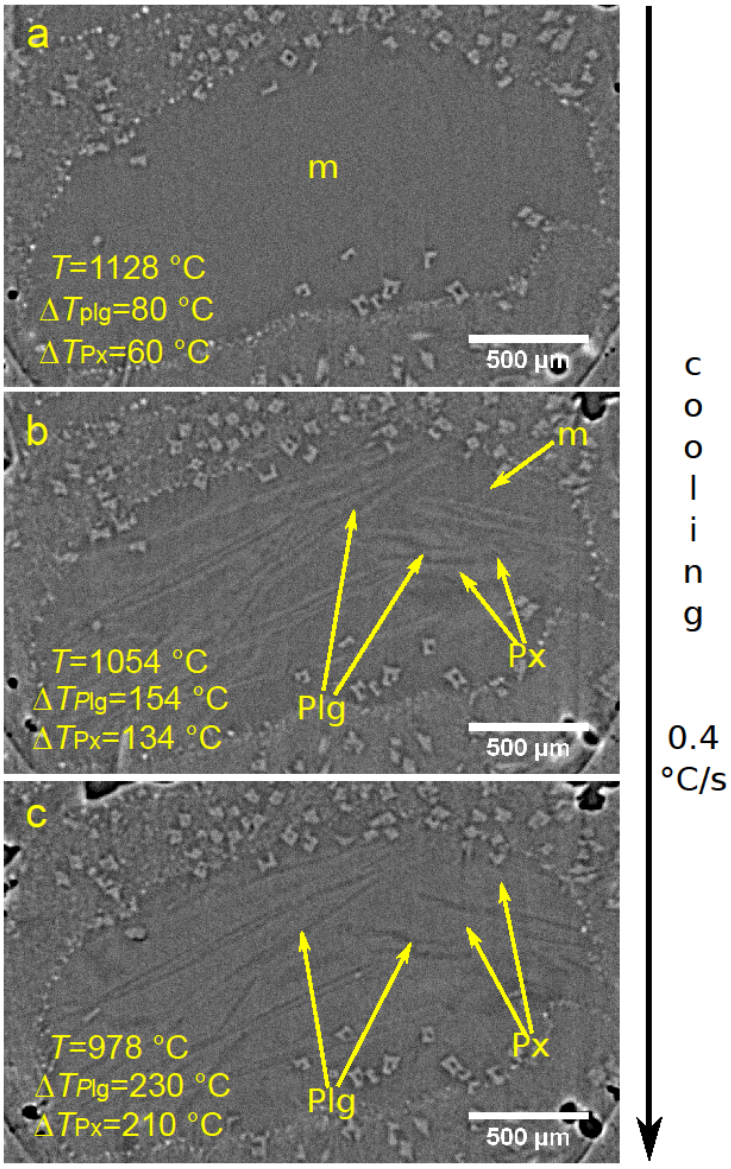
646 Figure 2. Plagioclase crystal morphology. (a) The 3D volume rendering of sample SS1150 shows
647 the morphology and the spatial distribution of plagioclase crystals that formed during the rapid
648 cooling at $96<\Delta T<155$ °C. (b) 3D view of the plagioclase with swallow-tailed crystal morphology.
649 (c) Back scattered electron image of plagioclase with swallow-tailed crystal morphology. (d) Back
650 scattered electron image of plagioclase with swallow-tailed crystal morphology produced during the
651 Etna 122 B.C Plinian eruption. Note heterogeneous nucleation of pyroxene around plagioclase, seen
652 as a light-coloured halo, and similar to that seen in figure 1b.

653

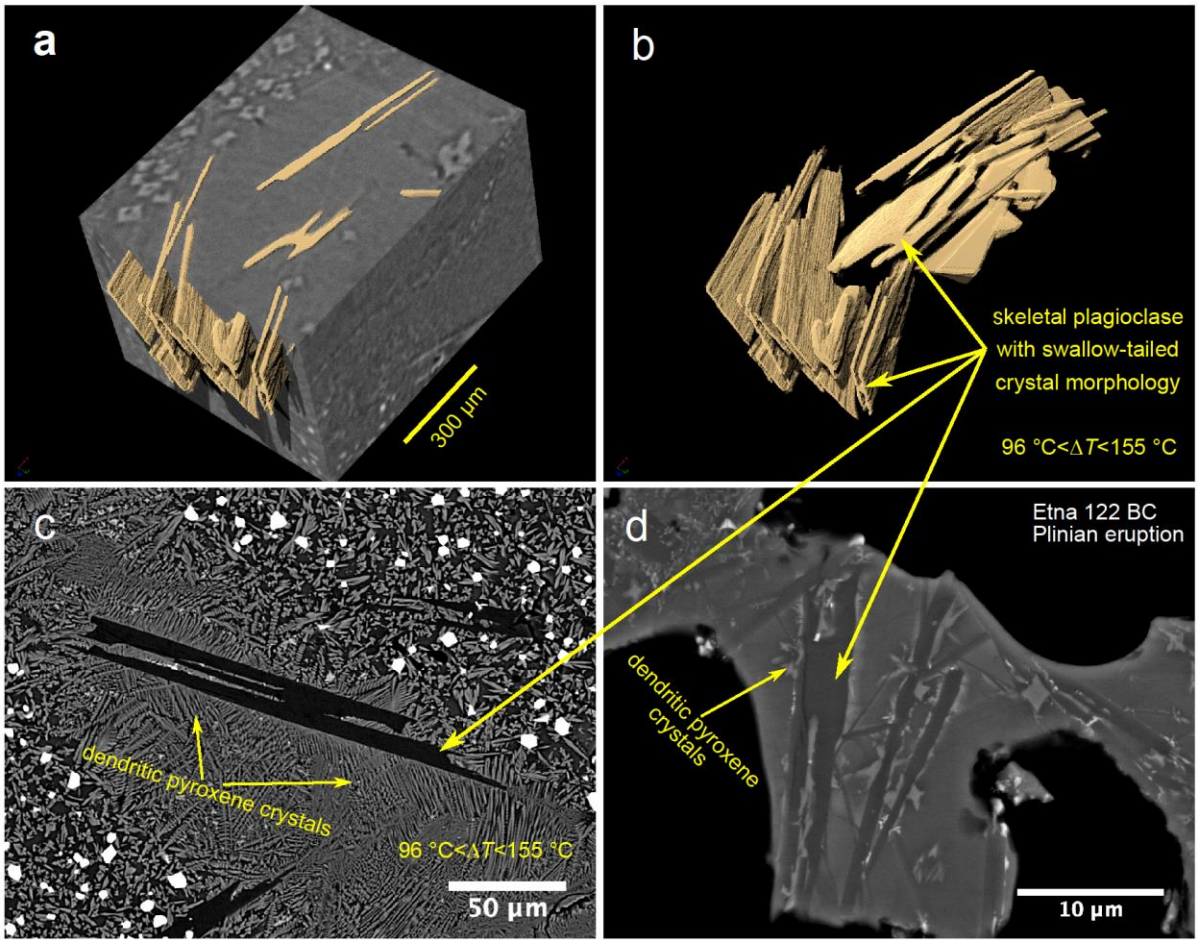
654 Figure 3. Model results during magma ascent. (a) Undercooling as a function of depth, calculated
655 for $\tau(c) = 10$ (blue) and 1000 s (red). Cooling is driven by adiabatic expansion of gas, mitigated by
656 latent heat of crystallisation particularly in the fast crystallising case. (b) Crystal content in vol%,
657 demonstrating the rapid increase in crystal load when $\tau(c) = 10$ s. (c) Magma viscosity,
658 demonstrating that the higher crystal load produces 3-4 order of magnitude increase in viscosity,
659 leading to fragmentation.

660

Figure 4. Relationships between characteristic time, initial temperature, initial H_2O content of the magma, syn-eruptive crystal content and magma viscosity and the undercooling of the system at the fragmentation level. These figures were calculated using repeated runs of the model while changing individual parameters to reveal the sensitivity of the system to each parameter. Likelihood of explosive eruption as a function of a specific parameter arises from the ratio between the number of model runs producing explosive eruptions divided by the total number of model runs used to test that parameter. Therefore, this is not a probabilistic assessment of eruption risk, but instead depends on the critical model parameters, which control when fragmentation occurs, and the calculated probabilities depend on the choice of upper and lower limits chosen for each investigated parameter. (a) Sobol index. (b) Undercooling vs magma temperature before ascent. (c) Undercooling vs the initial H_2O content of the magma (dissolved and exsolved). For a given input value, the solid lines are the mean of a given output parameter obtained from all the simulations assuming that input value. The error bars, instead, represent the spread of that output values with respect to the corresponding mean. (d) Frequency of explosive eruptions vs magma temperature before ascent. (e) Explosion frequency vs initial phenocryst content. (f) Frequency of explosive eruptions vs syn-eruptive crystal content at the fragmentation level. (g) Frequency of explosive eruptions vs magma viscosity at the fragmentation level.


a
m
T=1128 °C
ΔTplg=80 °C
ΔTPx=60 °C
500 µm
b
m
Px
T=1054 °C
ΔTPlg=154 °C
ΔTPx=134 °C
Plg
500 µm
c
Px
T=978 °C
ΔTPlg=230 °C
ΔTPx=210 °C
Plg
500 µm
cooling
0.4
°C/s

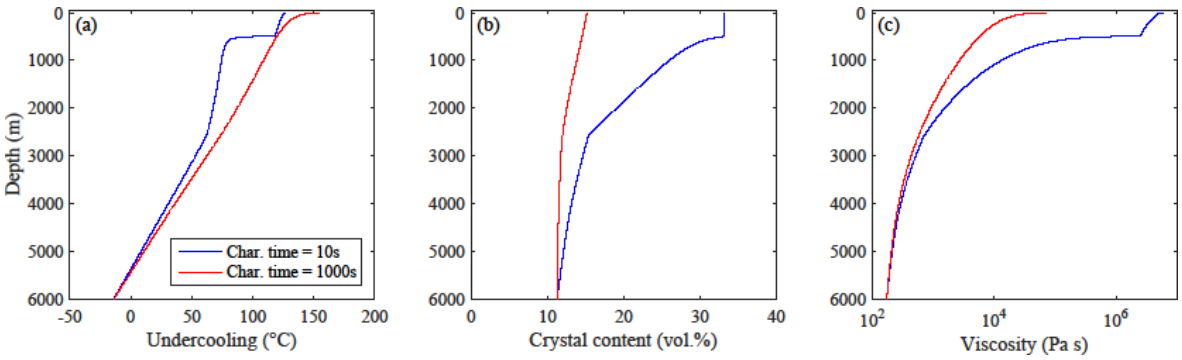
687 Figure1



688

689 Figure 2

690



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692 Figure 3

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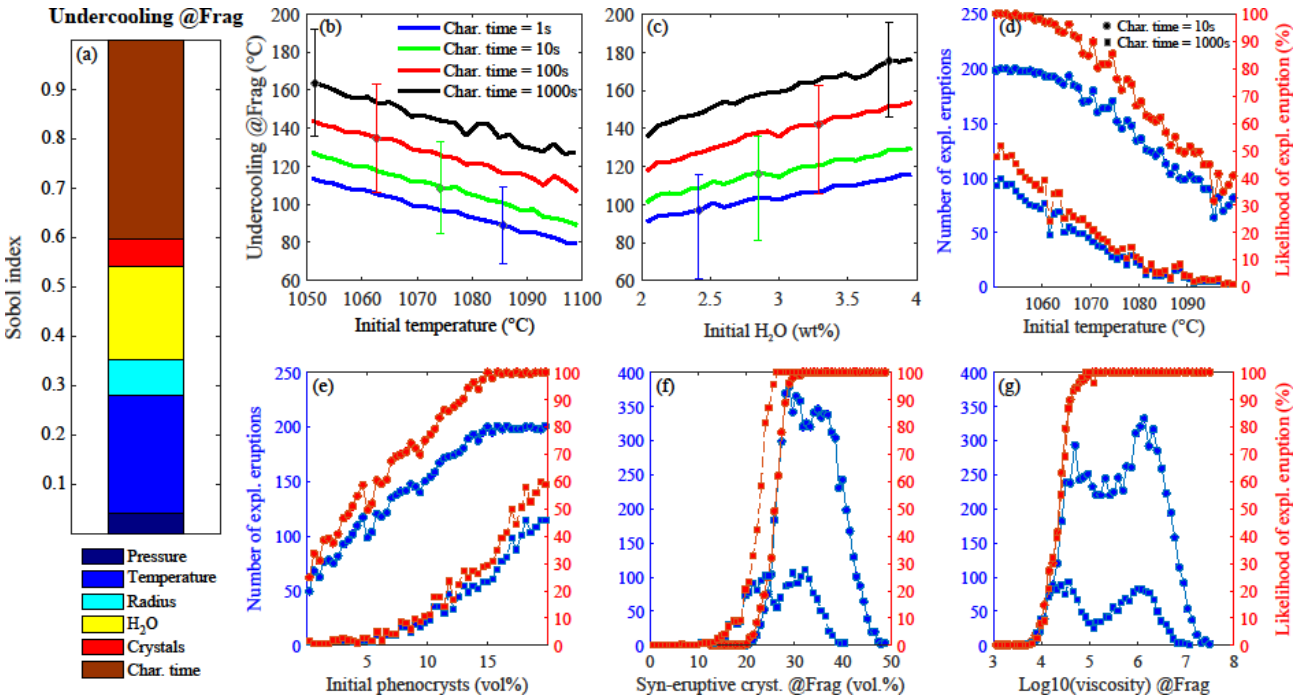
Undercooling @Frag
(a)
Sobol index
Pressure
Temperature
Radius
H2O
Crystals
Char. time
(b)
Char. time = 1s
Char. time = 10s
Char. time = 100s
Char. time = 1000s
Undercooling @Frag (°C)
Initial temperature (°C)
(c)
Initial H2O (wt%)
(d)
Char. time = 10s
Char. time = 1000s
Number of expl. eruptions
Likelihood of expl. eruption (%)
Initial temperature (°C)
(e)
Number of expl. eruptions
Initial phenocrysts (vol%)
(f)
Syn-eruptive cryst. @Frag (vol.%)
(g)
Likelihood of expl. eruption (%)
Log10(viscosity) @Frag

Figure 4