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The lateral extent of volcanic interactions during unrest and eruption

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Volcanic eruptions often occur simultaneously[1-4] or tap multiple magma reservoirs[5,6]. Such lateral interactions between magmatic systems are attributed to stress changes[7,8] or hydraulic connections[4,8-10] but the precise conditions under which coupled eruptions occur have yet to be quantified. Here we use InSAR to analyse the surface deformation generated by volcanic unrest in the Kenyan Rift. We identify several magma sources located at depths of 2 to 5 km; importantly, sources that are spaced less than 10 km apart interact, whereas those spaced more than 25 km apart do not. However, volcanoes up to 25 km apart have interacted in the geological past[1,11]. Thus, volcanic coupling is not simply controlled by the distance between the magma reservoirs. We then consider different tectonics settings globally, including intraplate volcanoes such as Hawaii and Yellowstone, arc volcanism in Alaska and Chile, as well as other rift settings, such as New Zealand, Iceland and Afar. We find that the most closely spaced interactions are controlled by the extent of a shallow crystal mush layer, stress changes can couple large eruptions over distances of 20 to 40 km, and only large dyke intrusions or subduction earthquakes could generate coupled eruptions over distances of 50 to 100 km.

Volcanic eruptions commonly involve multiple chemically distinct sources, different vents and even multiple volcanoes[2-5,12]. Multiply sourced eruptions can be explained by the emerging paradigm of magma reservoirs comprising stacked sills or chemically distinct melt lenses within crystalline mush zones[13]. The presence of multiple melt batches within a single subvolcanic system can produce protracted eruptions that progressively tap individual lenses, which are either vertically stacked[5,14] or laterally distributed,[5] and may ultimately lead to caldera collapse[13]. External factors, such as changes in the crustal stress regime, seismic waves and the interactions between mafic and silicic systems[8,9,15] also have the potential to affect multiple volcanic systems simultaneously[2,16].

We start by examining the geometry of magma storage within the Kenyan Rift (Figure 1). Here isotopically-distinct melt batches reside within isolated melt pockets at depths of 4-8 km[11,17] overlain by shallow geothermal reservoirs (1 – 3 km)[18]. Neighboring volcanic systems can be laterally connected, at least intermittently, as illustrated by lateral magma withdrawal during caldera collapse at Silali and Suswa[17], intermixing between comendites from Olkaria with peralkaline trachytes at Longonot[11] and interbedded ignimbrites at Suswa and Longonot[1]. To constrain short-term interactions among neighbouring volcanoes, we use satellite radar interferometry (InSAR) to map ground displacement in the Kenyan Rift[19] (see Methods). We first examine Paka and Silali, and then discuss interactions within the cluster at Longonot, Suswa, and Olkaria (Figure 2).

Unrest at Paka between 2006 and 2010 comprised temporally-variable deformation, including five phases that can be modeled by inflation and deflation of four distinct but simple sources beneath the northeastern and southern flanks of the volcano (Figure 2). Seismic and magnetotelluric studies indicate a hot body at 2.5 – 5 km overlain by a hydrothermal system[18]. With a source depth of only 1-2 km, the initial phase of deformation is consistent with pressure changes within the hydrothermal system under the southern flank. The next phase was located at ~5 km underneath the northeastern flank, and likely represents input of new material (magma and/or magmatic volatiles). The remaining phases had source depths of 2-5 km, consistent with magma migration through an interconnected series of sills or melt lenses. At the same time, neighbouring Silali volcano exhibited long-term, linear subsidence at rates of 1-2 $cmyr^{-1}$ and source depths of 4-5 km. This subsidence is consistent with volume decrease of a magmatic source associated with cooling and crystallization[20], or by degassing[21].

At Longonot, rapid uplift in 2004-2006 was followed by slow subsidence from 2007-2010. The initial uplift has a source depth of ~4 km directly beneath the edifice; subsequent subsidence was shallower and located south of the volcano (Figure 2), where geothermal exploration has identified a hydrothermal reservoir extending in the direction of groundwater flow[18]. Conceptual models involving direct fluid transfer or stress-induced permeability changes have been proposed to explain observations of this type elsewhere[21,22]. We see no evidence for deformation above 0.3 $cmyr^{-1}$ or unusual magmatic activity at nearby Suswa or Olkaria, yet both are thought to host active magmatic systems: Suswa experienced 2-3 cm of subsidence during 1997-2000[19], and Olkaria supports Africa's largest geothermal plant.

Is there evidence of interaction among any of these restless volcanic systems? Silali and Paka volcanoes are <25 km apart (Figure 1) and their deformation patterns suggest that both are underlain by complex and recently resupplied magmatic systems. Yet, although the activity at Paka triggered a response from other sources beneath Paka, there was no change in either deformation pattern or rate at Silali. From this, we infer that there is currently no lateral interaction between the two systems, either as hydraulic links by which material can move between mush zones, or through stress changes. The minimum dimensions of interaction between the two systems can be constrained by the distance between melt lenses at Paka (~10 km) and the diameter of the inferred source beneath Silali (~9 km). Given that the two systems behave independently, the maximum extent of connectivity must be <25 km. Similar arguments can be used to place constraints on the extent of interaction beneath Suswa-Longonot-Olkaria. Despite active systems at Suswa[19] and Olkaria[18], neither were affected by the pulse of uplift at Longonot in 2004-2006. The minimum dimension of the source at Longonot is ~12 km, with neighboring systems at Olkaria and Suswa only 15 km and 30 km away, respectively. Here again, lateral connectivity appears limited to distances of ~10 km.

Although these periods of unrest suggest that neighboring volcanic systems in the Kenyan Rift do not interact over distances >~10 km, the prehistoric record of simultaneous eruptions and intermixing indicates that in some circumstances, at least, adjacent systems can interact. To explain this apparent dichotomy, we hypothesise that the small increases in magma supply that cause unrest produce interactions over short distances only (<~10 km); large magma volumes produced during major eruptions, in contrast, cause large, rapid stress changes which are capable of influencing magma systems over larger distances (>~25 km).

We explore this hypothesis by considering scaling relations and drawing on evidence from eruptions elsewhere. Mechanisms proposed to explain volcano-volcano and volcano-earthquake interactions include stress changes due to permanent displacements[7] or the passage of seismic waves[23], lateral hydraulic connections such as dyke intrusions[9] and a common asthenospheric magma supply[10]. Both magmatic systems and stress fields vary with time, so here we adopt a first order approach and develop simple scaling relations to estimate the characteristic volumes and length-scales associated with each mechanism.

Stress changes, σ, caused by a pressure change, ΔP, or volume change, ΔV, of a point source decay with distance as ${}^{\Delta V}/_{r^3}$ (See Methods). Values of critical stress required to propagate magma to the surface are typically assumed to be ~ 1-10MPa, but several mechanisms have been proposed by which smaller stress changes could trigger a response[23]. Although volume changes, stress thresholds and magma geometries are notoriously difficult to quantify, the observations above provide some constraints (Figure 3). Unrest signals within the Kenyan Rift suggest that to produce a response among sources at Paka (Pa-Pa) without a corresponding response at neighbouring Silali (Pa-Si) requires a stress threshold of 1-10 kPa (assuming a point source within an elastic half-space of μ = 3-30 GPa). This threshold is much lower than previously thought, and suggests that an alternative mechanism might operate within individual magmatic systems.

For large erupted volumes ($V=10^9$-$10^{11}m^3$), stresses of 1-10 MPa extend 20-50 km from the source. These distances are within the range observed for "super-eruptions" within the Taupo Volcanic Zone, New Zealand, that have recently been interpreted as multiple near-simultaneous eruptions from along-axis melt lenses[2,5]. For example, simultaneous eruptions from Rotorua and Ohakuri calderas (30 km apart) have been used to suggest disturbance of the local stress field by evacuation of one magma batch from within a continuous intermediate mush zone, which activated regional faults and triggered additional eruptions[2]. Here we find that for such large volume changes, stress changes alone could explain the eruption coupling not only at Rotorua-Ohakuri (Ro-Oh), but also at Suswa-Longonot (Su-Lo), Novarupta-Katmai (No-Ka), and Kidnappers (Kid) (Figure 3).

Simple scaling relations for direct, hydraulic connections are more difficult to estimate, as the dyke width or conduit radius will depend on the relationship between pressure gradient and viscosity, and the travel distance is limited by the velocity required to avoid cooling and solidification. For simplicity, we assume that the volume must fill a constant cross sectional area, A, leading to a linear relationship between volume, V, and distance, r ; $V = Ar$.

Sub-aerially exposed spreading centres such as Dabbahu in Afar[24], Krafla[24] and Bárðarbunga[25] in Iceland, have propagated dyke intrusions over distances of 50-100 km and typically intersect multiple volcanic centres[24]. At

Dabbahu, for example, a 60-70 km-long dyke intersected three distinct magma sources, including the shallow Gabho and Dabbahu chambers and a deeper source at Ado ‘Ale[24] (AA-Ga). These intrusions were 2-8 m wide and spanned the seismogenic layer[25,26], giving A=10^4–10^5 m^2 (Figure 3). In non-rift settings, dyke intrusions are typically limited in extent, but have been implicated in paired eruptions such as Karymsky-Academy Nauk (Ka-AN), Kamchatka in 1996, Novarupta-Katmai (No-Ka), Alaska in 1912[4].

Shallow magmatic systems shared over small distances (<~10 km) often show linked eruptions. Recent eruptions of Alu-Dalafilla, Afar in 2008[27] and Eyjafjallajökull, Iceland in 2010[14] demonstrate that small eruptions can tap multiple laterally or vertically distributed magma sources without affecting nearby volcanic systems. At Alu-Dalafilla, pre-eruptive inflation was observed under Alu only, but co-eruptive subsidence extended 10 km towards Dalafilla[27] (Al-Da). Following the eruption, similar replenishment patterns indicated the systems had become hydraulically connected. Meanwhile, the Erte Ale lava lake, which is only 30 km away, displayed no significant response (AD-EA). The Eyjafjallajökull eruption began with a flank eruption at Fimmvörðuháls, located 8 km to the east (Fi-Ey); the eruption subsequently tapped a series of sills at different depths but within <6 km horizontally, suggesting extensive vertical but limited lateral propagation[14]. The neighboring system of Katla, which is only 25 km away, is seismically active but produced no obvious response (Ey-Ka). Similarly, recent larger eruptions, such Okmok, Alaska[6] and Cordon Caulle-Puyuehue (CC-Pu), Chile[28] have tapped multiple sources over distances of <10 km.

The above examples suggest hydraulic connections may exist within shallow magma-crystal mush zones spanning ~10 km, where the thermal regime supports porous flow, or prevents conduit freezing (Figure 3). However, under extreme circumstances, interactions may occur over greater distances; at Yellowstone, interactions over 30-40 km are likely mediated by the shallow geothermal reservoir[22]. Additionally, in Hawaii, inflation-deflation patterns accompanying eruptions at the Puʻu ʻŌʻō vent are reproduced several hours later at Kīlauea's summit, suggesting pressure waves travel along a 20 km conduit maintained by a high magma flux[29].

An additional mechanism for linked unrest involves pressure transients within an asthenospheric source feeding independent mid-crustal magma storage regions; this mechanism has been invoked to explain correlations between unrest and eruptions at Vesuvius and Campi Flegrei (Ve-CF)[30], separated by ~20 km, and at Kīlauea and Mauna Loa (Ki-ML)[10], separated by ~35 km. The geodetic record shows correlated inflation pulses at both pairs; periods of more frequent eruption at Vesuvius, however, concur with episodes of subsidence at Campi Flegrei[30] while at Ki-ML, the eruption histories are anti-correlated[10]. Finally, the typical ~70 km spacing of arc volcano clusters exceeds the spatial limits of stress changes expected for the assumed dimensions of deep magma supplies, but may still respond to dynamic triggering associated with M8-9 earthquakes[16].

Overall, we might expect 1) coupled unrest and small eruptions to be common at established volcanoes with well-developed mush zones capable of supporting lateral connections or porous flow (<10 km); 2) interactions over

20-40 km when triggering volumes are large or the volcanoes share a deep source; 3) extensional tectonic stresses to facilitate large dyke intrusions (<100 km) which intersect multiple magmatic reservoirs, and 4) a simultaneous response from clusters of arc volcanoes following large subduction earthquakes (<1000 km).

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Author Contributions.

J.B. conceived the project, with help from E.R., and ordered the data. E.R. analysed the data with help from J.B. J.B. wrote the manuscript with help from E.R. and K.C.

Figure Captions

Figure 1: Sketch of the Kenyan Rift showing caldera volcanoes (yellow) and non-caldera volcanoes (pink). Boxes show two clusters of volcanoes discussed here Silali-Paka and Olkaria-Longonot-Suswa. Inset shows East African Rift with red triangles marking Holocene volcanoes and black arrowing showing the relative motion of Somalian Plate to Nubian Plate in $mmyr^{-1}$. AR: Afar Rift; MER: Main Ethiopian Rift; KR: Kenyan Rift; WB: Western Branch; MR: Malawian Rift.

Figure 2: Satellite radar images of deformation at Paka (A-E), Silali (F) and

Longonot (G-H) volcanoes in the Kenyan Rift with corresponding models (I-P). Each fringe (colour cycle) represents 2π radians of phase change corresponding to 2.8 cm of range change in the line-of-sight direction for Envisat (A-C, G, I-K, O) and 11.3 cm for ALOS (D-F, H, L-N, O). (Q-X) are schematics showing the source models for each time period (active source filled, inactive sources empty). Vertical arrows show surface displacement rates in $cmyr^{-1}$ along the satellite line-of-sight direction. Vertical labels show source depths in km.

Figure 3. Relationship between volume and distances for coupled eruptions and unrest. Values for examples and details of scaling laws are provided in Methods. Abbreviations are defined in the main text and symbols are the same as figure 1. Lines: black - stress changes decay as $\Delta V/r^3$; blue – intrusion volumes for a constant area, $V \propto r$. Regions: grey – stress exceeds 0.1-1 MPa; blue – intrusion volumes and distances for $A = 10^4$-$10^5 m^2$; orange – distances < 10 km, the estimated extent of shallow mush zones; yellow- distances <40 km, the estimated extent of asthenospheric sources.

Methods

1. InSAR

We use 18 SAR images from the European Space Agency Envisat satellite from 2003-2010 and 13 images from the Japan Aerospace Exploration Agency ALOS-1 satellite spanning 2007-2011. Interferograms were created using JPL's ROI_PAC Software and topographic corrections applied using a 90m SRTM DEM. For each of the 13 volcanoes in the Kenyan Rift, we (a) stack interferograms to increase the signal-to-noise ratio; (b) calculate time series to identify the temporal evolution of ground displacement[31]; and (c) invert for source parameters using analytical solutions for deformation within an elastic half-space caused by a point source or penny-shaped crack (representing a sill)[32]. The volcanoes are generally low relief (< 600 m) – we see no obvious influence of stratified water vapour fields on our interferograms and assume that a half-space solution is sufficient to model the deformation sources. Table 1 compares the depths and misfits to the best-fitting point and penny-shaped crack models.

As source parameters are non-unique, we test the bound of each parameter using a Monte Carlo method and quote the mean and standard deviation of a normal or log-normal distribution in Figure 3[19]. In cases where the difference in misfit between solutions is small, the difference in depth between solutions is included in the range quoted in the main text. Errors on displacements and velocities are calculated during the time series inversion using the assumption of 1 cm uncorrelated noise on each interferogram.

	Point		Penny	
	Depth (km)	Misfit (mm)	Depth (km)	Misfit (mm)
Silali (2007-2010)	**3.86**	**9.36**	4.41	9.39
Longonot (2004-2006)	3.7	7.04	**3.69**	**5.84**
Longonot (2007-2010)	**2.26**	**8.41**	0.88	15.08
Paka (Phase 1)	2.46	5.12	**1.3**	**4.93**
Paka (Phase 2)	**5.2**	**3.58**	7.98	4.4
Paka (Phase 3)	**2.57**	**4.83**	3.73	6.78

Method Table 1: Comparison between best-fitting point and penny-shaped source models for each phase. The lowest misfit solution for each is given in bold.

It is interesting to note that no deformation is seen at Olkaria, where geothermal extraction was ongoing during our survey period, because geothermal plants can cause rapid subsidence when reinjection does not keep pace with extraction[33,34]. Although localised subsidence of 2-3 $cmyr^{-1}$ is detected associated with a flower farm within the volcanic complex, we see no deformation attributable to magmatic or hydrothermal processes above a threshold of 0.3 $cmyr^{-1}$. For no ground displacement to occur during geothermal extraction, pore pressures must remain high enough to maintain the lithostatic load. At Olkaria, the reservoir rocks are basalts, trachytes and tuffs, which are relatively incompressible, particularly compared to the lake sediments associated with high rates of geothermal subsidence elsewhere[33].

The code used to generate the interferograms can be accessed at .
http://roipac.org/cgi-bin/moin.cgi.

The code used to generate the source models can be accessed at
http://www.rsmas.miami.edu/personal/famelung/geodmod/geodmod.html

2. Volume-distance relationships

2.1. Stress. The radial stress change, σ_{rr}, due to a pressure change, ΔP, of a point source in an elastic full-space is given by,

$$\sigma_{rr} = -\frac{\Delta P a^3}{r^3}$$

where *a* is the radius of the source and *r* is the distance from it[7]. This can be converted into a volume change, ΔV, using the relation

$$\Delta V = \pi \frac{\Delta P a^3}{\mu}$$

where μ is the shear modulus, giving

$$\sigma_{rr} = -\frac{\mu \Delta V}{\pi r^3}$$

The geometry of the source will affect this relationship. Numerical models show that over distances of 10 km, the lateral stress change caused by a sill-like reservoir would be significantly smaller and that a point lying in the direction of dyke opening will experience similar stress change as from a spherical reservoir while a point lying along strike of the dyke will experience almost no stress change[35].

The rheology of the surrounding material will also influence both the stress and displacement fields. In particular, long-lived magma bodies are likely to be surrounded by a thermal aureole, which is often represented by a viscoelastic shell[7]. For a source of radius R_1 within a shell of radius R_2, the radial stress is given by

$$\sigma_{rr} = -\frac{\Delta P {R_1}^3}{r^3}\left[e^{-t/t_r} + \frac{R_2^3}{R_1^3}\left(1-e^{-t/t_r}\right)\right]$$

where t_r is the characteristic relaxation time. This has the same spatial form as for the uniform elastic full space, such that stresses decay with distances as $1/r^3$. For $t << t_r$, R_1 is the effective source radius and the stress will be equivalent to that in equation [1], but when $t>>t_r$, R_2 is the effective source radius and the stress will be greater at a given distance.

Figure 3 shows the relationship between log *V* and log *r* such that relationships of the form $V \propto {}^{1}/_{r^3}$ appear as a straight line of gradient 3. For a spherical body, the intercept is given by $\log \pi\sigma/\mu$. The value of shear modulus, μ, is usually taken to be ~20-30GPa for unaltered crust, but can be as low as 3 GPa in volcanic settings[36]. The overpressure required to generate tensile deviatoric stresses sufficient to allow a dyke to form, and magma to propagate to the surface without freezing, is thought to be 10–100 MPa for silicic magmas and <1 MPa for basaltic magmas[23]. However, changes in volcanic and geothermal systems are often seen in response to much smaller stresses, such as the passage of seismic waves and tidal loading, suggesting that alternative mechanisms (e.g. rectified diffusion, advective overpressure, bubble nucleation and dislodging crystal aggregates) may play a role[23]. To account for this, we extend our stress field by an order of magnitude to cover critical stresses as low as 0.1MPa.

2.2. Lateral Transport

Simple scaling relations for direct, hydraulic connections are more difficult to estimate, as the dyke width or conduit radius will depend on the relationship between pressure gradient and fluid viscosity[37]. Dyke propagation may be arrested by cooling and solidification or by intersection with a stress barrier[37]. For simplicity, we use a scaling law based on the conservation of volume such that there must be sufficient volume, V, available to fill a constant cross sectional area, A, for a distance, r, giving $V = Ar$. This leads to a linear relationship between volume, V, and distance, r which plots as a straight line of gradient 1 in log V-log r space and the intercept is controlled by log A. We use $A=10^5$ m^2 as an upper limit, which corresponds to a 10m wide dyke cutting through a seismogenic layer 10 km thick. Plume-related dykes can be up to several hundred metres wide, but these are not considered here[38]. Cyclindrical conduits would have a much smaller cross-sectional area; for example, A = 1 m^2 corresponds to a radius of 56 cm. The length of such conduits would be limited by the balance between heat input by the magma flow and heat lost to the surrounding material. Within a semi-crystalline mush, high temperatures could support narrow conduits over greater distances, but porous media flow would be a more appropriate conceptual framework[39].

3. Paired Eruptions.

Table M2 contains details of the paired eruptions shown in Figure 3 of the main text, including references. While distances between volcanic pairs are relatively simple to estimate, volumes are taken from a range of sources, including geodetic observations, and ash and lava volumes, each of which has separate uncertainties and biases. For dyke intrusion and unrest, we use the subsurface volume change rather than the erupted volume. For geodetic observations, the accuracy of volume estimates is limited by tradeoffs between depth and volume[19], and estimates of compressibility[36]. For recent eruptions, lava areas are relatively simple to calculate but simply multiplying this by a thickness - the planimetric approach - often leads to high uncertainties and direct measurements of volume change through a topographic approach are still relatively rare[40]. Tephra volumes are measured by interpolating field-based measurements of deposit thickness, which commonly show an exponential decay with distance[24]. Volumes are converted to mass using an average deposit density.

	Volcano1	Volcano2	r km	V m^3	Volume Type	Type	
Lo-Su	Longonot	Suswa	15	7×10^5	Geodesy, Unrest	No response	*
Lo-Ol	Longonot	Olkaria	30	7×10^5	Geodesy, Unrest	No response	*
Pa-Pa	Paka	Paka	10	1.1×10^6	Geodesy, Unrest	Geodetic Response	*
Pa-Si	Paka	Silali	10	2×10^6	Geodesy, Unrest	No Response	*
Al-Da	Alu	Dalafilla	10	2.5×10^7	Geodesy, co-erupt.	Eruption	[27]
AD-EA	Alu-Dalafilla	Erte Ale	30	2.5×10^7	Geodesy, co-erupt.	No response.	[27]
Fi-Ey	Fimmvörðuháls	Eyjafjallajökull	8	2×10^7	Lava vol.	Eruption	[41]
Ey-Ka	Eyjafjallajökull	Katla	25	1.8×10^8	Ash vol.	Possible jökullhaup	[41]
No-Ka	Novarupta	Katmai	10	6.9×10^9	Tephra volume	Caldera Collapse	[9]
Ve-CF	Vesuvius	Campi Flegrei	20	Correlated inflation; Bradyseism at CF reponse to VEI5 at Vesuvius.			[30]
Pu-Ki	Puu'Oo	Kilaeua	20	3×10^5	Geodesy	Hydraulic Connection	[29,42]
Ki-ML	Kilauea	Mauna Loa	35	Correlated inflation and anti-correlated eruption history.			[10]
AA-Da	Ado Ale	Gabho	35	2.5×10^9	Geodesy, Intruded Vol.	Hydraulic Connection	[26]
Ba-Ho	Bárðarbunga	Holuhraun	50	1.6×10^9	Lava volume	Hydraulic Connection	[10]
CC-Pu	Puyehue–Cordón Caulle	Puyehue–Cordón Caulle	10	4×10^8	Geodesy, co-eruptive subsidence	Hydraulic Connection	[28]
Ka-AN	Karymsky	Academy Nauk	9	4×10^7	Tephra	Hydraulic Connection	[4]
Su-Lo	Suswa-	Longonot	15	1.1×10^{10}	Tephra	Interbedded Tephra	[1]
Ro-Oh	Rotorua	Ohakuri	30	2.5×10^{11}	Tephra	Interbedded Tephra	[2]
Kid	Kidnappers	Kidnappers	30	4×10^{11}	Tephra	Interbedded. Tephra	[5]

*Methods Table 2. Values of volume and distance used for Figure 3 including measurement types and references. * This study.*

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