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The trigger mechanism of low-frequency earthquakes on Montserrat

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Abstract

A careful analysis of low-frequency seismic events on Soufrièere Hills volcano, Montserrat, points to a source mechanism that is non-destructive, repetitive, and has a stationary source location. By combining these seismological clues with new field evidence and numerical magma flow modelling, we propose a seismic trigger model which is based on brittle failure of magma in the glass transition. Loss of heat and gas from the magma results in a strong viscosity gradient across a dyke or conduit. This leads to a build-up of shear stress near the conduit wall where magma can rupture in a brittle manner, as field evidence from a rhyolitic dyke demonstrates. This brittle failure provides seismic energy, the majority of which is trapped in the conduit or dyke forming the low-frequency coda of the observed seismic signal. The trigger source location marks the transition from ductile conduit flow to friction-controlled magma ascent. As the trigger mechanism is governed by the depth-dependent magma parameters, the source location remains fixed at a depth where the conditions allow brittle failure. This is reflected in the fixed seismic source locations.

Keywords: seismic trigger; low-frequency events; conduit resonance; Soufrière Hills; Montserrat

1. Introduction and background

1.1. Low-frequency seismic earthquakes

One type of seismic event proved to be pivotal to modern volcano seismology: the *low-frequency earthquake*. Several examples have been quoted where these events have been successfully used to forecast volcanic eruptions (e.g. Chouet, 1996). Their occurrence alone has been used as an indicative tool even though a full understanding of their source mechanism has not yet been achieved. On Galeras volcano, Colombia, the low-frequency earthquake, there referred to as a *tornillo* (Spanish for screw), played a major role in the heated scientific and public debate regarding whether lives would have been saved had their occurrence been recognised (Gil Cruz and Chouet, 1997).

Low-frequency seismic events at Soufrièere Hills volcano on Montserrat, West Indies, occupy a spectral range of approximately 0.2–10 Hz and form a spectral continuum between the following two end-members of seismic event classes: the long-period earthquakes, or LPs for short, and so-called hybrid events. Hybrid events are LPs with an additional high-frequency onset; however, the amount of high frequency energy observed in the seismic signal on Montserrat can vary from station to station, and also at one seismic station from event to event. Therefore, LPs and hybrids on Montserrat are grouped together as low-frequency earthquakes (Neuberg, 2000). They have been observed at many volcanoes and are associated empirically with

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the pressurization of the volcanic system (Lahr et al., 1994; Voight et al., 1998; Neuberg et al., 1998).

Low-frequency earthquakes differ significantly from any other seismic signal on a volcano because, unlike tectonic earthquakes, they originate from a boundary between a fluid, such as magma or gas, and the solid surrounding rock (e.g. Ferrazzini and Aki, 1987; Chouet, 1988; Neuberg et al., 2000). Most of the seismic energy is trapped in the fluid-filled conduit–hence the term *conduit resonance*–and only a small part (dependent on the contrast of elastic parameters across the boundary) can escape from the conduit, propagate through the solid medium, and be recorded at a seismic station. Hence, low-frequency events are interface waves similar to surface waves generated by an earthquake. They are dispersive, i.e., their propagation velocity along the interface depends on their wavelength, the width of the dyke or conduit, and the contrast of elastic parameters across the interface. Therefore, these are the quantities that can be determined by studying low-frequency events. It is the dispersive property that explains their low frequency content (Ferrazzini and Aki, 1987).

In recent studies, low-frequency event modelling falls into two groups: the events are modelled by a pressure transient in a resonating crack of only a few centimetres width (e.g. Chouet, 1996) or, in contrast, a resonating conduit section of approximately 30 m width and several hundred meters length, where magma properties are depth and time-dependent (e.g. Neuberg, 2000). No matter which of these models are used, it appears that this type of seismic signal provides one of the few direct links between observations at the surface and the physical processes operating in magma-filled conduits and dykes. Therefore, its analysis is an essential component in any attempt to quantify and predict internal volcanic activity.

One exciting aspect of low-frequency seismic earthquakes on Montserrat and other volcanoes is their potential use as a forecasting tool for volcanic eruptions. On Montserrat, most major collapses of an andesitic lava dome are preceded by swarms of low-frequency earthquakes (Miller et al., 1998). However, these earthquake swarms are not always followed by a dome collapse. Galeras volcano in Colombia shows a similar pattern of monochromatic low-frequency earthquakes that are observed before eruptions (Hellweg, 2003). However, also at Galeras, the occurrence of these events does not guarantee an eruption. This demonstrates that the mere occurrence of low-frequency earthquakes, tornillos, or LPs is not yet a reliable forecasting tool. Unless the actual trigger mechanism is understood no direct link can be made between seismograms and the internal state of volcanic activity.

1.2. Conduit flow models

Several trigger mechanisms for low frequency events have been proposed, including magma–water interaction (Zimanowski, 1998), stick–slip motion of magma plugs in the conduit with shear fracture (Goto, 1999), magma flow instabilities (Julian, 1994), and periodic release of gas-ash mixtures into open cracks (Molina et al., 2004). Two studies at Soufrière Hills volcano on Montserrat were based on the correlation of seismicity with tilt cycles observed in 1997: Denlinger and Hoblitt (1999) used a conceptual model of Newtonian flow of compressible magma through a conduit combined with a stick–slip condition at the conduit wall, in analogy to the behaviour of industrial polymers. For a magma supply at a constant rate they obtained a cyclic behaviour of flow and pressure through a hysteresis in flow resistance controlled by the slip conditions at the conduit wall. Based on the same data set from Montserrat, Wylie et al. (1999) explained the instability of magma flow through volatile-dependent viscosity changes. They obtained a cyclic flow behaviour as well, but ignored the simultaneous occurrence of seismic swarms completely. More recent models by Melnik and Sparks (2002) are dedicated to unsteady magma flow in explosive eruptions after unloading through dome collapses. The behaviour of real-time seismic amplitude measurements (RSAM) has been mentioned in the same context but not used in any way to constrain the flow models (Melnik and Sparks, 2002).

New field evidence for brittle failure of magma was found by Tuffen et al. (2003) and fuelled new interest in studies of a wide variety of phenomena including flow banding, magma degassing through cracks, and also triggering of low frequency events. Tuffen et al. (2003) suggested, for the generation of low-frequency earthquakes, a mechanism where stress accumulation in a viscous magma leads to the formation of shear fractures through which a certain amount of gas and ash particles can escape. Subsequent healing of these fractures results in a repeated stress build-up with minimum repeat times in the order of a few tens of seconds. A study by Gonnermann and Manga (2003) focussed on magma rupture once a critical stress value of 10^8 Pa had been reached. The emphasis of that study was to demonstrate that magma fragmentation does not inevitably lead to explosive volcanism. Here the simultaneous generation of volcanic earthquakes is mentioned, but not incorporated, in the numerical modelling attempts.

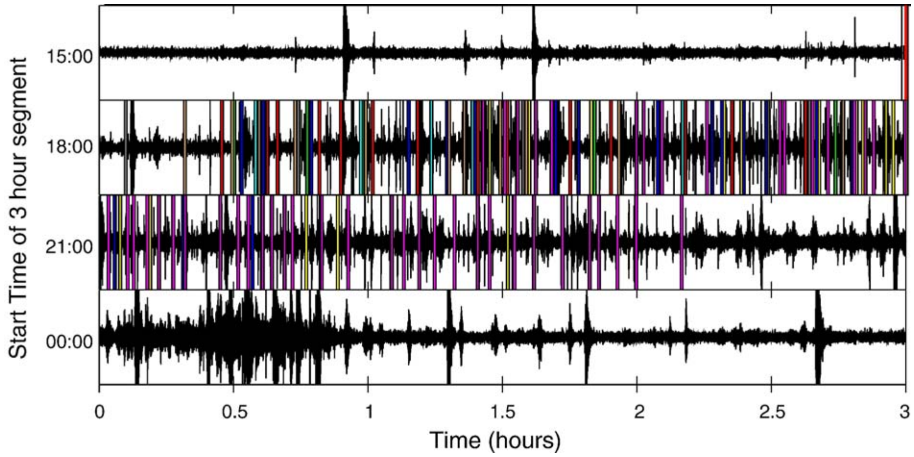

Fig. 1. Twelve hours of a low-frequency earthquake swarm on May 19 and 20, 1997, recorded on station MBGA in Montserrat in 4 consecutive time windows. Each vertical bar identifies a single event and the colour represents the family, which matches the colours in Fig. 3a.

1.3. Overview

Following from the investigations summarised above, we focus our study on a seismic trigger mechanism for low-frequency volcanic earthquakes. This is based on a combination of (i) seismological constraints from Montserrat, (ii) field evidence of examples of brittle failure in magmatic dykes and conduits, and (iii) modelling of conduit flow conditions leading to brittle failure of rising magma in the glass transition.

In the next section we describe the characteristics of low-frequency earthquakes on Montserrat and explain the seismological clues one can derive from those observations. In Section 3 we report the field observations which form together with the seismological arguments the framework of constraints upon which our trigger model is based. Section 4 introduces the numerical flow model we have employed and the two conditions necessary for the generation of low-frequency earthquakes: brittle failure in magma and the existence of a low viscosity fluid nearby to account for the seismic resonance. In Section 5 we present our trigger model and put it in a wider context of cyclic episodes of seismic swarm activity and dome growth. Finally, in Section 6 we compare our new model to existing models before concluding.

2. The seismological clues

2.1. Characteristics of low-frequency seismic events on Montserrat

In the following list we summarise some of the characteristics of low-frequency events on Montserrat, as well as some implications. A more detailed description can be found in Miller et al. (1998) or Neuberg et al. (1998).

- Low-frequency seismic events on Montserrat have a spectral contents of 0.2–10 Hz and a duration ranging from 10 to 30 s. They have a weak *P*-wave onset and the long-period coda shows elliptical particle motion.
- In general, low-frequency events occur in swarms lasting between a few hours and several days. Almost all of the large dome collapses were preceded by swarms of low-frequency events with identical waveforms, often occurring in cycles of 8–12 h for several weeks. This is indicative of a re-chargeable source mechanism, tapping into a limited energy reservoir.
- Throughout the eruption many episodes of volcanic tremor have been recorded at Montserrat. These usually follow swarms of low-frequency earthquakes with individual events becoming closer and closer together in time until they merge into tremor, demonstrating the close relationship between low-frequency events and this particular type of volcanic tremor.
- The occurrence of low-frequency swarms correlates with tilt measurements made on the flank of the volcano, which indicates that the dome inflates during a swarm and deflates when it finishes—this last change is often associated with dome collapse and pyroclastic flows (Voight et al., 1998). This strongly supports the hypothesis that low-frequency events are linked to a pressurisation process.
- The highly regular occurrence of seismic events has a spectral signature of integer harmonic spectral lines if the time window from which the spectral slice

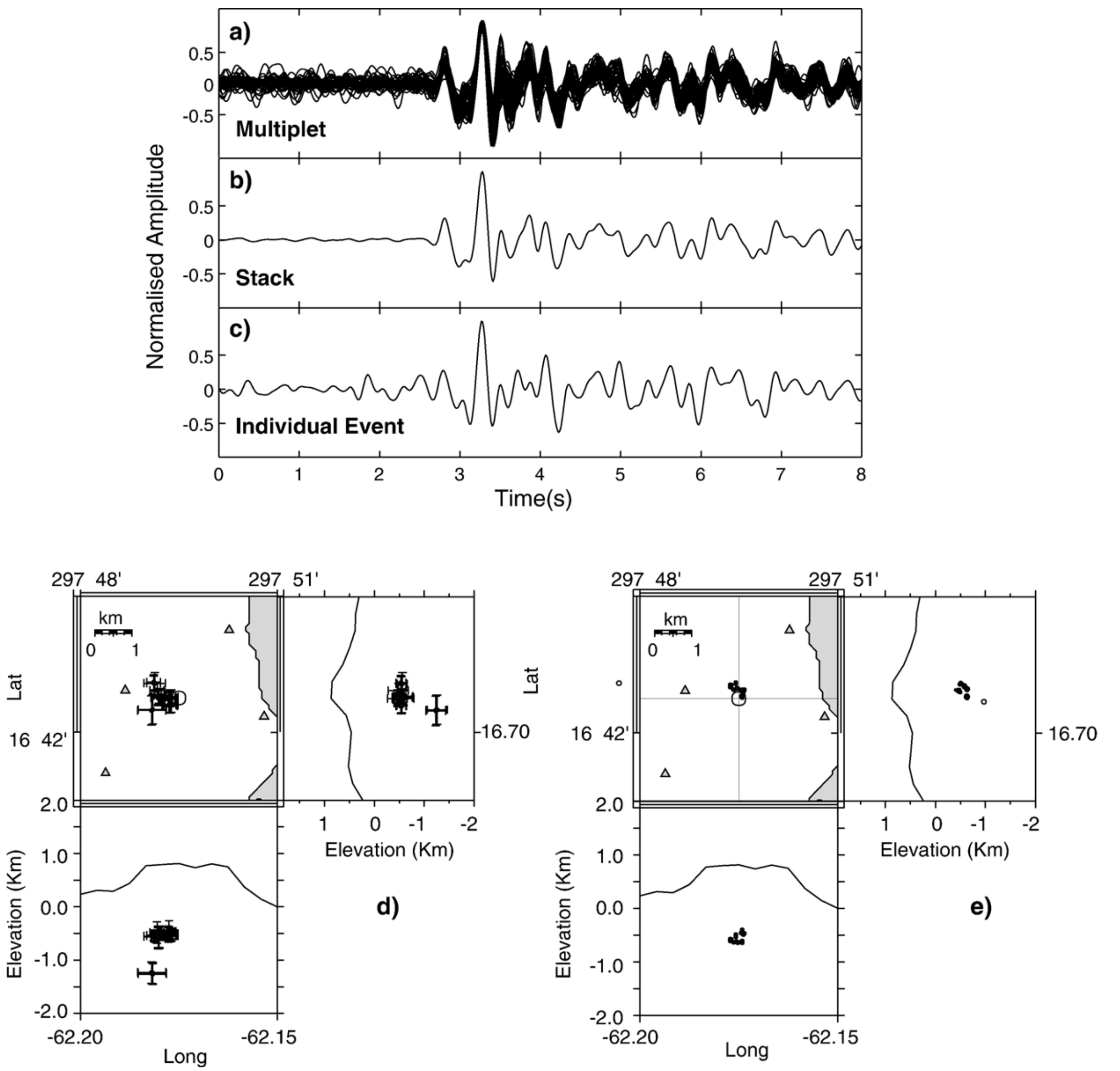

Fig. 2. Event family recorded in June 1997 on station MBGA in Montserrat; (a) 19 individual events, (b) stacked trace after cross correlation and aligning of individual events, and (c) individual event. Note the difficulty which would be encountered in isolating the correct *P*-phase from this individual event. (d) Hypocentre location of a series of 55 events using HYPOELLIPSE and a 3-layer velocity model; (e) hypocentral estimates for the same series of 55 events using VELEST where simultaneously a 3-layer velocity model is estimated. The average rms residual of the events is 0.013 s or one sample at 75 Hz, translating to ± 40 m. The source locations cluster within a constant velocity zone and are approximately 500 m from any velocity boundary.

computed is wide enough to cover several events. Shifting dominant frequencies which have been observed on many other volcanoes indicate the change of magma properties within a few tens of seconds (Powell and Neuberg, 2003).

- For most swarms the individual low-frequency earthquakes are very similar to each other in both amplitude and waveform, having cross-correlation coefficients as high as 0.9 ± 0.1. During a swarm, events can occur at very regular intervals, with the time between consecutive events varying by less than 2% for several hours on end (White et al., 1998). This points to a stable, non-destructive source mechanism.

2.2. Identifications of families in earthquake swarms

When looking at a low-frequency earthquake swarm as depicted in Fig. 1 it is not always obvious that many single low-frequency events show some similarity. By

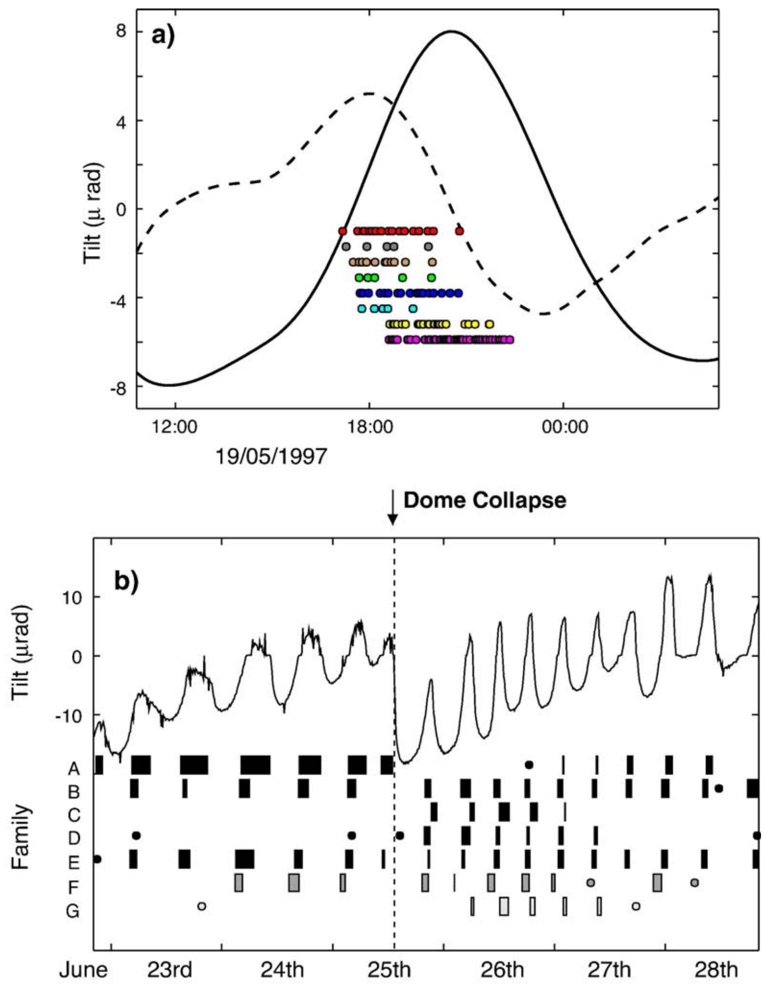

Fig. 3. (a) Families as shown in Fig. 1 together with tilt (solid line) and its time derivative (dashed line), both low-pass filtered. (b) Waveform families compared to the tilt record over the 6 days between June 23 and June 28, 1997. Note that different families represent different source locations: the oblongs represent the families, with different shades and patterns distinguishing between the different waveform groups A to G. The start of the oblong is positioned where the first event of that family occurs in that swarm and the end coincides with the last event. The dome collapse occurred at approximately 12:45 on June 25, 1997.

using cross-correlation methods, however, we found that up to 70% of all events fall into one of several different 'families' of similar waveforms (Green and Neuberg, 2006-this issue). The similarity has been defined through a correlation coefficient of 0.70 and higher and the entire waveform seen in a time window of 8 s has been used to compute the cross-correlation. Each family represents a set of events which have been produced within a small source volume from which the energy has been radiated along an identical path to the receiver. The remaining 30% of low-frequency events do not show any similarity and must be interpreted as signals originating from different non-stationary sources. Families can be consistent regarding their waveform over several days, i.e. the source location and mechanism is stable and can be reactivated again and again.

2.3. Quantifying the similarity in waveform and source location

We have performed numerical tests employing a finite difference method similar to that used in Neuberg (2000) where we use a simple Küepper wavelet with the dominant frequency of 1 Hz to kick start a conduit resonance. This trigger wavelet represents an explosion inside the fluid-filled conduit which is embedded in a solid. By varying the location of the trigger and cross-correlating the resulting synthetic signals we get an idea of the range across which the position of the source location can vary before similarity breaks down. Assuming an average acoustic velocity inside the conduit of 1200 m s^{-1}, we determine this range to be less than 120 m. This is significantly smaller than predicted by the quarter-wavelength hypothesis applicable for general seismic body waves (Geller and Mueller, 1980).

The onset of low-frequency events is generally very weak and classic location techniques using the *P* and *S*-phases do not work well. However, a much better absolute earthquake location can be achieved by using the following steps: (i) stack all members of one family after aligning the signals according to maximum cross-correlation, (ii) find the *P*-wave onset for the stacked 'master event', (iii) use this *P*-wave pick for individual events, and (iv) repeat the procedure for each station and use a classic *P*-wave travel time method to find the source location. This method uses the fact that a trigger signal at a fixed source location produces exactly the same resonance, which controls the long-period coda of the low-frequency events. This coda, in turn, dominates the cross-correlation. Fig. 2 shows the results for two different location methods: (i) using the standard Montserrat velocity model, and (ii) by determining the velocity model at the same time. The locations of all families fall into the narrow range of 500 ± 100 m below sea level, i.e. approximately 1500 m below the active dome surface. Each family forms a narrow cluster without showing any significant structure. These results are in good accordance with Rowe et al. (2004).

2.4. Correlation between earthquake swarms and tilt cycles

Voight et al. (1998) pointed out that a radial tilt signal recorded in 1997 on Chances Peak, an old dome adjacent to the active lava dome, correlated with the seismicity of Soufrière Hills volcano. A much more detailed picture emerges when only the low-frequency earthquake families are plotted together with the tilt signal and its time derivative (Fig. 3). Seismic activity commences when the tilt goes through a turning point, or its derivative reaches a maximum. Furthermore, low-frequency seismicity ceases when the tilt goes through a second turning point, i.e. its derivative reaches a minimum. Tilt, the rigid rotation of the volcano's flank, is a manifestation of pressure changes reaching the shallow (<1000 m) part of the plumbing system. We observe a very good match between the switching on and off of seismicity and the turning points in the tilt signal, which points to magma movement at depth: as soon as pressurisation has reached a certain threshold, magma starts to move, depressurising the system, and causing the tilt behaviour to turn. As soon as the magma has stopped the system begins to build up pressure due to the diffusive growth of gas bubbles.

The seismic sources are extremely stable and do not evolve significantly through several tilt cycles, or even through a major dome collapse (Fig. 3b). This places constraints on the location and mechanism for low-frequency seismicity. In particular, the repeatability of the patterns of waveform families between swarms indicates that the mechanism for shallow pressurisation is extremely well ordered with different seismic sources being reactivated in the same manner for each deformation cycle. After the dome collapse pressure conditions in the conduit have changed, and the sources most active prior to the collapse pause while a subset of families, previously showing minor activity, now dominates the generation of seismic energy. After a few cycles the system has recovered and the former source locations are again active. This observation will be essential for our trigger model described below.

2.5. Summary of seismological constraints

A low-frequency seismic signal on Montserrat comprises a *trigger* part at the high frequency onset and a *resonance* section forming the low-frequency coda. In events with high amplitudes a clear *P*-wave can be identified at the onset, which is followed immediately by interface waves with elliptical particle motions (e.g. Neuberg et al., 2000; Jousset et al., 2003). These events often occur in swarms comprising a limited number of *families* with distinct waveforms. Events with identical waveforms originate from identical source locations. Therefore, the existence of event families demonstrates that there is a number of fixed sources at about 500 m depth below sea level (1500 m below the dome) with a repetitive mechanism. In the case of single events merging into tremor the repeat time is as short as a fraction of a second. The correlation of low-frequency seismicity with inflation and deflation cycles reveals the role of low-frequency earthquakes as an indicator of magma movement and depressurisation. Furthermore, if pressure conditions change, the waveform and, hence, the seismic source locations change as well (Fig. 3b). This indicates a pressure dependent parameter window in which the trigger mechanism can act.

3. Field evidence for brittle failure in magma

Field evidence for a potential trigger mechanism for low-frequency earthquakes was recently discovered in dissected rhyolitic conduits at Torfajökull in Iceland (Tuffen et al., 2003; Tuffen and Dingwell, 2004). A summary of the new evidence, its implications for low-frequency trigger mechanisms, and its applicability to

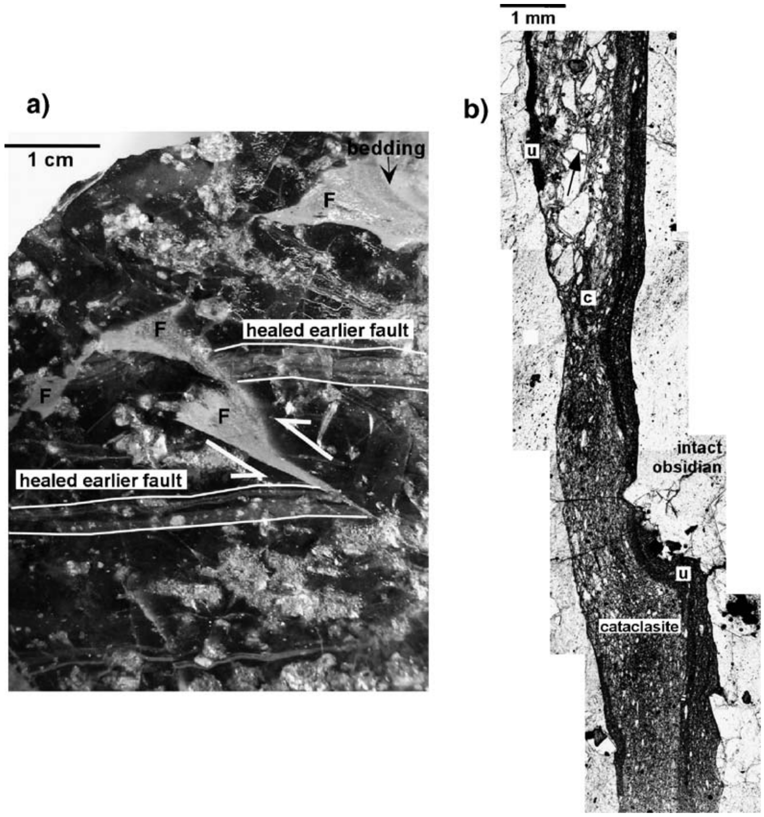

Fig. 4. (a) Part of a network of irregular fractures in obsidian at Skridugil Conduit, Raudufossafjöll, Iceland. The fault network (f) is filled with pale grey fragmented material, which locally exhibits bedding (indicated). An earlier healed fault zone has an apparent offset of 2 cm where cut by a fracture. Modified from Tuffen and Dingwell (2004). (b) Photomicrograph of a fault zone within obsidian. The zone is between 1 and 2 mm thick and filled with cataclasite (pale grey, labelled c) and ultracataclasite (dark grey, labelled u), which have the same composition as the surrounding obsidian (o). The arrow indicates where an obsidian clast within the cataclasite (near-white) is cut by a small fracture, indicating in-situ grain size reduction. Modified from Tuffen and Dingwell (2004).

Montserrat is given in this section. The dyke-like conduits, which fed small-volume lava flows, have widths of 10–20 m and are dissected to a depth of about 40 m. Complex brittle–ductile textures in the low-vesicularity conduit-filling obsidian record repeated fracture and healing of magma during ascent.

3.1. Fracture and faulting of highly viscous magma in conduits

In the Torfajökull case, networks of angular fractures, which have typical lengths of 0.1–2 m and apertures of 1–80 mm, cut the obsidian close to the conduit walls. They consist of shear fractures with displacements of several millimetres to centimetres, smaller tensile fractures which propagate from the walls of shear fractures, and irregular voids. The fractures are filled with fine-grained fragments of obsidian and broken phenocrysts with an identical composition to the surrounding obsidian. These fragments are formed by abrasion of fracture surfaces. There are complex sedimentary structures in the fragmented material, including cross-bedding and erosional surfaces (Fig. 4a). Beds range from 1 to 5 mm in thickness and are poorly to well sorted. Locally, beds can be traced from shear fractures into tensile fractures and voids, where their thickness greatly increases. Sedimentary structures in the fragmented material indicate that it was transported through the fracture systems by gas released from vesicles and pore space intersected by fractures. As argued in Tuffen et al. (2003), the seismic energy released during formation of such fractures is consistent with recorded low-frequency earthquakes at Montserrat and elsewhere. The occasional occurrence of pumice clasts in fractures cutting low-vesicularity magma of the conduit walls indicates lateral transport of gas and ash from the conduit interior. Similar bedded material is found in tuffisite veins that cut the country rock and infilling lava of conduits elsewhere (e.g. Stasiuk et al., 1996), and demonstrates that magma fracture can play an important role in degassing within ascending magma (Jaupart, 1998; Gonnermann and Manga, 2003).

As described in Tuffen and Dingwell (2004), fracture networks develop through coalescence and rotation into near-planar fault zones during subsequent upward flow of magma. Fault zones are near-vertical, up to 5 m in length, 1–50 mm thick, and have maximum measured shear displacements of 13 cm. The fragmented material within fault zones resembles that in the initial fracture networks, but has been modified by repeated friction-controlled slip events with the grinding of grains in localised slip bands. Textural associations closely resemble seismogenic tectonic faults as described by Chester and Chester (1998) and others. It is therefore likely that movement of each fault zone generated a number of similar seismic events (Tuffen and Dingwell, 2004).

Some fault zones have undergone an additional phase of healing and distributed ductile deformation, in which the fragmented material is thoroughly welded and eventually forms flow bands in the obsidian (Tuffen et al., 2003; Tuffen and Dingwell, 2004). Healing, which is mediated by frictional heating and/or reduced strain rate, leads to aseismic viscous deformation. In this way, seismogenic fractures and faults in magma may develop and heal during conduit flow.

4. Modelling of magma flow in brittle conditions

The fracture networks are thought to be formed by shear failure of high viscosity magma in the glass transition (Tuffen et al., 2003). When the product of shear strain rate and shear viscosity exceeds a critical value (approximately 10^7 Pa), deformation becomes unrelaxed and shear stress accumulation eventually leads to failure (Dingwell and Webb, 1989). In this section we consider a combination of high melt viscosity and strain rate that is necessary for failure to be achieved during conduit flow of silicic magma.

4.1. The viscosity problem

Low-frequency earthquakes consist of an onset which sometimes has high-frequency spectral components, and a low-frequency resonance coda produced by seismic energy that has been efficiently trapped in a fluid-filled part of the volcanic plumbing system. This poses a problem, which in turn will provide an important constraint for the following magma flow modelling attempts: if brittle failure of magma provides the trigger mechanism that sets off conduit resonance, then the *melt* must be viscous enough to allow the critical stress value to be exceeded, on the other hand the *magma* viscosity must be low enough to support seismic conduit resonance with low damping. This problem can be solved by considering high viscosity gradients across the conduit controlled by high gradients in temperature and gas content. Another option to reach the critical shear stress while maintaining a low viscosity is to consider high flow velocity gradients or strain rates as produced by e.g. geometry changes in the plumbing system with depth.

4.2. The critical shear stress

In this section we employ a finite element method to compute 2D magma flow in a dyke or conduit leading to rupture conditions. The aim is to study the interaction of governing parameters and, particularly, to identify locations in the conduit or dyke where the critical shear stress is exceeded resulting in brittle failure of the magma and, consequently, the triggering of low-frequency seismic signals. Similar to Gonnermann and Manga (2003) we solve the compressible Navier-Stokes equation for a mixture of gas, melt and crystals ascending at the same rate, however, we neglect crystal growth during ascent. Dimensions of the conduit/dyke are 30 m × 5000 m and the magma chamber depth is at 5000 m (Barclay et al., 1998), the pressure in the magma chamber is given by the lithostatic country rock pressure plus an excess pressure, P_{ex}. We include water solubility using Henry's Solubility Law (Burnham, 1975), and assume an ideal gas and constant bubble number density. As explained below (Section 5.1) the critical shear stress is reached at 830 m depth. Therefore, we include horizontal gas loss through permeable magma and through the conduit wall (Jaupart and Allègre, 1991; Jaupart, 1998), once a plug has formed at 830 m. We use a constant magma supply which is controlled by the boundary conditions for the excess pressure, P_{ex}, and no-slip boundary conditions at the conduit wall. However, at 830 m depth we change the boundary conditions at the conduit walls to friction-controlled slip, simulating the transition of the flow from a ductile to a friction-controlled mechanism. Boundary conditions for velocity and pressure at the exit are $u=0$ and $P=1\times105$ Pa (atmospheric pressure), respectively. The viscosity of the melt is given by Hess and Dingwell (1996), the volume viscosity of the magma is adopted from Prud'Homme and Bird

(1978), and the crystal effect on viscosity by Lejeune and Richet (1995):

$$\eta_{m+x} = \eta_m \left(1 - \frac{\phi_{\mathrm{cryst}}}{0.6}\right)^{-2.5} \tag{1}$$

where η_{m+x} is the shear viscosity of the crystal-melt mixture, η_m the shear viscosity of the melt and ϕ_{cryst} is the volume fraction of crystals (Table 1). We consider that bubbles alter the shear viscosity of the magma depending on capillarity number C_a (Llewellin et al., 2002). Using the 'minimum variation' given in Llewellin and Manga (2005) for $C_a<1$:

$$\eta = (1-\phi)^{-1} \tag{2}$$

For $C_a > 1$:

$$\eta = (1-\phi)^{5/3} \tag{3}$$

where η is the relative viscosity of the magma given by the ratio of the viscosity of the magma to η_{m+x}. ϕ is the gas volume fraction within the magma which increases with decreasing depth in the conduit due to bubble growth.

In order to produce high viscosity gradients (see above) we include a simulation of heat flux through the conduit walls into the surrounding country rock. Magma has a low thermal conductivity and therefore a thin thermal boundary layer develops where the temperature of the magma drops rapidly from the input temperature, T, (Table 1) to the temperature at the conduit wall, T_{wall} (Stasiuk et al., 1993). We adopt a critical shear stress, $\sigma_{\mathrm{melt}}=10^7$ Pa from Tuffen and Dingwell (2004) and define the criterion for brittle failure as

$$\frac{\eta_{\mathrm{wall}}\dot{\varepsilon}}{\sigma_{\mathrm{melt}}} > 1. \tag{4}$$

Here η_{wall} is the shear viscosity at the wall due to the reduced temperature and $\dot{\varepsilon}$ is the shear strain rate. All parameters used for the flow computation are listed in Table 1.

Fig. 5a shows the magma flow velocity in the centre of the conduit (solid line) and at the conduit wall (dotted line). From 5000 m to approximately 3500 m depth the flow maintains a parabolic structure similar to Poiseuille flow, due to the incompressibility of the magma below the bubble nucleation depth. Above nucleation the increasing gas volume fraction causes the magma to accelerate increasing the lateral velocity gradient due to the condition of no slip at the conduit wall. The shear strain rate of the magma is given by this lateral velocity gradient,

$$\dot{\varepsilon} = \frac{\mathrm{d}v}{\mathrm{d}x} \tag{5}$$

Due to the change in boundary conditions at 830 m the magma flow at the wall is now friction controlled and accelerates with increasing gas volume fraction in the magma. Due to mass conservation the centre velocity drops to a lower value resulting in a plug

Table 1
Parameters and constants used in the runs unless otherwise stated. Values are chosen to represent magma from Soufrière Hills Volcano, Montserrat, WI

Parameter	Value	Reference
Density of the melt, ρ_{melt}	2400 kg m^{-3}	Rivers and Carmichael (1987)
Volume % of crystals, ϕ_{cryst}	30%	Devine et al. (1998)
Density of the crystals (An_{53}), ρ_{cryst}	2680 kg m^{-3}	Carmichael (1990)
Density of country rock, ρ_{rock}	2300 kg m^{-3}	–
Ideal gas constant, R	8.3145 J mol^{-1} K^{-1}	general constant
Molecular weight of H_2O, M	0.018 kg mol^{-1}	general constant
Temperature of magma, T	1090 K	Barclay et al. (1998)
Temperature of melt at wall, T_{wall}	940 K	–
Total water content of H_2O, C_o	4.8 wt.%	Barclay et al. (1998)
Surface tension, Γ	0.05 Nm^{-1}	Lyakhovsky et al. (1996)
Bubble number density, n	1×10^{12} m^{-3}	Navon et al. (1998)
Solubility constant, K_h	4.11×10^{-6}	Burnham (1975)
Tensile strength of glass/melt, σ_{melt}	10 MPa	Tuffen and Dingwell (2004)
Slip resistance at conduit wall, β	5×106 Pa s m^{-1}	–
Excess chamber pressure, P_{ex}	20 MPa	–
Permeability for gas loss, K	5×10^{-11} m^2	–

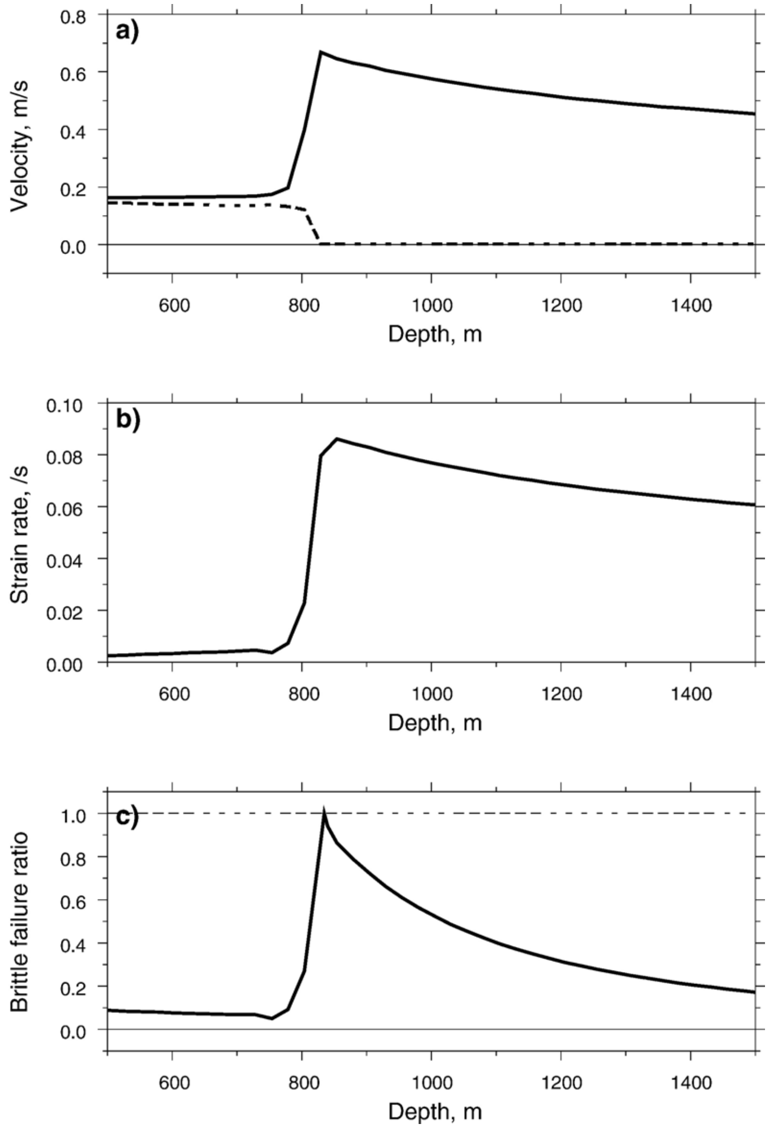

Fig. 5. (a) Variation of velocity with depth, solid-line is the velocity in conduit centre and the dashed line is the velocity at the conduit wall. (b) Variation of strain rate with depth 0.5 m from conduit wall. (c) Brittle failure ratio (Eq. (4)) 0.5 m from conduit wall plotted between the depths of 500–1500 m, brittle failure where ratio exceeds 1.

flow above 830 m depth. The brittle failure ratio (see Eq. (4)) 0.5 m from the conduit wall is plotted in Fig. 5c and its cross-section from conduit centre to wall at this depth in Fig. 6.

5. The model: Fracture of magma as a seismic trigger

5.1. The conceptual model

We propose a conceptual model for the generation of low frequency seismic events which is based on rupture of magma in the glass transition where the shear stress exceeds a critical value as in Goto (1999). This conceptual model is summarised in Fig. 7. Through magma flow modelling we explored the parameter space in which melt viscosity and strain rate determine where these conditions occur. As depicted in Fig. 6, the *brittle failure* ratio increases with the ascending magma and reaches the critical value of 1 near the conduit wall at a depth of 830 m. At this depth we change boundary conditions from no-slip to friction-controlled slip to enable the magma to ascend further by plug flow. In such a way, the trigger position indicated by the model remains at the same depth, as suggested by the seismic observations.

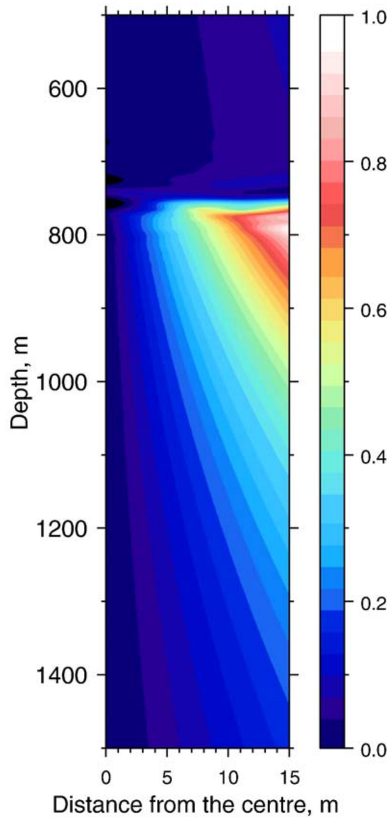

Fig. 6. Brittle failure ration as defined in Eq. (4) between conduit centre and conduit wall. Note that failure occurs near the conduit wall where ratio exceeds 1; magma ascends above this depth as a plug under friction-controlled conditions.

However, the depth at which critical stress conditions are reached does not match the depth determined seismically in Section 2. Left unconstrained, the depth at which the critical conditions for magma rupture occur can vary within a certain range due to trade-offs between several model parameters. Using the locations of seismic sources as a constraint we can choose the parameters in the flow models accordingly and force the depth to 1500 m.

This could be easily achieved by lowering the critical shear stress value to 10^6 Pa. Here further experimental results are necessary to test if the inclusion of gas bubbles and crystals could lower the critical shear stress to such an extent (Spieler et al., 2004). Alternatively, deviations in conduit geometry such as 'bottle necks' or the transition from a dyke to a cylindrical geometry could easily lead to an increase in magma flow rate, pushing the strain rate over the critical edge. An increase in magma flow rate to a metre per second rather than centimetres per second would be necessary. However, since certain magma processes such as crystallisation during magma ascent and vertical degassing are not yet part of the flow model, we refrain from forcing the model to match the seismically obtained depth and leave the parameters in an average range as listed in Table 1. Hence, we suggest a conceptual rather than quantitative model which, nevertheless, can explain the trigger mechanism for low-frequency events.

An additional and even stronger argument for the pressure- and, therefore, depth-dependence of a critical trigger region can be derived from the change in source locations of low-frequency earthquakes after a dome collapse as depicted in Fig. 3b. Right after the dome collapse and accompanying depressurisation a different subset of families (from different source locations) provide the observed seismic energy as described in Section 2.4. After several cycles the pressure has reached its former level and the same sources are active as before the collapse.

5.2. Trigger and resonance

So far we have concentrated on the actual trigger mechanism that sets off any elastic wave. In order to produce a low-frequency earthquake in a volcanic edifice a resonator needs to capture the seismic energy and trap it efficiently through a high impedance contrast between resonating container and surrounding country rock. It is not necessary for the trigger and resonance to be at the same location. Kumagai and Chouet (1999)

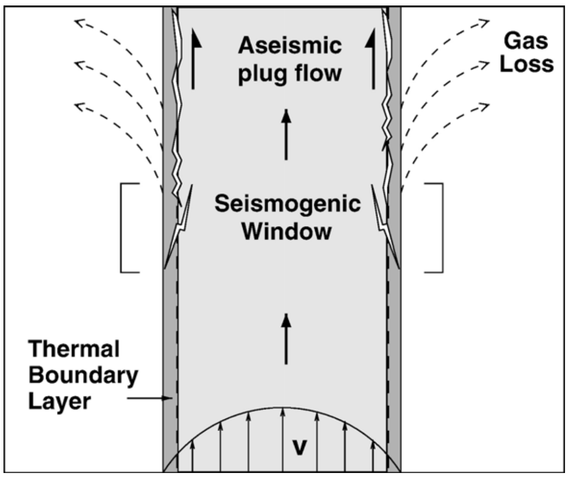

Fig. 7. A schematic cartoon illustrating the formation of shear fractures and their subsequent evolution into faults within a seismogenic window. Once formed the fractures move up due to the ascent of magma allowing friction-controlled plug flow to form. The opening of the fractures allows increased gas loss to the surroundings. The transition from seismogenic slip to aseismic plug flow may be due to frictional heating, as discussed in Section 3.1.

and Molina et al. (2004), suggest an ash-laden gas emission into an open crack, which acts then as a resonator. In our modelling framework we suggest a magma-filled dyke or conduit in close vicinity of the brittle fracture such that seismic energy released by the trigger mechanism can propagate to the resonator where it will be trapped through the high impedance contrast. A critical issue here is the damping which the seismic waves encounter traveling from the trigger location to the resonator, and then within the resonator. Collier et al. (2006-this issue) investigate the seismic damping issue in magma, particularly the impact of gas bubbles on damping in different frequency ranges. Their study takes diffusion and gas mass flux in and out of bubbles into account and demonstrates that the bubble nucleation level where diffusion is very effective acts as an efficient damping body. The results of this study suggest that the resonating parts (dykes or conduits) of the plumbing system on Montserrat are between the bubble nucleation level (at approximately 3500 m) and trigger level (at 1500 m) where conditions are such to allow conduit resonance without strong damping.

5.3. The model in a wider context

Low-frequency events very often occur in swarms with variable duration (e.g. Miller et al., 1998; White et al., 1998) and sometimes exhibit a regular excitation pattern which results in the observation of harmonic spectral lines (Powell and Neuberg, 2003). Seen in a wider context, our trigger model fits those earlier studies where the periodicity of earthquake swarms (and tilt cycles) is linked to the diffusion processes that dictate the supply rate of magma from a deeper reservoir once a magma plug has started to move and local pressure has begun to drop (e.g. Melnik and Sparks, 1999). Hence, the occurrence of low-frequency earthquake swarms indicates magma movement at depth which results in local depressurisation of the volcanic system below that depth.

The occurrence of harmonic spectral lines have been associated with a potential feed-back mechanism to explain the precise periodicity of the signal (Neuberg, 2000). The resonant part of a low-frequency earthquake forms spatially periodic pressure variations within the magma column. If this pressure variation is superimposed onto the local shear stress that is caused by viscous magma flow, the critical level for magma rupture is reached and another low-frequency earthquake is triggered. This earthquake results in further excitation of the conduit resonance and a feed-back system can be established until magma supply lags behind.

6. Discussion and conclusions

The field evidence for flow of gas/ash through a crack is also consistent with aspects of the crack resonance model, as developed and discussed by Chouet (1988), Kumagai and Chouet (1999), and Molina et al. (2004). Indeed, correlation between ash emission from domes and shallow seismic events at Santa Maria and Galeras lava domes indicates that release of gas and ash may be related to the seismic trigger mechanism. One problem is that the crack sizes required to produce the frequency content of measured low-frequency events are up to 100 times larger than the fracture networks observed by Tuffen et al. (2003). If earthquakes are triggered in conduits, the existence of cracks extending hundreds of metres into the country rock is geologically unfeasible. Despite this problem, Molina et al. (2004) suggest that changing amplitude decay of LP events during a swarm at Tungurahua, Ecuador can be produced by resonance of a crack that is gradually filling with ash. Per Collier et al. (2006-this issue) we suggest that a change in the damping properties of gas-charged magma could easily be achieved by slight pressure changes resulting in subsequent changes of gas-volume fraction.

Possible criticism of linking a trigger model for Soufrière Hills Volcano to field observations from an ancient rhyolitic conduit in Iceland could stem from uncertainty about whether similar fracture and faulting processes can occur in more crystal-rich or vesicular magma. However, two lines of evidence suggest that the model is indeed appropriate. Firstly, the conditions for shear fracture of magma are likely to be met during eruptions of crystalline andesite and dacite, as well as crystal-poor rhyolite (Goto, 1999; Gonnermann and Manga, 2003; Tuffen et al., 2003). Secondly, there is textural evidence for mixed brittleductile behaviour (i.e. fracture and healing) within many (recently) active lava domes, including Soufrière Hills Volcano, Unzen and Lascar (e.g. Sparks, 1997; Smith et al., 2001). It is therefore likely that shear fracture and faulting of magma play an important role in triggering seismicity and allowing gas escape at Montserrat and elsewhere.

In conclusion, we base the proposed trigger model on the following pieces of evidence:

- Low-frequency earthquake swarms occur in groups with similar waveforms indicating a stationary source location over a period of several days, as well as a repetitive source mechanism.
- These low-frequency seismic sources are located at a narrow depth range of 500 ± 100 m below sea level

or approximately 1500 ± 100 m below the active lava dome.

- Magma flow modelling predicts a critical shear stress to be reached at 830 m where magma parameters are chosen in an average range that reflects realistic conditions for Soufrière Hills volcano on Montserrat.
- To increase the modelled critical depth of 830 m to match the seismically determined depth of 1500 m, a shear stress as low as 10^6 Pa would be required.
- Field studies from Iceland provide evidence of shear fracture and faulting of magma.

The characteristics of the trigger model for low-frequency events can be summarised as follows:

- Conditions under which magma ruptures can be met at a depth where the product of melt viscosity and shear strain rate exceeds a critical value. This value is experimentally determined for pure glass to be 10^7–10^8 Pa.
- Shear cracks open at this depth and are taken upwards by the ascending magma. At this stage magma flow switches from ductile to a friction controlled regime. These cracks of the dimension of a few metres allow increased degassing and clog up by the deposition of ash.
- While magma ascends, the source depth remains constant where conditions for magma rupture are met. This provides repeatability of the source mechanism shorter than the time span needed for healing of the cracks.
- The seismic energy released is trapped in the part of the conduit where magma has a high impedance contrast to the surrounding country rock. The viscosity in this part must be low enough to prevent attenuation. This leads to conduit resonance providing the low-frequency character of the seismic signal.

Future work will focus on the amplitudes and seismic moments of low-frequency events in order to quantify magma motion at depth. Further magma flow models will help to explore the model space of magma parameters to gain a better insight into the physical processes related to degassing and viscosity changes at depth. Experimental work on glass with vesicles and crystals is required to determine the conditions under which gas-charged magma will fracture.

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