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Key Points:

- Seismic wave attenuation can be used to determine gas-charged magma ascent prior to eruption
- 3D imaging of vesiculation and wave attenuation modeling show crystal-poor magmas are detected at larger depth than crystal-rich ones
- Low-frequency seismic attenuation can be used to pinpoint the gas accumulation in shallow magma reservoirs prior to eruption

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Seismic Attenuation During Magma Vesiculation: A Combination of Laboratory Constraints and Modeling

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Abstract Assessing the potential and likelihood of explosive eruptions is vital for almost 800 million people living near active volcanoes. Magma reservoirs are not directly accessible, limiting our capability to understand the prerequisites for an explosive eruption and our ability to perform experiments that simulate magmatic processes at depth. Therefore, deciphering volcanic precursors is key to forecasting volcanic eruptions. In this study, we used synchrotron-based X-ray tomographic microscopy to observe felsic magma vesiculation at high temperature (1,100°C) to establish the pre-eruptive conditions that favor explosive activity. Specifically, we studied the increase of seismic attenuation with increasing gas content during magma vesiculation. These results are applied to the natural range of volcanic tremors (0.2–5 Hz) to detect the proximity of gas-charged magma to the surface prior to volcanic eruption: gas-rich, crystal-free felsic magmas are detected at larger depth (40 MPa) than gas-rich, crystal-bearing systems (10 MPa).

Plain Language Summary Forecasting the likelihood of explosive eruptions is vital for almost 800 million people living near active volcanoes. To this goal, volcano monitoring that uses seismic waves to image magmas at depth is crucial for deciphering the causes driving an explosive eruption. As we have no direct access to magmatic processes at depth, here we use very high-resolution 3D imaging techniques to observe the formation of gas bubbles in a lab-scale magma to establish the conditions that favor explosive activity. In particular, we correlate how the presence of excess gas in magmas can modulate the seismic imaging of magmas prior to their eruption. These results offer the opportunity to better illuminate the causes of explosive eruptions and how to apply seismic imaging for detecting the proximity of gas-rich magma to the surface prior to a volcanic eruption.

1. Introduction

Deciphering volcanic precursors (ground deformation, gas emissions, seismic activity) is key to forecasting volcanic eruptions. A longstanding task of volcanology is to detect the stress changes occurring around volcanoes. These stress changes can be detected by monitoring seismic attenuation with persistent seismic vibrations of low-frequency or tremors (Konstantinou & Schlindwein, 2003), typically appearing during unrest and eruption of a multiphase magma ascending in the volcanic conduit (Neuberg et al., 1998). Multiple sources can produce volcanic tremors: resonance within a magma-filled conduit, resonance of hydrothermal fluids within cracks, movement of fluids through a conduit, bubble growth and resorption in aquifers, brittle failure of magma undergoing high strain/strain-rate, and a combination of different low-frequency sources generating a single tremor waveform (Chouet & Matoza, 2013). The tremors associated with fluid dynamics in volcanic conduits are primary signals that shed light on the magma movement and vesiculation at depth prior to the eruption (Chouet & Matoza, 2013). These tremors were recorded in the period preceding eruptions at Soufrière Hills, Montserrat (Neuberg et al., 1998), Redoubt, Alaska (Chouet et al., 1994), and Galeras, Columbia (Fischer et al., 1994), and during magma unrest accompanied by wave-induced fluid flow (De Siena et al., 2017). Low-frequency seismicity as precursory signal of gas-driven eruptions (Dmitrieva et al., 2013; Varley et al., 2010) provides an important tool for assessing changes in the internal dynamics of volcanic conduits. Yet, no physical model that details the variation of seismic wave attenuation as a function of the presence of rapidly expanding gas during magma ascent and unrest is available.

The probability of an explosive volcanic eruption is inversely related to the efficiency of gas removal during magma ascent to the surface (Sparks, 2003). In high-viscosity systems, as magmas rise from depth, and overburden pressure decreases, volatiles exsolve forming bubbles, which grow and ascend with the magma (Navon et al., 1998). The decrease of H₂O concentration in magmas triggers crystallization (Hammer, 2008) and leads to an increase of magma viscosity (Caricchi et al., 2007). If volatiles are not readily removed from magma during this runaway process, the increase of viscosity, excess volatiles, and gas pressure may lead to explosive eruptions (Klug & Cashman, 1996; Melnik et al., 2005). Determining the details of numerous factors interacting to regulate the causes of explosive eruptions is inherently complex. Thus, in an attempt to anticipate eruptive behavior, we focus on understanding magma properties in the pre-eruptive stage of a volcanic crisis.

Several experiments have been conducted to explore the mechanisms of gas exsolution and permeability in magmas at depth (e.g., Larsen et al., 2004; Takeuchi et al., 2005). Yet in these studies, only the final run products, representing the last snapshot of a dynamic process, were analyzed. Pioneering experiments, combining high-temperature devices, and X-ray tomographic microscopy, have significantly contributed to the understanding of bubble formation and growth in mafic (Bai et al., 2008; Baker et al., 2012) and felsic systems (Gondé et al., 2011; Pistone et al., 2015). In this study, we designed a laboratory experiment to monitor the in situ generation of excess gas upon vesiculation in a time-elapsd manner. This included the bubble growth kinetics and the response of melt to gas expansion under isobaric and isothermal conditions. Specifically, we observed the growth of bubbles at magmatic temperatures using 3D and time-lapse analyses (Fife et al., 2012) and traced the physical properties during the experiment. The dynamics of high-temperature vesiculation in the selected sample is described in detail in Pistone et al. (2015), who performed several tests that demonstrate experimental reproducibility at identical conditions of sample heating and quenching and data acquisition at the TOMCAT beamline. Here we focus on how gas exsolution proceeds during rapid vesiculation with slow, diffusion-controlled bubble growth in a non-crystallizing melt and correlate it to the resulting seismic attenuation in a gas-charged magma-filled conduit resonator (Neuberg et al., 1998) at different depths using existing theoretical models (Collier et al., 2006), recently confirmed experimentally (Tisato et al., 2015). Here we do not consider other possible tremor sources (e.g., fluid-filled crack resonator; Molina et al., 2004) that could generate volcanic tremors.

2. Materials and Methods

A cylindrical (2 × 2 mm) crystal-free rhyolitic obsidian from Mayor Island (New Zealand) was used as a bubble system (with exsolved gas content expressed as bubble volume fraction or porosity, $\beta = 0.02$) at ambient pressure. Non-destructive X-ray tomographic microscopy was conducted at the TOMCAT beamline of the Swiss Light Source of the PSI (Switzerland), with a sample–detector distance of 300 mm and a high speed pco. Dimax CMOS camera. Using a 5% filter on the polychromatic beam (Marone et al., 2020) and readout of individual projections with 3 μ m isotropic voxel size in ~ 1 ms, which is orders of magnitude faster than traditional CCD cameras, a complete 3D data set (1,000 projections over 180° rotation) was acquired in 1 s.

High-temperature conditions were generated using the laser-based heating system available at TOMCAT (Fife et al., 2012) in combination with a hot-pressed boron nitride (BN) holder hosting the sample and with a BN rod placed on the top of the sample. In this setup, the sample is allowed to vesiculate radially and vertically. From the sample glass transition of 592°C, as estimated using the model of Giordano et al. (2008), temperature increase was executed with a rapid heating rate (2°C/s) to 1,100°C. Gas exsolution was monitored under isobaric (1 bar) and isothermal conditions (1,100°C), which ensured to replicate the melt viscosity ($\eta_{\text{melt}} \sim 10^4$ Pa s) and the vesiculation process expected in natural H₂O-saturated (5.7 wt.% H₂O in the silicate melt) magmas ascending from shallow depth (6–7 km depth = 200 MPa) at 800°C. After a dwell time of 15 min at 1,100°C, recorded with multiple 3D tomographic renderings, sample quenching was performed by switching off the laser power. Note, due to the read-out of the camera, the microstructural evolution upon sample vesiculation was only captured every 2 min.

3D image volume reconstructions were performed using GRIDREC (Marone & Stampanoni, 2012), and a single-distance phase-retrieval algorithm (Paganin et al., 2002) to enhance the contrast between gas and melt. The reconstructed volumes were then processed using the Bruker Skyscan CTVOX software (UNIL).

A custom set of algorithms developed at the TOMCAT beamline (Mader et al., 2012) were used to segment the tomographic reconstructions into individual gas bubbles and quantify shape metrics based on the shape tensor computed for each bubble. To track the bubble growth and expansion during the experiment, the melt–bubble interface movement with time was monitored. This gas–melt interface velocity inspected along the sample radius (radial velocity, V_r) and length (vertical velocity, V_z) during vesiculation is scaled by a characteristic velocity based on the time-integrated variation of the sample length (surface area/unit volume). Null interface velocity indicates gas bubbles not moving, and positive and negative values indicate bubbles expand and impinge against other bubbles, respectively.

The quality factor (Q) or seismic attenuation ($1/Q$) as a function of increasing β during magma vesiculation was determined in the frequency range from 0.01 to 10 Hz using a modified version of the model of Collier et al. (2006), originally assuming no interactions between different bubbles. This modified version considers that attenuation is the result of dynamic dissolution, exsolution, and diffusion of H_2O in the magma and assumes a Neumann (no-flux) boundary condition to assess H_2O mass conservation (i.e., the gradient is zero at the external boundary; Lyakhovsky et al., 1996). The model is designed as a finite difference code encompassing a Euler time-marching explicit scheme that solves: (a) the rate of change of the bubble radius, (b) the H_2O concentration gradient at the bubble–melt interface, (c) the variation of density in the bubble, and (d) the diffusion of H_2O in the melt. The first three equations correspond to the equations B1, B2, and B3 from Lyakhovsky et al. (1996), while the latter is a 1D Fick's law in spherical coordinates. While there are other available models that include viscoelasticity and compressibility of the melt in a bubbly magma, which are known factors that significantly increase Q at frequencies >1 Hz (Kurzon et al., 2008, 2011), the model of Collier et al. (2006) still offers an acceptable computational analysis in our study that: (a) excludes bubble coalescence and considers only the cumulative increase of β ; (b) simulates the ascent of compressible magma by applying discrete estimates of melt density (ρ_{melt} ; Lange & Carmichael, 1987) and viscosity (η_{melt} ; Giordano et al., 2008), bubble-bearing magma density (Melnik et al., 2005), H_2O diffusivity (D_{H_2O} ; Ni & Zhang, 2008) and solubility in melt (H_2O_{melt} ; Papale et al., 2006), and gas overpressure (ΔP_g ; Pitzer & Sterner, 1994) (Table 1) at each experimental time frame and depth of conduit scenario; (c) simulates relatively slow, diffusion-controlled bubble growth in magmas.

3. Results

The 3D renderings illustrate the profound change of bubble texture during vesiculation (Figure 1). Gas expansion promotes measurable volumetric inflation of the sample, resulting in the formation of foam with polygonal structure typical of felsic frothy pumices. Initially, the sample expands radially and eventually starts rising lifting the BN lid (Figures 1 and 2a). The bulk volumetric sample expansion produced by vesiculation reaches $\sim 30\%$ with β increasing locally up to ~ 0.9 in the sample location where the laser spots are focused (Figure 1). The lower portion of the sample is relatively cool as evidenced by the lack of vesiculation (Figure 1). The upper sample portion shows presence of gas bubbles with maximum diameters up to ~ 350 μm . This is the largest size achieved by bubbles prior to their reciprocal impingement along melt films and at the three neighboring bubble edges (see yellow arrows in Figure 1e). As the bubble growth proceeds with a rate of 1 $\mu m/s$ above 1,000°C (Pistone et al., 2015), the maximum size of ~ 350 μm is achieved ~ 6 min after attaining 1,100°C. After ~ 6 min, bubble impingement initiates with limited bubble deformation and, in places, with local bubble connectivity (Figures 1b–1d).

The Q modeled upon the experimental observations of the sample portion heated by the laser decreases by 2.5 orders of magnitude in the first 2 min of gas expansion above the glass transition temperature. Subsequently, Q decreases only by 10% during further course of the experiment (Figure 1f) due to the dominant bubble impingement that prevents any further gas expansion (Figure 1e). At frequency >3 Hz, Q is ~ 1 in the three experimental frames above the glass transition temperature (Figure 1). In the observation bandwidth (0.2–5 Hz), Q decreases as the bubble radius increases (Figures 1b–1d). Such a variation can be explained by the shift in Q peak as the bubbles change their dimension. Where the spatial resolution can resolve the presence of thin melt films separating the bubbles (Figure 1e), we can observe that the gas–melt interface velocity decreases when bubble impingement occurs at 7 min from the start of the experiment (Figure 1c) and increases during bubble coalescence and at the sample periphery (Figure 2). The bubble shape evolution is in line with the evolution of both V_r and V_z and the gas–melt interface moves with $V_r \sim 0$ and maximized

Table 1
Summary of All Input Values in the Seismic Attenuation model

In situ X-ray tomographic experiment								
This study			Pistone et al. (2015)	Lange and Carmichael (1987)	Giordano et al. (2008)	Pitzer and Sterner (1994)	Burnham (1975)	Ni and Zhang (2008)
T (°C)	r (μm)	β	H_2O_{melt} (wt.%)	ρ_{melt} (kg/m ³)	η_{melt} (Pa s)	ΔP_g (MPa)	H (mol/[Pa m ³])	D_{H_2O} (m ² / s)
592	93	0.02	0.24	2,341	1.00E + 12	0.10	5.63E – 03	2.55E – 13
1,100	107	0.33	0.06	2,353	2.13E + 06	1.04	9.00E – 02	5.30E – 11
1,100	108	0.45	0.06	2,353	2.13E + 06	1.04	9.00E – 02	5.30E – 11
1,100	110	0.67	0.05	2,354	2.28E + 06	1.04	1.30E – 01	5.28E – 11
1,100	111	0.74	0.05	2,354	2.28E + 06	1.04	1.30E – 01	5.28E – 11
1,100	111	0.84	0.05	2,354	2.28E + 06	1.13	1.30E – 01	5.28E – 11
1,100	113	0.87	0.04	2,354	2.46E + 06	1.13	2.03E – 01	5.27E – 11
Simulated Scenarios of 1996 Soufrière Hills ($r = 10$ μm) and 2008 Chaitén ($r = 100$ μm)								
This Study			Papale et al. (2006)	Lange and Carmichael (1987)	Giordano et al. (2008)	Pitzer and Sterner (1994)	Burnham (1975)	Ni and Zhang (2008)
P (MPa)	r (μm)	β	H_2O_{melt} (wt.%)	ρ_{melt} (kg/m ³)	η_{melt} (Pa s)	ΔP_g (MPa)	H (mol/[Pa m ³])	D_{H_2O} (m ² / s)
120	10 & 100	0.00	5.70	2,095	1.05 E + 04	1.00	9.97E – 06	4.24E – 11
110	10 & 100	0.07	5.63	2,097	1.11 E + 04	1.00	1.02E – 05	3.87E – 11
100	10 & 100	0.23	5.42	2,096	1.30 E + 04	1.00	1.10E – 05	3.55E – 11
90	10 & 100	0.25	5.19	2,105	1.56 E + 04	1.00	1.20E – 05	3.20E – 11
80	10 & 100	0.27	4.93	2,115	1.94 E + 04	1.00	1.33E – 05	2.85E – 11
70	10 & 100	0.29	4.65	2,127	2.46 E + 04	1.00	1.50E – 05	2.50E – 11
60	10 & 100	0.32	4.34	2,140	3.26 E + 04	1.00	1.72E – 05	2.17E – 11
50	10 & 100	0.36	4.00	2,154	4.50 E + 04	1.00	2.03E – 05	1.84E – 11
40	10 & 100	0.42	3.60	2,171	6.78 E + 04	1.00	2.50E – 05	1.52E – 11
30	10 & 100	0.49	3.13	2,191	1.15 E + 05	1.00	3.31E – 05	1.23E – 11
20	10 & 100	0.60	2.55	2,215	2.39 E + 05	1.00	4.98E – 05	9.39E – 12
10	10 & 100	0.80	1.77	2,245	8.06 E + 05	1.00	1.03E – 04	6.65E – 12

P = Pressure; β = Bubble Volume Fraction; r = Bubble Radius; H_2O_{melt} = Water Dissolved in the Melt; ρ_{melt} = Melt Density; η_{melt} = Melt Viscosity; ΔP_g = Gas Overpressure; H = Henry's Law Solubility Constant; D_{H_2O} = water diffusivity.

V_z (Figure 2). As evidenced by the superimposition of interface velocity areas for each radial sample portion (Figure 2a), V_z increases sharply in the first 3.4 min and, in the remaining 8.1 min of the experiment, decreases with increasing β and impingement hampering bubbles from expanding any further (Figure 2).

4. Discussion

The in situ time-elapsd observation applied in this study allowed to parametrize the kinetics of vesiculation that produced variations in β in the magma. We observed the evolution of an initial silicate melt becoming a foam ($\beta \approx 0.9$; Figure 1) that produces a major decrease of its bulk viscosity by four orders of magnitude and its density by >50% (Melnik et al., 2005). The decrease of these two parameters make the magma become largely buoyant and rheologically mobile with respect to the surrounding rocks.

Bubble growth is controlled by two main parameters (Navon et al., 1998): (a) DH_2O (diffusion-limited regime); and (b) η_{melt} (viscosity-limited regime). To define these two regimes of bubble growth, the timescales for diffusion (τ_d) and viscous deformation (τ_v) must be determined:

Figure 1. In situ lab-scale reservoir experiment of sample vesiculation. (a)–(d) Orthoslices from the 3D sample with gas bubbles (black) and glass/melt (light gray). (e) Exploded view of panel c with yellow arrows indicating representative regions of bubble impingement. (f) Seismic attenuation model based upon the experimental observations reporting the frequency range of natural volcanic tremors in gray. β = bubble volume fraction; r = bubble radius.

$$\tau_d = \frac{r^2}{D_{H_2O}} \quad (1)$$

$$\tau_v = \frac{\eta_{melt}}{\Delta P_g} \quad (2)$$

To assess whether bubble growth is diffusion- or viscosity-limited, we must determine the non-dimensional Péclet number (Pe) using Equations 1 and 2:

Figure 2. Gas–melt interface vertical (V_z) and radial velocity (V_r) distribution within the sample in each time frame, made of two successive 3D renderings. (a) Velocity distribution areas obtained by cumulative ten velocity distribution lines for each radial sample portion, measured every 50 μm from sample's inner portion to outer periphery. (b) Gas–melt interface velocity variations due to gas impingement and connectivity (blue), moderate gas expansion slowed down by bubble impingement (yellow), and large gas expansion within available space between pre-existing bubbles (brown).

$$Pe = \frac{\tau_d}{\tau_v} = \frac{\Delta P_g r^2}{\eta_{melt} D_{H_2O}} \quad (3)$$

Using the values in Table 1, the experiment simulated H_2O exsolution under conditions of $Pe \gg 1$, which suggests that H_2O diffusion cannot keep pace with the rate of expansion (1 $\mu\text{m/s}$) and the bubble growth is diffusion-limited. Gas expansion during exsolution produces “ H_2O -supersaturation” of the silicate melt due to short H_2O diffusion length scale ($l \sim 200 \mu\text{m}$) using $l = \sqrt{t \times D_{H_2O}}$, with t as the experiment timescale (11.5 min; Figure 2) and $D_{H_2O} = 5.3 \cdot 10^{-11} \text{ m}^2/\text{s}$ at 1,100°C (Table 1). Therefore, the limited D_{H_2O} produces a lower η_{melt} around an individual bubble and, consequently, gas bubbles impinge plastically without coalescence by brittle rupture of the melt film (Figures 1 and 2). The generation of impermeable foams tends to favor rapid magma decompression-induced explosive eruptions (Gonnermann & Manga, 2007).

Quantifying bubble growth kinetics unravels the physical and chemical causes modulating seismic attenuation induced by ascending magmas. In particular, at $Pe \gg 1$, in absence of gas coalescence and loss, and with r remaining fairly constant (93–113 μm ; Figures 1a–1d), we observe that an increase of β from 0.02 to 0.45 from 592°C to 1,100°C (Figures 1a–1d) results in a decrease of Q by 2.5 orders of magnitude within the frequency range of volcanic tremors (Figure 1f). The Q values are maxima here as the pressure wave amplification produced by viscoelasticity and compressibility of the melt (Kurzon et al., 2008) and presence of large growing bubbles in a bubbly magma (Kurzon et al., 2011) are not included in the model of Collier et al. (2006). But the effective amplification is characterized by a Q reduction of $\sim 30\%$ – $\sim 5\%$ at β values between 0.33 and 0.87, and mainly noticed at frequencies $> 1 \text{ Hz}$ (Kurzon et al., 2011). Moreover, microscopic gas bubbles (18 μm size) at $\beta < 0.01$ can have a large attenuative effect on the seismic wave propagation by decreasing Q by up to $\sim 40\%$ in shallow reservoirs located at $\sim 1 \text{ km}$ depth (Tisato et al., 2015). The model of Collier et al. (2006) is thus applicable to magmas that rapidly vesiculate at $Pe \gg 1$ and can offer plausible estimations of Q in the simulation of natural scenarios of ascending multiphase magmas.

5. Implications

One of the main challenges in active volcano monitoring is to correlate the low frequency (0.2–5 Hz) seismic events to the state of magma unrest preceding a volcanic eruption (Lahr et al., 1994; Neuberg et al., 1998). Such events are inferred to be related to the movement of highly viscous, gas-charged magma above H_2O -saturation depth where bubble nucleation initiates. Bubbles damp seismic waves (Chouet, 1996). While extrinsic seismic attenuation can be estimated from studying the propagation of seismic waves with distance (e.g., Gao et al., 1993), intrinsic attenuation can be calculated from magma properties and energy-absorbing processes. To assess whether attenuation can help forecast the formation and increase of bubbles before an eruption, we consider the wave-induced thermodynamic disequilibrium between magma and H_2O , causing elastic energy absorption in a gas-charged magma.

We applied the model of Collier et al. (2006) to two natural scenarios: 1996 activity of Soufrière Hills Volcano (Montserrat) and May 1, 2008 explosive eruption of Chaitén Volcano (Chile). The first scenario includes a crystal-rich (40 vol.%) andesite with a residual rhyolitic melt carrying small bubbles ($r = 10 \mu\text{m}$) entrapped within the crystal framework. The parametrization of the andesite ascent is from Collier et al. (2006). The second scenario includes a crystal-free rhyolite with average $r = 100 \mu\text{m}$ (from this study) and β estimated using the model of Papale et al. (2006) at H_2O -saturated conditions, in absence of other major volatile components (i.e., CO_2 and S species). Assuming slow, near-isothermal (800°C) magma decompression that minimizes delayed nucleation of bubbles that, in turn, favor bubble growth only at shallow level within the conduit (Mangan & Sisson, 2000), the ascent of the H_2O -saturated (5.7 wt.%) magma is modeled from a reservoir located at $\sim 6 \pm 0.5$ – ~ 0.5 km below the surface. During the simulated magma decompression, β increases with increasing the number of bubbles having constant radius (r) due to a combination of H_2O_{melt} decrease (from 5.7 to 1.8 wt.%), η_{melt} increase ($\sim 10^4$ – $\sim 10^6$ Pa s), and relatively low DH_2O at 800°C ($> 7 \cdot 10^{-12} \text{ m}^2/\text{s}$), which favors diffusion-limited gas expansion (i.e., $Pe \gg 1$). Table 1 summarizes all input parameters used in the two simulated scenarios.

Both Soufrière Hills crystal-rich andesite and Chaitén crystal-poor rhyolite display a closely similar residual rhyolitic melt composition. Therefore, we are able to track the evolution of the physical properties of comparable magmatic systems, such as β , H_2O_{melt} , η_{melt} , and ρ_{melt} during the simulated magma ascent (Figures 3a and 3b). Both scenarios showcase a decrease in absolute Q peak and slight shift in frequency of the maximum Q value (Figures 3c and 3d). In the case of Soufrière Hills, the maximum β is ~ 0.3 in presence of 40 vol.% crystals, which causes an appreciable decrease of Q at relatively great depth (> 40 MPa or $> \sim 2$ km depth; Figure 3c). At shallow depth (10 MPa or ~ 0.5 km), the attenuation becomes more pronounced in the range $Q \approx 3$ to 10 at frequency > 1 Hz (Figure 3c). At 0.2–0.3 Hz, there is no clear distinction of Q values at different pressures (Figure 3c). Due to the large crystal content (40 vol.%), the frequency decreases from 5 to 1 Hz towards lower values of Q in the observation bandwidth (Figure 3c) at pressure > 40 MPa.

In the case of Chaitén, we observe no frequency shift within the range of natural volcanic tremors (Figure 3d). However, we observe a sharp decrease of Q peak by ~ 1 order of magnitude from 100 to 10 MPa, corresponding to an increase of β from 0.23 to 0.8 (Table 1). The most evident decrease of Q occurs from 40 to 10 MPa (Figure 3c), corresponding to an increase of β by 53% from 0.42 to 0.8 (Table 1). If we compare both scenarios, at 100 MPa, it is challenging to discriminate a crystal-bearing system with $\beta = 0.3$ from a crystal-free magma with $\beta = 0.23$ (Figures 3c and 3d). With decreasing pressure, the Q peak of Chaitén results lower than the Q peak of Soufrière Hills by 0.2 orders of magnitude (Figure 3d). While the attenuative response of Soufrière Hills magma is best detected ($Q < 1$) once the magma is at 10 MPa, Chaitén magma becomes already detectable ($Q < 1$) at 40 MPa in the frequency range of volcanic tremors (Figures 3c and 3d). At 40 MPa, a H_2O -saturated crystal-free magma with $\beta \approx 0.42$ (Table 1) can generate a Q decrease $> 75\%$ at 1 Hz (Figure 3d). This result highlights the potential detectability of gas-rich magmas at a depth that is not too proximal to the volcanic vent (> 0.5 km) before the magma becomes highly bubbly and prone to erupt.

This study highlights that the accumulation of volatiles in the shallow magma reservoir (< 100 MPa) exacerbate the low frequency range (< 10 Hz; Figures 3c and 3d) and, at equivalent volume of excess fluids, crystal-poor magmas can be detected at larger depth than crystal-rich systems. Laboratory testing and modeling are suitable tools that can offer aid to geophysical methods, such as time-resolved 3D seismic tomography (Patanè et al., 2006) using seismic attenuation to pinpoint which subvolcanic reservoir hosts gas-charged

Figure 3. Modeling of seismic attenuation of two ascending H₂O-saturated magmas: Soufrière Hills (Montserrat) and Chaitén (Chile). Decompression-induced change of the physical properties of both magmas having nearly identical melt composition but different crystal volume fraction (ϕ): (a) β = bubble volume fraction, H_2O_{melt} = residual H₂O dissolved in melt, (b) ρ_{melt} = melt density, η_{melt} = melt viscosity (Table 1), and (c)–(d) Q = quality factor.

magma. This critical evaluation of the actual depth of storage of gas-charged magma prior to unrest using seismic attenuation is key to the assessment of volcanic forecast and hazard to a specific well monitored active volcano and may be applied in a timely manner.

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Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

All data generated in this study are summarized in Table 1 available in this public domain data repository: https://zenodo.org/record/4651226#.YGYtSWiQi_s. Tomographic data as tomographic reconstructions are available in the PSI public domain data repository: doi.org/10.16907/a672b15a-9973-4d92-bf1e-3488cd-fb4d1b. The custom set of algorithms used to segment the tomographic reconstructions into individual gas bubbles and quantify shape metrics based on the shape tensor computed for each bubble is freely available in this public domain data repository: <https://zenodo.org/record/4660345#.YGde3WiQiMI>.

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