**Long-Period seismicity in the shallow volcanic edifice formed from slow-rupture earthquakes**

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**Despite recent technological advances in volcano monitoring, eruption forecasting is still inadequate. Improved forecasting requires a deeper understanding of when unrest will lead to an actual eruption. Shallow Long Period (low spectral frequency) seismic events are routinely employed as a primary tool in forecasting strategies as they often precede eruptions. They are universally explained as resonating fluid-filled cracks or conduits, indicating the presence of mechanically active near-surface fluids. We undertake very high resolution seismic field experiments at Mt Etna, Italy; Turrialba, Costa Rica and Ubinas, Peru, in which we find that seismogram resonance is propagation path related whilst the seismic sources comprise short pulses. Data analysis and numerical modelling show that slow-rupture failure in unconsolidated volcanic materials reproduces all key aspects of these new observations. Contrary to current interpretations, here we show that our observed Long Period events are not direct indicators of fluid presence/migration, but rather are markers for upper edifice deformation. This finding encapsulates this seismicity within growing observations of a spectrum of deformation rates in other non-volcanic environments, from slow-slip earthquakes through fast dynamic rupture. It calls for a reassessment of how low-frequency seismic signals are interpreted in their key role in eruption forecasting.**

A primary goal in volcano seismology is to find reliable precursory signals to volcanic eruptions. Hence it is of paramount importance to understand the physical processes which underlie the diverse range of seismic events supported by volcano edifices. Broadly these events can be divided into four groups (i) volcano tectonic (VT) earthquakes, (ii) Long

Period (LP) events, (iii) very long period (VLP) events and (iv) seismic tremor[1]. Whilst VTs resemble tectonic earthquakes in signal character and are thought to be caused by brittle failure of the edifice[2], LPs, VLPs and tremor are usually interpreted in terms of fluid driven processes. In particular, the current universally applied models for LP events require that they are necessarily associated with either fluid-filled crack[3,4] or conduit[5,6] resonance, triggered by rapid disturbance of the fluid-filled structure (e.g. a fluid pressure fluctuation in a sub-surface conduit or dynamic excitation by a local VT earthquake). As fluids are thought to participate actively in the source process, mechanistically this sets all LPs apart from both VT and tectonic earthquakes, forming a distinct earthquake class which is viewed with considerable interest in terms of our understanding of the shallow volcano 'plumbing system'[7,8,9,10]. Their increased frequency of occurrence is regularly used as a key component in eruption forecasting [2].

**Seismic source versus path effects**

LPs have lower seismic frequencies than VTs, typically 0.3 – 2 Hz central frequency, and are usually of narrow spectral band and of long duration (5-40 secs). Recently, full wavefield numerical seismic simulations in realistic volcano velocity models designed to constrain LP source inversions[11] demonstrated a surprisingly strong influence of path effects. Short duration low frequency input source pulses present as artificial long duration resonating LP-like signals when recorded more than a few hundred metres from the source. The effect is very pronounced for shallow sources (< 1 km depth) in low velocity near-surface layers. Sensitivity kernel[12] analysis demonstrates that these effects are 'global' (i.e. not local to individual recording stations, but seen throughout the numerical model) implying that they could be misinterpreted as seismic source effects. As such proximal field observations above the sources (e.g. stations < *c.* 500 m from the epicentre) are very rare in field studies, these numerical results motivated an uniquely high resolution field experiment on Mt Etna in 2008, in which two families (similar waveforms) of LP events were detected[13]. The wavefield distribution for LP events from that experiment (Fig. 1a) shows extreme spatial variability, with pulse-like low frequency signals at the summit stations that appear as classical resonating LP signals at off-summit stations, in excellent agreement with numerical predictions[11]. This unequivocally demonstrates that the apparent resonance in these low-frequency seismic events is caused by wavefield distortions as a consequence of wave propagation effects and is not source-related. Similar pulse-like LP events can be seen at other time periods at Mt Etna and other volcanoes, when short-range observations are available (Fig. 1b, Supplementary Fig. S1). In fact, we were unable to identify long duration LP events at summit stations in these datasets amongst many thousands of pulse-like LPs. Hence, on several volcanoes, we have identified short duration pulse-like LP populations which masquerade as classical long-duration LP events if recorded > *c.* 500 - 800 m from the epicentre. We conclude that the long seismogram durations in these data are purely path related. Based on waveform analysis, Harrington and Brodsky[14] arrived at the same conclusion for some hybrid volcano seismic events. A literature search reveals that shallow LPs are almost universally recorded at epicentral distances > 500 m on standard volcano networks, which usually lack summit stations for operational reasons. Hence non-resonating LP sources could be commonplace but further specific field experiments will be needed to detect them. Three location methods applied to the 2008 Mt Etna data[15] show that these LP events are shallow: < 800 m and < 600 m depth beneath the summit for families 1 and 2, respectively. Moment tensor solutions for both 2008 LP families favour a predominantly tensile crack mechanisms[16] with the possibility of a shear component. Both LP families occur on cracks with very similar strikes (approximately WSW-ENE) however the failure planes for both families are orthogonal in conjugate sets of fractures, which is why their seismograms differ leading to two distinct families (Supplementary Fig. S2). We cannot unequivocally distinguish between crack opening and closing, although displacement seismograms favour an opening mechanism.

**Dry failure of the edifice?**

The pulse-like nature of the LP signals in this study is difficult to reconcile in terms of a fluid-driven crack model. It would require specifically tuned choices of crack stiffness and/or seismic wave attenuation, to dampen resonance. This difficulty is compounded by independent results from laboratory experiments, which demonstrate that the presence of fluids in the source process leads to long duration monochromatic signals[17,18], which, as demonstrated, we do not see as a source signature in these short duration pulse-like field data. Specifically, 'wet' and 'dry' laboratory experiments designed to determine unequivocal seismological discriminators between fluid-driven LP seismicity and brittle (VT) events reveal that 'wet' experiments generate long duration resonating 'classical' LP-type signals whereas the 'dry' experiments produce VT-like brittle failure events[17]. Importantly, the source corner frequency (fc) for the 'dry' experiments scales with event size as $1/fc^3$ [(19)], consistent with brittle failure theory[20]. In contrast, the fluid driven laboratory events do not show any clear fc scaling with event size[19], consistent with current applications of a fluid driven crack model, where frequencies for a crack of a fixed size are controlled by the crack's geometry and stiffness, not seismic moment (if the crack size was allowed to vary in these models, scaling would be observed). Figure 2 shows an equivalent analysis for 2004 field LP data from Mt Etna. A clear scaling of fc with seismic moment is visible, strongly indicating that these shallow Mt Etna LPs are caused by 'dry' mechanical failure, consistent with laboratory observations. Given the body of observations above, it is prudent to seek an alternative explanation for LP events observed in this study. We investigate stress-driven edifice deformation as a possible cause of LP activity.

Active volcanoes have relatively high deformation rates. On Mt Etna, flank instability in the form of East-West spreading is by far the predominant type of deformation[21,22]. Rheological mechanisms for accommodating this deformation in the near surface (*c.* top 1 km) have not been addressed in the literature. In general VT and LP events on volcanoes separate spatially, with VT events located approximately > 1-2 km below the edifice surface and LP events located at depths of < 1 km[2,23-25]. Brittle failure VTs are markers for edifice fracturing. Hence the absence of near surface fracturing (i.e. very shallow VT events) seems to imply that deformation is almost entirely silent (aseismic) in the top 1 km or that seismic events other than brittle VTs are related to near surface deformation. As demonstrated in this study, LP events develop on populations of primarily tensile cracks (depths < 800 m) at WSW-ENE orientations, consistent with known East-West spreading and sliding of the eastern flank of Etna volcano. A key question is: do they play a role in accommodating some near surface deformation in the form of slow seismic rupture? That is, are LP events a weak brittle signal in a largely ductile deformation field? The top 800 m on many volcanoes contain poorly consolidated material with very low seismic velocities[26-29]. De Barros *et al* [13] determined a mean P-wave velocity of 1.8 km/s in the top 800 m of Mt Etna's edifice, giving a Rayleigh wave velocity ($V_R$) of *c.* 855 m/s (for a Poisson's ratio of 0.3 [(16)]). Source inversions constrain the 2008 LPs to have a predominantly tensile (mode I) crack mechanism[16]. Broberg[30] demonstrated that mode I brittle failure cracks (i) are theoretically forbidden to rupture at speeds > $V_R$ and (ii) in practice rupture at speeds << $V_R$ in 'low stress' environments. Hence rupture speed can be controlled by both the failure mechanism and the stress environment. We performed 2D molecular dynamic (MD) rupture simulations[31], using codes from O'Brien and Bean [32], for mode I cracks in a model with an upper edifice (top 800 m) $V_R$ = 855 m/s (Figs 3a and 3b). As our model 'fault' is idealised and smooth, in order to achieve sub shear-wave rupture speeds in the MD model we approximate the real structure as a shallow vertical tensile crack (5% weaker than the host material). The model is pre-stressed with vertical stress $\sigma_1$= 0 and horizontal stress $\sigma_3$= -0.0065 MPa (i.e. in tension, up to the approximate tensile strength of near-surface volcanic materials [33]). It is then allowed to equilibrate and failure is achieved by subsequent incremental increases in the horizontal extensional stresses until rupture initiates. Consistent with theoretical predictions [30], we observe sub-$V_R$ rupture speeds. These dynamic simulations demonstrate that slow rupture speeds (relative to failure in stiff materials with higher Rayleigh wave speeds) of *c.* 816 m/s on a 500 m long crack lead to 2 second long pulse-like LP seismograms, similar to those recorded at the summit stations on several volcanoes (compare Fig. 1b with Fig. 3a). Our kinematic simulations (not shown) produce equivalent pulse-like synthetic seismograms, for crack dips from ref 16. Even though the true stress regime is unknown and highly stylised in the 2D MD model, seismogram velocity amplitudes are of the same order of magnitude as our largest LP field observations.

## Volcano seismicity and other 'slow-rupture' observations

Crack dimensions for field data can be estimated using an expression for the radius of an expanding circular tensile crack[34] $r=1.99*V_s/(2\pi*fc)$. Taking $V_s$=950m/s and fc=0.72 (Fig. 2) gives a circular crack diameter of *c.* 800 m. This estimate is broadly consistent with the dynamic rupture simulation results (Fig. 3) and the vertical dimension of the near surface low velocity zone (< 1 km), which we argue herein limits rupture speeds and stress accumulation in quasi-brittle failure events leading to low frequency seismic radiation even for small faults. The past decade has seen the discovery of a wide range of slow-earthquake phenomena in a wide variety of environments from subduction zone interfaces to landslides to glaciers (*see 35 and references therein*). The controlling mechanism for rupture and/or slip rate (fast classical earthquakes to slow or even creeping) appears to relate to the local stress state and/or frictional properties of the material, with brittle failure as the underlying cause of seismic radiation. Peng and Gomberg[35] developed an integrated perspective on the continuum between classical earthquakes, slow-rupture and slow-slip phenomena. Within their framework we propose that volcano VT earthquakes morph into LP events in the shallow edifice. This VT-LP transition occurs because weak low-stiffness materials promote slower rupture speeds, both leading to low frequency seismic radiation. This implies that at least some populations of LPs (but almost certainly not tornillo[7] events, which do seem fluid-driven) are part of the same mechanical/dynamical class as VT events, in that they are caused directly by brittle failure. Hence, like VT events these LPs are 'stress driven' and do not require a 'fluid driven' source model. As slip rate ≈ [(stress drop x rupture velocity)/shear modulus] [36], the interplay between local stress conditions and local material properties controls failure details. Our estimated stress drops from waveform analysis[37] of the Mt Etna LP seismicity of *c.* 0.01 MPa are 2-3 orders of magnitude smaller than expected for tectonic seismicity. Possible causes of such low stress drops are discussed below. In Mt Etna's case, we propose that East-West extension leads to 'slow-tear' predominantly tensile failure. It should be noted that we cannot definitively rule out the presence of possible double-couple components in Moment Tensor inversions for small LP events. Crack apertures are poorly constrained, estimates vary from *c*. 2 mm using $\Delta u = (24/7\pi)(\Delta\sigma/\mu)r$ [(34)] to 0.5 mm taking the shear failure analogy of slip $=M_0/(r^2\,\pi\mu)$ , where $\Delta u$ = crack maximum aperture, $\Delta\sigma$ = stress drop, $\mu$ = shear modulus, $M_0$ = seismic moment. Hence these displacements are likely below the current detection threshold of earthquake related deformation studies.

Apuani *et al* [33] undertook comprehensive field studies and laboratory experiments to determine the physical and mechanical properties of near surface deposits at Stromboli volcano, Italy. Their key finding is that bulk deposits are both compliant and weak. Using the damage mechanics model of Amitrano[38] we simulate seismicity in an equivalent material, figure 4a. A surprising result is the emergence of swarms of very low stress-drop events rather than discrete higher stress-drop individual failures, very similar in character to our low stress-drop LP families (swarms). We also see diffuse damage in these seismicity simulations, consistent with the observed spread in LP hypocentres, which are not localised on an individual structure[13,15] – but are sufficiently close to display similar waveforms, when they share the same focal mechanism. Unexpectedly, model failure-size distributions (seismic '*b*-values') show non power-law scaling, with a deficit of larger events, which we determine is primarily controlled by the low internal friction angle of the material. While the stress-strain relationship in figure 4a has the characteristics of ductile behaviour, it is accompanied by seismicity, which demonstrates that the material sits at the brittle-ductile transition and appears to be unable to sustain the high stresses required for larger seismic events. Seismic '*b*-values' for LP seismicity at Mt Etna and Turrialba, Costa Rica (data collected during a summit deployment in 2009), exhibit similar behaviour (Fig. 4b) suggesting that the swarms (often forming families) of low stress-drop pulse-like LP events are a consequence of failure in material close to the brittle-ductile transition. Here, as in weakly cemented sand[39], we suggest that quasi-ductility in shallow (< 1 km depth) volcanic materials is primarily controlled by low internal friction angles of the volcanic deposits, not by high temperature and pressure. Numerical simulation tests demonstrate that model seismicity data show power law scaling for internal friction angles > *c*. 35 degrees. Drilling at Unzen volcano, Japan[40], where borehole washout and accidental side tracking occurred to several hundred metres depth, testifies to the unconsolidated nature of subsurface volcanic materials at depth – demonstrating that the physical properties determined by Apuani *et al* [33] and used in our simulations are not merely superficial. Consistent with our numerical seismicity results (Fig. 4a), laboratory experiments in gypsum at the brittle-ductile transition show swarms of low frequency acoustic emissions associated with the dynamic propagation of shear bands[41], showing that brittle-ductile materials can support both dynamic rupture and low frequency seismicity.

**Eruption forecasting: Reassessing the upper edifice**

We propose a new model for a class of LP volcano seismicity, where LP swarms are caused by slow, quasi-brittle, low stress drop failure driven by transient upper-edifice deformation. This model is not appropriate for tornillo-type LP events. Underlying stress fluctuations can be driven by gravity, gas influx or magma migration, whilst the consequent material failure is 'dry' mechanical, where fluids are not directly involved in the seismic signature of the source process. The absence of resonance in the source, non power-law *b*-value observations, swarms of very low stress drop LP events, LP source crack orientations consistent with known East-West spreading on Mt Etna, the lack of upper edifice VT earthquakes, $1/fc^{2\text{-}3}$ scaling of LP corner frequencies, dynamic rupture simulations and numerical modelling of upper edifice material failure, all call for a new paradigm which is consistent with this broad range of new observations. We hypothesise that LPs in this study represent a weak slow-rupture brittle signature in an otherwise ductile failure field. The proposed model predicts the temporal evolution of pre-eruptive seismicity as summarised by McNutt[42], particularly the relative timing of VT and LP events and the relative seismic quiescence often seen prior to eruptions, when magma or gas driven stress perturbations are concentrated in the uppermost part of the edifice. More broadly, this slow-rupture model is consistent with an emerging framework[35] for failure in Earth materials, from slow-slip phenomena through full dynamic rupture, across a wide range of geo-environments. In our proposed model high spectral frequency VT seismic events morph into low spectral frequency LP events as they attempt to initiate in very weak near-surface volcanic materials. In this model (non-tornillo) LP and VT seismicity are unified, forming part of the same 'stress driven' mechanical class - simplifying our understanding of these event types and integrating aspects of volcano seismology with wider seismological developments over the past decade. Finally, improved forecasting requires a deeper understanding of when unrest will lead to an actual eruption or when potential eruptions stall[43]. The model presented herein offers a new perspective on the workings of the uppermost edifice on volcanoes. Consequently it has significant implications for the way in which we apply seismology to forecasting and hazard estimation. Additional very near-summit observations of LP seismicity and high resolution surface deformation are required, at a variety of volcanoes, to further test this model.

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**Figure:**

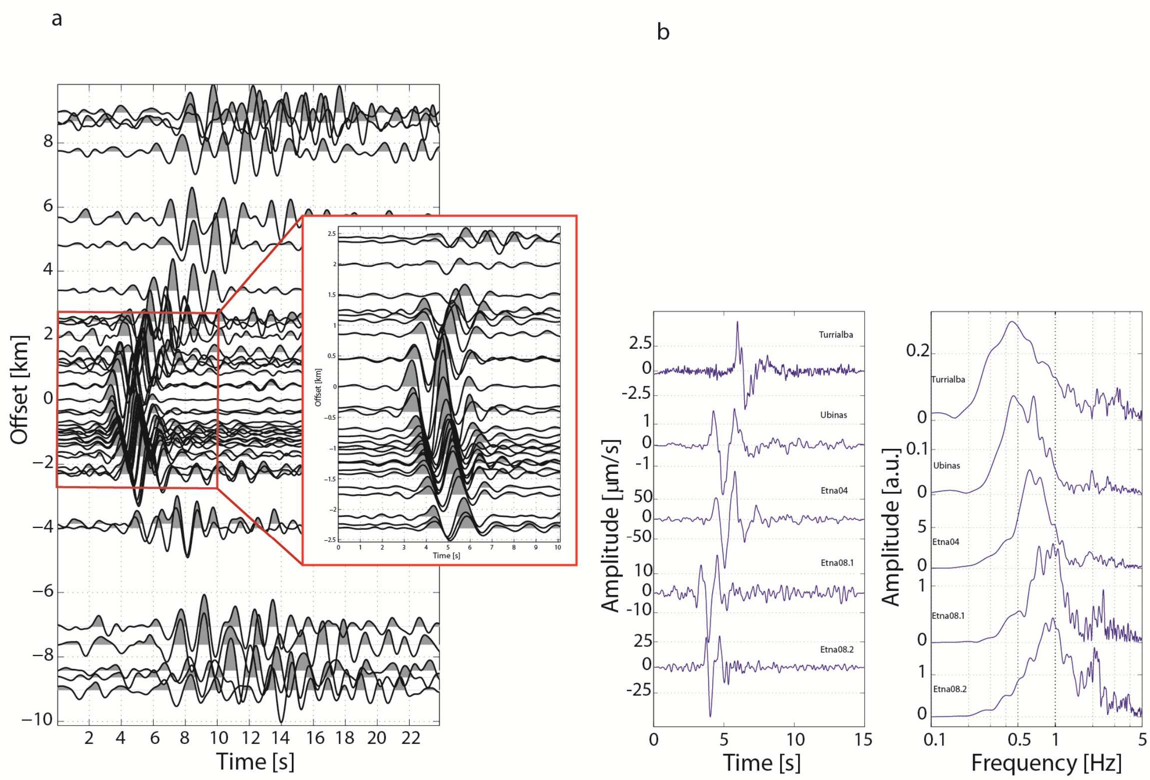



**Figure 1 Illustration of short duration LPs and strong propagation path effects**

**a**, Spatial distribution of the LP wavefield for Family #1, 2008 Mt Etna field data. Each seismic trace is a stack of *c*. 60 events in the family. Events are located beneath the summit at depths < 800m. Normalised vertical component traces are plotted as a function of the station's distance from the volcano summit. **b**, left panel, Shallow pulse-like LP events (vertical component) detected on near summit stations at Turrialba Costa Rica (2009); Ubinas, Peru (2009), and two different time periods on Mt Etna (2004 & 2008, Family 1 & 2); right panel, amplitude spectra for the data shown in the left panel.

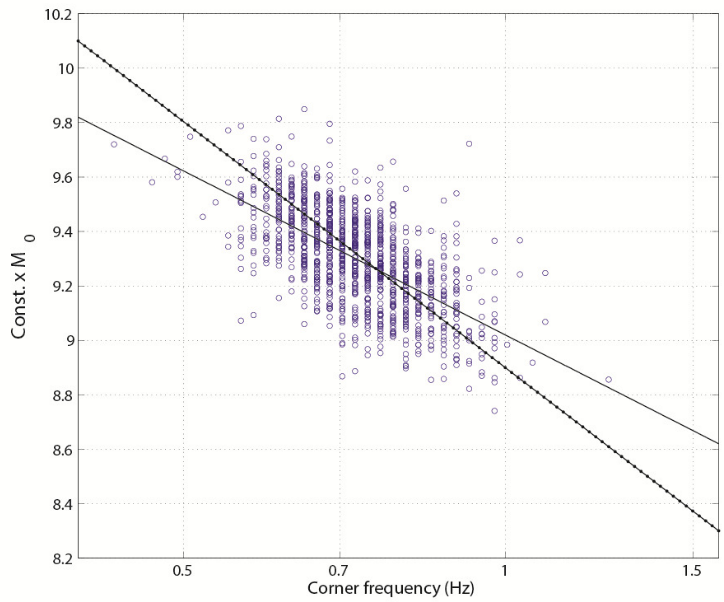
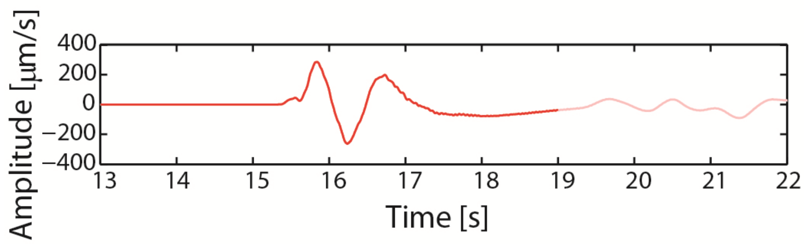



**Figure 2 Scaling of LP seismic moment magnitude versus corner frequency**

The amplitude spectra of 1150 LP events, recorded in March 2004 at Mt Etna *ECPN* near-summit station are fitted with an $\omega^{-2}$ source model, to determine the corner frequency. The y-axis is proportional to the seismic moment. The thin line is best fitting with slope -2.2; the thick line has slope -3, and is broadly consistent with the data. Note: We use 2004 data as the S/N ratio is not high enough in the 2008 Mt Etna data to perform this analysis.

a

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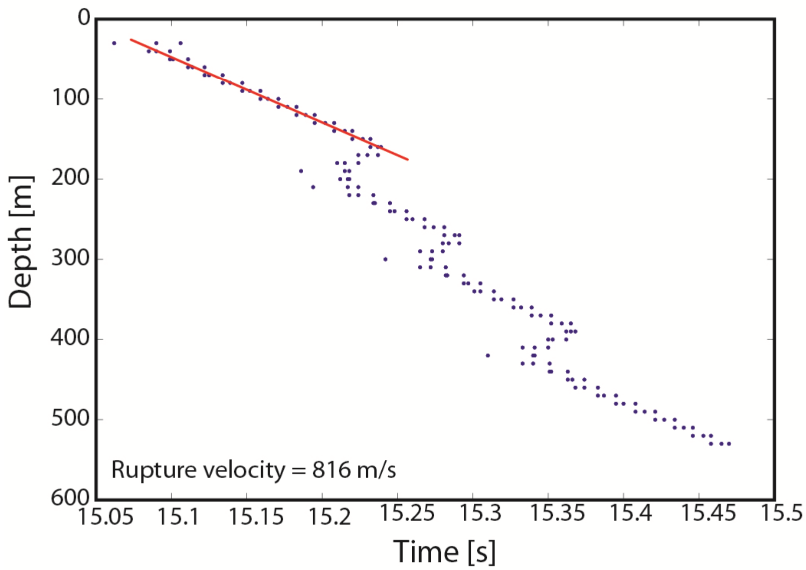


**Figure 3 2D Molecular Dynamic simulations of rupture propagation**

**a**, Illustrative example of slow rupture seismograms. Vertical component seismic trace caused by rupture on a 500 m long vertical tensile crack. The crack's top is 20 m below the surface. The seismic station is on the free surface, 500 m laterally displaced from a surface projection of the crack. Faded parts of traces represent edge effects from model boundaries/surface. **b**, Rupture velocity plot for simulated data in **a**. Blue dots represent failed bonds. The estimated (red line) sub-Rayleigh rupture velocity ~ 816 m/s.

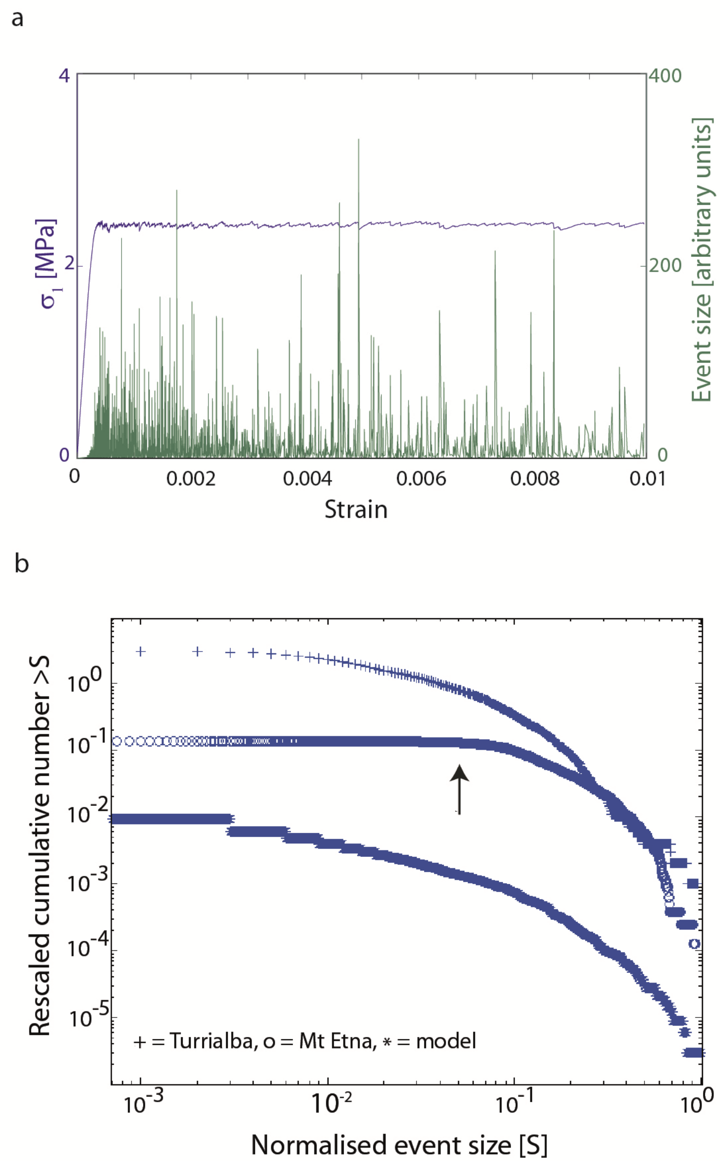


**Figure 4 Simulated seismicity in a weak volcanic upper edifice**

**a**, Stress-Strain (stress on left y-axis & blue data) and seismicity (seismic event sizes on right y-axis & green data) output from a simulated uni-axial deformation experiment using the damage mechanics model of Amitrano [38] and material properties for shallow volcano deposits[33]. Model parameters: Modulus of elasticity 0.8 GPa; Poisson's ratio 0.3; Apparent cohesion 0.6-1.1 MPa; Angle of internal friction 20°; **b**, Seismic '*b*-value' plots for model data from fig 4a (*), for LP events from Mt Etna (o) and for LP data from Turrialba in 2009 (+). Data are normalised, all catalogues have > 1100 events. Mt Etna magnitude of completeness is indicated by black arrow.

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C.J.B. initiated the concepts, analysed synthetic seismicity data and wrote the manuscript. L. De B. analysed the seismic data and with I.L. helped develop the concepts. J-P.M. provided data and intellectual input. G.O'B. carried out rupture modelling and S.M. made contributions on source modelling.

# Long-Period seismicity in the shallow volcanic edifice formed from slow-rupture earthquakes

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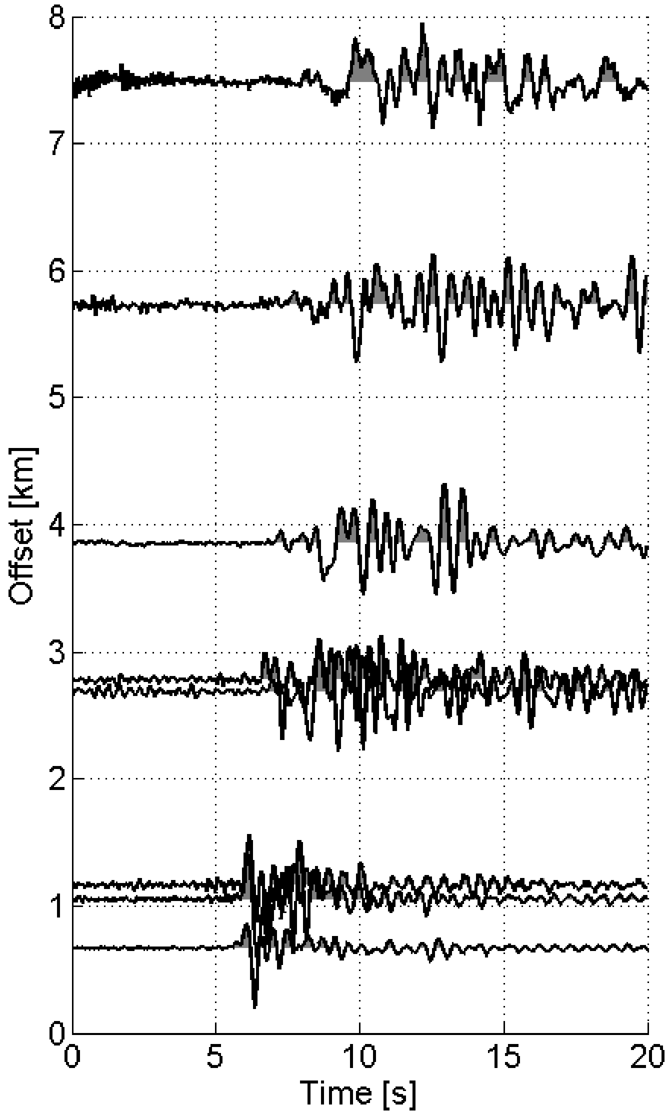
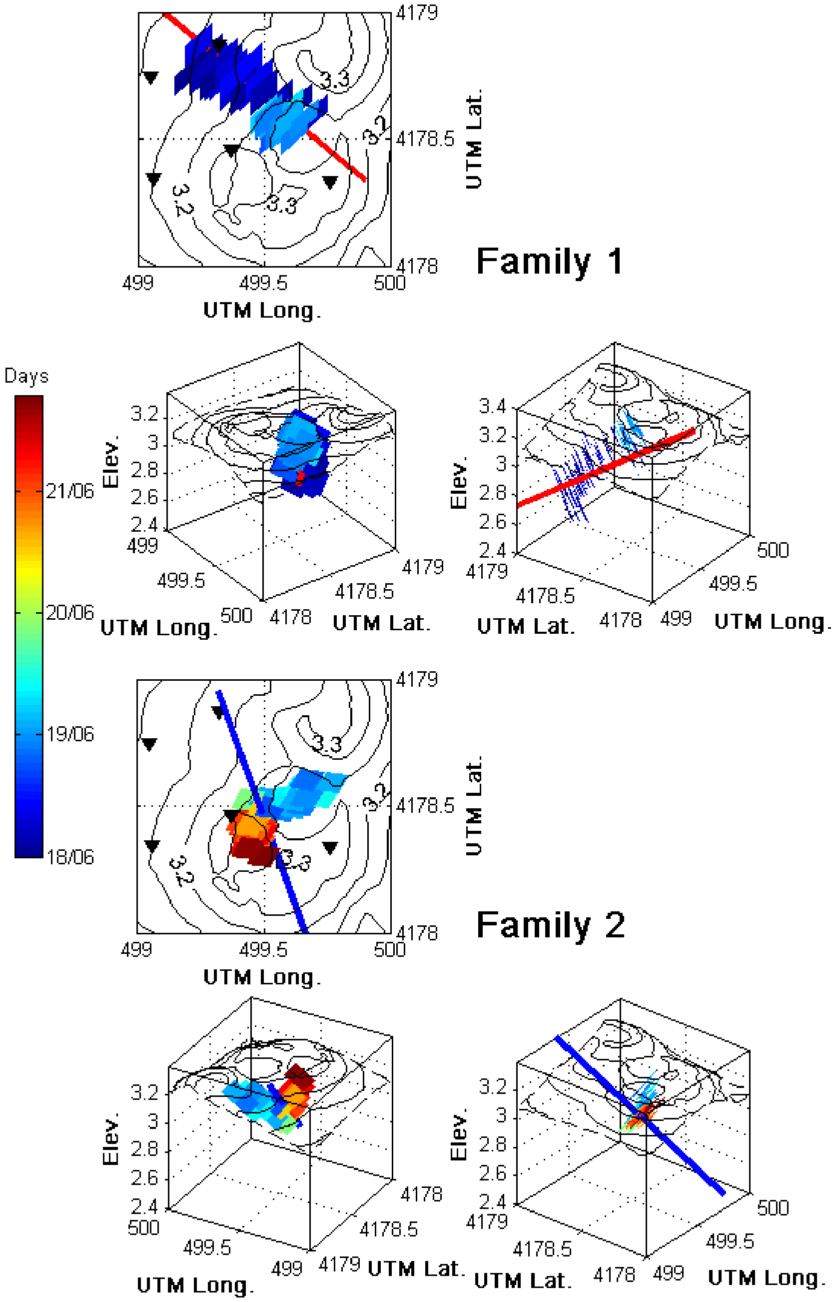
## Supplementary Information




**Supplementary Figure S1: Path related resonance for Long Period events at Turrialba volcano, Costa Rica**

Seismic traces plotted as a function of epicentral distance from the source for a LP event on Turrialba volcano, Costa Rica. The data are from a 2009 deployment specifically designed to capture near source seismograms. Data are filtered 0.5-5Hz. Traces are normalised individually. Near source seismograms are pulse-like, whilst more distal seismograms show strong path related resonance.




**Supplementary Figure S2: Crack geometries for Long Period events at Mt Etna, Italy**

LP event crack geometries obtained for the 2008 Mt Etna dataset through Moment Tensor inversion[16] are represented by the squares centred on the source locations obtained in reference 13. The solid lines indicate the normal axes (i.e. the tension axes) of the cracks. Geometries are represented in map and inclined views (left: from SE to NW, and right: from SW to NE) for both LP families. The colour scale represents the occurrence time of the events.