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Slow rock fracture as eruption precursor at Soufriere Hills volcano, Montserrat

Christopher R.J. Kilburn

Benfield Greig Hazard Research Centre, Department of Geological Sciences, University College London, London, U.K.

Barry Voight

Department of Geosciences, Pennsylvania State University, Pennsylvania

Abstract. Breakout of magmatic activity at Soufriere Hills volcano, Montserrat, was preceded by a tenfold increase in rate of earthquake occurrence. A new model of subcritical rock failure shows that this increase is consistent with the growth, possibly episodic, of the magma conduit at a rate controlled by progressive weakening of the host country rock. The preferred weakening mechanism is stress corrosion, by which circulating juvenile and hydrothermal fluids chemically attack the country rock and promote failure at stresses smaller than the rock's theoretical strength. The results illuminate the potential for slow-cracking models to enhance eruption forecasts using the inverse-rate technique combined with traditional monitoring methods.

Introduction

Following more than three years of elevated seismicity and four months of strong phreatic eruptions at the Soufriere Hills volcano [*Young et al.*, 1997], a tenfold increase in the daily rates of earthquakes immediately preceded the breakout of andesite lava in November 1995 (Figure 1). This increase was accompanied by accelerating rates of ground deformation [*Jackson et al.*, this issue] and was clearly related to propagation of a magmatic conduit through the volcanic edifice. Conduit propagation results from some combination of an increase in magmatically-induced stress and a weakening of host rock. Since magmatic stresses are often assumed to be the dominant factor at active volcanoes, comparatively little attention has been directed toward the influence of rock weakening [*Voight*, 1988; 1989]. The computed acceleration is here used instead to highlight the importance of slow rock fracture as a mechanism for limiting the rate at which an eruption is approached. We review pertinent aspects of time-dependent subcritical rock fracture, consider its occurrence in volcanic environments and, applied to Soufriere Hills volcano, evaluate its potential for forecasting eruptions.

Subcritical rock failure

Rocks can weaken and crack at stresses much smaller than their theoretical strengths [*Atkinson*, 1984; *Meredith et al.*, 1990]. Under conventional geologic conditions, the dominant weakening mechanism is considered to be stress corrosion, i.e. stress-enhanced chemical reaction [*Anderson and* Grew, 1977; *Atkinson*, 1984]. Circulating fluids, notably water which can readily corrode silicate materials [*Atkinson*, 1984], attack molecular bonds at the tips of existing flaws, promoting crack growth and reducing the bulk strength of a rock. (Mineral precipitation from fluids can also heal cracks and strengthen rock, but normally at timescales greater than those for stress-induced corrosion to take place.)

Cracks propagate by breaking a rock's molecular bonds. As the stretched bonds relax, they release elastic strain energy and this becomes available to drive further growth or to open new cracks [*Marder and Fineberg*, 1996]. The new and growing cracks increase the bulk volume of rock under stress and, eventually, they coalesce to form a major fracture [*Lockner et al.*, 1991; *Main and Meredith*, 1991].

As collections of cracks grow towards each other, their local stress fields may interact to inhibit coalescence until the number-density of cracks exceeds a critical value [*Main et al.*, 1993]. The bulk fracture resistance of the host rock may therefore increase temporarily until coalescence begins. Should this occur among small cracks ahead of a main fracture (in the so-called "damage zone"), then the main fracture may show intermittent rates of advance. The average growth rate of the main fracture, however, is expected to increase with time because the net rate of elastic-strain-energy release varies with the increase in total crack volume, while the net rate of energy loss depends on the increase in total crack area [*Griffith*, 1920]. Thus the mean growth rate accelerates until the deforming bulk stress is relaxed, or until the fracture is halted by stronger rock.

Fracturing at volcanoes

Conditions within and below volcanic edifices are particularly conducive to stress corrosion: volcanic rocks contain structural flaws for crack propagation, they are strongly strained by intruding magma, and they are subject to severe chemical attack from hot geothermal or magmatic fluids. Indeed, eruptions are commonly preceded by self-accelerating processes, including earthquake frequency and rates of ground deformation [*Scarpa and Tilling*, 1996]. Such accelerations can be described by [*Voight*, 1988]:

$$(d^2\Omega/dt^2) = A\ (d\ \Omega\ /dt)^\alpha \quad (1)$$

where A and α are constants, and Ω is the quantity whose rate of change measures the rate at which eruption is being approached. In this "failure forecast method" (FFM), the constants A and α are determined empirically from observational data and, when they are known, Equation (1) can be integrated to yield an estimate of the time to eruption. Applied to several pre-eruptive sequences, α has been found to lie between 1 and 2 (typically closer to 2), irrespective of the process which Ω describes [*Voight*, 1988; *Cornelius and Voight*, 1995].

Fracture studies in the laboratory suggest that rates of crack nucleation increase exponentially with time [*Lockner et al.*, 1991; *Main and Meredith*, 1991], but that rates of crack extension

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increase exponentially with the length of the crack [*Main et al.*, 1993]. Extrapolating laboratory results to large-scale failure, the bulk rock strain ε due to a *population* of fractures is expected to change with time as [*McGuire and Kilburn*, 1997]

$$d\varepsilon/dt = (d\varepsilon/dt)_0\, e^{\lambda(t-t_0)}\, e^{\alpha(\varepsilon-\varepsilon_0)} \tag{2}$$

where $(d\varepsilon/dt)_0$ is the bulk strain rate at time, t_0, λ is an empirical nucleation constant (1/time), $a = B\omega^2 S^2 L/YkT$ (dimensionless), L is the characteristic diameter of the zone of strained rock when slow cracking commenced, S is the remote applied stress (tensile or compressive), Y is Young's modulus, k is Boltzmann's constant, T is rock temperature (absolute), ω is an atomic stretching distance for breaking bonds at crack tips, and B (dimensionless) incorporates Poisson's ratio for the host rock, the coefficient of friction along the crack, and terms describing the geometry of the array [*Lockner*, 1993; *Main et al.*, 1993]; all components of a are assumed to be constant.

Seismic event rate can be related to bulk strain-rate by setting $\varepsilon = N\varphi/L$, where N is the number of recorded events and φ is the length extension per fracturing event averaged for the whole population (including both the opening and the extension of fractures). With this substitution, Equation (2) yields

$$dN/dt = (dN/dt)_0\, e^{\lambda(t-t_0)}\, e^{\gamma(N-N_0)} \tag{3}$$

where $\gamma = a\varphi/L$, and φ is assumed constant. The reference event rate $(dN/dt)_0$ describes the host rock's resistance to fracture, smaller values being linked with greater resistance. Derived from reaction-rate theory, γ measures the fraction of fluid molecules with energy sufficient to corrode country rock. As γ increases, the energy available for cracking increases and, hence, also the rate of crack acceleration. Notice that $(dN/dt)_0$ and γ need not vary together: once fracturing has started, the rate at which dN/dt *changes* (measured by γ) can be the same in rocks with different resistance, although the *actual* event rate at any given time will be smaller in rocks with greater resistance (smaller $(dN/dt)_0$).

Differentiating Equation (3) with respect to time leads to

$$d^2N/dt^2 = \lambda\, dN/dt + \gamma\, (dN/dt)^2 \tag{4}$$

(using, for convenience, initial values to define the reference parameters). Equation (4) shows, as expected, that at low event rates (and strain rates), fracturing is driven by the appearance of new cracks, such that $d^2N/dt^2 \approx \lambda dN/dt$ ($\alpha \to 1$, $A \to \lambda$ in (1)), while at large event rates, fracturing becomes limited by the rate of crack extension, for which $d^2N/dt^2 \approx \gamma(dN/dt)^2$ ($\alpha \to 2$, $A \to \gamma$). The limits to Equation (4) are identical to those reported for pre-eruptive episodes using FFM Equation (1), providing a physical interpretation of the FFM approach in terms of subcritical rock fracture.

Equation (3) indicates that the final stages of deformation before catastrophic failure (when dN/dt is very large) should be well-described by $d^2N/dt^2 \approx \gamma(dN/dt)^2$ which gives, after integration,

$$(dN/dt)^{-1} = (dN/dt)_0^{-1} - \gamma(t - t_0) \tag{5}$$

so that a plot of inverse event rate against time yields a negative linear trend. The time at failure (as $dN/dt \to \infty$) can then be estimated from the intersection between this trend line and the time axis ($(dN/dt)^{-1} = 0$). The linear nature of the trend greatly helps failure forecasting [*Voight*, 1988] and forms an important part in our analysis of events at Soufriere Hills.

Application to Soufriere Hills, November 1995

As a magmatic conduit propagates towards the surface, the rates of detected rock cracking are expected to increase because (1) the rate of fracture growth is accelerating, and (2) fracturing is occurring at shorter distances from the monitoring equipment, thereby increasing the proportion of small events (e.g., in a damage zone) which can be recognised. Of primary interest here is the increase in seismicity due to accelerated fracture growth, and so the effect of a changing source position must be filtered out.

Since larger events are most likely to be caused by extension of a principal conduit, an obvious strategy is to focus attention on the frequency of seismic events larger than a threshold value. Ideally, the threshold value is chosen by selective analysis of the total population; in practice, and especially during a crisis, it is determined by expediency. Even so, provided the absolute range and frequency distribution of fracture sizes show only modest variations with time, the frequency of events greater than the threshold should be proportional to the event-rate of the whole population. It is thus viable to seek evidence for fracture propagation using larger-event data.

Pre-eruptive seismic data at Soufriere Hills (Figure 1) were obtained from an array of short-period vertical-component seismometers, with triggered data obtained from the PC-SEIS acquisition program [*Aspinall et al.*, this issue]. The triggering algorithm counts the number of events during each 10-minute period by comparing successive 2.5-second average signal amplitudes, with thresholds set on ratio of compared amplitudes, and on absolute amplitude value [*Murray et al.*, 1996].

On an inverse-rate diagram (Figure 2a), the data show a broad inverse relation between rate of triggered events and time. Initially the trend is crudely-defined, but after 11-12 November it appears linear and converges upon the observed date of eruption (about November 15). The early fluctuations may reflect data uncertainty or truly intermittent rates of fracture growth. The main sources of uncertainty in measured event rates are (1) variation in the proportion of active cracks that trigger detected events, (2) clustering with time of detected event-rates, and (3) variability in the physical properties determining γ in Equation (5); instrumental error is expected to be negligible in comparison. Although lack of data prevents these uncertainties from being measured directly, first-order estimates can be made by analogy with laboratory data and from standard error propagation.

Laboratory measurements of slow cracking (for which errors on γ are assumed insignificant) show typical scatters of a few percent [e.g., *Meredith et al.*, 1990]; accordingly, σ, the combined error due to (1) and (2) above, is set at 5-10%. For uncertainties in event rate due to variations in γ, standard error propagation

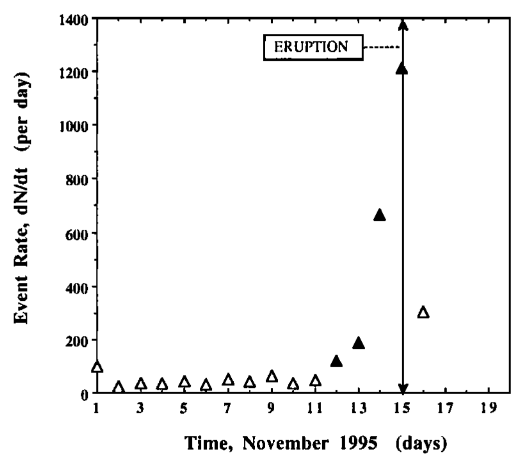

Figure 1. Acceleration in daily seismic event rate before eruption on 15 November 1995. Filled triangles as in Figure 2a.

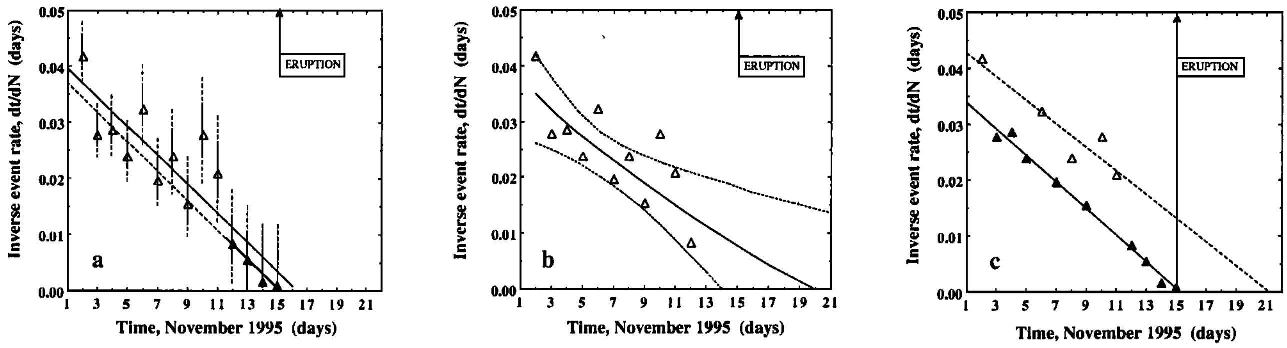

Figure 2. Inverse seismic event rates with time. *(a)* Treated as a single population, a simple linear trend is clear only after 11 November. Filled error bars, $\sigma = n = 0.05$; dashed error bars, $\sigma = n = 0.1$. The narrow solid line is a linear regression for the whole data. Backward extrapolation (dashed) of regression line (bold) for last four data points (filled triangles) runs close to five earlier points and suggests the possibility of two distinct trends (see *c*). *(b)* Simulated inverse event using nonlinear FFM analysis and extrapolating from 12 November. The preferred value of α is 1.88, yielding failure on 19-20 November (solid line). The eruption window is 14 November - 13 December (broken lines). *(c)* Treated as two populations, the data fall on near-parallel inverse-rate trends, perhaps due to episodic growth of the magmatic conduit (filled triangles, population 1; open triangles, population 2).

[*Barlow*, 1989] yields a fractional error on inverse event rate of $t\gamma n\sqrt{8}$, where n (assigned a single value for simplicity) is the fractional error on component terms for γ (except for k, which is constant). Figure 2a shows example variations in total error for $\gamma = 0.0024$ d^{-1} (see below) and $\sigma = n$ and set at 5 and 10%. The larger error bars ($\sigma = n = 0.1$) yield total uncertainties on inverse event rate from 13% to almost 400% as event rate increases. Nevertheless, a single line cannot be located through all the error bars (even the best solutions omit 2 of the 14 error bars); for $n < 0.1$, this condition can be achieved only when $\sigma > 0.2$. Although large, component errors of 20% or more cannot yet be discounted. Two evaluations thus remain for Figure 2a, according to whether or not the early fluctuations can be attributed to data uncertainty.

Treating the data as a single trend with large scatter, linear regression (Figure 2a) gives an eruption date of 16 November, but only a modest correlation coefficient ($r^2 = 0.79$). A nonlinear regression analysis used in FFM forecasting [*Cornelius and Voight*, 1995] yields $\alpha = 1.88$ (i.e., very close to the linear inverse-rate trend, $\alpha = 2$) and preferred eruption dates of 19-20 November (Figure 2b). Both analyses give preferred eruption dates within days of the actual event (15 November), supporting the association of accelerating event rate with propagation of a magmatic conduit.

Relaxing the single-trend assumption, Figure 2a shows that half of the ten "scattered" data points (before 12 November) lie along a linear extrapolation of the regression line ($r^2 = 0.94$) for the last four pre-eruptive event rates. The remaining data appear to define a second linear trend and, if two trends are assumed (Figure 2c), linear regression yields, for dN/dt in events per day and t in days,

$$(dN/dt)^{-1} = 0.037 - 0.0024\, t \qquad (6)$$

for the trend that includes the last four pre-eruptive data (population 1; $r^2 = 0.99$ for 9 points), and

$$(dN/dt)^{-1} = 0.045 - 0.0021\, t \qquad (7)$$

for the second trend (population 2; $r^2 = 0.87$ for 5 points); in both cases, $t_0 = 0$ corresponds to 01 November. Strong linear trends are implied, again supporting an interpretation in terms of conduit propagation, especially as the population 1 trend correctly indicates 15 November as the preferred date of eruption.

As expected for a common deforming source, the two trends share virtually the same gradient ($\gamma \approx 0.0024$ d^{-1}). Their relative displacement indicates a greater resistance to fracture growth for population 2 events. The simplest interpretation is that the populations are produced by the intermittent advance of a single fracture system, intermittency being induced either by alternating layers of rock with different fracture resistance, or by temporary increases in effective resistance within a damage zone.

A damage-zone control is attractive because it is an integral part of the fracturing process, and does not depend on variations in external conditions. The alternative interpretation would require a bimodal structure for Soufriere Hills. Even for fracture growth over kilometers, the layers of different strength must have thicknesses ~100 m or less, and so may be too thin to be resolved by standard geophysical surveys. However, if a damage-zone control was important, then the population 2 events might be distinguished from population 1 by a greater occurrence of smaller events as the damage zones form. Such analyses await future study. The key point is that the inverse event-rate data, whether treated as one or two populations, suggest that subcritical rock fracture is a feasible mechanism for limiting the rate of conduit propagation before an eruption.

Forecasting volcanic eruptions

A tantalizing feature of the inverse-rate method is its potential as a tool for forecasting some types of eruption [*Voight*, 1988; *Cornelius and Voight*, 1995, 1996]. As for Soufriere Hills, a common problem has been ambiguity in detecting an inverse-rate trend from apparently noisy data [*Cornelius and Voight*, 1995]. Recognition here that even a single fracture system might generate parallel inverse trends due to episodic advance offers new insight to analysing some inverse-rate data. In addition to seeking a single mean trend by statistical methods [*Cornelius and Voight*, 1995], parallel linear trends might also be considered using pattern-recognition techniques. Applied to the Montserrat seismicity data, which fortunately (and perhaps deceptively) appear to be simple, even judgment by eye might have raised suspicions of two trends by November 9. Extrapolating the two trends from this date would have revealed 15-21 November as a potential eruption window (smaller than the window from FFM analysis on the whole data set; Figure 2b).

Previous experience has shown that forecasters should never rely on a single data set, but should use all appropriate supplementary evidence. FFM analysis (analogous to Figure 2b) on November 12 for line-length displacements of Castle Peak dome [*Jackson et al.*, this issue] suggests an eruption about November 16, within an eruption window 13-27 November. Considered together, eruption windows from different measurements could provide information useful for hazard management decisions. The Montserrat data thus further support the use of inverse-rate forecasts at least for vent opening before the emplacement of andesitic-dacitic lava domes [*Cornelius and Voight*, 1995, 1996].

Conclusions

The pre-eruptive increase in seismic event rate at Soufriere Hills in November 1995 is consistent with growth of the fracture system along which magma reached the surface. The rate of growth was constrained by weakening of country rock around the tips of propagating fractures. The preferred weakening mechanism is stress corrosion, driven by the stress enhanced chemical attack of hydrothermal and juvenile fluids. Alternations in daily seismic event rate suggest that propagation was episodic. Recognition of episodic fracture advance can simplify interpretation of linear trends between inverse (seismic) event rate and time, encouraging the view that such trends may usefully contribute to eruption forecasts, especially when consistent with changes in associated precursors (e.g., ground deformation). The fracture control also highlights the forecasting potential of other seismological data, such as information on the size-frequency distribution of seismic events and high resolution monitoring of foci migration.

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C.R.J. Kilburn, Benfield Greig Hazard Research Centre, Department of Geological Sciences, University College London, Gower Street, London WC1E 6BT, U.K. (e-mail: c.kilburn@ucl.ac.uk)

B. Voight, Department of Geosciences, Pennsylvania State University, 116 Deike Building, University Park, PA 16802, U.S.A. (e-mail: voight@ems.psu.edu)

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