EARTHQUAKE–VOLCANO INTERACTIONS

New measurements, statistical analyses, and models support the conjecture that a large earthquake can trigger subsequent volcanic eruptions over surprisingly long distance and time scales.

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"Is this eruption related to that earthquake?" That's a question commonly asked of volcanologists and seismologists when a newsworthy volcanic eruption develops within days or weeks following a newsworthy earthquake somewhere else in the world. Accurate answers are important: Continued scientific advances in understanding tectonic and magmatic processes and their interaction are fundamental to providing reliable information on the nature of hazards that large earthquakes and volcanic eruptions pose to the many population centers at risk around the globe.

Traditional answers to the question are of the form, "Not likely. The stress changes from that earthquake are too small—smaller than stresses associated with the solid-earth tides (about 0.001 MPa)—to influence processes at this volcano. Furthermore, the temporal patterns of volcanic eruptions and earthquakes appear to be stochastic, and statistically independent sequences occasionally produce events that are nearly coincident in time." Such answers are based largely on calculations that model earthquakes as shear dislocation—that is, transverse displacements across faults—and Earth as a linearly elastic medium, and on estimates of statistical significance. Dislocation theory and linear elasticity are particularly appealing because of their simplicity and their remarkable power in explaining an earthquake's complex seismograms in terms of the detailed internal structure of Earth and such properties of the earthquake itself as the stress change and the dimensions of the fault where the shear dislocation occurs.

When the question is viewed over sufficiently long time and distance scales—over centuries to millions of years and over thousands of kilometers—the answer is decidedly affirmative. Earthquakes and volcanoes are inexorably linked through plate tectonics, with the major seismic and volcanic belts concentrated along the boundaries between plates. An affirmative answer at these scales, however, does not address the relative timing or possible causal links between individual events.

The answer is also affirmative at the other extreme. Large earthquakes are certainly capable of triggering eruptions within a matter of minutes or days at nearby volcanoes—those within a characteristic distance set by the length along the fault plane over which the slip occurs. For example, on 29 November 1975, an eruption at the summit of Kilauea in Hawaii began half an hour after the magnitude-7.5 Kalapana earthquake that occurred beneath the south flank of the volcano. And the Cordon Caulle volcano in central Chile erupted two days after the great magnitude-9.5 earthquake of 22 May 1960 off the coast of Chile.[1]

In this article, we focus on emerging evidence for earthquake–volcano interactions at intermediate distances, for which the answer remains, "well, maybe." Examples and plausible models are accumulating for interactions in which large earthquakes, of magnitude 6.5 or greater, are followed—mere minutes to a few hundred years later—by episodes of volcanic unrest or eruptions located from one to many fault lengths away. (Fault lengths of magnitude-6.5 to magnitude-9.0 earthquakes range from about 20 to more than 1000 km, respectively.) Research on earthquake–volcano interactions is in its infancy, and limitations on available data constrain our present ability to decide between competing models for interaction processes. Progress will depend to a large degree on the whims of nature for "repeat experiments" at suitably instrumented volcanoes.

Volcanic eruptions and unrest

As described in the box on page 43, volcanic eruptions are the culmination of a complex ensemble of processes. An eruption is preceded by the buildup of pressure in the chamber in which magma has pooled within the crust. As the magma in the chamber cools, crystals form and volatile compounds trapped within the remaining melt exsolve to form bubbles. When the pressure in the chamber approaches a critical value, the volatile-rich magma at the top of the chamber will begin breaching the chamber's roof with sheetlike intrusions called dikes, some of which may reach the surface in a volcanic eruption. The events leading to an eruption are manifested as episodes of volcanic unrest, characterized by swarms of earthquakes, ground deformation, and excitation of the hydrothermal system as the magma heats groundwater in the upper part of the crust. Multiple episodes of volcanic unrest of increasing

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frequency and intensity typically precede eruptions. Unrest episodes outnumber eruptions by as much as an order of magnitude.[2]

Volcanic systems that have evolved to a critical state are most susceptible to small perturbations produced by a distant earthquake. For that reason, seismologists and volcanologists are interested not only in the triggering of volcanic eruptions but also in the triggering of episodes of volcanic unrest, which may be early symptoms of an impending volcanic eruption.

Interaction modes

Interest in interactions between earthquakes has grown dramatically over the past two decades,[3] and the associated research provides the foundation for studies of earthquake–volcano interactions. Models for earthquake–earthquake interactions are based on three stress transfer modes: (1) static stress changes, that is, the difference in the stress field from just before an earthquake to shortly after the seismic waves have decayed; (2) quasi-static stress changes associated with slow viscous relaxation of the lower crust and upper mantle beneath the epicenter of a large earthquake; and (3) dynamic stresses induced by the seismic waves from a large earthquake. (For more on earthquakes, see the article by Hiroo Kanamori and Emily Brodsky in PHYSICS TODAY, June 2001, page 34.)

Static stress changes decay rapidly, as $1/r^3$, where r is the distance from the earthquake epicenter. Interactions by this mode are generally limited to distances of just a few fault lengths from the initial earthquake and lag times of months to years. Quasi-static stress changes decay more slowly with distance—as roughly $1/r^2$ to $1/r$—and involve lag times of years to centuries; the delays are governed by the effective viscosity of the lower crust and upper mantle. Both interaction modes are based on incremental stress changes: The first earthquake nudges the stress field in the vicinity of the second fault closer to its threshold for slip.[3]

Research on long-range interactions by dynamic stresses was spurred by the remarkable onset of earthquake activity across much of the western US immediately following the magnitude-7.2 Landers earthquake of 28 June 1992 in the Mojave Desert of southern California (see PHYSICS TODAY, September 1993, page 17).[4] Dynamic stresses remotely triggered earthquakes, with lag times of minutes to hours, at distances up to 1200 km (about 10 fault lengths) from the Landers epicenter. The amplitudes of dynamic stresses generally decay with distance as roughly $1/r^2$ to $1/r$ for seismic body waves and surface waves, respectively. The largest amplitudes usually propagate with the shear wave (a body wave with transverse polarization) and the surface waves in the upper 20 km of the crust. The physics behind earthquakes that are remotely triggered by dynamic stresses remains an area of active research.

The distinction between static and dynamic stress transfer modes breaks down at distances less than a fault length from a given earthquake. That region is the "nearfield" and the domain of aftershocks, a sequence of smaller earthquakes that begin immediately following the initial (and usually largest) earthquake and gradually decay in frequency and magnitude over time. Aftershock sequences are the most familiar form of earthquake–earthquake interaction.

Static stress triggering

Evidence continues to mount that static stress changes from earthquakes can trigger volcanic eruptions at intermediate distances (up to a few hundred km) and delays of months to years. Examples include the last major eruption of Japan's Mount Fuji in 1707, about one month after an offshore earthquake of magnitude 8.2, and the 1991 eruption of Mount Pinatubo in the Philippines following the magnitude-7.7 earthquake of 1990.[5] Both volcanoes were located about 100 km from the epicenters.

Explanations invoked for earthquake–volcano interactions due to static stress changes are based on pressure changes in a magma body induced by the isotropic, compressional component of the stress field in the vicinity of the volcano. Increased compressional stress in the crust surrounding a magma chamber that is close to its critical state may squeeze magma upward;[5] a decrease in compressional stress can promote additional melting, the formation of bubbles as volatiles exsolve, and the "unclamping" of conduits above the magma chamber. The observed delays of months to years between the earthquake and the eruption are poorly understood, although diffusion of interstitial crustal fluid, mostly water, in response to static stress changes is thought to play an important role.

Interactions mediated by static stress changes between nearby earthquakes and volcanoes appear to be bidirectional. Jim Savage and Wayne Thatcher of the US Geological Survey describe cases in California and Japan in which magma chamber inflation and pressurization may have triggered moderate earthquakes at regional distances (a few hundred km) with delays of a few years. Concetta Nostro of Italy's National Institute of Geophysics and Volcanology (INGV), working with Ross Stein of the USGS and others, examined both the triggering of volcanic events by earthquakes and the converse for several earthquake–volcano pairs in Italy over the past 600 years. They inferred a correlation between the times of earthquakes and the times of major eruptions of Mount Vesuvius. Pairs of events usually occurred within 10 years or less for cases in which, according to calculations, pressure changes in the magma chamber nudged the stress field near a neighboring fault closer to the condition for brittle failure.

Viscoelastic relaxation

In most tectonically active regions, the continental lithosphere is well approximated as a stratified viscoelastic medium in which the upper 15–20 km of the crust deforms elastically and, because of higher temperatures, the underlying lower crust and mantle deform plastically; the transition corresponds to the 350–400°C isotherm.[6] Earthquakes are confined to the upper crust. After an earthquake, the shear stresses that were induced in the underlying plastic zone will gradually relax. The elastic crust, coupled to the plastic crust below, continues to deform quasi-statically with time until the viscoelastic system reaches a new state of equilibrium. Recent studies[7] have invoked this process for earthquake–earthquake triggering over periods of 7–50 years and distances of 30–150 km.

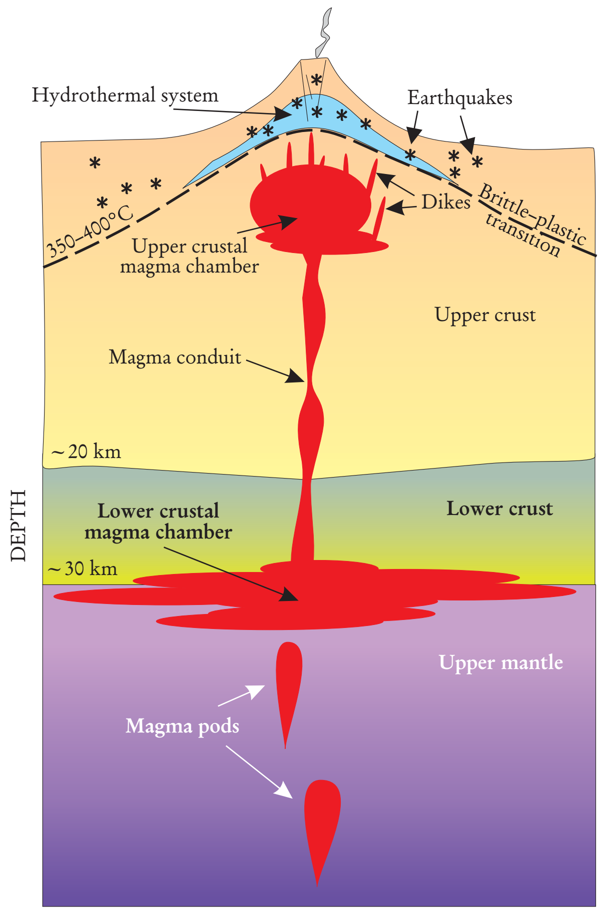
Does this process of postearthquake stress diffusion lead to earthquake–volcano interactions as well? In a statistical analysis of the occurrences of large explosive eruptions following large earthquakes, Warner Marzocchi of INGV has found evidence for a correlation of eruptions following earthquakes by 0–5 years and 30–35 years at distances up to 1000 km; he attributes the correlation to stress diffusion. Potentially favorable settings for such an interaction mode involve chains or arcs of volcanoes that lie inland from convergent boundaries between major tectonic plates, such as the Cascadia subduction zone shown in figure 1a.

Convergent plate boundaries are capable of producing

Anatomy of a Continental Volcano

The continental crust is typically 30–40 km thick. As shown in the figure, partial melting of the upper mantle at depths of about 100 km produces basaltic magma (45–52% SiO_2 by weight) that rises buoyantly into the base of the crust, where it may pool in lower-crustal magma chambers and reside for extended periods. That magma is hot enough to partially melt surrounding rocks of the lower crust, which are also of basaltic composition. The initial products of this melting are rhyolitic—having a composition similar to that of granite with 70–80% SiO_2. Further melting yields magmas of intermediate composition, from 70% down to about 52% SiO_2. At the same time, heat loss from the original basaltic magma induces its partial crystallization. Residual melt fractions become progressively richer in SiO_2. Magmas of various compositions move upward through the crust in dikelike cracks or strawlike conduits to accumulate in mid- or upper-crustal magma chambers.

Magmas also contain volatile components (mainly water, carbon dioxide, and sulfur dioxide) that are dissolved in silicate melt at low concentrations and high pressures (deep in the crust) and exsolve to form bubbles at higher concentrations and lower ***OK?***pressures (in the shallow crust). Formation and rise of bubbles pressurize magma chambers. These pressures and forceful ascent of magma will induce small earthquakes in the brittle upper 10–15 km of the crust. The magma may also inflate slightly, which will cause localized distension of Earth's surface. When the pressure approaches a critical value, magma may break through to the surface and a volcanic eruption will begin. Eruptions of basaltic magma are typically effusive and produce fluid lava flows as in Hawaii. Silica-rich magmas are cooler and up to five orders of magnitude more viscous, and their eruptions are typically explosive.



so-called megathrust earthquakes, which have magnitudes of about 9 and involve meters of slip along 1000-km stretches of the plate boundary. Since the slip also extends down 50–60 km through the elastic crust to the underlying plastic crust, a megathrust event approximates a two-dimensional stress source spanning dozens of volcanic centers within the adjacent volcanic arc. Smaller, more frequent earthquakes with magnitudes of 8 or less, such as have been documented in Japan for several centuries, involve fault lengths of at most a few hundred km. Such earthquakes thus behave more nearly as point sources, when viewed from individual volcanoes along the adjacent volcanic arc. The associated stress changes are therefore smaller and decay more rapidly with distance than those for mega-earthquakes.

Of the few well-documented examples of magnitude-9 earthquakes, data for the Cascade volcanic arc show a tantalizing relation between the Cascadia megathrust earthquake[8] in 1700 and an apparent peak in eruption frequency[9] during the 1800s (see figure 1b). The behavior of the region can be parameterized by the volumetric strain $\Delta(t)$. Stress diffusion following the earthquake increases $\Delta(t)$, whereas the reloading of stress along the fault will tend to return $\Delta(t)$ to its initial value. Figure 1b shows the theoretical variation of $\Delta(t)$ at a depth of 5 km beneath the Cascade volcanic arc. Our $\Delta(t)$ calculation, which assumes a 600-year cycle of strain accumulation between mega-earthquakes, is based on a model of viscoelastic coupling similar to the more detailed model of Kelin Wang, of the Geological Survey of Canada, and his colleagues.

In the Cascadia example, the peak in $\Delta(t)$ leads the peak in eruption rate by about 50 years, which we speculate may be linked to an activation process in which vertical dike propagation from deep magma chambers is enhanced by a local increase in volumetric strain and a corresponding reduction in isotropic, compressional stress.

Volcanoes in Cook Inlet, Alaska, show an apparent acceleration in rate from 1965 through at least 1996 following the magnitude-9.2 Alaskan earthquake[10] of 1964, but the trend may only reflect improved eruption reporting. Evidence has yet to emerge for increased eruption rates in the volcanic arcs behind the magnitude-9.5 earthquake off the coast of Chile in 1960 or the magnitude-9.0 earthquake off Kamchatka in 1952. Perhaps, as suggested by the Cascadia data, 50–60 years is a minimum time interval for postearthquake stress diffusion to significantly influence eruption rates in those volcanic arcs.

The examples illustrate the precarious footing of our current understanding of earthquake–volcano interactions through quasi-static stress triggering. Although the Cascadia result is tantalizing, we are dealing with only a single cycle, and thus have no basis for assessing its statistical significance. Furthermore, the written record of Cascade eruptions is incomplete before the large influx of European settlers in the late 1800s. This limitation underscores the challenge endemic with the Earth sciences: extending a uniform record back in time. Improving our understanding will require advances in techniques for identifying and dating

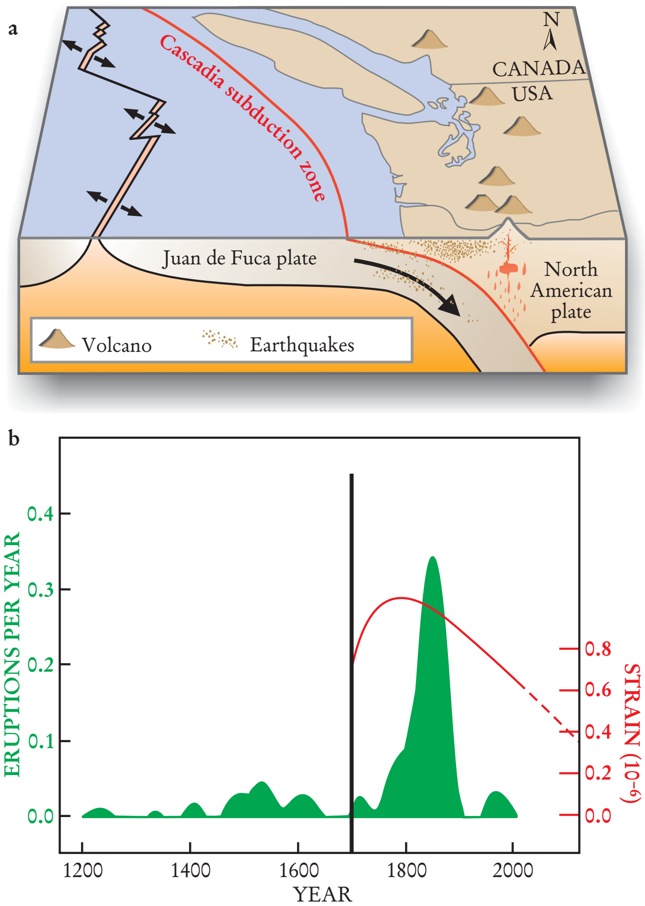


FIGURE 1. QUASI-STATIC STRESS CHANGES are one mechanism that might connect eruptions to earthquakes. **(a)** As the Juan de Fuca and North American plates converge, strains gradually accumulate along the Cascadia subduction zone. The accumulated strains are released every few hundred years by the abrupt thrusting of the oceanic plate beneath the overriding continental plate to produce a magnitude-9 "megathrust" earthquake.[18] Following the megathrust earthquake, changes in stress quasi-statically diffuse inland toward nearby volcanoes. Background earthquake activity (small dots) occurs during the interval between megathrust earthquakes. **(b)** The composite annual eruption rate of Cascades volcanoes[9] (smoothed using a uniform moving window) shows a marked increase following the magnitude-9 earthquake on the Cascadia subduction zone in 1700.[8] The red curve shows calculations of the volumetric strain below the Cascade volcanoes that followed the earthquake. The peak strain precedes the peak eruption rate by about 50 years.

prehistoric earthquakes and volcanic eruptions, and the careful scrutiny of the eruption histories for volcanic arcs adjacent to convergent plate boundaries with well-documented magnitude-9 earthquakes.

Dynamic triggering

After the 1992 Landers earthquake, answers to the question of earthquake–volcano interactions became distinctly more interesting and complex.[4] The abrupt onset (within minutes to hours) of seismicity at some 14 sites at distances as great as 1200 km from the main shock epicenter put the issue of statistical significance to rest. It also underscored the limitations of linear elasticity theory in understanding Earth processes, because that theory yields stress changes far below those thought capable of triggering the remote seismicity. Many remotely triggered sites were located in areas of volcanic and geothermal activity that were geologically young (less than 10 000 years old).

The Landers triggering immediately raised the question of other such occurrences. The ensuing search for more examples of dynamic triggering uncovered a number of candidates from around the globe. Joan Gomberg and Scott Davis of the USGS documented eight instances of abrupt seismicity increases at the Geysers geothermal field in northern California following earthquakes with magnitudes from 6.5 to 7.9 and at distances from 200 to more than 4000 km. Examining global catalogs for large earthquakes (above magnitude 7) and explosive volcanic eruptions (with erupted volumes above 10^6 m^3), Alan Linde and Selwyn Sacks of the Carnegie Institution of Washington found that there are many more eruptions within a day or two following large earthquakes than would be expected without dynamic triggering. Surprisingly, a near repeat of the Landers "experiment" occurred on 16 October 1999, when the magnitude-7.1 Hector Mine earthquake ruptured a fault parallel with and just 35 km east of the fault that produced the Landers earthquake seven years earlier. The Hector Mine earthquake was followed by an abrupt increase in seismicity at distances of 110–270 km, principally at sites within the Salton Trough south of the epicenter.[11]

Long Valley caldera (figure 2), a young volcanic system located in eastern California some 420 km north-northwest of the Landers and Hector Mines epicenters, responded to both earthquakes with local strain transients that were an order of magnitude larger than can be accounted for by the cumulative slip from the triggered local earthquakes. Of all the recognized instances of remote triggering, however, only the Long Valley site had high-resolution strain instrumentation capable of tracking continuous changes in ground deformation (see figure 2b). Whether similar strain transients occur in other episodes of remote triggering is an important open question.

Several models have been proposed to explain the phenomenon of dynamic triggering by distant earthquakes. We focus here on models in which fluids are active in the triggering process, because they play an obvious role in geothermal and magmatic systems and because of the growing recognition of their important role in active Earth processes in general. Indeed, that the groundwater level often responds to distant earthquakes is well established.[12] With this focus on fluids, however, we omit an intriguing class of explanations based on nonlinear friction models explored by Gomberg and her USGS colleagues.[13]

▶ **Advective overpressure and rectified diffusion.** These two models consider bubbles in a two-phase fluid and apply to both hydrothermal and magmatic systems. The models are appealing because bubbles are important in eruption dynamics and in certain types of seismic sources commonly detected in active volcanic areas.[14] Bubbles are formed as volatile components in the magma exsolve. The most common volatiles in magma, in order of decreasing abundance, are water, carbon dioxide, and sulfur dioxide; of those, carbon dioxide is the least soluble in magma. Carbon dioxide is also relatively insoluble in water and, together with steam, may be the dominant

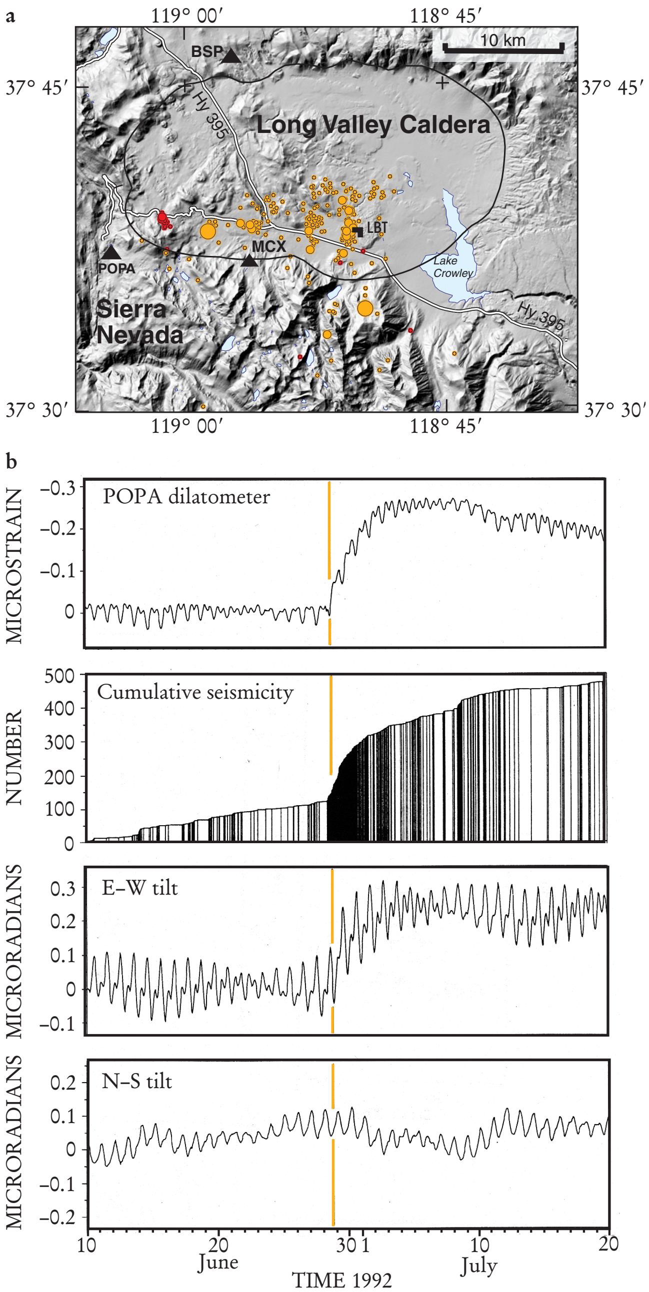


FIGURE 2. SEISMICITY AND VOLCANIC UNREST were triggered in Long Valley caldera in eastern California by the magnitude-7.3 Landers earthquake of 28 June 1992[4] and the magnitude-7.1 Hector Mine earthquake of 16 October 1999.[11] **(a)** Earthquakes triggered by the Landers and Hector Mine earthquakes are indicated by orange and red circles, respectively, whose sizes scale with increasing magnitude from 1 to 3. Borehole dilatometers that measure volumetric strain in the crust are indicated by the solid triangles and the long-base tiltmeter (LBT) by the inverted L. **(b)** Time histories from 10 June to 20 July 1992 show the volumetric strain recorded by the POPA dilatometer, cumulative number of earthquakes, and the east–west and north–south components of the LBT. Negative strain is compressional; positive tilt is down to the east and north. The high-frequency wiggles on the dilatometer and tiltmeter records are solid-earth tides. The time of the Landers earthquake is marked by the orange line. Similar but somewhat smaller effects were seen after the Hector Mine earthquake.

gaseous component in a hydrothermal system.

Advective overpressure models examine the pressure increase as bubbles rise in a confined fluid. In the highly idealized case of an incompressible fluid confined in a rigid, sealed container, the pressure increases as $\rho g h$ as a bubble of perfect gas rises a distance h, where ρ is fluid density and g is the acceleration of gravity; in physically realistic systems with finite compressibility, the pressure gain will be substantially smaller.[15] Taking reasonable values of compressibility for rocks and magma, Linde and his colleagues argue that the pressure increase in a compliant crust should be roughly one-third to one-fourth that of the idealized case. They conclude that ***OK?*** bubbles rising in a chamber of basaltic magma with a diameter of roughly 1 km provide a sufficient pressure gain—7 to 10 MPa—to explain the deformation observed in Long Valley caldera following the Landers earthquake. In basaltic magma, which has a viscosity of about 100 Pa s, bubbles of radius 5 mm should take about four days to rise 1 km or so; that time frame is consistent with the growth time of the deformation transient in Long Valley caldera. Linde and colleagues propose that the seismic waves from distant earthquakes can trigger an episode of advective overpressure by dislodging bubbles that are held by surface tension to the floor and walls of a magma chamber.

Rectified diffusion results when a gas-saturated fluid with pre-existing bubbles is subjected to pressure oscillations induced by seismic waves from a distant earthquake. The surface area of a bubble is larger during the dilatational phase of the pressure wave, when the volatiles exsolve from the fluid into the bubble, than during the compressional phase, when the volatiles dissolve back into the fluid. Thus there is a net flux of gas into the bubble and a corresponding increase in pressure over multiple cycles of the seismic wave. The efficiency of this process depends, among other things, on bubble size (smaller bubbles favor a more rapid pressure rise), the concentration and diffusivity of the gas dissolved in the liquid adjacent to the bubble, and the frequency of the waves. Brodsky (now at UCLA), working with Kanamori and Brad Sturtevant at Caltech, calculated that, through rectified diffusion in submillimeter-sized bubbles, the strong seismic waves from the Landers earthquake could produce a pressure increase of about 0.01 MPa in an H_2O–CO_2 hydrothermal system pervading the seismogenic crust—the upper, brittle section of the crust where earthquakes nucleate—in Long Valley caldera or the Geysers geothermal system. Such a pressure increase may be sufficient to trigger earthquakes by reducing fault strength but is too small to account for a deformation transient of the sort observed in Long Valley caldera following the Landers earthquake.

Quite possibly, rectified diffusion may work in tandem with advective overpressure by increasing bubble size, thus increasing buoyancy and helping to dislodge bubbles that adhere to rock surfaces.

▶ **Relaxing magma body.** As it evolves, a magma chamber in the crust can reach a partially crystallized state either by the gradual cooling of a previously molten volume or by the partial melting of a previously solid volume. The resulting interconnected solid matrix will sustain stress differences that accumulate in the region. Strong seismic waves from a distant earthquake might disrupt the integrity of the solid matrix and release any accumulated

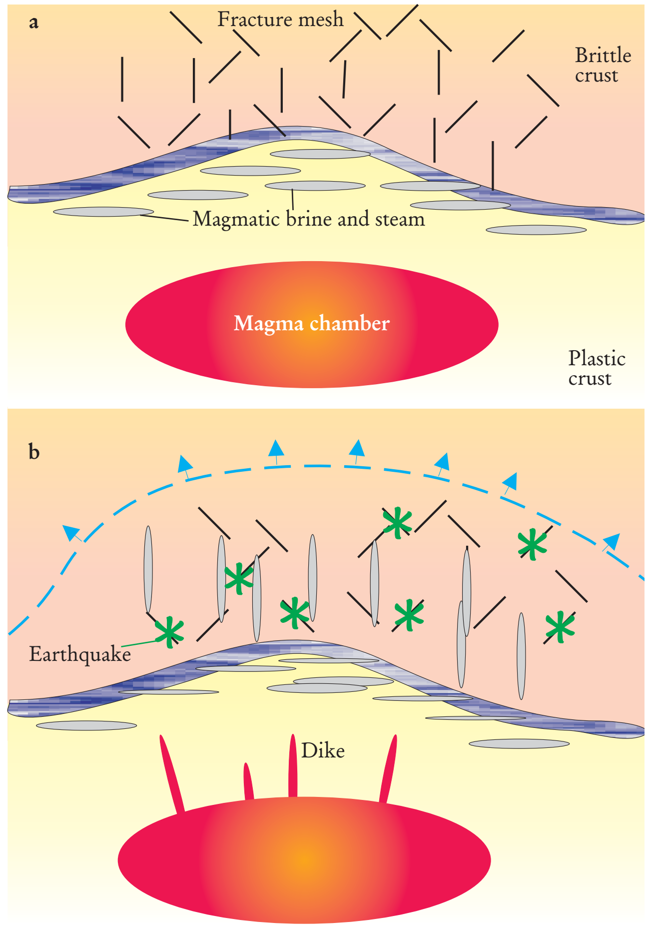


FIGURE 3. HYDRAULIC SURGES are the basis of one model for dynamic triggering. **(a)** This cross section through the base of a hypothetical geothermal system shows the brittle crust (orange) and the underlying plastic crust (yellow). Hydrothermal fluids circulate by thermal convection through a fracture mesh in the brittle crust; the fracture permeability is maintained by recurring earthquakes. The hydrothermal system is heated from below by conductive heat flow across the impermeable, brittle–plastic transition zone (purple). The transition from brittle to plastic deformation coincides with the 350–400°C isotherm in silica-containing rocks. Below this transition zone, rocks deform plastically, which reduces the permeability effectively to zero. Volatile magmatic fluids released by the underlying magma chamber (red) are trapped in isolated pods (gray) under high pressure. **(b)** Large dynamic strains accompanying the seismic waves from a distant earthquake disrupt the brittle–plastic transition, allowing fluids within this zone access to the fracture mesh in the overlying brittle crust. The resulting pressure surge (blue) propagates through the fracture mesh, triggering earthquakes (green) and deforming the brittle crust in an episode of volcanic unrest.

stress differences, thereby deforming the surrounding crust and triggering local earthquakes. To first order, this process would produce an observable surface deformation that would evolve exponentially with a time constant $\tau \approx \gamma/k$, where γ is the magma's effective viscosity and k is the elastic stiffness of the surrounding crust. If one makes the reasonable assumption of a subcrustal basaltic magma chamber with an effective viscosity of about 10^{15} Pa s (roughly four orders of magnitude below that for the surrounding mantle) and a depth of around 60 km, the relaxing magma body model can match the Long Valley caldera deformation transient that followed the Landers earthquake.[4]

▶ **Sinking crystal plumes.** Relatively dense, loosely held masses of crystals can accumulate on the ceiling and sides of a crystallizing magma body and might be dislodged by passing seismic waves. As crystal-rich plumes sink, less dense and crystal-poor magma is forced upward, producing convection. Magma that is forced upward will contain volatiles, and the ascent can trigger bubble formation. Further ascent of the bubbles will induce advective overpressure. Calculations specific to dynamic triggering have yet to be carried out, but some estimates can be extracted from the high rates of sulfur dioxide flux measured at nonerupting volcanoes. Those measurements suggest that magma can convect at a rate of 0.01 to 10 cm/s. At 1 cm/s, gas-rich magma in a preexisting conduit can be pushed upward from subvolcanic storage to eruption in less than a week.

▶ **Hydraulic surge.** In this model, the seismogenic crust is separated from the underlying plastic crust and an embedded magma body by an impermeable transition zone,[16] as illustrated in figure 3a. A hydraulic surge occurs when the large dynamic strains associated with seismic waves from a large, distant earthquake disrupt the impermeable seal. Fluids once trapped under high pressure in the plastic crust can then escape into the overlying hydrothermal system (figure 3b). This process may culminate in a steam blast if water at shallow depths is heated to its flash point. A magmatic eruption may follow if the reduced confining pressure on the magma starts a runaway process of bubble formation, ascent, and growth.[14]

Each of these models requires time for the respective systems to return to a near-critical state after a triggering episode. The recharge time may vary from days to weeks for the bubble-based models in a hydrothermal system and from decades to centuries for the relaxing magma body model. In principle, volcanic unrest triggered by each of these interaction modes has the potential for evolving to a volcanic eruption, provided the associated magmatic system is sufficiently close to a critical state.

So what's the answer?

The accumulating evidence indicates that, under the right circumstances, stress changes associated with a large earthquake are indeed capable of triggering volcanic unrest or an eruption over a wide range of distances and times. This connection, however, only increases the challenge in providing informative answers to the question of whether a given eruption is related to a given earthquake. The challenge begins with establishing statistical significance. The global rates for earthquakes of magnitude 6.5 and higher and for reported volcanic eruptions are comparable, each averaging 50–60 per year. Reported episodes of volcanic unrest number several hundred per year. Such high numbers of events provide fertile ground for the uncritical enthusiast looking for physical links where none may exist. The problem is compounded by the woeful lack of a global archive for instrumental data on volcano unrest and the incomplete record of prehistoric earthquakes and volcanic eruptions. Archival efforts such as that being developed under the World Organization of Volcano Observatories[17] are attempting to redress the first limitation; continued careful research in extracting the locations, sizes, and dates of prehistoric earthquakes and volcanic eruptions from the geologic record is needed for the second.

For those potential earthquake–volcano interactions that stand out statistically or by otherwise especially compelling data, the challenge lies in determining the physics and chemistry behind the triggering process. Advances in understanding the triggering process will, in turn, lead to a better understanding of the magmatic system and tectonic–magmatic interactions. Currently, however, even the data available for the Landers triggering, the most compelling instance of volcanic unrest triggered by a distant earthquake, are not adequate to distinguish among competing models for the triggering process. A more definitive answer must thus await the response to future large earthquakes of multiple volcanic systems that are well-monitored with multi-parameter instrumentation.

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