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Conduit dynamics of the Rungwe Pumice eruption (Tanzania): From storage to fragmentation of phonolitic-trachytic magmas

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Abstract

The eruptive style of magma is shaped by both storage conditions and ascent processes. Peralkaline melts, with total alkalis exceeding total aluminium, retain relatively high water concentrations and low viscosity and are therefore expected to better resist fragmentation than more common subalkaline melt compositions. However, trachytic and phonolitic magmas can still generate highly explosive eruptions, as demonstrated by the ~4 ka Rungwe Pumice Plinian eruption (Tanzania). This VEI 5 event involved a crystal-poor, microlite-free phonolitic/trachytic magma stored at high temperatures and, in comparison to most peralkaline magmas, with moderate water concentrations. 2D and 3D textural analyses, coupled with embayment speedometry, reveal a delayed homogeneous bubble nucleation event (ΔP_{sat} ~50 MPa) at shallow depths. Rapid bubble nucleation and growth during fast ascent (~6 MPa s^{-1}) prevented outgassing through a highly vesicular foam and consequently promoted strong melt-gas coupling which, combined with a sudden rheological shift, ultimately led to fragmentation. This eruption underscores the critical role of conduit dynamics in peralkaline magma explosivity, beyond storage conditions alone.

Muhitasari

Mtindo wa mlipuko wa magma huundwa na hali ya uhifadhi na michakato yake ya upandaji. Magma za peralkali zilizoyeyuka, ambazo jumla ya alkali zake inazidi jumla ya aluminiamu, hubakiza ndani yake viwango vya juu vya maji na mnato (viscosity) mdogo, kwa hivyo inatarajiwa kuwa na uwezo wa kustahimili kupasuka vipande vipande zaidi kuliko uyeyukaji wa magma za kawaida za subalkali. Hata hivyo, magma za trakitiki na fonolitiki bado zinaweza kusababisha mlipuko mkubwa sana, kama ilivyoainishwa na mlipuko wa Plinian Pumice wa Rungwe (Tanzania) uliotokea takriban miaka 4,000 iliyopita. Tukio hili lililokuwa na kiwango cha 5 (VEI 5) cha milipuko ya volcano, lilihusisha magma ya fonolitiki/trakitiki isiyo na fuwele (crystals) nyingi wala fuwele ndogo ndogo (microlite), iliyohifadhiwa katika halijoto ya juu sana, ikilinganishwa na magma nyingi za peralkalini, zinazokuwa na kiwango cha wastani cha maji. Uchambuzi wa muundo, kwa kutazama katika pande mbili na pande tatu (2D na 3D), pamoja na upimaji wa kasi ya ghuba (embayment speedometry), unaonyesha uchelewaji katika uundaji wa mfuko wa povu ulio sawa (ΔP_{sat} ~50 MPa) katika kina kifupi. Uundaji na ukuaji wa haraka wa povu (buble nucleation) wakati wa kupanda kwa kasi (~6 MPa s^{-1}) ulizuia kutolewa kwa gesi kupitia povu lenye vifereji vingi vya hewa ya ndani na hivyo kusababisha muunganiko mkubwa wa magma zilizoyeyuka na gesi, ambao, ulijumuishwa na mabadiliko ya ghafla ya rheolojia, hatimaye ukasababisha kupasuka vipande vipande. Mlipuko huu unasizitiza kuwa uwezo wa kulipuka wa magma za peralkaline hautokani na hali ya uhifadhi pekee bali pia na mienendo ya mipasuko ya miamba.

Keywords: Explosive eruptions; Peralkaline magmatism; East African Rift; Rungwe volcano.

1 Introduction

Plinian-style explosive eruptions rank among the most severe natural hazards, with impacts ranging from regional to global scales [e.g. Carey and Sigurdsson 1989; Cioni et al. 2015]. These eruptions are governed by a combination of pre-eruptive magmatic conditions—particularly water concentrations and melt composition [e.g. Wilson et al. 1980; Cioni 2000]—and the dynamic processes occurring within the conduit during magma ascent. Pre-eruptive factors, such as volatile saturation, magma storage depth, and temperature, set the stage for eruption potential by controlling bubble nucleation, crystallisation, and magma rheology [Cashman and Mangan 1994; Dingwell 1996; Gonnermann and Manga 2007; Cassidy et al. 2018]. For Plinian eruptions, however, it is within the conduit that magma undergoes a dramatic transformation, from a hot, pressurized silicate melt to a high-energy bubble suspension capable of sustaining explosive fragmentation [Sparks 1978; Wilson et al. 1980; Cashman and Mangan 1994].

As magma ascends and decompresses within the conduit, volatile solubility—primarily H_2O and CO_2—decreases significantly, with volatiles eventually reaching saturation. However, because nucleation is energetically demanding, bubbles will

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form only after a threshold level of volatile supersaturation is exceeded, and this threshold depends on magma properties such as surface tension and volatile diffusivity in the melt [e.g. Mourtada-Bonnefoi and Laporte 2004; Shea 2017, among others]. When suitable nucleation surfaces are available, such as those of micro- or nanolites inherited from the reservoir or generated during ascent, bubbles can nucleate at relatively lower supersaturation pressures (i.e. the difference between magmastatic pressure at saturation and the pressure at nucleation), resulting in heterogeneous nucleation [Shea 2017; Cáceres et al. 2022]. In contrast, when such surfaces are absent, bubbles nucleate spontaneously through homogeneous nucleation, but only at substantially higher degrees of supersaturation [Mourtada-Bonnefoi and Laporte 2004; Shea 2017; Buono et al. 2020]. In more exceptional cases, if significant crystallisation occurs at the reservoir level, incipient degassing may initiate magma ascent at greater depths due to second boiling [Edmonds and Woods 2018].

Following nucleation, bubbles grow by volatile diffusion and decompression and if exsolved volatiles cannot efficiently escape the system—as commonly occurs in viscous and/or rapidly ascending magmas [e.g. Cassidy et al. 2018]—they drive progressive vesiculation of the melt-gas-crystal mixture [e.g. Sparks 1978; Proussevitch et al. 1993]. This process further accelerates the magma and enhances bubble overpressure, shear stresses and strain rates within the conduit, ultimately promoting magma fragmentation [e.g. Dingwell 1996; Zhang 1999].

Understanding the conduit dynamics that dictate the transition between effusive and explosive eruptive style is therefore crucial for predicting eruption explosivity and associated hazards. These dynamics have been extensively studied across a variety of magmatic compositions. For fast-ascending, highly viscous silicic magmas, homogeneous nucleation under high supersaturation conditions is commonly inferred [e.g. Shea 2017, and references therein], which eventually leads to strain rate-induced brittle failure [Gonnermann 2015, and references therein]. In contrast, for less-viscous basaltic magmas, heterogeneous nucleation, abrupt rheological changes, or external perturbations (e.g. magma-water interactions) have been suggested to govern eruptive explosivity [e.g. Giordano and Dingwell 2003; Houghton and Gonnermann 2008], with ascent rate having a dominant control on preventing outgassing despite high bubble connectivity and permeability [e.g. Houghton et al. 2004; Burgisser et al. 2017; Bamber et al. 2024].

Peralkaline felsic magmas exhibit distinctive physicochemical properties that strongly influence their eruptive behaviour [e.g. Polacci et al. 2004; Shea 2017]. With high silica contents ($SiO_2 \geq 60$ wt.%) and substantial dissolved water (up to 8 wt.% at 200 MPa [e.g. Carroll and Blank 1997; Di Matteo et al. 2004]), these magmas can ascend rapidly within the conduit through density-driven isostatic processes [Browne and Szramek 2015]. Their ascent is further facilitated by viscosities comparatively lower than those of calc-alkaline melts (up to 2-3 orders of magnitude) under comparable T–P–H_2O conditions [Di Genova et al. 2013]. This reduced viscosity reflects the high abundance of alkalis and halogens, which weaken the silicate network by breaking Si–O bonds and thus increase the proportion of non-bridging oxygens, producing a depolymerisation effect that is more pronounced than in peraluminous melts [Di Genova et al. 2013]. In addition, potentially high dissolved volatile contents, owing to their elevated solubility, further enhance melt depolymerisation. Increased melt depolymerisation enhances volatile diffusivity, facilitating bubble nucleation and growth and thereby reducing supersaturation pressures and bubble overpressures; however, it also promotes vigorous magma foaming, leading to substantial acceleration as vesiculation increases, further favoured by reduced frictional resistance along conduit walls.

Collectively, these conditions can favour efficient outgassing and, potentially, effusive to mildly explosive activity. Yet rapid ascent rates may suppress outgassing efficiency, instead promoting more energetic explosive behaviour, consistent with the widespread occurrence of sub-Plinian to Plinian eruptions of peralkaline felsic magmas [e.g. Pappalardo and Mastrolorenzo 2012; Hughes et al. 2017; Shea 2017; Pappalardo et al. 2018; Stabile et al. 2021; Wallace et al. 2025]. Such explosive events are frequently accompanied by coeval lava effusion, underscoring the need to investigate the eruptive dynamics governing fluctuations in eruptive style within these distinctive magmatic systems [Andújar and Scaillet 2012; Shea 2017].

Numerical conduit models provide a means to describe and quantify variations in key physical parameters during magma ascent, including pressure, temperature, viscosity, density, ascent velocity, bubble volume fraction, and bubble overpressure within the melt-gas-crystal mixture [e.g. Campagnola et al. 2016; La Spina et al. 2017; Aravena et al. 2018]. These parameters ultimately govern magmatic fragmentation and eruptive style. However, the development and calibration of such models require critical constraints that can be obtained from detailed analyses of natural and experimental products of explosive fragmentation, such as pumiceous rocks [e.g. Shea et al. 2010; Campagnola et al. 2016]. These products, formed by rapid quenching during fragmentation, preserve in their chemistry and microstructures—particularly vesicularity and crystallinity—a snapshot of the final conduit conditions immediately prior to fragmentation [Cashman and Mangan 1994; Shea et al. 2010]. Their detailed investigation enables the reconstruction of conduit processes, including magma decompression rates, supersaturation pressures, and fragmentation criterion.

Consequently, the methods commonly used to investigate the mechanisms and timescales of magma ascent in volcanic conduits are those based on bubble formation or crystallisation kinetics, as well as on the diffusion of molecules or elements in the melt or in crystals. These approaches rely primarily on the textural and geochemical investigation of eruptive products, respectively [e.g. Costa et al. 2020; Toramaru 2025]. Within this framework, several recent studies have focused on the diffusivity of fast-diffusing volatile elements (such as H and Li) in minerals or melt embayments (i.e. open melt pockets within crystals). In particular, the study of H_2O diffusion in embayments provides an effective way to investigate magma ascent timescale and fragmentation depth [e.g. Costa et al. 2020; Hosseini et al. 2023; Hosseini and Myers 2024]. At the same time, textural characterisation of bubbles or microlites (e.g. number density, size distribution, shape and orientation; porosity and permeability) allows the determination of magma ascent rates [e.g. Toramaru 2006; Toramaru et al. 2008; Toramaru 2025]; additionally, pumice textures provide valuable indicators of several mechanisms such as degassing dynamics, outgassing efficiency, or fragmentation—information otherwise inaccessible—and allow recognition of changes in intra-conduit vesiculation processes that produce distinct pumice types. Such insights are especially important for kinetically reactive magmas like those of peralkaline compositions, where vesiculation and outgassing processes evolve rapidly during ascent [Shea 2017].

This study focuses on conduit processes, highlighting their role in shaping the eruptive behaviour of the Rungwe Pumice (RP) eruption and providing insights into the mechanics of sustained explosive activity of phonolitic-trachytic magmas. The RP eruption offers a unique opportunity to explore the interplay between magmatic dynamics and eruptive processes.

We employ an integrated approach combining traditional 2D textural analysis methods [Cashman and Mangan 1994; Shea et al. 2010; Gurioli et al. 2015] with advanced 3D imaging techniques using X-ray computed microtomography (μXCT [Polacci et al. 2010; Giachetti et al. 2011; Hughes et al. 2017; Pappalardo et al. 2018; Valdivia et al. 2022; Buono et al. 2023; Pappalardo et al. 2023; Torres-Orozco et al. 2023; 2024]. While 2D imaging methods provide valuable insights into vesicle characteristics, they are limited in their ability to represent real textural features and require stereological corrections [Shea et al. 2010]. Moreover, 2D analyses cannot fully resolve permeability or vesicle network connectivity [Giachetti et al. 2011]. In contrast, μXCT allows for direct, non-destructive 3D observations of these features. However, μXCT accessibility may restrict the number of analysed samples and limit representativity. In complement to these imaging analyses, we utilised embayment water diffusion speedometry [Liu et al. 2007] to cross-check ascent rates estimated from textural data. By combining these techniques, this study aims to provide a comprehensive understanding of the conduit dynamics that sustained the explosive activity of peralkaline phonolitic-trachytic magma during the RP eruption.

2 Geological Background

The Rungwe Volcanic Province (RVP), covering over 1500 km^2, is located in southern Tanzania, within the Western branch of the East African Rift (EAR), and comprises three main Holocene volcanic centres—Rungwe, Ngozi, and Kyejo—surrounded by several smaller eruptive centres [Figure 1A; Harkin 1960; Ebinger et al. 1993; Fontijn et al. 2010a]. Volcanic activity in the RVP began at approximately 9 Ma, alternating between effusive and explosive eruptions of high-alkali magmas [Fontijn et al. 2010b; 2012].

Rungwe, the largest volcanic edifice in the RVP, is a relatively young stratovolcano (0.25 ± 0.01 Ma, based on whole-rock K-Ar dating on a lava flow at the base of the edifice [Ebinger et al. 1989]) centrally located in the province (Figure 1B). It predominantly produced magmas of basaltic, phonolitic, and trachytic compositions. Over its history, Rungwe is estimated to have experienced at least one explosive eruption every 500 years, ranging from violent Strombolian to Plinian events [Fontijn et al. 2010b]. Among these, the Rungwe Pumice Plinian eruption stands out as the largest event in the province's Holocene history [Fontijn et al. 2011].

2.1 Rungwe Pumice Eruption

The RP deposit consists of a massive pumice lapilli fall deposit that blankets the entire RVP. Radiocarbon dating and sedimentary records from Lake Masoko place the eruption at ~4000 cal BP [Garcin et al. 2006; Fontijn et al. 2010b]. The deposit is lithic-poor and predominantly consists of cream-coloured, highly vesiculated pumice lapilli of high-alkaline, trachytic composition (whole-rock [Fontijn et al. 2013]). Alkali feldspar is the primary free mineralogical component, followed by subordinate amounts of biotite, clinopyroxene and Fe-Ti oxides. Notably, cyan-coloured haüyne occurs in the RP deposit, distinguishing it from most other RVP deposits and serving as a field marker for identifying RP outcrops [Fontijn et al. 2011].

The deposit extends on land up to 28 km from the volcano summit, where a 30 cm-thick outcrop has been documented [Fontijn et al. 2011]. More distal RP tephra has been recovered from lake sediment cores up to ~115 km SE of Rungwe [Garcin et al. 2006; Fontijn et al. 2011; 2012]. On land, outcrops are radially distributed around the summit, and near-circular isopleths indicate wind-still conditions during the eruption [Fontijn et al. 2011]. A peak eruptive column height of 30-35 km was inferred using the maximum lithics' size method [Carey and Sparks 1986; Fontijn et al. 2011], and the absence of associated pyroclastic density current deposits suggests that this column was sustained for most of the eruption. Peak mass discharge rates of 2.8–4.8 × 10^8 kg s^{-1} were calculated using a range of inferred column heights, and the minimum erupted volume was estimated at 1.4 km^3 dense rock equivalent [Fontijn et al. 2011]. Based on these parameters, the eruption was classified as Plinian, with a Volcanic Explosivity Index of 5 [Fontijn et al. 2011], following the classification of Newhall and Self [1982].

A type section for RP deposit was identified at ~11.7 km SSE of the Rungwe summit and designated as KF176 in previous studies (Figure 1B [Fontijn et al. 2010b; 2011; 2013]). The section comprises a ~2.5 m-thick, massive pumice lapilli breccia deposit reversely graded at the base and bounded by overlying and underlying palaeosols. The deposit was previously sampled through its entire thickness by Fontijn et al. [2011] and subdivided into 14 samples every 20–25 cm from base to top and labelled sequentially from KF176-B to KF176-O, whereas KF176-A corresponds to the basal palaeosol.

The RP plumbing system has been described as a relatively hot reservoir, ranging from 925 to 975 ± 22 °C [Fontijn et al. 2013; Cappelli et al. 2025], of trachytic-phonolitic composition (based on haüyne-hosted melt inclusions major element concentrations). Prior to the eruption, it ponded at shallow crustal depths (at ~3.5 km minimum depth based on saturation pressure models derived from haüyne-hosted melt inclusions; [Cappelli et al. 2025]) and yielded on average ~4.82 ± 0.58 wt.% of dissolved water, while CO_2 was below FTIR detection limit (<10–100 ppm [Cappelli et al. 2025]). The system was likely destabilised by an input of volatile-rich magma, which in-

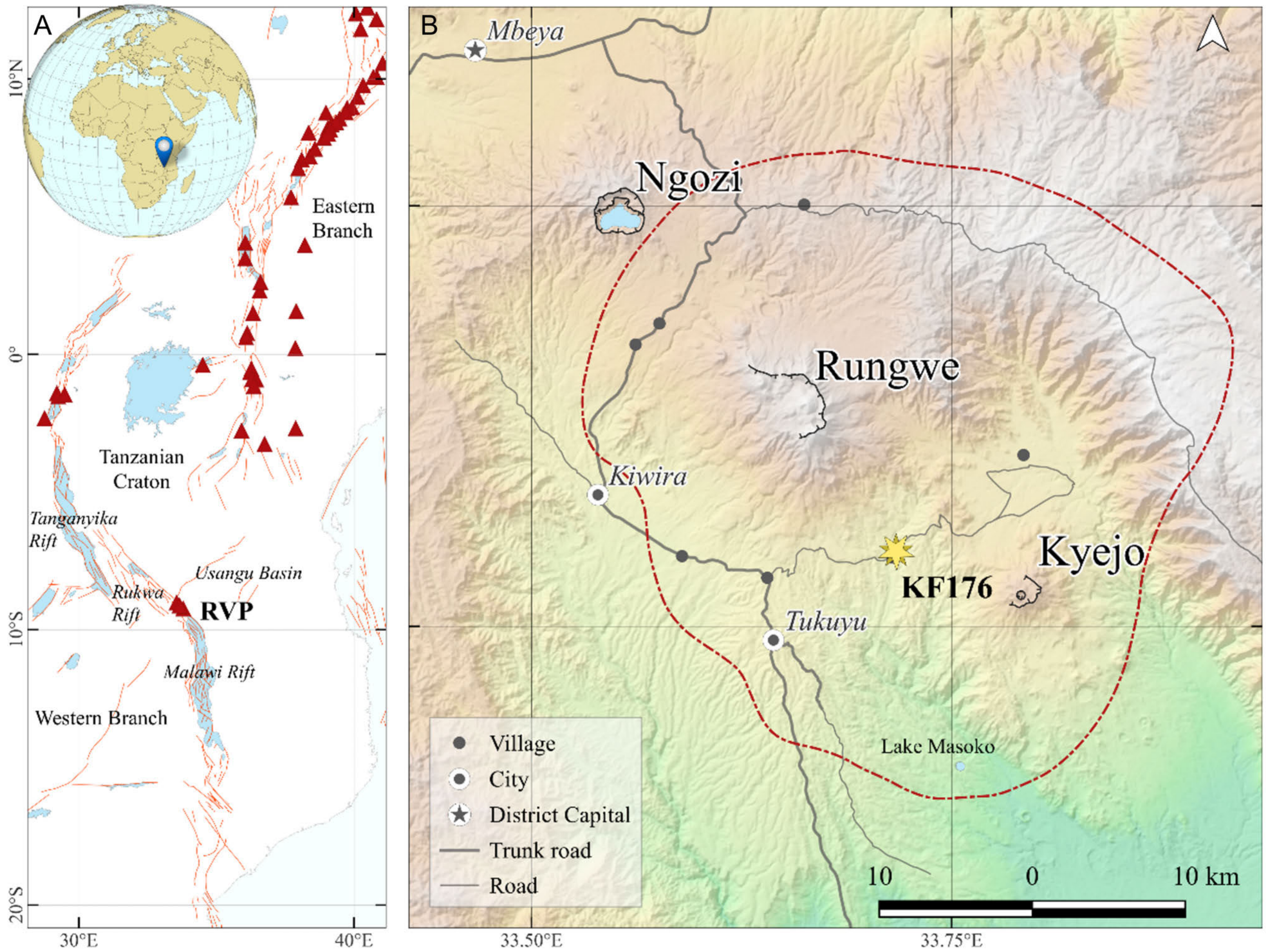

Figure 1: [A] Overview of the East African Rift, showing the location of the Rungwe Volcanic Province (RVP) along the Western Branch. Major rift faults (red lines) and volcanic centres (red triangles) are indicated. [B] Close-up of the main volcanic centres within the RVP, highlighting the location of the type section KF176 (yellow star). The red dashed contour represents the 25 cm isopach of the Rungwe Pumice deposit (modified from Fontijn et al. [2011]).

creased oxygen fugacity, ultimately leading to the eruption [Fontijn et al. 2013; Cappelli et al. 2025].

3 Methods

Samples from the Rungwe Pumice eruption type section KF176 were dry-sieved down to 63 µm and subdivided into granulometric size classes at ϕ/2 intervals ($\phi = -\log_2[diameter(\mathrm{mm})]$). To facilitate data interpretation, samples were grouped into five stratigraphic horizons: base (KF176 C-D), bottom half (KF176 E-F-G), middle (KF176 H-I-J), top half (KF176 K-L-M), and top (KF176 N-O). Subsequent analyses kept this partitioning scheme and targeted specific samples as representative of each horizon.

3.1 Embayment diffusivity speedometer

Magma decompression rates were estimated based on the decompression-induced diffusivity-dependent decrease in volatile concentrations within crystal-hosted melt (glass) embayments. Volatile concentration tends to decrease from the innermost portions of embayments to their outlets, which remain in contact with the external melt and undergo re-equilibration of volatile saturation during ascent [e.g. Liu et al. 2007; deGraffenried and Shea 2021; Geshi et al. 2021; Hosseini et al. 2023].

Accurate inspection of glass embayments was conducted in haüyne crystals, the mineralogical phase hosting the largest number and best-developed melt inclusions among all crystals from the RP eruption [Cappelli et al. 2025]. Glass embayments that best preserve volatile concentration gradients tend to exhibit a cylindrical shape, with minimal necking near the outlet and no significant irregularities [deGraffenried and Shea 2021; Hosseini et al. 2023]. They should also lack decompression bubbles in their internal regions while maintaining contact with an external bubble (or its remnant shape) at the outlet [Hosseini et al. 2023]. Approximately 25 crystals, handpicked from crushed pumices, were initially selected for measurement. These were collected from the bottom half, middle, top half and top stratigraphic horizons of the RP deposit. Suitable embayments could not be retrieved from the base of the deposit due to the scarcity of haüyne crystals in this stratigraphic portion.

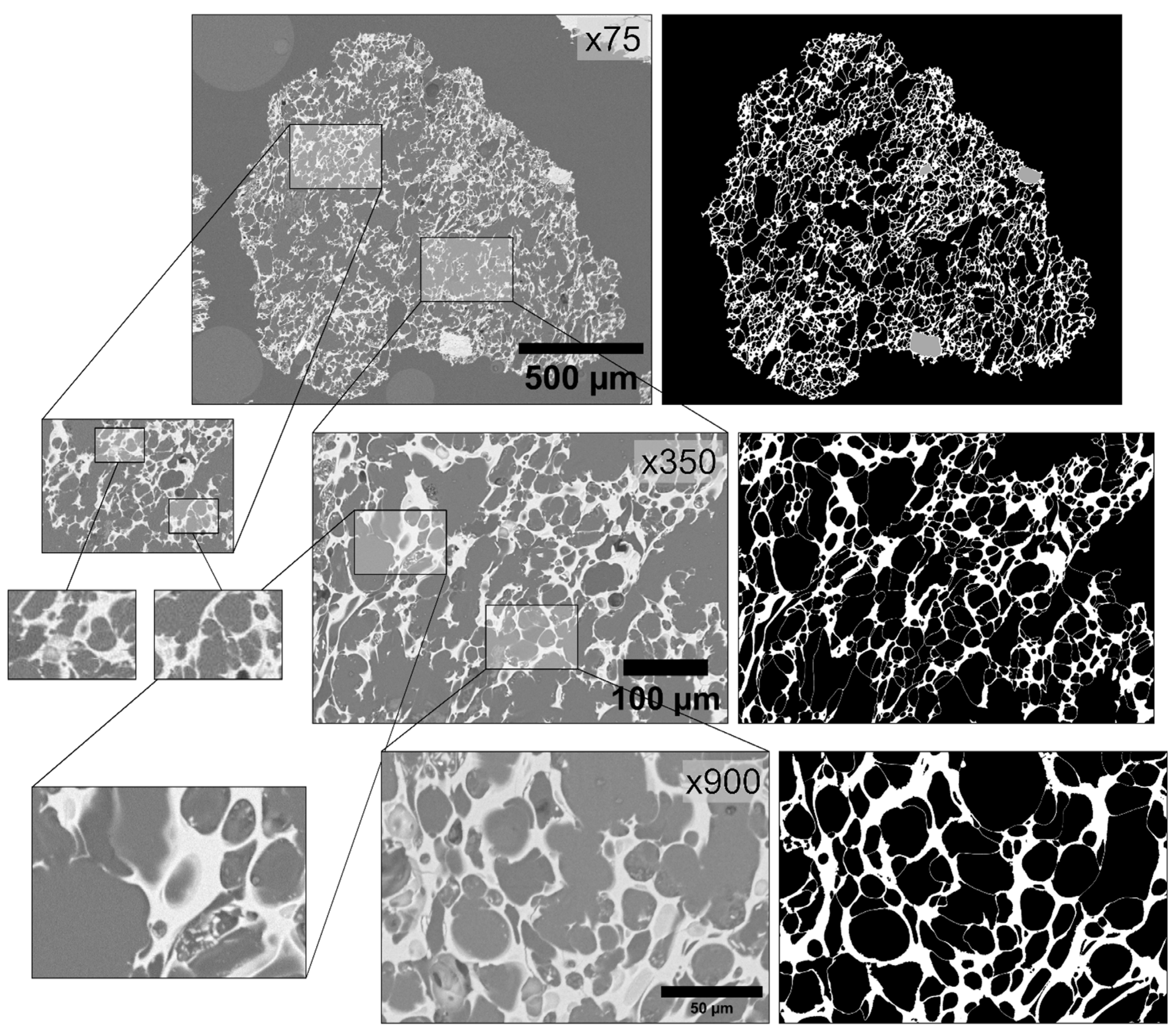

Figure 2: Image nesting method adopted for stereological conversion in FOAM, following the approach suggested by Shea et al. [2010] over a standard RP pumiceous clast from the middle stratigraphic horizon. For each clast, two images were selected at ×350 magnification, with four additional images collected at ×900 magnification within these regions. Grey-scale SEM images are shown alongside their binarized, decoalesced counterparts, where white represents pumice glass and black denotes vesicles (voids). In the ×75 magnified inset, also the exterior background appears black (removed during FOAM processing), and phenocrysts are displayed in grey.

The crystals were individually embedded in Crystalbond™ thermosetting resin and ground to expose the embayment(s). Afterwards, the resin was melted, and crystals were cast all together in an epoxy resin mount and mechanically polished using diamond pastes down to 1 µm. Water in embayments' glass was analysed with Raman spectroscopy using a Horiba Jobin LabRAM HR Evolution at KU Leuven (Belgium). Samples were irradiated with an Nd-YAG-sourced laser, maintaining low laser power (≤50 %) to prevent overheating of the resin—particularly in thin embayments—and to avoid potential damage to the crystals. Scattering spectra were acquired in the 150– 4000 cm^{-1} wavenumber range and the `SilicH2O` open-source software [Van Gerve and Namur 2023] was used to perform the baseline correction and to extract the peak areas. To account for potential interference in cases where embayments were too thin, and noise from underlying phases was recorded, spectra of the host crystal and pristine resin were also acquired and used to correct the glass spectra [Van Gerve and Namur 2023]. Then, water concentrations in the glass were quantified using the calibration feature of `SilicH2O`. A calibration curve was constructed by correlating the areas of the silica peaks (200–600 cm^{-1} and 800–1300 cm^{-1} ranges) with the area of the water vibrational peak (~3500 cm^{-1}). This calibration was based on spectra of melt inclusions previously analysed by FTIR spectroscopy, for which water contents were independently determined [Cappelli et al. 2025]. In total, only nine glass embayments from across the deposit—mostly from its upper portion,

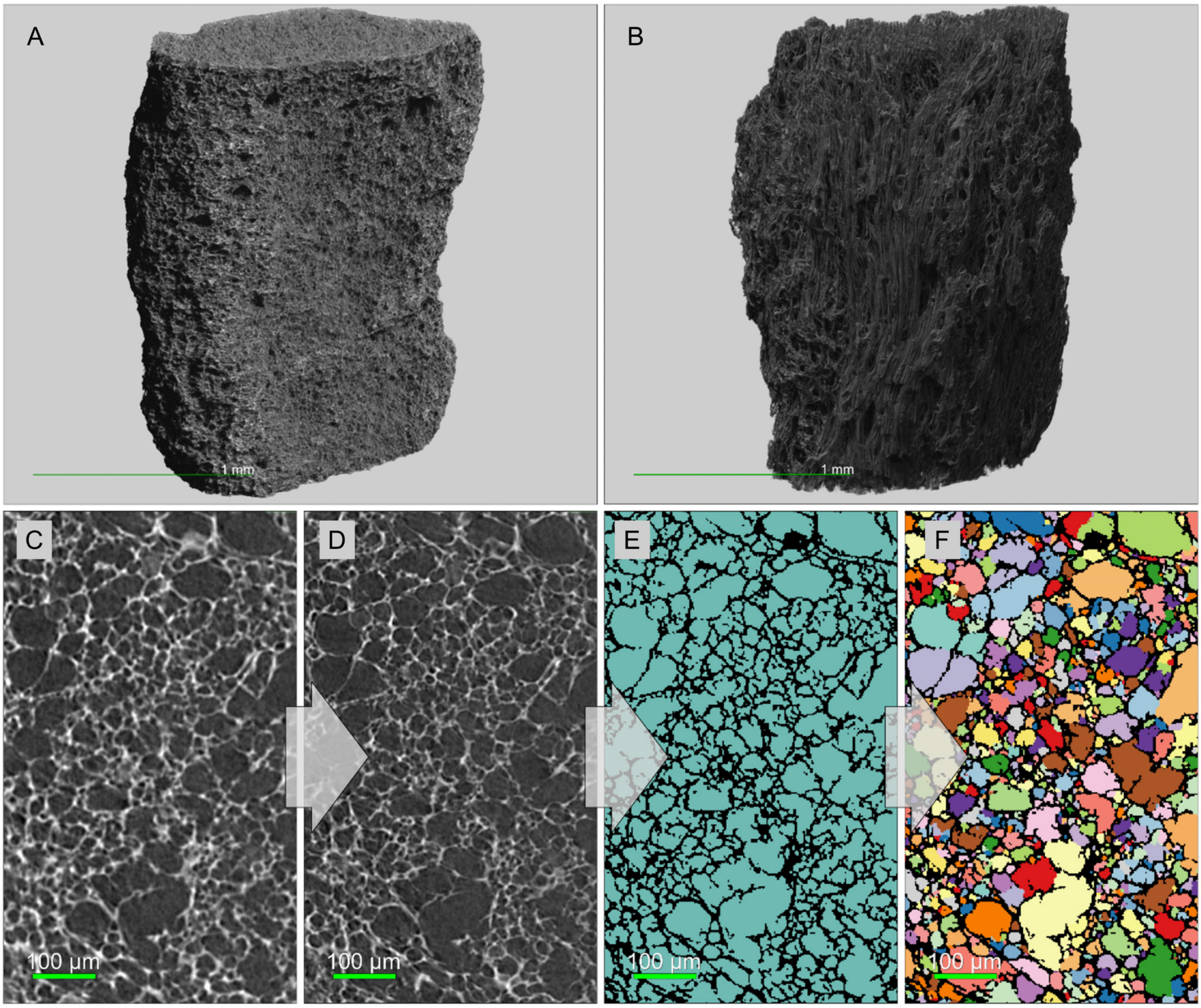

Figure 3: [A] and [B] µXCT volume reconstructions of a "standard" pumice from the middle horizon and the tube pumice, respectively, created in Dragonfly; [C] 2D slice of a sample volume, where glass appears as light grey and vesicles as dark grey, processed through the following steps: [D] application of the `U-Net2D` super-resolution model, [E] segmentation using the `U-Net2.5D` segmentation model (vesicles shown in cyan and glass in black), and [F] vesicle separation using the watershed algorithm.

where haüyne is more abundant, but also from the middle and bottom half intervals—were successfully measured.

An embayment speedometer was developed by Liu et al. [2007] and subsequently elaborated by several authors using different input parameters and coding languages. In this study, we adopted the publicly available `EMBER` software, written in `MATLAB` [Georgeais et al. 2021]. `EMBER` estimates decompression rates by comparing modelled volatile diffusion profiles with the measured concentration gradients in embayments, thereby determining the best fit for decompression rates, initial dissolved concentrations, and initial exsolved gas content. It can determine decompression rates from H_2O, CO_2, and S concentration gradients, however, in our case only water concentrations were available and therefore other volatiles were excluded. Notably, RP glass, both in haüyne-hosted melt inclusions and embayments, was depleted in CO_2, falling below the FTIR detection limit, indicating a minimal effect on water solubility [Cappelli et al. 2025]. Input parameters include: i) temperature, assumed constant throughout the ascent and equivalent to RP storage temperature, previously estimated at 975 °C on average [Cappelli et al. 2025]; ii) initial pressure, defined as the saturation pressure corresponding to the H_2O concentration plateau in the innermost portions of each embayment, calculated using the trachytic solubility model of Di Matteo et al. [2004]; and iii) the final pressure before melt quenching, which was iteratively varied between 0.5 and 25 MPa to achieve the best correlation (minimum squared error). Degassing paths provided as inputs for `EMBER` were calculated for seven initial exsolved gas contents (between 0 and 3.2 wt.%; see also Georgeais et al. [2021]) under closed-system conditions using `VESIcal v1.2.6`. [Iacovino et al. 2021] and `Thermoengine Python` libraries [Ghiorso and Gualda 2015], both accessible on ENKI servers. A wide range of exsolved gas contents was considered to evaluate any relevant effect on

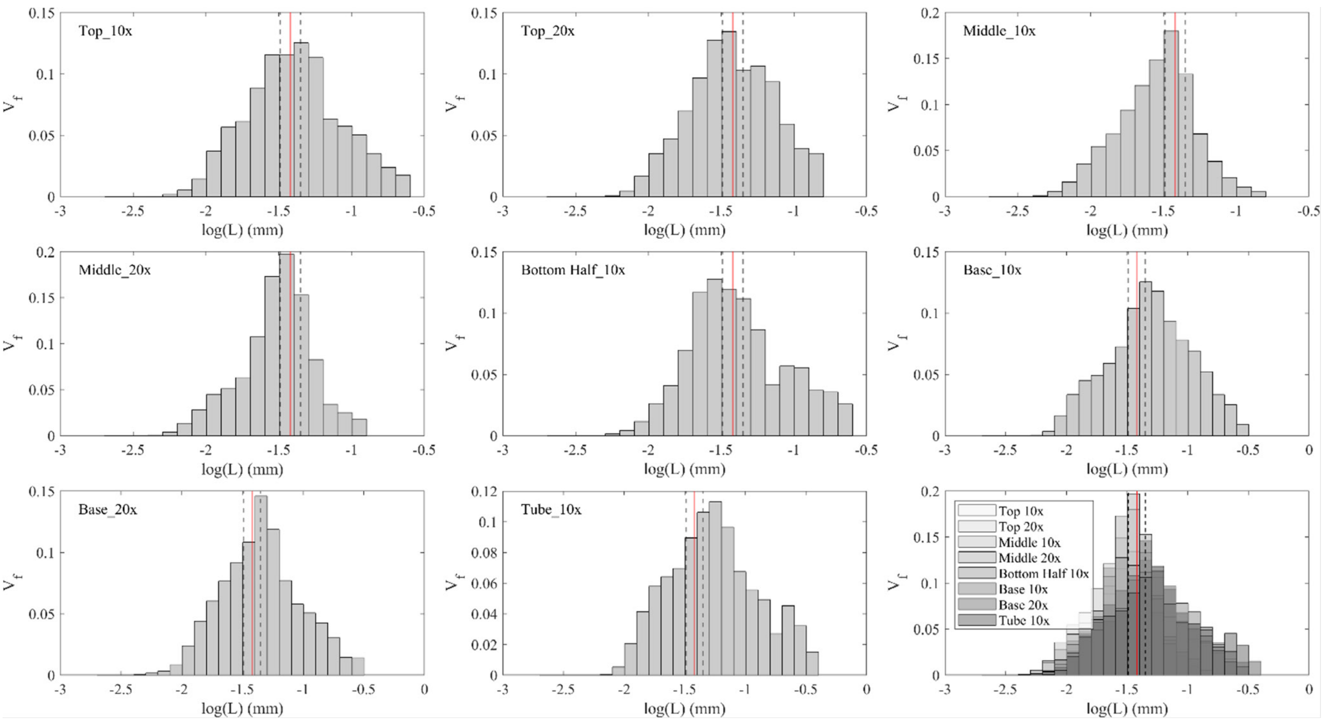

Figure 4: VVD plots for each μXCT dataset, presented individually and collectively for comparison. Most datasets exhibit lognormal, unimodal, distributions except for bottom half horizon and the tube pumice, which display a mild bimodality characterized by a secondary mode in the larger size range. This secondary mode may indicate the influence of coalescence effects on size distributions (see text for further details). The average of principal modes (red lines) corresponds to an equivalent diameter of 39 μm, with standard deviation indicated by grey dashed lines.

interpolations as in Georgeais et al. [2021], however, as suggested by melt inclusion analyses [Cappelli et al. 2025], a 0 wt.% initial gas was preferred.

Water diffusivity in EMBER is calculated with three different models according to glass composition: basaltic, rhyolitic and intermediate [Georgeais et al. 2021]. For the RP composition, we applied the diffusivity model for intermediate compositions using the equations from Ni and Zhang [2018], which are tested for calc-alkaline compositions. We acknowledge that this assumption inevitably introduces some uncertainty; however, diffusivity values calculated under identical conditions using equations more appropriate for trachytic [Fanara et al. 2013] or phonolitic compositions [Fanara et al. 2013; Schmidt et al. 2013] produced results of the same order of magnitude (10^{-10} $m^2\,s^{-1}$). To our knowledge, no published embayment speedometer specifically tested for (per)alkaline magmas currently exists. Decompression rates estimated with EMBER were then compared with textural methods (see Section 3.4).

Silica content (wt.%), and other major element concentrations, in embayment glasses were measured using a Tescan Mira 4 FEG scanning electron microscope equipped with an Oxford Xplore30 EDX detector at the KU Leuven Core Facility (Belgium). Samples were carbon-coated to a precise thickness of 10 nm, and analysed using an acceleration voltage of 20 keV and a beam current of 6 nA. The Beam Measurement calibration routine integrated into the Oxford Aztec software was used to calibrate Mn elemental concentrations on standard material, ensuring accurate absolute concentrations of major elements. Measurements were performed over areas rather than single spots to minimise the loss of alkali elements, and each embayment was measured 3–10 times to ensure consistency and account for potential variations in glass composition near the embayment outlet. Measurement accuracy was validated by analysing the ATHO-G, StHs6/80-G, and T1-G glass standards [Jochum et al. 2006] multiple times at the start and end of the analytical session. Additionally, glasses of RP crystal-hosted melt inclusions previously analysed with electron probe microanalysis [Cappelli et al. 2025] were used as reference material. For silica concentrations, standards and reference materials yielded a maximum percentage difference to expected values of 0.7% and 0.3% respectively, with a relative standard deviation between measurements of no more than 0.9% (Supplementary Material 1).

3.2 2D pumice textures

Textures of RP pumice were first evaluated on 2D images of polished sections. Ash particles within the 0/−0.5ϕ (1–1.4 mm) grain size range from each horizon were rinsed in an ultrasonic bath and then embedded in epoxy resin. A fine grain size range minimises the potential effect of post-fragmentation inflation of gas bubbles that could alter original vesicularity in slower-cooling, larger clasts [Kaminski and Jaupart 1997; Pappalardo et al. 2018]. Moreover, isolated voids in coarser samples often prove difficult to fully impregnate with resin, complicating image processing [Shea et al. 2010]. For comparison, coarse pumice lapilli vesicularity was also inves-

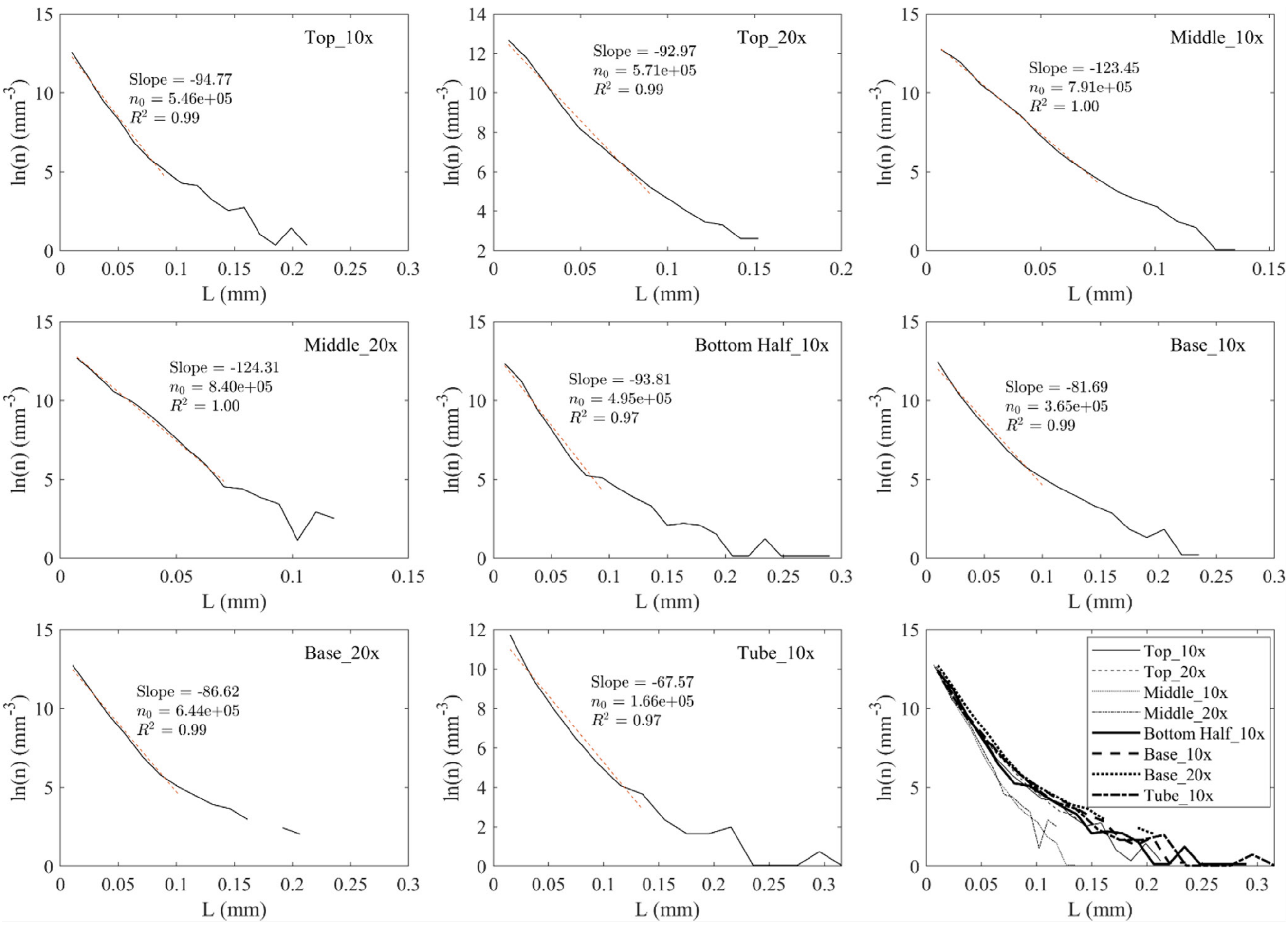

Figure 5: VSD plots for each µXCT dataset, presented individually and collectively for comparison. The slope, intercept at n_0, and goodness of fit (dashed red lines) for the linear interpolation within the smallest size range are provided for each sample.

tigated by combined gas and water displacement pycnometery (based on Archimedes's principle; methodology and results are presented in Supplementary Material 2). Particles were ground until they were sliced to about half their original volume, then polished with diamond pastes down to 1 µm and finally carbon-coated. Images were acquired in backscattered electron mode (BSE) using an SEC SNE-4500M Plus B scanning electronic microscope equipped with a Bruker EDS Quantax detector at the Laboratoire G-Time of the Université libre de Bruxelles (Belgium). Measurements were performed at an accelerating voltage of 15 keV. For each stratigraphic horizon, one BSE image at ×350 magnification was captured for ten randomly selected particles, resulting in a total of 50 analysed particles. The same magnification was used for all particles to ensure representativity—after confirming they yielded similar vesicle size distributions—and to capture larger vesicles while keeping the error for vesicles down to 1 µm equivalent diameter (~22 pixels in area) below 4.5 % for a single misrepresented pixel, following the approach of Shea et al. [2010].

Greyscale BSE images were processed using the opensource software `Fiji` (`ImageJ` [Schindelin et al. 2012]). A preliminary manual rectification was conducted on images to account for vesicles not filled with resin. The `Trainable Weka Segmentation` tool [Arganda-Carreras et al. 2017] was then applied to classify pumice glass, vesicles and, when present, phenocrysts, producing binary images. Finally, the distance transform watershed algorithm provided in the `MorphoLibJ` library [Legland et al. 2016] was used to reconstruct vesicle walls lost during sample preparation or due to bubble coalescence during final stage of magmatic ascent.

From the processed binary images, 2D vesicularity (percentage of the area occupied by vesicles), the number of vesicles per unit area (N_a), and the average bubble area were calculated. Each clast was then scored based on the root-sum-of-squares of the variances of these parameters relative to the average value calculated for clasts within the same stratigraphic horizon (i.e. Euclidean distance). The clast bearing the minimum score, representing the least deviation from the average, was identified as the most representative of its stratigraphic horizon and selected for further investigation.

Additional BSE images were collected for these five selected clasts at varying magnifications following the nested image strategy proposed by Shea et al. [2010] for stereological conversion. Specifically, one image was acquired at a magnification sufficient to capture the entire clast (×67–×100), then two images at ×350 were taken from areas exhibiting the greatest vesiculation difference based on visual inspection. Within

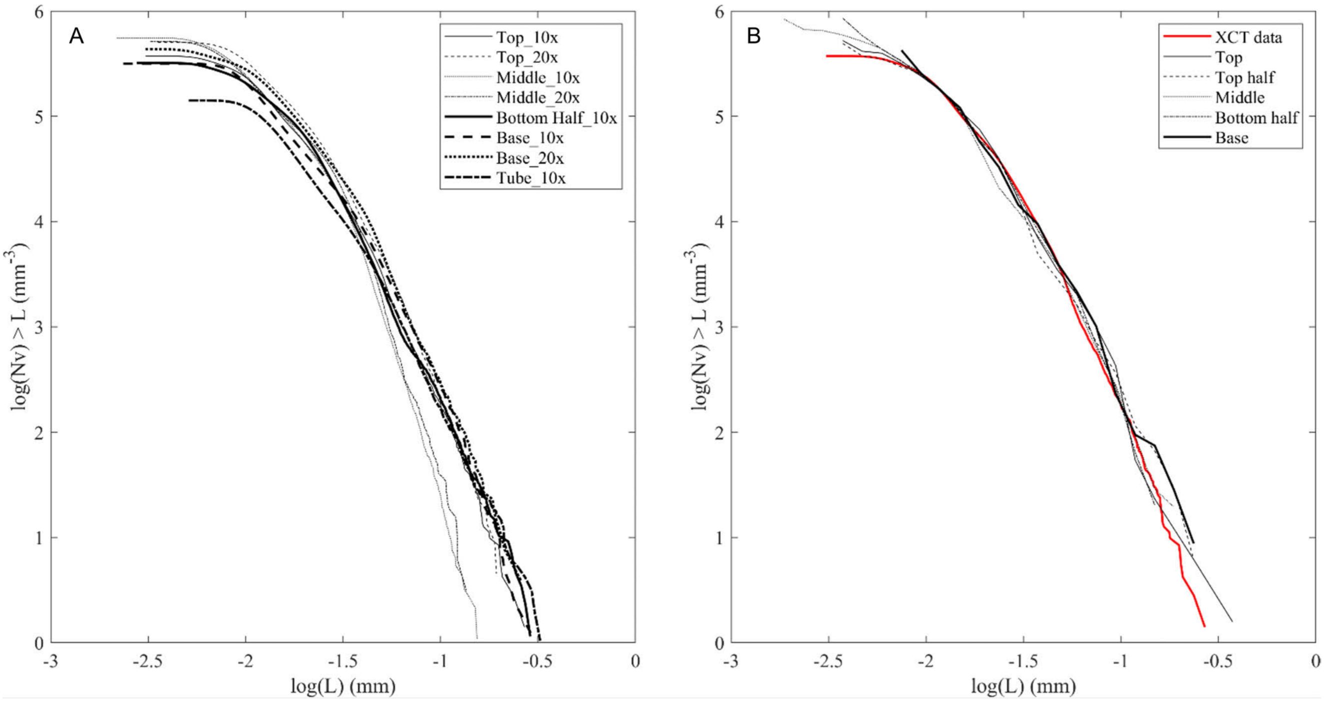

Figure 6: CVSD plots for [A] μXCT datasets and [B] 2D images stereologically converted with FOAM, where the μXCT trend of the base horizon is included for comparison.

these areas, two images each were captured at ×900 (Figure 2). Images were processed in Fiji as previously described. Using FOAM software [Shea et al. 2010], we then obtained statistical descriptors of shapes and sizes of vesicles corrected for magnification choice, cut-effect and intersection probability [Cashman and Mangan 1994; Shea et al. 2010, and references therein]. Stereological conversion in FOAM utilises vesicularity values constrained by three-dimensional reconstructions of particles (Section 3.3 and Section 3.4), as vesicularity estimates from the stereological conversion are prone to overestimation [Shea et al. 2010].

3.3 3D pumice textures

3D maps of pumice clasts were generated using μXCT with a ZEISS Xradia Versa 410 at the Istituto Nazionale di Geofisica e Vulcanologia-Osservatorio Vesuviano (Italy). A single ash particle within the 0/−0.5ϕ (1–1.4 mm) grain size range was randomly selected from each of the bottom (KF176-C), bottom half (KF176-F), middle (KF176-I), and top (KF176-O), horizons of the type section. Additionally, a tubular pumice particle [Martí et al. 1999] was selected from the top half horizon (KF176-L) for textural comparison (Figure 3).

The particles were cleaned using an ultrasonic bath before being scanned with μXCT. The field of view was configured to capture nearly the entire volume of each clast, and the working distance was adjusted to maximise resolution. Additionally, a 10× magnification lens was positioned before the detector to optically enhance the resolution, resulting in a final pixel size of 2 μm/px (8 μm^3/voxel). A total of 4001 bidimensional X-ray absorption projections were collected during a 360° rotation of the sample at 80 kV and 7 W. For samples KF176-C, KF176-I, and KF176-O, an additional scan was performed at 150 kV and 10 W decreasing the working distance and using a 20× magnification lens—and consequently reducing the field of view—to obtain an improved resolution of 1.1 μm/px (~1.3 μm^3/voxel). When necessary, a low-energy (LE1) filter was used to minimise beam hardening. The scans were then reconstructed into tomographic volumes using the integrated XRM Reconstructor software.

Tomographic volumes were processed using Dragonfly software [Dragonfly 2024]. Initially, a U-Net2D super-resolution model [Ronneberger et al. 2015], based on deep-learning neural networks, was trained using correspondent high-resolution (1.1 μm/px) and low-resolution (2 μm/px) volumes and applied to all lower-resolution datasets (2 μm/px). This approach significantly improved these datasets which, due to their larger field of view, captured more extensive and representative volumes of the original particles while maintaining practical scanning times [Buono et al. 2023]. The effectiveness of this approach was validated through visual inspection (Figure 3).

Subsequently, glass, vesicles (voids), and phenocrysts were labelled using the U-Net2.5D segmentation model (Figure 3 [Ronneberger et al. 2015]). For each dataset, the model was trained on at least five 2D areas that had been manually segmented assigning grey-scale values (proportional to phase X-ray absorption) to the corresponding phases. This approach, considering both grey-scale values and shape factors, significantly improves segmentation accuracy, particularly for thin glass walls. Model performance was evaluated by comparing the automated segmentation to manual segmentation, ensuring that a minimum of ~50 two-dimensional patches (64^2 voxels each) per sample were included in the validation set. A maximum binary cross-entropy loss of 0.013 was consid-

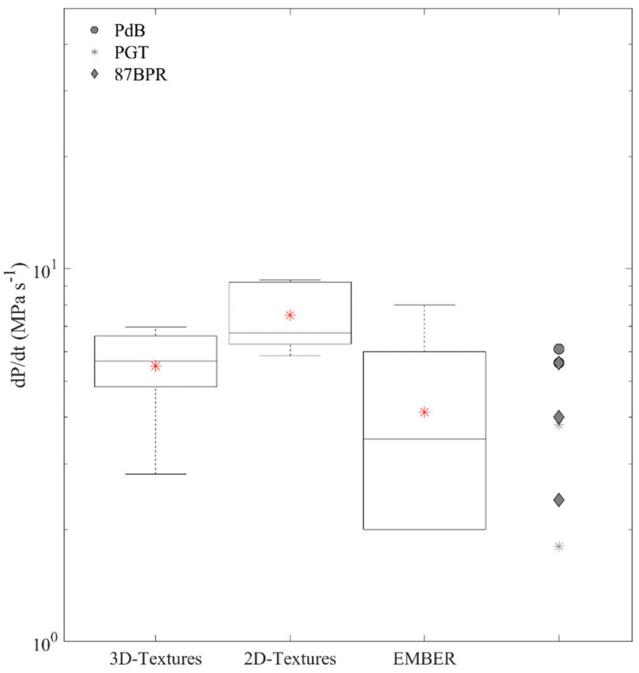

Figure 7: Box and whisker plots of decompression rates estimated by applying Shea [2017] models on 3D and 2D textural data, and the embayment diffusivity speedometer (EMBER). Whiskers indicate range limits, black horizontal lines medians, and red asterisks mean values. Previous data for explosive peralkaline magmas are included for comparison. PdB: Somma-Vesuvius Pomici di Base trachytic eruption [Pappalardo et al. 2018]; PGT: Pantelleria Green Tuff trachytic eruption [Campagnola et al. 2016]; 87BPR: ~87 ka Baricha peralkaline rhyolitic eruption [Tadesse et al. 2024].

ered acceptable (average across samples: 0.009 ± 0.002) for the training, while the validation Dice coefficient—computed as 2TP/(2TP+FP+FN), where TP, FP, and FN represent true positives, false positives, and false negatives voxels, respectively—ranged between 0.50 and 0.84. To account for open vesicularity, vesicles open to the clast exterior were virtually closed at their outlet using a maximum width of 70 µm as a threshold, corresponding to the upper quartile of the volume-weighted distribution of vesicle equivalent diameters. This approach minimized artificial alterations of the vesicle network, avoiding overestimation of naturally open vesicle volumes while retaining a significant portion of the total vesicle volume.

Once glasses and vesicles were segmented, the distance-transformed watershed algorithm was applied to separate individual bubbles that had become connected during the final stages of bubble growth or due to the scanning resolution being insufficient to capture extremely thin glass walls (< 2 µm; Figure 3). This procedure successfully identified most bubbles down to 2 µm (~2 px) in equivalent diameter. Labelled objects smaller than this threshold were excluded from the dataset, to prevent potential noise artefacts [Pappalardo et al. 2018; Liedl et al. 2019; Buono et al. 2020].

3.4 Conduit dynamic modelling

From the textural datasets, we retrieved several parameters that were used as inputs for models or as quantitative descriptors of conduit dynamics. Pumice vesicularity (φ) was defined as the percentage of volume occupied by vesicles relative to the total volume (i.e. voids + glass – phenocrysts), while vesicle connectivity as the percentage of a connected vesicle network (pre-watershed, see Section 3.3) relative to the total vesicle volume. Volumetric descriptors of the shape and size of each vesicle (post-watershed) were used to construct vesicle size distribution and vesicle population density trends that can be used to describe the nucleation process [Cashman and Mangan 1994; Shea et al. 2010, and references therein]. Vesicle number density (VND), a key parameter for estimating magma decompression rates from vesicle textures [e.g. Toramaru 2006; Shea 2017], was calculated by dividing the total number of vesicles (post-watershed) by the glass volume [Proussevitch et al. 2007]. Decompression rates were computed using the model developed by Toramaru [2006] in its simplified version proposed by Shea [2017]:

$$\frac{dP}{dt} = \left(\frac{N_V}{A \times 10^4}\right)^{\frac{2}{3}} \tag{1}$$

where N_V corresponds to VND (mm^{-3}) and A is a composition-dependent fitting constant (3 ± 1.8 for phonolites and trachytes [Shea 2017]).

3.4.1 *Supersaturation pressure*

The supersaturation pressure of bubble nucleation (ΔP_{sat}), and the related nucleation pressure (P_n), were estimated for a given decompression rate by integrating nucleation rates for progressive decompression steps (t_i in Equation(3) in Mourtada-Bonnefoi and Laporte [2004]) and iteratively recalculating the bubble number density (Equation 4 in Mourtada-Bonnefoi and Laporte [2004]) until the latter overpassed the value of 1 mm^{-3} [Mourtada-Bonnefoi and Laporte 2004; Shea 2017]. The nucleation rate of bubbles (J) at a given melt pressure (P_M) was calculated following the classical nucleation theory [Hirth et al. 1970; Shea 2017]:

$$J = \frac{2n_0^2 DV}{a_0} \cdot \sqrt{\frac{\varsigma}{kT}} \cdot exp\left(-\frac{16\pi\varsigma^3}{3kT\,(P_B - P_M)^2}\theta\right), \tag{2}$$

where θ is a geometric factor equivalent to 1 or $0 < \theta < 1$ for homogeneous and heterogeneous nucleation respectively [Shea 2017], and k is the Boltzmann constant (i.e. 1.38×10^{-23} $m^2\,kg\,s^{-2}\,K^{-1}$). T (K) is the temperature of the melt at the reservoir conditions (1248 ± 22 K [Cappelli et al. 2025]). The water saturation pressure, P_{SAT} (Pa), was set to 92 ± 15 MPa, determined using saturation models in crystal-hosted melt inclusions (see Cappelli et al. [2025] for a detailed definition of the pressure parameter), and used to calculate P_B (Pa) which is the internal pressure of an incipient bubble nucleus, iteratively calculated for each step of P_M [Shea 2017; Buono et al. 2020]. D ($m^2\,s^{-1}$) is the diffusivity of the volatile phase in the melt, calculated for water in phonolitic and trachytic melts using the model of Fanara et al. [2013].

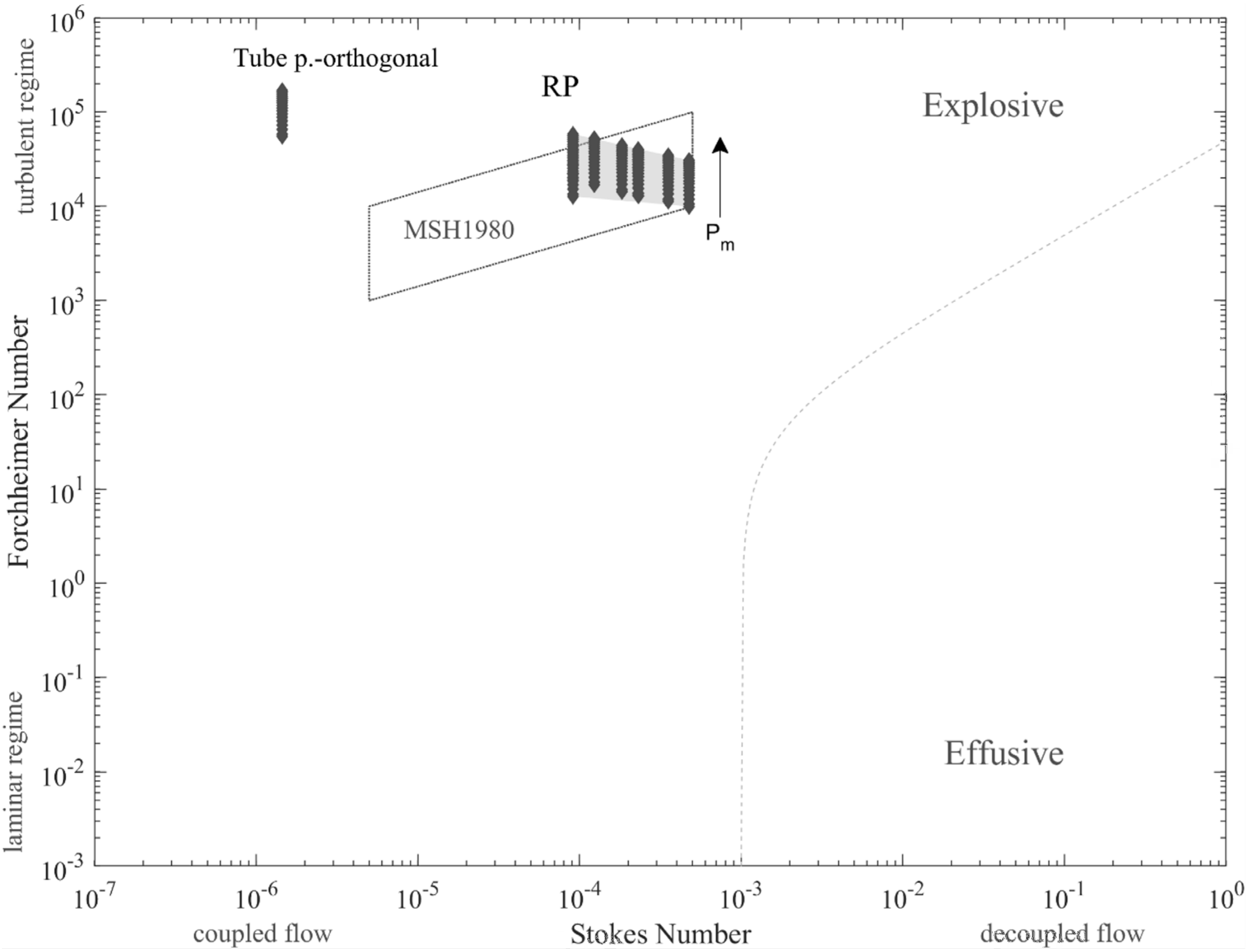

Figure 8: Stokes and Forchheimer numbers for the Rungwe Pumice (RP) samples represented as grey diamonds over a magmastatic pressure (Pm) range from 13 to 40 MPa (black arrow). The grey-shaded region represents the range for standard RP pumices, while values for the tube pumice across directions orthogonal to the main vesicle elongation are shown for comparison. Additionally, the range for the 1980 Mount St. Helens Plinian eruption (MSH1980 [Degruyter et al. 2012]) is indicated (dotted contour). The dashed line marks the critical Stokes and Forchheimer numbers defining the transition between effusive and explosive regimes for MSH1980 rheology [Degruyter et al. 2012]; although not directly applicable to RP, a variation of maximum one order of magnitude is expected [Valdivia et al. 2022].

The volume of volatile molecules V (m^{-3}) was calculated according to Equation 5 in Shea [2017]. The mean distance between volatile molecules in the melt, a_0 (m), was derived as $n_0^{-1/3}$, where n_0 ($mol\,m^{-3}$) represents the number density of volatile molecules. n_0 was determined using Equation 6 in Shea [2017] with inputs derived from haüyne-hosted melt inclusion data [Cappelli et al. 2025] which provided initial dissolved water concentrations (i.e. 4.82 ± 0.58 wt.% on average, converted to a single-oxygen mass fraction) and melt density ($2250 \pm 10\ kg\,m^{-3}$ on average). Finally, ς ($N\,m^{-1}$) corresponds to the surface tension that can be either calculated for homogeneous nucleation adopting the Equation 13 from Shea [2017] or fixed at the average value of 0.025 $N\,m^{-1}$, considering magnetite microlites as primary nucleation sites [Shea 2017]. A Monte Carlo simulation with 5000 iterations was run separately for homogeneous and heterogeneous nucleation to assess the statistical representativeness of ΔP_{sat} and P_n and to quantify their uncertainty. For each iteration, random values of T, P_{SAT}, X_{H2O}, X_{SiO2}, ρ, and dP/dt were sampled from normal distributions defined by their measured means and standard deviations, then ΔP_{sat} was calculated accordingly. For heterogeneous nucleation, the wetting angle was additionally sampled from the range appropriate for Fe–Ti oxides (i.e. 90–160°), which are the mineral phases most effective at promoting bubble nucleation [e.g. Shea 2017].

3.4.2 *Volatile outgassing in a porous magma*

Tortuosity measures the deviation of the flow path between two bubbles from a straight line along the flow direction. It is a critical parameter reflecting the roughness of a porous medium and is therefore directly related to its permeability. However, due to the highly intricate network of vesicles, it was computationally impractical to quantify tortuosity directly using `Dragonfly`'s skeletonization on a volume sufficiently large

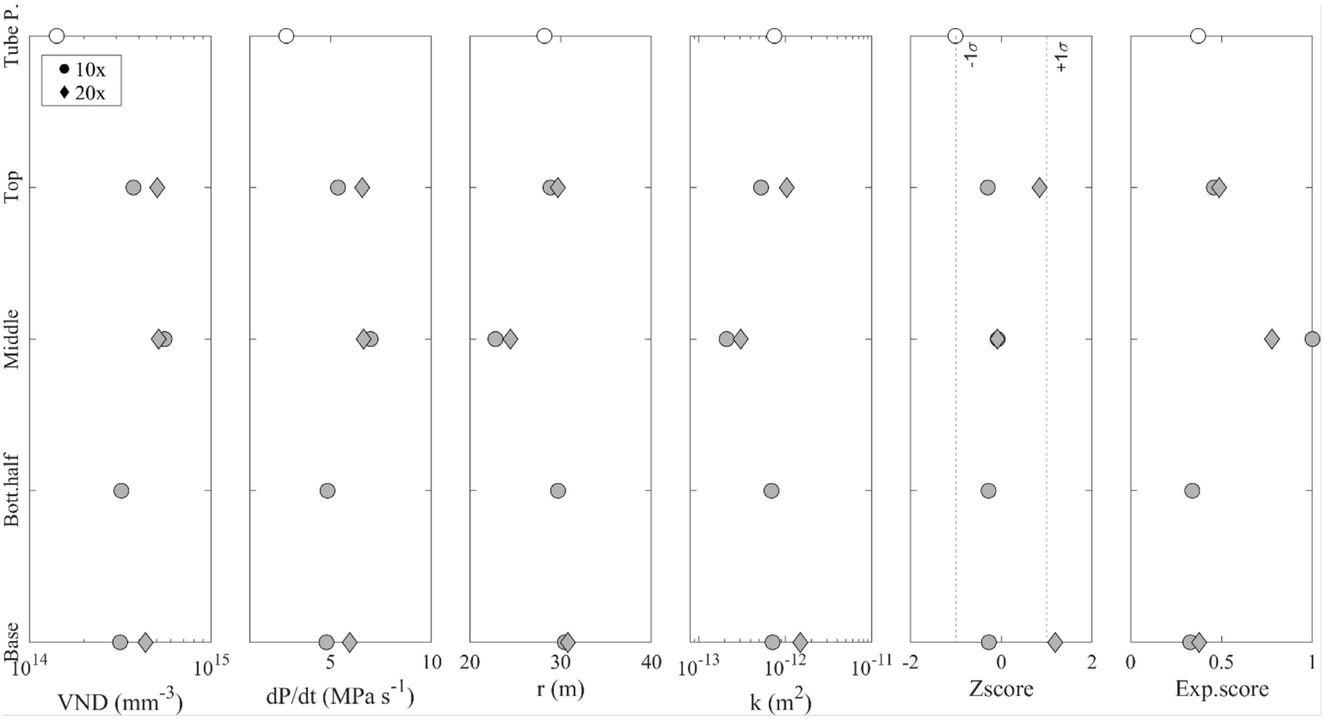

Figure 9: Comparison of key conduit parameters for different stratigraphic horizons analysed by µXCT: vesicle number density (VND), decompression rate (*dP/dt*), conduit radius (*r*), and Darcian permeability (*k*). An overall relative explosivity score—where 1 represents the maximum potential for explosivity—is calculated for each sample by summing the min-max normalised values of interdependent parameters promoting explosivity such as *dP/dt* and *k*. The reciprocal value of *k* was used in scoring, as it is inversely proportional to explosivity potential. Additionally, a normalised *z* score was calculated to express the interrelated consistency of these parameters.

to be representative of the datasets. As an alternative, we utilized the tortuosity factor ($\tau*$), which quantifies the difference between diffusivity under laminar conditions (fully conductive) and the actual diffusivity through the porous medium [Epstein 1989; Cooper et al. 2016]. Tortuosity factor can be related to tortuosity through Archie's law (Equation 4 in Degruyter et al. [2012]). To calculate $\tau*$, we used the `MATLAB` application `TauFactor`, which computes $\tau*$ along three mutually perpendicular directions on cuboid volumes [Cooper et al. 2016]. These cuboids were extracted from labelled vesicle volumes (pre-watershed), selecting the largest possible cuboid for each dataset, always ensuring it was several orders of magnitude larger than the largest vesicle present in the dataset [Pappalardo et al. 2018]. When present (Section 4.1), the preferential orientation of vesicles, indicated by the direction of their equivalent ellipsoid major axes, was used as a proxy for the primary flow direction, and the y-axis of each cuboid was oriented accordingly.)

Darcian permeability (m^2) across the above-mentioned cuboid volumes was calculated using the Kozeny-Carman relation [Rust and Cashman 2004; Degruyter et al. 2012; Wei et al. 2018; Valdivia et al. 2022] along each of the three perpendicular directions:

$$k_D = \frac{\varphi^3}{cS^2\tau^2}, \tag{3}$$

where the square of tortuosity is calculated through Archie's law, φ is vesicularity, S (m^{-1}) is the surface area per unit volume of the vesicle network, and c is the Kozeny constant for pores-controlled media (set to 8 [Degruyter et al. 2012]).

Key parameters describing the flow of volatiles through porous media (outgassing) and their coupling with magma ascent include the dimensionless Forchheimer (Fo) and Stokes (St) numbers [Rust and Cashman 2004; Degruyter et al. 2012; Zhou et al. 2019; Valdivia et al. 2022, and references therein]. The Forchheimer number is defined as the ratio between the inertial and viscous forces resisting the flow. Similar to Reynolds number, it characterises flow behaviour, with lower Fo values indicating laminar flow. It can be expressed as [Rust and Cashman 2004; Degruyter et al. 2012, and references therein]:

$$\mathrm{Fo} = \frac{\rho_g v}{\mu_g} \cdot \frac{k_D}{k_I}, \tag{4}$$

where ρ_g is the density of the volatile phase defined as $P_M/(RT)$ with R as the specific gas constant for water (461.4 J kg^{-1} K^{-1} [Degruyter et al. 2012]) and P_M the pressure in the conduit at a given depth, iterated during decompression; μ_g is the viscosity of the volatile phase (15×10^{-5} Pa s [Degruyter et al. 2012]); v (m s^{-1}) is the average magma ascent velocity, calculated as the ratio of the mean decompression rate to the magmastatic gradient in the conduit (approximated by the lithostatic gradient, 0.027 MPa m^{-1} [Browne and Szramek

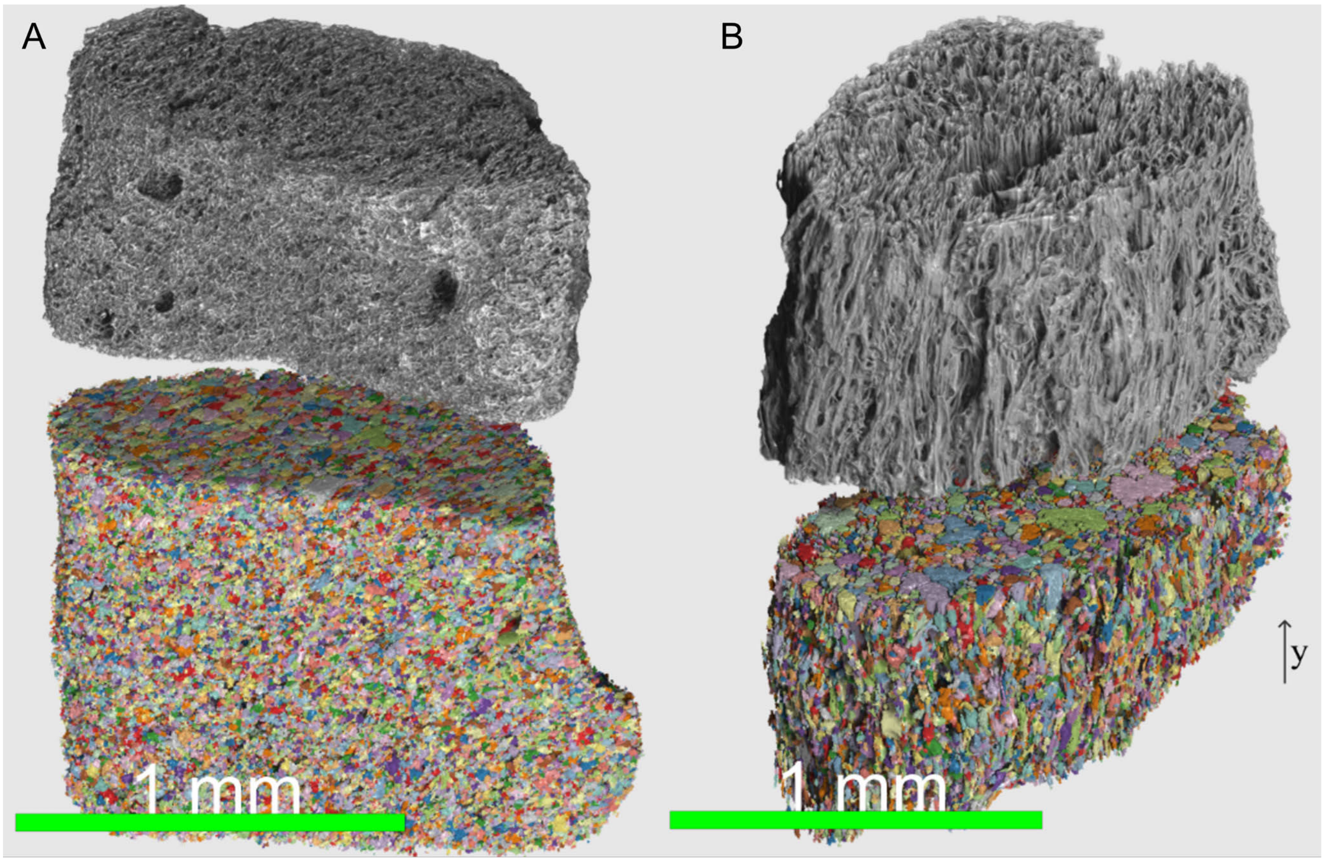

Figure 10: Grey-scale three-dimensional reconstructions and segmented vesicle volumes of a [A] "standard" pumice and the [B] tube pumice, providing a visual comparison of vesicle textures. In [B] the y-axis indicates the direction of vesicle preferential orientation. Both particles were scanned with a 10× magnification lens.

2015]); and k_I is the inertial permeability [Rust and Cashman 2004; Zhou et al. 2019] derived from k_D using the relationship proposed by Gonnermann et al. [2017, Equation B1].

The Stokes number is defined by the ratio of the response time of magma to the flow time of volatile phases [Degruyter et al. 2012] and at low values indicates strong coupling between volatiles and the ascending magma. St can be expressed as [Degruyter et al. 2012]:

$$\mathrm{St} = \left(\frac{\rho_B k_D}{\mu_g}\right) \Big/ \left(\frac{r}{v}\right) \tag{5}$$

where ρ_B (kg m^{-3}) is the bulk density (melt + bubbles) and r is the radius of the conduit estimated from eruption mass discharge rate (Section 5.1).

4 Results

4.1 Vesicularity and vesicle metrics

Vesicularities determined from three-dimensional imaging ($\bar{x}$ ~63 % ± 5 %) are, on average, lower than values obtained using 2D analyses ($\bar{x}$ ~74 % ± 6 %). However, both 2D and 3D datasets yield consistent vesicularities when comparing samples within or across different stratigraphic horizons (Table 1). Despite this discrepancy between 2D and 3D datasets, the consistent values of vesicularity and VND in pumiceous ash (%RSD ~7 %) observed from 2D imaging across a wide selection of clasts confirms that randomly selected single clasts were sufficiently representative for µXCT analysis. Connected vesicle networks consistently account for 99.9 % of the total vesicularity.

Vesicularity estimated from 3D reconstructions is influenced by the selection of grey-scale threshold during segmentation; however, conservative manual segmentation demonstrates that this effect on total vesicularity is minimal (~5 vol %) and deep learning models effectively delineate vesicle contours, preserving even the thinnest vesicle walls detectable at the applied resolution. The method's reliability is further supported by the small differences (5–8 %) between scans collected at 20× and 10× (improved with the super-resolution model) magnifications, confirming the robustness of the approach despite a slight tendency to underestimate vesicularity at the lower resolution. The discrepancy between 2D and 3D analyses thus likely arises from the limitations of bidimensional sectioning and sample preparation (e.g. glass breakage), which struggles to accurately represent irregular vesicles, risking overestimation of vesicularity [Shea et al. 2010]. Vesicularity of lapilli-sized clasts analysed with pycnometry methods ranges around 0.85 ± 0.04 (Supplementary Material 2). This high degree of vesicularity may result from post-depositional bubble inflation, which is more likely in larger clasts due to prolonged cooling times [Thomas et al. 1994; Kaminski and Jaupart 1997]. This phenomenon is visually evident from the

Table 1: Vesicularity, VND, and geometric parameters of pumice textures analysed with 2D and 3D methods.

Sample	3D Textures							2D Textures†		FOAM§
	Lens	Envelope volume° (mm^3)	Vesicle number ($\times 10^5$)	VND (m^{-3})	Vesicularity	Connectivity	Surface Area per unit volume (mm^{-1})	Vesicularity*	VND** ($\times 10^{14}$ m^{-3})	VND ($\times 10^{14}$ m^{-3})
Top	10×	1.884	2.65	3.73	0.62	0.9986	145	0.76 ± 0.05	4.06 ± 2.10	5.23
Top	20×	0.749	1.11	5.04	0.71	0.9999	160			
Top half								0.77 ± 0.04	4.92 ± 3.60	4.91
Middle	10×	2.089	5.13	5.53	0.56	0.9993	182	0.72 ± 0.08	2.38 ± 1.55	8.34
Middle	20×	0.791	1.65	5.14	0.59	0.9997	188			
Bottom half	10×	2.312	2.79	3.21	0.62	0.9996	150	0.76 ± 0.04	3.61 ± 2.38	8.53
Base	10×	2.227	2.54	3.15	0.64	0.9998	158	0.71 ± 0.05	3.84 ± 3.00	4.24
Base	20×	0.914	1.14	4.35	0.71	0.9999	151			
Tube pumice	10×	2.155	1.36	1.42	0.56	0.9985	129			
Average‡				4.31 ± 8.95	0.64 ± 0.05	0.9995 ± 0.0004	162 ± 15	0.74 ± 0.06	3.76 ± 2.75	6.25 ± 1.81

°Volume of glass+vesicles. Open vesicles with an outlet diameter ≤70 μm were closed by a wrapping surface;
‡average of parameters for 3D datasets is calculated considering only "standard" pumices;
†for each horizon, 2D Vesicularity and VND are given by the average of processed images for 10 different samples;
*2D vesicularity is computed as fraction of vesicle area (excluding vesicles at borders) over total area (corrected for border vesicles);
**2D VND is estimated from number per the unit area: Na/L where L is the average vesicle dimension, which can be written as $Na/\sqrt{(A/n)}$ with A: total vesicle area and n: number of vesicles;
§2D VND stereologically corrected using FOAM.

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Table 2: Indicators derived from 3D size distribution trends and shape parameters. The slope and intercept of VSDs are provided, along with the estimated average growth rate.

Sample	Lens	VSD Slope (mm^{-1})	VSD no ($\times10^5$ mm^{-3})	Average bubble growth rate ($\times10^{-4}$ $mm\,s^{-1}$)	VVD equivalent diameter mode (μm)	Average sphericity
Top	10×	−95	5.46	6.81	45	0.74 ± 0.09
Top	20×	−93	5.71	6.94	36	0.70 ± 0.09
Middle	10×	−123	7.91	5.23	36	0.74 ± 0.10
Middle	20×	−124	8.40	5.19	36	0.70 ± 0.10
Bottom half	10×	−94	4.95	6.88	28	0.73 ± 0.10
Base	10×	−82	3.65	7.90	45	0.71 ± 0.10
Base	20×	−87	6.44	7.45	45	0.71 ± 0.10
Tube pumice	10×	−68	1.66	9.55	57	0.68 ± 0.12
Average				7.22 ± 1.37	39 ± 6∗	

∗Average VVD mode is calculated considering only "standard" pumices.

Table 3: Estimates of decompression rate calculated using both Shea [2017] equation for pumice textures and the embayment speedometer for comparison. Modelled fragmentation conditions are also reported.

Sample	3D Textures				2D Textures†
	Lens	Decompression rate ($MPa\,s^{-1}$)	Ascent rate ($m\,s^{-1}$)	Conduit radius (m)	Decompression rate ($MPa\,s^{-1}$)
Top	10×	5.4	199	28	6.7
Top	20×	6.6	243	29	
Top half					6.5
Middle	10×	7.0	259	23	9.2
Middle	20×	6.6	246	24	
Bottom half	10×	4.9	180	30	9.3
Base	10×	4.8	178	30	5.9
Base	20×	5.9	220	31	
Tube pumice	10×	2.8	104	36	
Average‡		5.9 ± 0.9	218 ± 30	28 ± 3	7.5 ± 1.5

Sample	Embayment speedometer			
	Decompression rate ($MPa\,s^{-1}$)	Final pressure (MPa)	Final water concentration (wt.%)	Initial water concentration (wt.%)
Top	1.67	10	0.63	4
Top	2.78	10	0.63	5
Top	2.78	25	1.51	5
Top half	2.78	5	0.32	4
Top half	0.56	15	0.93	5
Top half	2.10	5	0.32	5
Top half	0.20	15	0.93	3
Middle	0.80	15	0.93	4
Bottom half	2.00	20	1.22	4
Average‡	1.74 ± 0.95	13 ± 6	0.82 ± 0.37	4.3 ± 0.7

‡Average of 3D datasets is calculated considering only "standard" pumices;
† obtained from VND values corrected with stereological conversion in FOAM.

occurrence of millimetric to centimetric vesicles on clast surfaces, which can influence and potentially alter original vesicularity determinations (Supplementary Material 2).

Table 4: Model outgassing parameters. Parameters estimated across three mutually orthogonal and randomly oriented directions.

Sample	Lens	Darcian permeability (m^2)			Inertial permeability (m^2)			Stokes number		
		x	y	z	x	y	z	x	y	z
Top	10×	4.84×10^{-13}	4.36×10^{-13}	6.56×10^{-13}	3.26×10^{-9}	2.83×10^{-9}	4.91×10^{-9}	6.38×10^{-5}	5.75×10^{-5}	8.65×10^{-5}
Top	20×	1.09×10^{-12}	8.77×10^{-13}	1.14×10^{-12}	9.76×10^{-9}	7.28×10^{-9}	1.04×10^{-8}	1.10×10^{-4}	8.83×10^{-5}	1.15×10^{-4}
Middle	10×	3.45×10^{-13}	1.27×10^{-13}	1.60×10^{-13}	2.06×10^{-9}	5.34×10^{-10}	7.28×10^{-10}	6.61×10^{-5}	2.44×10^{-5}	3.06×10^{-5}
Middle	20×	2.53×10^{-13}	2.14×10^{-13}	4.55×10^{-13}	1.35×10^{-9}	1.08×10^{-9}	3.00×10^{-9}	4.14×10^{-5}	3.50×10^{-5}	7.46×10^{-5}
Bottom half	10×	6.35×10^{-13}	7.90×10^{-13}	6.57×10^{-13}	4.71×10^{-9}	6.32×10^{-9}	4.92×10^{-9}	7.94×10^{-5}	9.87×10^{-5}	8.21×10^{-5}
Base	10×	7.73×10^{-13}	6.41×10^{-13}	7.30×10^{-13}	6.13×10^{-9}	4.76×10^{-9}	5.68×10^{-9}	9.05×10^{-5}	7.50×10^{-5}	8.55×10^{-5}
Base	20×	1.57×10^{-12}	1.47×10^{-12}	1.42×10^{-12}	1.60×10^{-8}	1.46×10^{-8}	1.40×10^{-8}	1.44×10^{-4}	1.34×10^{-4}	1.30×10^{-4}
Tube pumice	10×	1.18×10^{-14}	7.49×10^{-13}	5.15×10^{-14}	2.15×10^{-11}	5.87×10^{-9}	1.57×10^{-10}	1.44×10^{-6}	9.13×10^{-5}	6.28×10^{-6}

VNDs derived from both 2D and 3D imaging datasets yield values on the order of (10^{14} m^{-3}), and are again consistent across different stratigraphic horizons. VND values from both datasets are reported inTable 1. In 3D reconstructed particles, vesicle volume distributions (VVDs) predominantly exhibit lognormal distributions, with unimodal modes—indicating the volumetrically most represented size—ranging from 28 μm ($\log(L) = -1.55$) to 57 μm ($\log(L) = -1.25$) equivalent diameters (Figure 4; Table 2). Vesicle size distributions (VSDs) exhibit curved trends with one or more break points and tend to level off at larger sizes (Figure 5). Such distributions are strongly influenced by the choice of bins, which, in this case, are linearly spaced and equivalent in number to the geometric binning used for VVDs [Shea et al. 2010]. Artefact spikes or broken segments may appear as a result. However, VSD trends of the finest vesicles allow us to extrapolate angular coefficients and intercepts at $L = 0$ mm (Figure 5; Table 2), which are useful for estimating vesicle growth at late nucleation and growth conditions (Section 5.2). Cumulative vesicle size distributions (CVSDs) follow exponential trends but exhibit slope breaks at the largest sizes (Supplementary Material 2). These may reflect difficulties in accurately representing the largest vesicles, potentially due to the effects of bubble coalescence [e.g. Blower et al. 2003; Pappalardo et al. 2018; Liedl et al. 2019]. Size distribution descriptors derived from the stereological conversion of 2D textures are consistent with those obtained from 3D reconstructions (Figure 6; Supplementary Material 2).

In 2D sections, vesicles are generally subrounded; however, larger vesicles tend to deform and flatten at contacts with neighbouring vesicles or adopt irregular polylobate shapes especially when coalesced (Figure 2). The average vesicle sphericity of 3D datasets ranges between 0.64 and 0.74 with a narrow spread ($\pm0.09 < 1\sigma < \pm0.12$). Sphericity is also negatively correlated with vesicle size (equivalent sphere diameter), and larger vesicles tends to be more irregular (Supplementary Material 2 Table S2). While no significant differences are observed across the other horizons, the minimum sphericity value is observed in the tube pumice, where vesicles predominantly exhibit elongated shapes. The preferential orientation of vesicles is evident in pole figures representing the orientation density of the vesicle's major axis relative to three mutually perpendicular axes (Supplementary Material 2). Although most pronounced in tube pumice, other samples also show a mild preferential orientation.

4.2 dP/dt and ΔP_{sat}

Estimates of 3D VND-based decompression rates [Shea 2017] are consistent across stratigraphic horizons, ranging between 4.8 and 7 MPa s^{-1}, with an average of 5.9 ± 2.4 MPa s^{-1}. The relatively large uncertainty partly reflects propagation of the ~40 % relative error associated with the compositional constant A in Equation 1. In contrast, when model uncertainty is excluded, the average decompression rate across all samples yield a narrower ±1 σ of 0.9 MPa s^{-1}, with the tube pumice being the only outlier, recording a comparatively lower value of 2.8 MPa s^{-1} (Table 3). These values fall within the same order of magnitude—though slightly lower—than those obtained from stereologically converted 2D VND, which yield an average of 7.5 ± 3 MPa s^{-1}. Decompression rates obtained using the embayment speedometer were evaluated across a range of initial dissolved water contents and final quenching pressures, according to the minimum least error [Georgeais et al. 2021]. The resulting rates are generally of the same order of magnitude as those derived from vesicle texture analyses ($\bar{x}$ = 1.74 ± 0.75 MPa s^{-1}; Table 3; Figure 7). An initial dissolved water content of ~4.3 wt.% consistently provided the best fits, falling within the uncertainty range of initial water contents inferred from melt inclusions (4.82 ± 0.58 wt.% [Cappelli et al. 2025]), with final quenching pressures ranging around 13 ± 6 MPa (Supplementary Material 2 Table S1).

The Monte Carlo simulation for homogeneous nucleation yields a median ΔP_{sat} of ~52 MPa (with a 16th–84th percentile range of 51–55 MPa) corresponding to an average nucleation pressure of ~40 MPa. Under a heterogeneous nucleation regime, the simulation produces a substantially lower ΔP_{sat} of ~16 MPa (16th–84th percentile range of 17–18 MPa), and an average Pn of ~76 MPa, primarily due to the reduced surface tension (ς in Equation 2). Detailed simulation outputs are provided in Supplementary Material 2.

4.3 Volatile outgassing

The tortuosity factor for "standard" clasts ranges from 1.70 to 3.78 and is consistent for the three directions investigated ($\pm 1\sigma$ = 0.12 to 0.72). In contrast, the tube pumice exhibits much greater heterogeneity, with a relative standard deviation of 50 % between the directions orthogonal to the major elongation of vesicles and the one parallel to it. A similar observation comes from Darcian permeability of clasts, which overall ranges from 1.27×10^{-13} and 1.27×10^{-12} m^2, and between 1.18×10^{-14} and 7.48×10^{-13} m^2 for the tube pumice, showing a relative standard deviation of 125 % (Table 4).

Stokes and Forchheimer numbers are calculated using the average of Darcian permeability across each direction, except for the tube pumice where the maximum value was used to account for its anisotropy. For the other samples, Stokes numbers range from 2.44×10^{-5} and 1.44×10^{-4} (Table 4), indicating significant gas-melt coupling [Figure 8; Rust and Cashman 2004; Degruyter et al. 2012]. Forchheimer number calculations were iterated between the lowest estimated nucleation pressure (40 MPa) and a minimum quenching pressure of 13 MPa (Section 5.2), and consistently exceeded a value of 10^4, suggesting a predominantly turbulent and hindered flow of volatiles through magma vesicularity [Degruyter et al. 2012].

5 Discussion

5.1 Sustained activity during the RP eruption

Plinian-style eruptions are characterised by the violent ejection of pyroclastic materials and could last for hours or even days, potentially maintaining a quasi-stable mass discharge rate (MDR [Cioni et al. 2015, and references therein]). Changes in MDR during a prolonged eruption may occur due to factors such as reservoir replenishment, conduit widening, or variations in volatile concentrations [e.g. Carey and Sigurdsson 1989]. MDR fluctuations can be identified at the outcrop scale—where they may manifest as grading in fallout deposits or the onset of pyroclastic fountaining, which generates pyroclastic density currents—but are also preserved in textures of pumiceous products.

The pyroclastic fall deposit of the RP eruption is reversely graded at the base, suggesting an initial intensification of eruptive explosivity [Fontijn et al. 2011]. Above this interval, the deposit remains massive to the top and lacks evidence of widespread PDCs descending the volcano's flanks. This stratigraphy suggests sustained eruptive activity throughout the event [Fontijn et al. 2011], although we cannot exclude that poorly preserved or unidentified small PDCs may have been deposited close to the volcano summit (within a travel distance of 1-2 km [Fontijn et al. 2011]). To evaluate whether these stable eruption conditions are also reflected in steady decompression rates and dynamics of bubble nucleation and growth, we analysed eruptive products collected from different horizons of the deposit, spanning its vertical extent.

VND, vesicularity, inferred magma decompression rate, and outgassing parameters all remain within the same order of magnitude across the deposit, and each horizon yields average z-scores within 1 σ (Table 3; Table 4; Figure 9). The consistency of these conduit parameters suggests that bubble evolution and fragmentation mechanisms remained largely unaltered throughout the RP eruption. One exception is the central horizon, represented by sample KF176I. This sample displays the highest VND—and thus the highest inferred decompression rate—yielding a greater proportion of smaller vesicles relative to the other samples (Figure 5). It also produces the highest explosivity score, calculated as the average of min–max normalised values of interdependent explosivity indicators (decompression rate and Darcian permeability; Figure 9). This modest deviation may reflect either a brief peak in ascent rate or, more plausibly, natural variability within the magmatic foam.

By inverting Equation 16 for MDR from Shea [2017], we estimated conduit radius (assuming a cylindrical conduit) as a function of magma bulk density—derived from glass density and pumice vesicularity—and decompression rate. Using a representative value from the peak MDR range reported by Fontijn et al. [2011] (i.e. 3.8×10^8 kg s^{-1}), based on maximum column-height estimates, we calculated a conduit radius for each stratigraphic horizon, yielding a skewed distribution with an average of 28 ± 3 m (Table 3). These values are systematically smaller than previous crater radius estimates of 50–60 m [Fontijn et al. 2011], which instead possibly represent shallower vent conditions [Woods 1998], above the magma fragmentation level, where erosion process in the conduit wall reaches maximum efficiency [e.g. Macedonio et al. 1994]. Given the stable decompression rates independently inferred throughout the sequence, we find no evidence for progressive conduit widening when assuming a constant peak MDR. Conduit widening during sustained Plinian eruption is considered a common process, as conduit erosion can increase conduit width by several meters over timescales of hours [e.g. Carey and Sigurdsson 1989; Cioni et al. 2015]. Under a constant peak MDR, such a few-meter radius increase would reduce decompression rate only slightly (< 1 MPa s^{-1}). Although

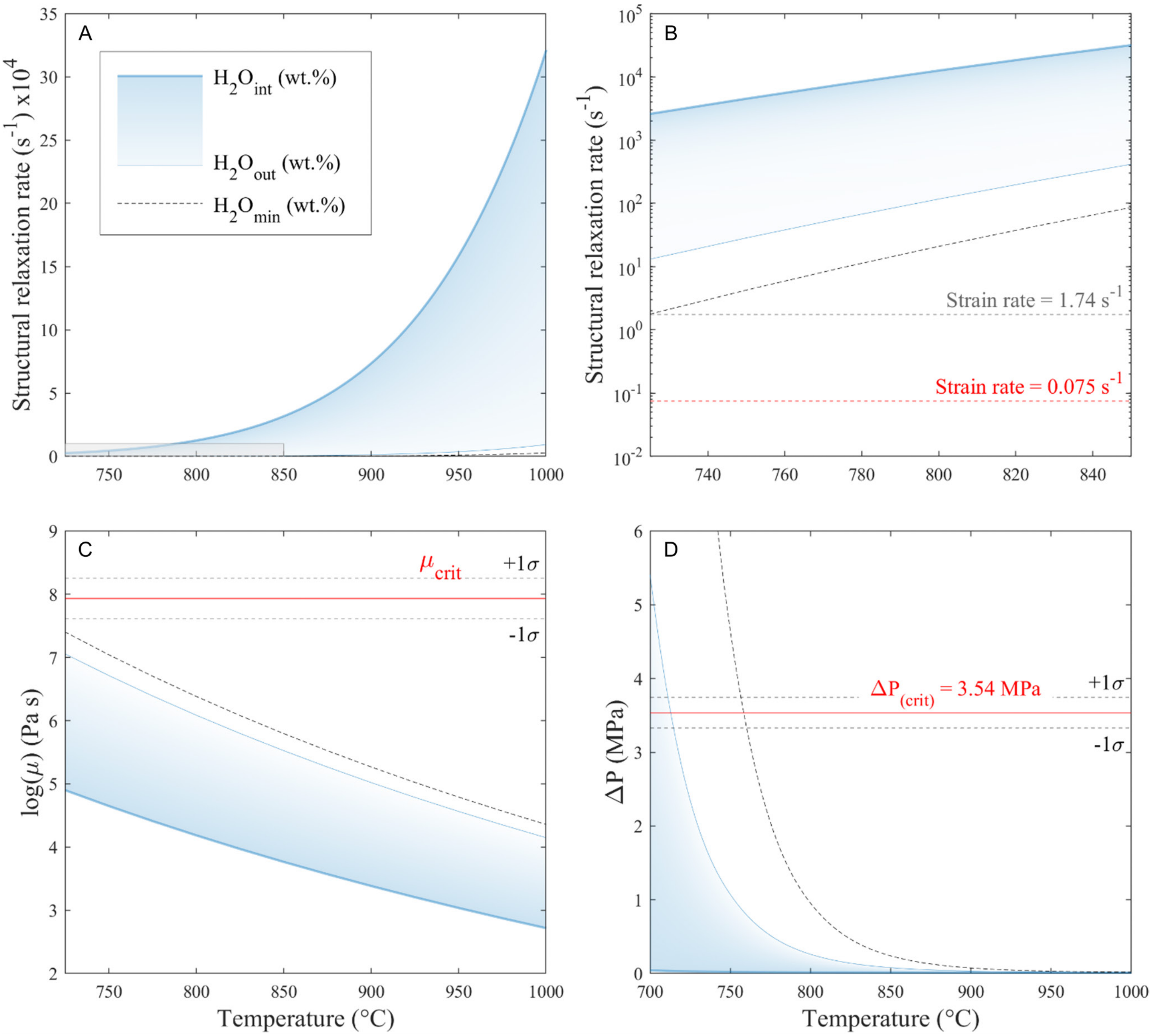

Figure 11: Different fragmentation criteria applied to the melt compositions of melt embayments spanning their water concentration range—from innermost (H_2O_{int}; thickest line) to outmost (H_2O_{out}; thinnest line) portions. The minimum water concentration inferred for a quenching pressure of 13 MPa (black dashed line) is also shown. The criteria have been evaluated for a range of temperatures (700–1000 °C) to represent the temperature drop due to the gas expansion upon ascent (see text for further details). Only one sample from the middle horizon is shown for clarity. [A] Strain rate criterion is represented by melt structural relaxation drops caused by viscosity increase due to magma degassing and cooling. The grey inset highlights the close-up in the low-temperature region shown in [B], where the criterion is satisfied when the critical strain rate (0.075 s^{-1}) exceeds the structural relaxation time. The hypothetical strain rate (1.74 s^{-1}) resulting from magma acceleration only within the last 90 m of ascent is also shown; [C] shear-induced fragmentation criterion expressed as viscosity increase, where fragmentation is expected when the viscosity (μ) exceeds the critical viscosity calculated for RP conditions (μcrit). Average of μcrit from the Monte Carlo simulation is indicated with a red line together with $\pm 1\sigma$ (grey dashed lines); [D] increase in bubble internal gas overpressure with increasing melt viscosity upon ascent where fragmentation is expected when the overpressure exceeds the average critical value of 3.54 MPa (red line with $\pm 1\ \sigma$ indicated by grey dashed lines).

we do not observe this trend, a small widening of a few meters may fall within the uncertainty of our estimates ($\pm 1\ \sigma \simeq 3$ m) and therefore remain unresolved. In contrast, if MDR had progressively increased toward a peak—though not supported by field data—then, for a constant decompression rate, the conduit would need to widen substantially during the eruption, as a twofold increase in MDR would enlarge the radius by ~10 m.

The 3D textures of vesicles in the tube pumice show slightly lower VND, though still within the same order of magnitude, and surface area per unit volume, resulting in an estimated decompression rate approximately half that of the "standard" ash samples. The tube pumice sample also displays the strongest iso-orientation of vesicles (Figure 10; Supplementary Material 2), which are predominantly larger in size, along with the highest variability in permeability and outgassing efficiency across the three investigated directions (Table 4). These characteristics are associated with high shear conditions occurring along the conduit walls, where friction with the bedrock is maximal and promotes elongated bubble channels, particularly at high ascent rates [Martí et al. 1999; Mastin 2005; Torres-Orozco et al. 2023]. However, these frictional forces also act to slow the ascent, generating a radial gradient in ascent velocity, explaining the lower estimated decompression rate [Gonnermann and Manga 2003]. As tube pumices are found throughout all horizons (on average 5 ± 2 % of components [Fontijn et al. 2011; Cappelli et al. 2025]), they are interpreted as a product of localised enhanced shear stresses rather than evidence of changing ascent conditions.

5.2 Bubble nucleation and growth dynamics

The absence of microlites and the very low phenocrysts abundance (≤0.03 vol.%) suggest that the bubble nucleation in the RP magma was most likely dominated by homogeneous processes. Iron-titanium oxide nanolites, which would be below the spatial resolution of the μXCT datasets, could in principle act as heterogeneous nucleation sites [Shea 2017; Cáceres et al. 2022]. However, no nanolites were observed even at the highest magnification during SEM imaging (0.04 μm pixel^{-1}), nor was the diagnostic Raman peak at ~690 cm^{-1} associated with Fe-bearing nanolites [Di Genova et al. 2017] detected at embayment outlets, where melt is in direct contact with the external bubble and water loss—and thus nanolite formation—would be expected to be most pronounced. Although the presence of undetected nanolites cannot be entirely excluded, there is no evidence that such phases formed prior to, or triggered, the onset of significant bubble nucleation. Instead, nanolite crystallisation, if present, may be promoted by melt dehydration following bubble formation, rather than preceding it. In addition, heterogeneous nucleation may also be favoured by structural heterogeneities in the melt inherited from reservoir conditions [Shea 2017, and references therein]. However, the relatively low viscosity of hot, hydrated phonolitic melts likely facilitates rapid structural relaxation and chemical homogenisation thanks to higher element diffusivity, reducing the persistence of melt heterogeneities compared to more polymerised silicic melts. This behaviour would tend to shift nucleation towards more homogeneous conditions, even though complete structural homogenisation is challenging in natural systems lacking prolonged superliquidus annealing [Shea 2017]. Therefore, while minor heterogeneous nucleation cannot be ruled out, the available textural, spectroscopic, and compositional evidence supports homogeneous nucleation as the predominant mechanism in the RP magma, occurring relatively late during ascent in the conduit under conditions of higher supersaturation pressure and shallow depths, following chamber rupture and initial dyke propagation driven by magmatic overpressure and buoyancy forces [e.g. Browning et al. 2015]. Such conditions were likely promoted by the input of a volatile-rich magma from depth [Fontijn et al. 2013; Cappelli et al. 2025].

Ascent velocity of magma in the conduit is a pivotal factor influencing processes ranging from bubble nucleation to outgassing efficiency [e.g. Gonnermann and Manga 2007; Cassidy et al. 2018, and references therein]. Decompression rates derived from the VND-based model (5.9 ± 2.4 MPa s^{-1}) align with values reported for violent explosive Plinian-style eruptions of similar compositions (e.g. Vesuvius 79$_{\text{CE}}$; Campanian Ignimbrite, Campi Flegrei; Pomici di Base, Somma-Vesuvius; Green Tuff, Pantelleria; Baricha ~87 ka; [Campagnola et al. 2016; Shea 2017; Cassidy et al. 2018; Pappalardo et al. 2018; Tadesse et al. 2024, Figure 7]). The VNDs used here—on the order of 10^{14} m^{-3}—fall within the typical range for Plinian deposits [e.g. Humphreys et al. 2008; Rust and Cashman 2011; Shea 2017; Buono et al. 2020, and references therein]. However, we note that the automated watershed algorithm may underestimate VND by failing to fully reconnect bubble walls lost during coalescence; however, even doubling the number of bubbles would not shift the order of magnitude of either VND or the resulting decompression rates. This methodological limitation therefore has minimal influence on our estimates, particularly given the consistency with 2D datasets. These VND-derived values probably represent peak decompression rates [Shea 2017], occurring during the late conduit acceleration when bubble expansion is most vigorous until fragmentation and quenching, preserving in pumice textures a maximal VND.

Embayment speedometers, in contrast, cannot record the highest decompression rates because during the final acceleration stage the timescale of H_2O exsolution can become too short for concentration gradients to track the pressure drop [Humphreys et al. 2008; Hosseini et al. 2023]. Experiments demonstrate that above ~0.25 MPa s^{-1} equilibrium degassing cannot be maintained [Gardner et al. 1999], promoting H_2O supersaturation and potentially leading to overestimation of decompression rates inferred from embayment profiles [Humphreys et al. 2008]. Our embayment-derived values (1.74 ± 0.95 MPa s^{-1}) are lower than the VND-based rates, and fall within the upper range of values reported so far in the relatively few studies using similar models (e.g. up to 1.6 MPa s^{-1} [Shea 2017; Hosseini et al. 2023]). These values are either similar (e.g. 0.9–1.6 MPa s^{-1} for the 1980 Mount St. Helens eruption [Humphreys et al. 2008]) or higher (e.g. 0.008–0.25 MPa s^{-1} for the 3.6 ka Santorini eruption [Myers et al. 2021]) than values previously reported for other Plinian eruptions. Such variations may result from the natural variability of Plinian eruptions or from the lack of standardised procedures for determining decompression rates using this method. Four our estimates, this difference may reflect our choice of starting conditions in `EMBER`, in particular allowing initial H_2O contents to vary around the observed embayment plateau concentrations rather than fixing them to melt inclusion-derived values and, accordingly, setting initial pressure equal to the saturation pressure of the plateau concentrations. Such assumptions have been shown to yield decompression rates up to 20 times higher than previous estimates [Georgeais et al. 2021] and are consistent with the expectation that ascent rates are not constant. For example, an initial slow, near-equilibrium ascent stage may allow embayments to re-equilibrate in their innermost portions and allow a diffusion gradient to start forming at pressures lower than reservoirs. Such multistep decompression paths have been invoked to reconcile discrepancies among different speedometers—VND, microlite, and embayments—within single eruptions [e.g. Harris et al. 2024].

Because the quenching plateau pressures (84 ± 16 MPa) overlap within uncertainty of melt inclusion–derived saturation pressures (92 ± 15 MPa), a prolonged initial slow-ascent phase is unlikely or negligible. Therefore, embayment-derived values are interpreted to record an average, conduit-scale decompression rate, whereas VND–based rates capture the final and fastest decompression pulse approaching fragmentation, when the magma reaches peak acceleration. Assuming a constant magmastatic gradient equivalent to the average lithostatic gradient (0.027 MPa m^{-1}), the peak decompression rate implies a maximum apparent ascent velocity of 218 m s^{-1} at fragmentation. Such high velocities are consistent with gas–magma mixture exit velocities at the vent, as inferred from explosive eruption modelling [e.g. Degruyter et al. 2012; Cassidy et al. 2018]. However, decompression rates inferred at fragmentation correspond to transient, local pressure release at the fragmentation front and cannot be directly interpreted as bulk magma ascent velocities. Instead, these values likely reflect near-vent outlet conditions dominated by rapid gas expansion, permeability increase, and shear-induced pressure relaxation, rather than vertical magma movement [e.g. Gonnermann 2015]. In contrast, the average decompression rate derived from embayment speedometry (e.g. 1.74 MPa s^{-1}) yields an ascent velocity of ~64 m s^{-1} which is more representative of conduit-scale magma transport. Such rapid ascent can suppress microlite crystallisation, favouring homogeneous nucleation during ascent.

We estimated a homogeneous supersaturation pressure for bubble nucleation of approximately 52 MPa, based on an initial reservoir pressure of 92 MPa, which corresponds to a nucleation pressure of 40 MPa. This significantly limits the time available for nucleation and growth. Decompression experiments demonstrated that a supersaturation pressure of ~50 MPa is sufficient to trigger homogeneous nucleation in evolved alkaline melts, as opposed to much higher ΔP_{sat} required in rhyolitic subalkaline or calcalkaline melts (100–180 MPa [Shea 2017; Buono et al. 2020]). In alkaline melts, the bubble number density peak (i.e. VND final order of magnitude) is almost reached within the first nucleation event at $\Delta P_{sat} \leq 50$MPa [Mourtada-Bonnefoi and Laporte 2004; Buono et al. 2020]. Further decompression may increase the VND by 1-2 orders of magnitude through multiple or continuous nucleation events [Mourtada-Bonnefoi and Laporte 2004; Gonnermann and Manga 2007; Buono et al. 2020], which coupled with bubble growth, eventually culminates in bubble coalescence and fragmentation.

The available time for bubble growth is therefore constrained between the nucleation depth (~1.9 km equivalent to a magmastatic pressure of 40 MPa) and the fragmentation level. In principle, water concentrations measured at the embayment mouths should reflect the final water saturation state of the melt prior to quenching and thus provide an estimate of the fragmentation depth. However, surface irregularities at the rims of polished glass embayments limited Raman measurements to a few micrometres inward from the embayment mouths. As a result, saturation pressures derived from these measurements (30 ± 7 MPa on average; Supplementary Material 2) are interpreted as maximum estimates of fragmentation pressure. An alternative constraint is obtained by extrapolating the best-fitting decompression path modelled to zero distance in water concentration profiles. This approach yields an average final quenching pressure of 13 ± 6 MPa, implying a relatively short interval for bubble growth. Given that nucleation initiated near 40 MPa, this corresponds to 8–42 s, assuming an average decompression rate of 1.74 ± 0.95 MPa s^{-1}.

Nevertheless, the differing timescales of decompression and volatile diffusion suggest that complete re-equilibration was unlikely. Over the inferred decompression interval, only the outer 4–10 μm within embayments would have enough time to re-equilibrate by diffusion (using $t = L^2/2D$ [Buono et al. 2020], with D from Fanara et al. [2013]), given the minimum residual ~1.3 wt.% H_2O observed at embayment mouths. Consequently, the melt likely remained supersaturated in H_2O and, also considering `EMBER` is calibrated for comparatively water-poorer calc-alkaline melts, the quenching pressure estimate may overestimate the fragmentation depth, implying that fragmentation could have occurred at slightly shallower levels.

VSDs and VVDs indicate continuous nucleation and growth under disequilibrium degassing conditions [Cashman and Mangan 1994; Klug and Cashman 1994; Blower et al. 2001; 2003; Shea et al. 2010], with minor bubble coalescence effects that flatten the VSD curves in the larger size regions, particularly evident in bottom half horizon (Figure 5). The small equivalent diameter of the modes (39 ± 6 μm) indicates a fine vesicularity (Table 2), consistent with the estimated limited time available for bubble growth. Using the time for bubble nucleation and growth (t), calculated assuming a constant decompression rate between 40 and 13 MPa, a minimum average growth rate (G) of ~7×10^{-4} mm s^{-1} was estimated via the relation $a = -1/Gt$, where a is the angular coefficient of higher VSD trends (Figure 5 [Cashman and Mangan 1994; Klug and Cashman 1994; Klug et al. 2002]). Peaks in bubble growth rates have been modelled at up to 0.1 mm s^{-1} for rhyolitic melts [Proussevitch and Sahagian 1998]. In contrast, the lower growth rates modelled for the RP suggest that bubble nucleation dominated over bubble growth even during the last stages of ascent. The exponential form of CVSDs further supports continuous bubble nucleation during ascent, driven by decompression-induced disequilibrium [Blower et al. 2001; 2003]. However, as these CVSDs are not able to develop a full power-law trend, it implies a limited number of nucleation events (< 5 [Blower et al. 2001]), likely due to insufficient time or diffusion-limited processes to allow formation of new bubbles to fill void spaces within growing ones in the final ascent stages. This suggests that VND was rapidly established mainly during early nucleation events of bubbles, which had limited time to grow. As bubbles came into contact with each other, they began to coalesce resulting in a poorly organised packing of approximately equally sized bubbles [Blower et al. 2001]. This ultimately produced a relatively low-vesicularity bubble suspension, falling below the 74–85 vol.% vesicularity range typical of polydisperse spherical foams produced during Plinian eruptions [Proussevitch et al. 1993; Cashman and Mangan 1994].

The size distribution derived from de-coalesced datasets represent bubble conditions prior to the latest stages of coalescence [Klug and Cashman 1994], whereas earlier stages cannot be resolved by watershed segmentation if merged bubbles reacquire a near-subspherical shape. The prevalent unimodal distribution of VVDs and VSDs therefore suggests that coalescence played a limited role during the early stages of bubble formation. As bubbles continued to grow, however, they eventually interconnected to form a complex vesicle network. Despite this interconnection, the short time available for bubble growth inhibited the development of sufficient permeability, preventing significant outgassing (Figure 8). Notably, permeability pathways did not develop even near conduit margins, where shear-induced bubble stretching would be expected to enhance gas escape through the alignment and channelling of elongated vesicles. Tube pumice outgassing parameters (Section 3.4.2) estimated along the primary elongation axis are comparable to those of "standard" subcircular pumice. As a result, closed-system degassing persisted, maintaining coupled magma-gas ascent that intensified the internal pressure buildup, directly contributing to the violent fragmentation observed during the RP eruption. The combination of rapid nucleation, limited time for growth, and restricted permeability created a system primed for explosive fragmentation.

5.3 Fragmentation criterion

We exclude the influence of external factors such as the input of external water or sudden decompression due to edifice collapse in promoting magma fragmentation and eruption explosivity, as no evidence for these processes exists in the RP tephrostratigraphic record [Fontijn et al. 2011; 2013]. For viscous magmas unaffected by external gases or liquid water inputs, several fragmentation criteria have been proposed to produce explosive eruptions. However, fragmentation of peralkaline magmas is complex to evaluate with conventional criteria [e.g. Shea et al. 2017] and has been associated with a combination of localised strain, bubble overpressure, and a critical bubble volume fraction [e.g. Polacci et al. 2004; Hughes et al. 2017; Pappalardo et al. 2018; Stabile et al. 2021]. To explore the conditions that led to explosivity during the RP eruption, we discuss three key fragmentation criteria: i) the strain-rate criterion [Dingwell and Webb 1989; Dingwell 1996; Papale 1999]; ii) the conduit-walls shear zone criterion [Gonnermann and Manga 2003]; and iii) the bubble overpressure criterion [Zhang 1999; Spieler et al. 2004; Mueller et al. 2008].

- i) The strain rate criterion is based on the effects of rapid elongational strain, induced by magma acceleration during ascent, on the structural response of magma [Dingwell and Webb 1989; Dingwell 1996; Papale 1999]. In this context, fragmentation is associated with crossing the glass transition threshold which occurs when the strain rate exceeds the structural relaxation time of the magma [Dingwell and Webb 1989; Dingwell 1996; Papale 1999; Gonnermann 2015]. This transition marks a shift from viscous to elastic behaviour, eventually leading to brittle failure [Gonnermann 2015]. Fragmentation can result from either a drop in temperature below the glass transition temperature or from a sudden increase in the deformation rate [Dingwell and Webb 1989]. The strain rate criterion can be described through the Maxwell relation, where the critical strain rate (γ_{crit}) is related to the reciprocal of the relaxation time (t) by the constant $a = 0.01$ [Papale 1999]:

$$\gamma_{crit} = a\frac{1}{t}. \tag{6}$$

During ascent, the elongational strain rate depends on magma acceleration and is described as dv_z/dz, or rather the dependency of magma ascent rate (v) with depth (z), where $z = 0$ m at the base of the conduit and increases upward. The structural relaxation time is linked to magma viscosity (μ) through the elastic modulus $G = 10$ GPa [Gonnermann and Manga 2003], therefore the fragmentation criterion can be written as [Papale 1999]:

$$\frac{dv_z}{dz} > a\frac{G}{\mu}. \tag{7}$$

For the RP, assuming a linear acceleration of magma from $z = 0$ m (P ~92 MPa) with an initial velocity $v_0 = 0$ m s^{-1} to peak ascent rate of 218 m s^{-1} (based on pumice textures recording the hypothetical maximum ascent rate at fragmentation) at z = 2296–2926 m (P ~30–13 MPa; Section 5.2), we can estimate a conduit-scale strain rate of 0.075–0.095 s^{-1}. Considering the relaxed viscosity (10^2–10^3 Pa s), calculated for the RP compositions using the model of Giordano et al. [2008], at initial water concentrations (i.e. 4.82 wt.%) and temperatures (i.e. 925–975 °C), the criterion cannot be satisfied (Figure 11A–11B). We acknowledge that the assumption of linear acceleration is a simplification, as conduit flow models predict strongly nonlinear increases in ascent rates at shallow depths [Papale 1999; Gonnermann and Manga 2003; 2007; Campagnola et al. 2016; La Spina et al. 2021], driven by rapid increases in gas volume fraction and buoyancy. Nevertheless, the criterion cannot be satisfied for the given relaxation state, even if strain is assumed to localize within the shallowest few metres of ascent (corresponding to the last ~0.1 MPa of decompression).

However, upon ascent, magma rheology is expected to change markedly due to processes such as water exsolution and bubble expansion cooling [Mastin and Ghiorso 2001; Gonnermann and Manga 2007], potentially increasing viscosity by 2–3 orders of magnitude. Different processes significantly influence magma temperature [Mastin and Ghiorso 2001; Gonnermann and Manga 2007, and references therein]: while friction at the conduit walls may heat the system—potentially inducing local bubble nucleation [Lavallée et al. 2015]—cooling effects like melt and gas expansion, as well as gas exsolution, dominate, especially in fast ascending magmas with large conduit radius. The system therefore can be subjected to an overall cooling trend, with gas temperature decreasing by up to 500 °C over a 50 MPa decompression [Mastin and Ghiorso 2001]. A simplified temperature balance can be derived by considering a single-phase, perfect gas system, where only the gas expansion effect is accounted for, as outlined in Equation 21 of Mastin and Ghiorso [2001]. This simplification excludes the contributions from melt expansion and gas exsolution, which together account for only 10–15 % of the total cooling. Although decompression-induced microlite crystallisation could theoretically generate a heating effect counteracting gas exsolution undercooling [Mastin and Ghiorso 2001; Blundy et al. 2006], syn-eruptive microlite crystallisation in RP magmas was extremely limited, preventing any significant crystallisation-related heating. Using this approach, we estimated a minimum temperature drop up to ~200 °C for magma ascending from pressures of 40 MPa to 13 MPa.

We therefore tested the strain rate criterion for a temperature drop down to 725 °C and a gradient of water content spanning the concentrations recorded in melt embayments. Additionally, we also evaluated the water concentration potentially reached by the melt at the inferred quenching pressure of 13 MPa (i.e. ~0.83 wt.%), estimated by Di Matteo et al. [2004] solubility model. Nevertheless, to satisfy the criterion under the most extreme conditions—lowest water content and highest degree of cooling—a minimum strain rate of 1.74 s^{-1} must be exceeded, which would require an unlikely localisation of magma acceleration within the uppermost ~90 m beneath the fragmentation surface (Figure 11A–11B). This may indicate that the magma viscosity of these compositions is generally too low to induce strain rate-dependent fragmentation caused solely by ascent-induced elongation.

- ii) The conduit wall shear zone criterion governs fragmentation at the melt-bedrock interface due to frictional forces created by the viscous flow of magma [Gonnermann and Manga 2003]. Shear-induced fragmentation occurs locally at the conduit walls when stress concentrates inducing a non-Newtonian response in the liquid and leading to the breakup of molecular bonds in a shear-thinning process [Dingwell 1996; Gonnermann and Manga 2003; Gonnermann 2015]. However, this localized process does not always generate explosive eruptions. In some cases, for highly viscous magma, it can increase permeability through melt and bedrock fractures, which facilitate outgassing and may promote effusive rather than explosive behaviour [Gonnermann and Manga 2003].

Like the strain-rate criterion, the shear zone criterion is dependent on melt viscosity at zero frequency and shear-strain rate, which is linked to conduit geometry, as postulated by Gonnermann and Manga [2003]:

$$\frac{Q}{\pi R^3} \approx C G \mu_{\text{crit}}^{-0.9}, \tag{8}$$

where Q is the volumetric flow rate ($m^3\,s^{-1}$) through a cylindrical conduit of radius R (m), C is a fitting parameter equal to 0.01 $(\mathrm{Pa\,s})^{-0.1}$, and G = 10 GPa is the elastic modulus at infinite frequency. The fragmentation criterion is satisfied when melt viscosity exceeds the critical viscosity (μ_{crit}).

To account for uncertainty in μ_{crit}, we performed 5000 Monte Carlo simulations, randomly sampling from the distributions of each input parameter. Q was obtained by dividing the MDR by the magma bulk density (melt+bubbles), which was derived from the melt density ($2250 \pm 10\ \mathrm{kg\,m^{-3}}$ [Cappelli et al. 2025]) multiplied by the complementary value of ash average vesicularity (0.63 ± 0.05). MDR values were randomly sampled from the peak MDR range estimated by Fontijn et al. [2011] (2.8–4.8 × 10^8 $\mathrm{kg\,s^{-1}}$). A statistical distribution of conduit radii was then obtained via 5000 Monte Carlo simulations of Equation 16 in Shea [2017], incorporating the given range of decompression rates ($5.9 \pm 2.4\ \mathrm{MPa\,s^{-1}}$), bulk densities, and MDRs.

The resulting critical viscosity is extremely high (~10^8 Pa s) and cannot be exceeded even for minimum water contents and a temperature drop to 725 °C (Figure 11C). Furthermore, fragmentation at conduit walls could have been mitigated by the cogenetic effect of viscous dissipation heating [Mastin 2005] or by the lubricating action of compressible fusiform bubbles—as indicated by the presence of tube pumices in the deposit—which dissipated shear stress. Additionally, no evidence of shear-produced banded obsidians, indicative of prolonged cycles of fragmentation, deformation, and reannealing, [Gonnermann and Manga 2003], was found in the deposit. While we cannot exclude that such a process occurred locally at conduit margins, it was most likely ineffective in triggering the fragmentation of the entire system.

- iii) The bubble overpressure criterion predicts magma fragmentation when the gas pressure inside bubbles exceeds the tensile strength of the magma, causing bubble-wall failure [Zhang 1999; Gonnermann and Manga 2003; Spieler et al. 2004; Mueller et al. 2008]. During magma ascent, bubbles expand due to volatile diffusion and decompression. However, if this expansion is hindered or retarded by factors such as viscous resistance, surface tension, or volumetric constraints, bubbles can develop internal overpressure in disequilibrium with the melt pressure [Zhang 1999]. In viscous magmas, bubbles rise at the same rate as the surrounding melt, limiting the overpressure dissipation through outgassing. This buildup of overpressure can lead to an energetic release once it surpasses a critical instability threshold, triggering fragmentation and propagating a shock decompression wave downward through the conduit [Gonnermann 2015]. The growth of gas overpressure depends on the efficiency of outgassing, which is related to the permeability (k_D) and vesicularity (φ) of the magma. These parameters have been experimentally linked to the critical overpressure necessary for fragmentation by the relationship [Mueller et al. 2008]:

$$\Delta P_{\text{crit}} = \frac{a\sqrt{k_D} + \sigma_m}{\varphi}, \tag{9}$$

where a and σ_m are fitting parameters equivalent to 8.21 × 10^5 $\mathrm{MPa\,m^{-1}}$ and 1.54 MPa respectively.

To calculate the pre-fragmentation gas bubble overpressure for the RP eruption, we used a simplified Rayleigh-Plesset equation [Lensky et al. 2001; Gonnermann 2015]:

$$\Delta P = \frac{2\sigma}{r} + 4\mu\frac{G}{r}, \tag{10}$$

which relates the bubble average radius r (m) to bubble growth rate G ($\mathrm{m\,s^{-1}}$), where ς is the bubble surface tension, assumed to be ~0.1 $\mathrm{N\,m^{-1}}$ [Gonnermann 2015]. We tested the criterion across a range of viscosities (μ), iteratively adjusting water concentration and temperature until the bubble overpressure exceeded the critical threshold of 3.54 ± 0.21 MPa derived from Equation 9. The criterion is satisfied for the minimum water concentrations recorded in the embayment glasses and for temperatures ≤725 °C, while for water contents at quenching pressure it results satisfied for temperatures of ~760 °C (Figure 11D).

Bubble growth during magma ascent is controlled by diffusion-driven volatile exsolution and decompression-induced gas expansion. Water concentration gradients in embayments indicate that diffusion was not particularly effective in promoting exsolution. This is especially relevant considering that exsolution itself increases the viscosity of the melt around bubbles, thereby inhibiting further diffusion. As a result, gas expansion likely played a dominant role. However, we propose that growth was eventually limited, potentially restricting undercooling of the entire melt system. Instead, intense localised thermal gradients potentially formed at the bubble-melt interface inducing bubble wall rupture [Mastin and Ghiorso 2001; Hughes et al. 2017], further supporting the bubble overpressure criterion.

In summary, we find that the bubble overpressure fragmentation criterion is satisfied, driven by drastic changes in melt viscosity induced by water exsolution and melt cooling. The conditions for RP explosivity are therefore linked to pre-eruptive water concentrations. Although these concentrations are lower than those observed in other phonolitic and trachytic systems [Carroll and Blank 1997; Berndt et al. 2001; Romano et al. 2022], their combination with rapid ascent rates and initial overheating prevented microlite crystallisation and induced delayed bubble nucleation capable of forming an extremely energetic bubble suspension. We infer that the interplay between relatively shallow storage conditions and rapid ascent velocity was critical in governing magma behaviour. RP reservoir was likely destabilised by input of a volatile-rich magma from beneath [Fontijn et al. 2013; Cappelli et al. 2025]. Such high ascent rates were possibly facilitated by the reservoir's relatively low density and viscosity, which allowed the rapid rise of hot, microlite-free, and pressurised magma. Shallow magmatic conditions are not unusual for phonolitic-trachytic reservoirs associated with explosive eruptions [e.g. Andújar et al. 2008; Scaillet et al. 2008]. It has been documented how explosivity in such magmas is marked by the positive correlation of pressure and water concentration of reservoirs [Andújar and Scaillet 2012], highlighting an interdependence between shallow conditions and low water content in driving explosive activity, even under undersaturated conditions [Andújar and Scaillet 2012]. The case for the RP eruption aligns with this trend while remaining unique in its specific conditions, offering a valuable comparison for other little-studied alkaline systems along the EAR that might share similar pre-eruptive shallow conditions.

6 Conclusions

In this study, we combined 2D and 3D textural methods with embayment water diffusivity-dependent speedometers to unravel the conduit processes driving the explosivity of the Rungwe Pumice eruption. We demonstrated how, during unrest, the rapid ascent of hot magma from shallow crustal levels inhibited microlite crystallisation. The absence of nucleation sites delayed bubble formation, leading to spontaneous nucleation at high supersaturation pressures. Energetic bubble nucleation outburst further accelerated magma ascent. However, bubble growth was constrained by the packing of bubbles into an intricate pore network and the limited time available before brittle fragmentation. This process led to bubble overpressure, which could not dissipate via outgassing through the vesicle network due to rapid ascent and insufficient time for the development of permeability pathways. Eventually, bubble overpressure overcame the critical threshold when rheological changes induced by temperature drops and water exsolution (likely concentrated at bubble interconnections) allowed the brittle failure of the melt. Our model highlights the challenges in applying conventional fragmentation criteria to microlite-free and phenocryst-poor, relatively low-viscosity peralkaline magmas. In such systems, significant rheological alterations must be inferred to explain fragmentation. These findings suggest that further experimental work is essential to better characterise the fragmentation dynamics of these magmatic systems.

Author Contributions

This study was conceptualized by LC and KF. Data were collected by LC together with LP, GB, TDG, and VNS. Fieldwork was conducted by KF and facilitated by EM with contributions from SK, EA, and GGJE. LC handled data curation and drafted the manuscript. KF reviewed the first draft. All authors contributed to scientific discussion and manuscript revision.

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Data Availability

The raw 3DXCT scan data are available in the GFZ Data Services repository at: https://doi.org/10.5880/fidgeo.2025.023.

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