

# Seismic Triggering of Eruptions in the Far Field: Volcanoes and Geysers

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## Key Words

rectified diffusion, advective overpressure, magma chamber convection, liquefaction, mud volcanoes, triggered seismicity

## Abstract

Approximately 0.4% of explosive volcanic eruptions occur within a few days of large, distant earthquakes. This many “triggered” eruptions is much greater than expected by chance. Several mechanisms have been proposed to explain triggering through changes in magma overpressure, including the growth of bubbles, the advection of large pressures by rising bubbles, and overturn of magma chambers. Alternatively, triggered eruptions may occur through failure of rocks surrounding stored magma. All these mechanisms require a process that enhances small static stress changes caused by earthquakes or that can convert (the larger) transient, dynamic strains into permanent changes in pressure. All proposed processes, in addition to viscoelastic relaxation of stresses, can result in delayed triggering of eruptions, although quantifying the connection between earthquakes and delayed, triggered eruptions is much more challenging. Mud volcanoes and geysers also respond to distant earthquakes. Mud volcanoes that discharge mud from depths greater than many hundreds of meters may be triggered by liquefaction caused by shaking, and may thus be similar to small mud volcanoes that originate within a few meters of the surface. Changes in permeability of the matrix surrounding main geyser conduits, by opening or creating new fractures, may explain the observed changes in their eruption frequency.

## 1. INTRODUCTION

Catastrophic geological events are rare, and when two very different ones occur closely spaced in time, it is natural to wonder whether there is a causal relationship. For the case of large earthquakes and volcanic eruptions, a causal relationship might be expected given that an eruption requires a change in stress, and earthquakes cause rapid and sometimes large changes in stress. Moreover, a spatial relationship exists: Most explosive eruptions and large earthquakes occur at subduction zones.

Because an eruption requires deformation to permit the movement of magma to the surface, local earthquakes within or below the volcanic edifice should be expected prior to and during the eruption. For example, Kilauea, Hawaii, erupted in November 1975 within one half hour of a magnitude 7.2 earthquake and large mass movements on the flank of the volcano (Lipman et al. 1985). How volcanic eruptions can be triggered by much more distant earthquakes is less obvious.

Several volcanoes have, nevertheless, erupted within days of large, distant earthquakes many hundreds of kilometers away. Notable examples include Cordón Caulle, Chile, which erupted 38 h after the magnitude 9.5 1960 Chilean earthquake 240 km away (Lara et al. 2004), and Minchinmavida and Cerro Yanteles, which erupted within a day of a magnitude 8.1 1835 earthquake between 600 and 800 km away (Darwin 1840). These events are suggestive of a connection, but quantifying the relationship is not straightforward and only recently has been substantiated for distant triggering (Linde & Sacks 1998). Many large earthquakes have no immediate effect on volcanoes. For example, the December 26, 2004, magnitude 9.3 Sumatra earthquake occurred in the most volcanically active region of the world and the only new eruption within days was a mud volcano on Andaman Island. Manam volcano, which had been active, showed no change and, importantly, no increase in activity within many days of the earthquake. However, within two days of a magnitude 6.7 aftershock on April 12, 2005, three Indonesian volcanoes, Talang, Krakatoa, and Tangkubanparahu, all showed signs of unrest and small eruptions. Understanding these types of distant interactions—real or not—is the subject of this review.

Other processes and events external to volcanoes besides earthquakes are known to modulate volcanic activity. At short timescales, volcanic eruptions may be triggered by Earth tides (e.g., Johnston & Mauk 1972, Sparks 1981), although Mason et al. (2004) contest these claims. Other short-term phenomena that have been proposed to influence eruptions include daily variations in atmospheric pressure and temperature (e.g., Neuberg 2000). Over timescales greater than hundreds of years, volcanism may be influenced by changes in sea level (e.g., McGuire et al. 1997) and by ice loading (e.g., Jellinek et al. 2004). By comparison with all these external influences, earthquakes cause very rapid changes in stress, although the magnitude of the stress changes may be similar.

Some of the processes invoked to explain triggering of eruptions involve magma movement, the large-scale overturn of magma in magma chambers, or the relaxation and diffusion of stresses induced by the earthquakes. These processes may take many days to many decades, so if unrest at a volcano is initiated by an earthquake, the actual surface expression in the form of an eruption will be delayed. One example

of a triggered, delayed eruption might be the December 1707 eruption of Mt. Fuji, Japan, 50 days after a magnitude 8.2 earthquake (Schminke 2004). Another might be the 1991 eruption of Pinatubo 11 months after the magnitude 7.8 Luzon earthquake (Bautista et al. 1996). Barren Island, India, which had not erupted for a decade, began erupting on May 28, 2005, five months after the 2004 Sumatra earthquake. For delayed responses, identifying triggered eruptions is especially problematic, and for this reason we emphasize in this review eruptions within days of the earthquake.

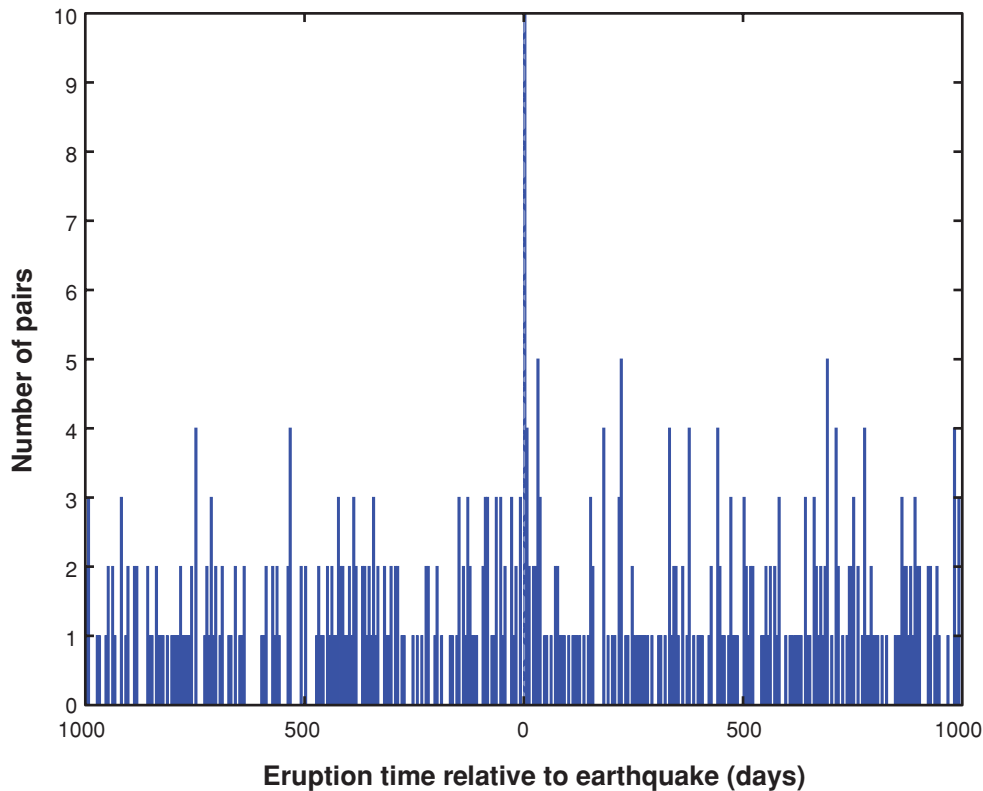
Magmatic systems and processes are complicated, and because of their large size are difficult to characterize and monitor. We thus consider other erupting systems, specifically mud volcanoes and geysers, that in principle are simpler and involve a smaller number of interacting processes

## 2. IS THERE A CORRELATION?

Linde & Sacks (1988) examined the historical record of large earthquakes and explosive eruptions, and they concluded that eruptions occur in the vicinity of a large earthquake more often than would be expected by chance. In **Figure 1** we reproduce the analysis of Linde & Sacks (1988) with an updated catalog of earthquakes (Dunbar et al. 1992; Advanced National Seismic System catalog) and volcanic eruptions until the end of 2004 (Siebert & Simkin 2002; <http://www.volcano.si.edu/world/>); in **Table 1** we list eruptions that occurred within five days of large earthquakes. In **Figure 1** and **Table 1** we consider only eruptions with  $VEI \geq 2$  and located within 800 km of the (inferred) earthquake epicenter. The VEI (volcanic explosivity index) is a measure of the size of an eruption and is based primarily on the volume erupted (Newhall & Self 1982). A VEI of 2 corresponds to a moderate explosive eruption that produces  $O(10^7)$  m<sup>3</sup> of tephra and eruption columns with heights >1 km. The conclusion that large earthquakes can trigger explosive eruptions with a short time lag (less than a few days) is unchanged for our updated analysis.

We limit the analysis to  $VEI \geq 2$  because the catalog of eruptions appears to be complete for eruptions of this size. The relationship between number of eruptions and magnitude of the eruption (in this case, VEI) appears to follow a power law (Simkin & Siebert 2000). **Figure 2** shows the relationship between the numbers of eruptions and VEI for the eruption catalog used in the analysis. **Figure 2** suggests that the catalog is statistically complete since 1500 AD for  $VEI \geq 2$ . Although the catalog of analyzed events may appear to be statistically complete, it is difficult to ascertain whether increased awareness of events following large earthquakes might influence, temporarily, the number of reported eruptions. Of some concern to us in believing the analysis too strongly is the apparent absence of triggered eruptions for the most recent large earthquakes (Alaska 1964, Sumatra 2004).

Linde & Sacks (1998) also noted that paired eruptions of volcanoes separated by up to a couple hundred kilometers (e.g., Williams 1995), that is, volcanoes that erupt within two days of each other, occur more often than the average rate. **Figure 3** reproduces their analysis with the updated earthquake and eruption catalogs. Linde & Sacks (1998) note that the enhancement of paired eruptions disappears for separations



**Figure 1**

Histogram showing the number of eruptions as a function of time relative to  $M > 8$  earthquakes. Negative times correspond to eruption prior to the earthquake. Earthquake catalog from Dunbar et al. (1992) supplemented with ANSS catalog data; eruption catalog from Siebert & Simkin (2002). Only eruptions located within 800 km of the earthquake epicenter are included. Bins are 5 days wide. This figure is similar to figure 1 in Linde & Sacks (1998) but updated with more recent (and revised) earthquake and eruption catalogs.

$>200$  km, suggesting a common regional trigger for the eruptions because local earthquakes caused by eruptions tend to be small, typically much less than magnitude 6. Alternatively, because eruptions themselves generate shaking (e.g., Kanamori & Givens 1982), paired eruptions might reflect volcano triggering by the other volcano.

**Figure 1** indicates that only a small fraction of eruptions are triggered directly and immediately by large earthquakes. The same weak correlation appears to be true for nonexplosive basaltic eruptions, for example, in Iceland (Gudmundsson & Saemundsson 1980). This implies that the stress changes caused by earthquakes are small compared to the stresses needed to initiate volcanic eruptions.

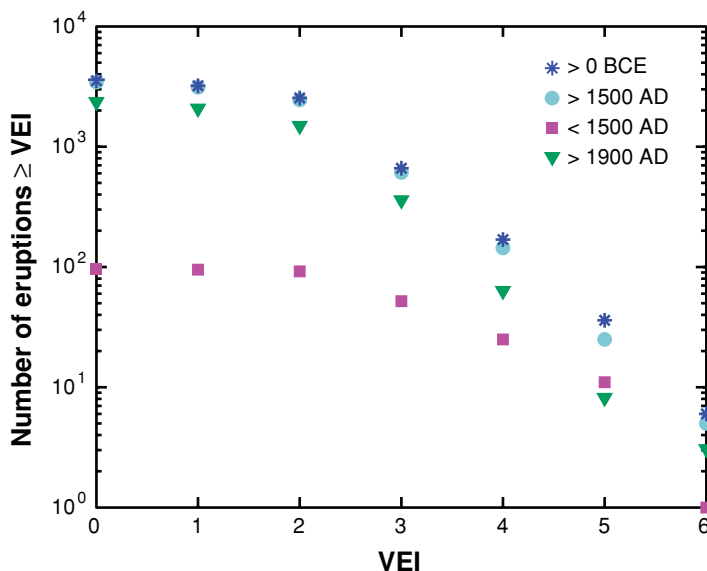
**Table 1** Table of eruptions (VEI  $\geq 2$ ) following major earthquakes (magnitude  $M > 8$ )

Earthquake	Latitude	Longitude	M	Eruption	Date	Latitude	Longitude	VEI
1730 7 8 10:00	-33.10	-71.60	8.7	Villarrica	1730 7 8	-39.42	-71.93	2?
1822 11 19 15:00	-33.50	-71.80	8.	San Jose	1822 11 19	-33.782	-69.897	2
1822 11 19 15:00	-33.50	-71.80	8.5	Villarrica	1822 11 19	-39.42	-71.93	2
1837 11 7 11:30	-39.80	-73.20	8.0	Osorno	1837 11 7	-41.1	-72.493	2
1837 11 7 11:30	-39.80	-73.20	8.0	Villarrica	1837 11 7	-39.42	-71.93	2
1913 3 14 08:45	4.50	126.50	8.3	Awu, Indonesia	1913 3 14	3.67	125.50	2
1913 10 14 08:08	-19.50	169.00	8.1	Ambrym	1913 10 14	-16.25	168.12	3
1939 4 30 02:55	-10.50	158.50	8.1	Kavachi	1939 4 30	-9.02	157.95	2
1950 12 2 19:51	-18.30	167.50	8.1	Ambrym	1950 12 6	-16.25	168.12	4
1957 3 9 14:22	51.30	-175.80	8.6	Vsevidof	1957 3 11	53.130	-168.693	2
1960 5 22 19:11	-39.50	-74.50	8.5	Cordon Caulle	1960 5 24	-40.52	-72.2	3

Mud volcanoes and geysers also appear to respond to distant earthquakes. Observations are too few, however, to perform a comparable analysis for mud volcanoes and geysers.

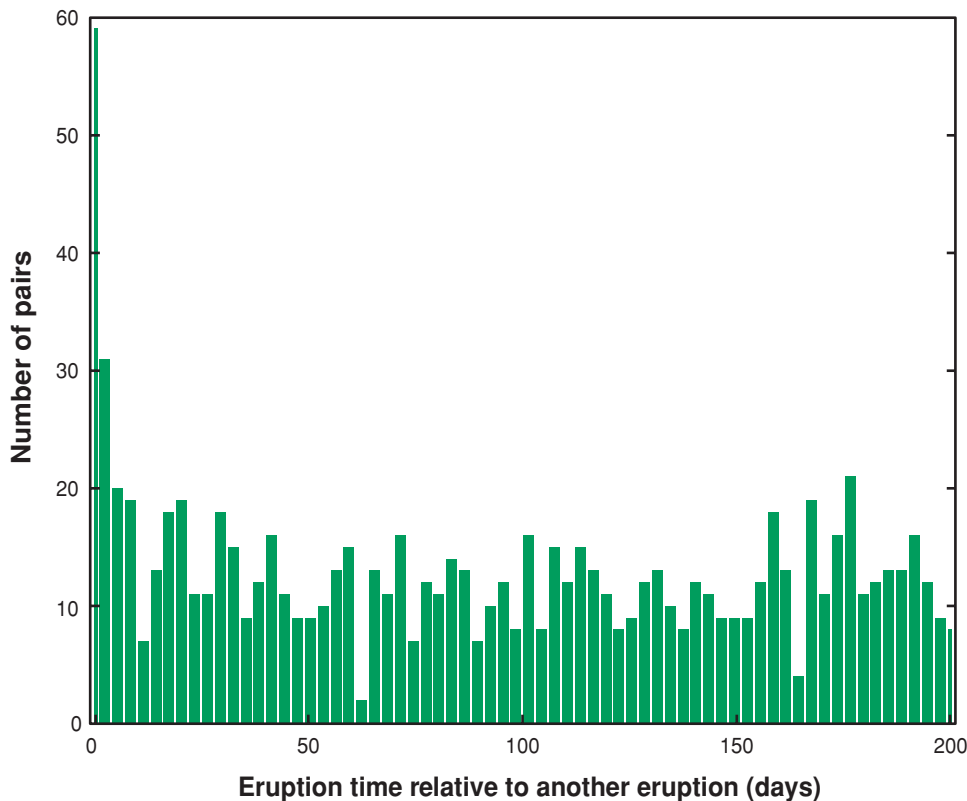
### 3. STRESSES CAUSED BY EARTHQUAKES

Earthquakes can stress magmatic systems either through static stresses, i.e., the offset of the fault generates a permanent deformation in the crust, or dynamic stresses from the seismic waves. Both stresses increase as the seismic moment of the earthquake,



**Figure 2**

Relationship between the number of eruptions and the “magnitude” of the eruption, which is characterized by the volcanic explosivity index, VEI (Newhall & Self 1982). Historical eruptions are divided into three groups: all eruptions, eruptions since 1500 AD, eruptions before 1500 AD, and eruptions since 1900 AD. Assuming that a power-law relationship exists, we infer the catalog of eruptions is statistically complete and meaningful for the time period since 1500 AD for eruptions with VEI  $\geq 2$ .



**Figure 3**

Temporal correlation of volcanic eruptions located within 10–200 km of each other. The horizontal axis shows the time interval between eruptions. Bins are 3 days wide.

but they decay very differently with distance  $r$ . Static stresses fall off as  $1/r^3$ , whereas dynamic stresses fall off more gradually. Dynamic stresses are proportional to the seismic wave amplitude. A standard empirical relationship between surface wave amplitude and magnitude has dynamic stress falling off as  $1/r^{1.66}$  (e.g., Lay & Wallace 1995). Other factors, such as directivity, radiation pattern, and crustal structure, also influence the amplitude of shaking. However, the significant difference in dependence on distance is a robust feature distinguishing static and dynamic stresses.

The difference in decay rate helps identify which stresses are important for triggering eruptions. **Table 2** lists and compares the magnitude of earthquake stresses with other external stresses that have been proposed to influence volcanoes. Both static and dynamic stresses may be significant in the nearfield (within up to a few fault lengths), but only dynamic stresses are larger than other potential triggering stresses, such as tides, in the far field (**Table 2**). For instance, Nostro et al. (1998) shows that eruptions at Vesuvius are preceded by earthquakes 50–60 km away on Apennine

**Table 2** Amplitude and period of stress changes of earthquakes compared with other forcings

		Stress (MPa)		Period
Solid Earth tides		$10^{-3}$		12 h
Ocean tides		$10^{-2}$		12 h
Hydrological loading		$10^{-3}$ – $10^{-1}$		days–years
Glacier loading		$10^1$ – $10^2$		$10^3$ years
	$10^2$ km	$10^3$ km	$10^4$ km	
Static stress changes, M8	$10^{-1}$	$10^{-4}$	$10^{-7}$	NA
Dynamic stress changes, M8	3	0.06	0.001	20 s

normal faults. The coupling is two-way in that eruptions at Vesuvius also have a (weaker) influence on Apennine earthquakes. This example can reasonably be associated with static stresses. On the other hand, the long-range triggering identified by Linde & Sacks (1998) is likely caused by the seismic waves.

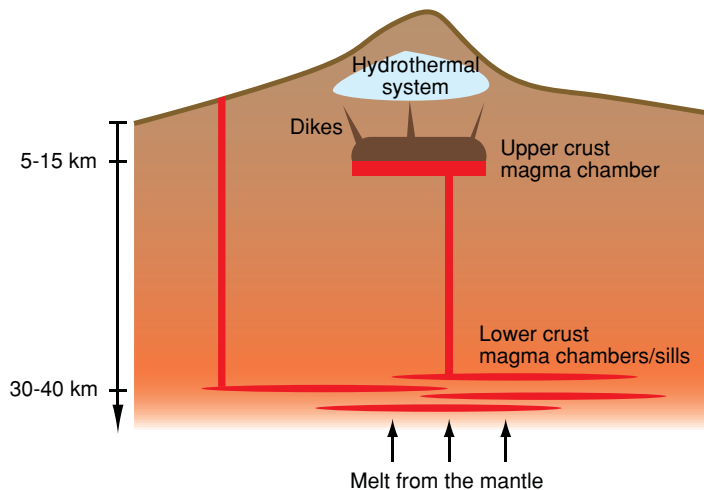
#### 4. VOLCANOES

It is possible to initiate an eruption both through changes that occur outside magma bodies and changes within the subsurface magma itself (**Figure 4**). In the former category are changes in the regional stress field due to tectonic events (e.g., Hill 1977), glacial unloading (Jellinek et al. 2004), sea level change (McGuire et al. 1997), or local changes caused by large landslides. Alternatively, changes in the overpressure in the magma, stored either in a conduit or a magma chamber, initiates the eruption. Overpressure can increase owing to injection of fresh magma (e.g., Sparks et al. 1977), accumulation of bubbles (e.g., Woods & Cardoso 1997), and crystallization that induces vesiculation (e.g. Blake 1984). The overpressure  $\Delta P_c$  required to generate tensile deviatoric stresses sufficient to allow a dike to form, and magma to propagate to the surface without freezing, is estimated to be 10–100 MPa for silicic magmas and smaller than 1 MPa for basaltic magmas (McLeod & Tait 1999, Tait et al. 1999, Jellinek & DePaolo 2004).

For eruptions caused by overpressure, if a critical overpressure of 10 MPa is a reasonable estimate and the overpressure builds up from 0 at a constant rate, it is not surprising that there are so few triggered eruptions. The probability of having a triggered eruption in a given year is the product of the probability  $\pi_v$  of the volcano being near enough to failure to be triggered and the probability  $\pi_{EQ}$  of having a large earthquake. The fraction of triggered eruptions is the ratio of the probability of having a triggered eruption to the probability of having an untriggered one. Assuming that the volcano pressurizes at a constant rate,

$$\pi_v = \Delta P_{EQ} / \Delta P_c, \tag{1}$$

where  $\Delta P_{EQ}$  is the extra pressure generated by the earthquake. The probability of getting a sufficiently large earthquake is



**Figure 4**

Schematic illustration of a volcanic system. Basaltic melts generated in the mantle initially may accumulate in magma chamber or sills at the base of the crust. These may replenish more shallow chambers in the mid to shallow crust where differentiation and/or assimilation of the surrounding crust can form more silicic magmas. A hydrothermal system may exist above such shallow regions of magma storage.

$$\pi_{EQ} = 1/T_{EQ}, \quad (2)$$

where  $T_{EQ}$  is the time between large earthquakes affecting the volcano, and, similarly, the probability of having an untriggered eruption is  $1/T_v$ , where  $T_v$  is the ordinary recurrence time of volcanic eruptions. The fraction of eruptions that are triggered  $X_t$ , assuming the stress changes caused by the earthquakes are simply added to the overpressure, is

$$X_t = \Delta P_{EQ} T_v / \Delta P_c T_{EQ}. \quad (3)$$

The static and dynamic stresses  $\Delta P_{EQ}$  caused by earthquakes are typically  $10^{-2}$ – $10^{-1}$  MPa (Table 2). Thus, the overpressure must be within 99%–99.9% of the maximum overpressure for the earthquake to initiate an eruption. Typical recurrence intervals  $T_v$  for VEI 2 and 3 eruptions are 1– $10^2$  years (Simkin & Siebert 2000). The recurrence time  $T_{EQ}$  for large magnitude  $>8$  earthquakes near a given volcano is  $10^2$ – $10^3$  years. Thus, only a very small fraction ( $\ll 1\%$ ) of eruptions will be triggered at a given volcano.

The recurrence time of eruptions may be difficult to measure independently from triggering, and both earthquake and eruption recurrence intervals may be poorly defined in a given region. Nevertheless, we can estimate the fraction of triggered eruptions. In the catalogs we used, since 1547 there are on average five VEI  $\geq 2$  recorded eruptions per year and 0.5 magnitude  $\geq 8$  recorded earthquakes per year. If  $\Delta P_E / \Delta P_c \sim 0.1\%$ – $1\%$ , then we would expect 0.01%–0.1% of the catalog to be triggered. In fact, we observe 10 events since 1547, i.e., 0.4% of the catalog. The observation



of triggered eruptions may thus place a constraint on the magnitude of the stresses generated by the earthquake. To explain the number of triggered eruptions observed, the pressures generated by the earthquakes must be closer to those generated by the dynamic stresses than the static stresses. However, seismic waves are merely transitory phenomena, and an additional mechanism is necessary to maintain the increased pressure in the magmatic system.

Fortunately, from the perspective of trying to explain triggered eruptions, there are mechanisms that may increase the stress changes caused by distant earthquakes or allow dynamic stresses to be converted into permanent stress changes, and this is the topic reviewed next. We first consider mechanisms that can change the magma overpressure, and then mechanisms that might promote eruption through changes surrounding stored magma.

## 4.1. Magmatic Processes

Overpressure within a magma chamber can increase if bubbles nucleate or grow, owing to convection, which may promote bubble nucleation and growth, or from the injection of fresh magma. Next, we consider how dynamic strain can influence bubbles in stored magma and magma chamber dynamics.

**4.1.1. Bubbles.** Bubbles are critical because they provide the driving force for triggering eruptions, either by increasing buoyancy or by increasing magma chamber overpressure. Huppert & Woods (2002) point out the importance of bubbles for sustaining eruptions—bubbles increase compressibility significantly, and greater compressibility permits the eruption of more magma at a given overpressure. Thus nucleation of new bubbles directly by seismic waves, or indirectly through nucleation of crystals, which increases the volatile content of the remaining melt, enhances the ability of a chamber to initiate and sustain an eruption.

Two mechanisms that involve bubbles directly have been invoked to explain distant volcano (and also earthquake) triggering; we also discuss the possibility that the nucleation of new bubbles by seismic waves may trigger an eruption.

**Rectified diffusion.** A bubble immersed in a volatile-saturated magma undergoing periodic variations in pressure will expand and contract. When the bubble contracts, the magma is undersaturated and vapor will diffuse into the melt. Conversely, when the bubble expands, vapor diffuses into the bubble. Because the surface area is smaller for the contracted bubble, the diffusive losses will be less than the diffusive gains, and there can be a small net addition of mass to (and hence volume of) the bubble following the passage of seismic waves. The increase in volatiles inside the bubble implies an increase in magma pressure in a closed system. This process is termed rectified diffusion.

Brodsky et al. (1998) used a linear approximation to model this process and concluded that the net pressure changes are at most  $10^{-2}$  MPa in basalt and  $10^{-4}$  MPa in rhyolite (in both cases, assuming the volatile phase is water) and can only be achieved if the magmatic system is sufficiently oversaturated prior to the earthquake. If such an

oversaturated condition exists, then a delicate balance must have existed between pressure increase through gas escape and bubble growth through ordinary diffusion. Rectified diffusion merely upsets the balance. Ichihara & Brodsky (2006) further reduce the plausible parameter space for rectified diffusion by solving the nonlinear problem in which resorption of gas as the pressure increases is considered in a self-consistent way. Linear estimates greatly over-predict the pressure rise caused by rectified diffusion.

**Advective overpressure.** One mechanism for triggering an eruption is through the increase in pressure caused by rising bubbles in a closed system (Steinberg et al. 1989). Sahagian & Proussevitch (1992) proposed that rising bubbles can generate several hundred MPa of overpressure, sufficient to fracture the rocks surrounding the magma if the bubbles can rise and separate from the magma fast enough (Woods & Cardoso 1997). If the liquid surrounding the bubble is incompressible, the mass of the bubble does not change, the walls of the container (e.g., a conduit or magma chamber) are not deformable, and there is a vertical pressure gradient in the magma, then a rising bubble will not change its internal pressure as it rises. Consequently, it can carry a large pressure from near the bottom of a magma chamber to the surface, and the pressure increase would promote eruption. Assuming a magmatic pressure gradient, the change in pressure is given by

$$\Delta P_{\max} = \rho g \Delta H, \quad (4)$$

where the subscript max indicates that this is the maximum possible pressure change and  $\Delta H$  is the change in vertical position of the bubble. This model has been verified with lab experiments (e.g., Sahagian & Proussevitch 1992, Pyle & Pyle 1995), and the pressure changes are found to be reversible (Pyle & Pyle 1995). If  $\Delta H$  is 1 km, then pressure changes can be as large as 30 MPa, and thus similar to the overpressures needed to erupt silicic magmas.

Linde et al. (1994) appealed to this mechanism to explain both triggered seismicity (Hill et al. 1993) and deformation measurements in Long Valley Caldera following the Landers earthquake. Linde et al. (1994) propose that bubbles were shaken loose by the passage of seismic waves.

The problem with this mechanism is that almost all the assumptions listed previously are poor approximations for magmas, bubbles in magmas, and the containers in which magmas are stored. Bubbly magmas are compressible, which has a large effect on the pressure change. Gases can diffuse in and out of bubbles on timescales of seconds to hours in response to pressure changes. Conduits and chambers can change their volumes in response to pressure changes. Bagdassarov (1994) and Pyle & Pyle (1995) reanalyzed the advective overpressure model, as it applies in magmas, and concluded that, in general, vertical transport of bubbles should induce negligible changes in pressure. Pyle & Pyle (1995) further show that for advective overpressure to play a role in triggering an eruption, the transport of bubbles must occur by convective overturn of the magma, but even in this case, bubbles are likely to act primarily as tracers of the flow.

**Creating new bubbles.** The nucleation of a bubble requires a supersaturation pressure  $\Delta P_s$  to overcome the energy barrier provided by surface tension. Classical nucleation theory (e.g., Hirth et al. 1970) predicts a very strong dependence of the homogeneous bubble nucleation rate  $\mathcal{J}$  (number of bubbles per unit volume per unit time) on the supersaturation pressure  $\Delta P_s$ ,

$$\mathcal{J} \propto \exp \left[ \frac{-16\pi\sigma^3}{3kT\Delta P_s^2} \right], \quad (5)$$

where  $\sigma$  is surface tension,  $k$  is the Boltzmann constant, and  $T$  is temperature. Small changes in pressure, for example, those induced by passing seismic waves, can thus potentially lead to large changes in nucleation rate.

Most nucleation studies with magmas are conducted using decompression experiments, and these determine the critical supersaturation pressure  $\Delta P_{s_{crit}}$  to trigger bubble nucleation. The barrier for homogeneous nucleation in crystal-free rhyolite is large,  $\Delta P_{s_{crit}} > 100$  MPa (e.g., Mangan & Sisson 2000, Mourtada-Bonnefoi & Laporte 2004). Heterogeneous nucleation, in which the presence of certain crystals (e.g., Fe-Ti oxides) provide nucleation substrates because they lower surface tension, can greatly reduce  $\Delta P_{s_{crit}}$ , to less than a few MPa (e.g., Hurwitz & Navon 1994). Increases in water content and concentration of mafic crystals both reduce  $\Delta P_{s_{crit}}$  (Mangan & Sisson 2005). Indeed, the potential for large  $\Delta P_{s_{crit}}$  may be limited to rhyolitic melts (Mangan et al. 2004). The importance of large  $\Delta P_{s_{crit}}$  for ascending, supersaturated magmas is that once bubbles do form they grow rapidly because of both large gradients in volatile content and short diffusion length scales. Rapid increases in vesicularity promote explosive eruption.

The likelihood of an earthquake triggering an eruption depends on the pressure disturbance generated by the earthquake, the critical supersaturation, and the timescale for supersaturation to increase in the magma, in much the same way as Equation 3. Dissolved gas is added to the magma due to crystallization, and if the partial pressure of this gas is close to the critical supersaturation, then the small pressure change due to the seismic waves will result in significant excess nucleation. In a crystal-poor system that can sustain a high supersaturation, when the bubbles nucleate, the large excess gas pressure will make the new bubbles grow rapidly and perhaps the volcano respond explosively. In a system with a small critical supersaturation, the nucleation from the seismic waves may have a less dramatic effect. If the rate of dissolved gas addition from crystallization is the same in both cases, then, from Equation 3, triggering a nucleation event is equally probable in both cases.

**4.1.2. Falling roofs.** Hill et al. (2002) suggest that loosely bound aggregates (crystal mush) of dense crystals that crystallize below the roof of magma chambers might be dislodged by passing seismic waves. Dense crystals can accumulate if the solidification front advances faster than individual crystals can sink (e.g., Sparks et al. 1993). The formation of a gravitationally unstable layer is possible because of a finite yield strength provided by crystal-crystal bonds or connections (e.g., Marsh 2000). If dynamic strains break these bonds and connections, the yield strength decreases and the gravitational instability of the crystal mush creates sinking plumes of crystals.

The rising magma that replaces the sinking magma may then vesiculate as it ascends, assuming it is volatile-saturated at depth, increasing the overpressure in the magma chamber. Bubble exsolution, following intrusion of new magma into a reservoir or crystallization, is sometimes invoked to initiate eruption. Vesiculation accompanying convective overturn may similarly increase overpressure and trigger an eruption.

It is conceivable that overturn is rapid enough to explain the triggering within days implied by **Figure 1**. In more detail, the size of crystal-laden plumes can be predicted from the Rayleigh-Taylor instability of two superposed fluids with different densities (for a review, see Johnson & Fletcher 1994). We assume that the mush layer with thickness  $b$  behaves as a Newtonian fluid and that the layer has uniform properties (whereas there should be a gradient in crystallinity and melt viscosity owing to the temperature gradient that causes crystallization). We let  $\gamma$  denote the viscosity ratio of the mush layer to that of the underlying magma. For  $\gamma \gg 1$  and  $b$  much smaller than the thickness of the magma body, the most unstable wavelength is (Ribe 1998)

$$\lambda = \frac{1}{2}b(180\gamma)^{1/5}. \quad (6)$$

Although the creation of the layer requires a yield strength so that the mush layer is not Newtonian, the strain (dynamic or static) caused by the earthquake is assumed to significantly reduce this strength. The predictions of stability analyses that allow for depth-dependence of viscosity and density (Conrad & Molnar 1997) or non-Newtonian viscosity (Houseman & Molnar 1997) produce results that are not different from an order-of-magnitude perspective.

If we assume the crystal mush has a yield strength  $\sigma_y$  of  $10^4$ – $10^5$  Pa (e.g., McBirney & Murase 1984, Ryerson et al. 1988, Hoover et al. 2001), and  $\Delta \rho/\rho$  is 0.01 (where  $\Delta \rho$  is the density difference between the crystal mush and the underlying magma with density  $\rho$ ), then the yield strength can support a crystal mush with thickness  $b \sim \sigma_y/\Delta \rho g$  in the range of 30–300 m. If the viscosity ratio  $\gamma$  is  $10^2$ , then  $\lambda \sim 10^2$ – $10^3$  m. The size of sinking plumes, once formed, also scales with  $\lambda$ . Because  $\lambda$  is so large, the Reynolds number can be large,  $>10^1$ , even in silicic magmas (assumed to be water saturated at typical magma chamber pressures). The sinking velocity of fully formed plumes can thus be estimated as  $\sim(\Delta \rho g \lambda/\rho)^{1/2} \sim 10^1$  m/s. Plumes can potentially travel fast enough, that is they traverse a magma chamber within days, to induce sufficient overturn that they trigger eruptions within days.

We conclude that sinking crystal-rich plumes may be a viable mechanism for rapid triggering if dynamic strain can significantly reduce the yield strength of crystal mushes, that these crystal mushes exist and are thick (so that the sinking velocity of plumes is large), and the magma chamber is volatile saturated. This conclusion assumes that the classic Rayleigh-Taylor analysis correctly predicts both the wavelength of the instability and the size of plumes that form from a layer in which the density differences arise from the presence of dense crystals. Voltz et al. (2001) found excellent quantitative agreement between wavelengths and growth rates in experimental measurements and the Rayleigh-Taylor predictions. Their experiments were for volume fractions of particles less than 6.5%. In contrast, Michioka & Sumita (2005) performed laboratory experiments with much higher concentrations (more realistic

for magma chambers) and found that the apparent layer thickness  $b$  that matches a Rayleigh-Taylor prediction is not the thickness of the layer of crystal-rich magma. Instead,  $b$  scales with the size of the crystals and is only a few crystal radii. The plumes that form are thus much smaller and travel much more slowly. Further work is clearly needed to understand what governs the size of plumes formed from a crystal mush.

It is possible to increase the rate of magma chamber overturn if the vesiculation of rising volatile-saturated magma results in a density difference greater than that for the sinking plumes. In this case, the triggering is a two-stage process in which the dynamic strain causes the sinking of crystal-rich plumes, which in turn triggers vesiculation, convection and eruption.

**4.1.3. Timescales.** The period over which overpressure develops is critical because the stresses must be generated sufficiently rapidly that they cannot relax. Major relaxation mechanisms include viscoelastic relaxation and gas loss.

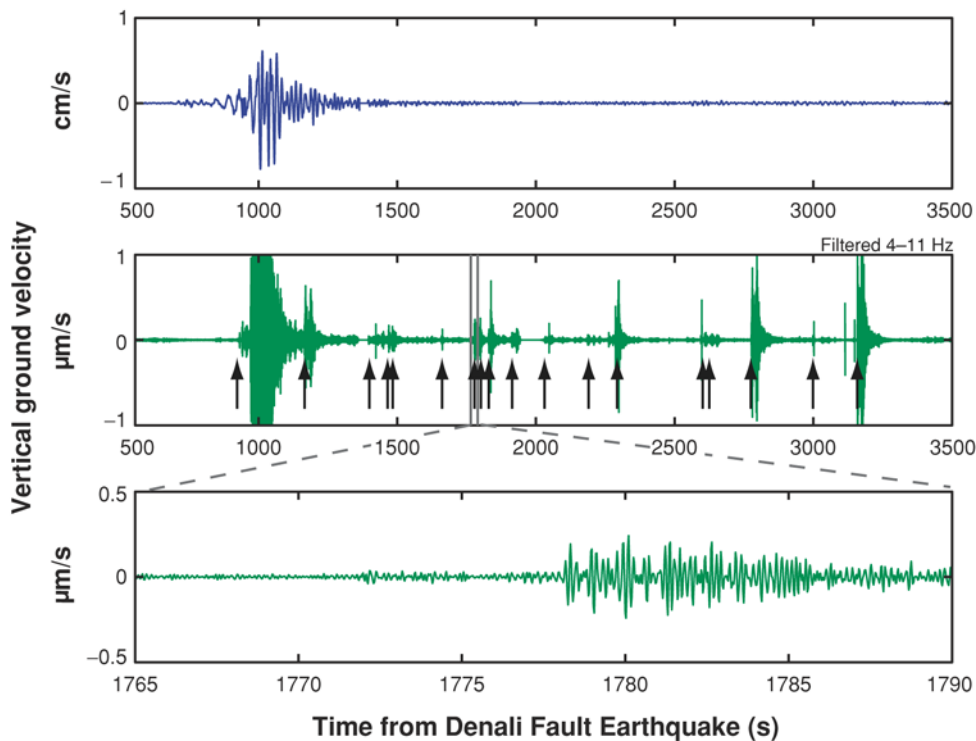
The relaxation timescale for viscoelastic deformation is  $\sim(2E/\mu)^{-1}$ , where  $E$  and  $\mu$  are the elastic modulus and effective viscosity of the wall rocks of the magma chamber. This timescale is likely to be several months to many decades for most magmatic systems, a timescale that is usually long compared to the timescales that characterize the processes discussed in this section.

On the other hand, the timescales of gas loss may be rapid. The timescale of percolation of gas out of a chamber of size  $L$  is  $L^2/c$ , where  $c$  is the hydraulic diffusivity of the gas. We estimate that typical values of  $c$  range from  $10^{-3}$ – $10$  m<sup>2</sup>/s, assuming that bubbles are connected and that permeabilities measured on vesicular rocks (e.g., Rust & Cashman 2004) are appropriate. At the higher end of this range, gas can percolate out of a 100 m body of magma in  $10^3$  s, i.e., a timescale not much longer than the duration of a wavetrain of a very large earthquake. Therefore, a larger or more tightly sealed magmatic system is necessary to permit the pressure to build to levels needed for eruption.

## 4.2. Triggers External to the Magma

The eruption of juvenile magma requires opening pathways for magma ascent. Thus, any process that influences stresses around a magma chamber, such as local earthquakes (e.g., Hill 1977), or that creates fractures may initiate an eruption. In some cases there is evidence that local tectonic earthquakes cause stress changes that initiate eruption (e.g., La Femina et al. 2004).

**4.2.1. Triggered seismicity.** Earthquake swarms are a common precursor of eruptions and may trigger further unrest in a volcanic system. One feature of volcanic and geothermal systems is that they often experience triggered seismicity from other earthquakes hundreds and even thousands of kilometers from the original mainshock (e.g., Hill et al. 1993, Brodsky et al. 2000, Gomberg et al. 2001, Prejean et al. 2004). Based on a compilation of reports in the literature, 85% of cases of triggered eruptions

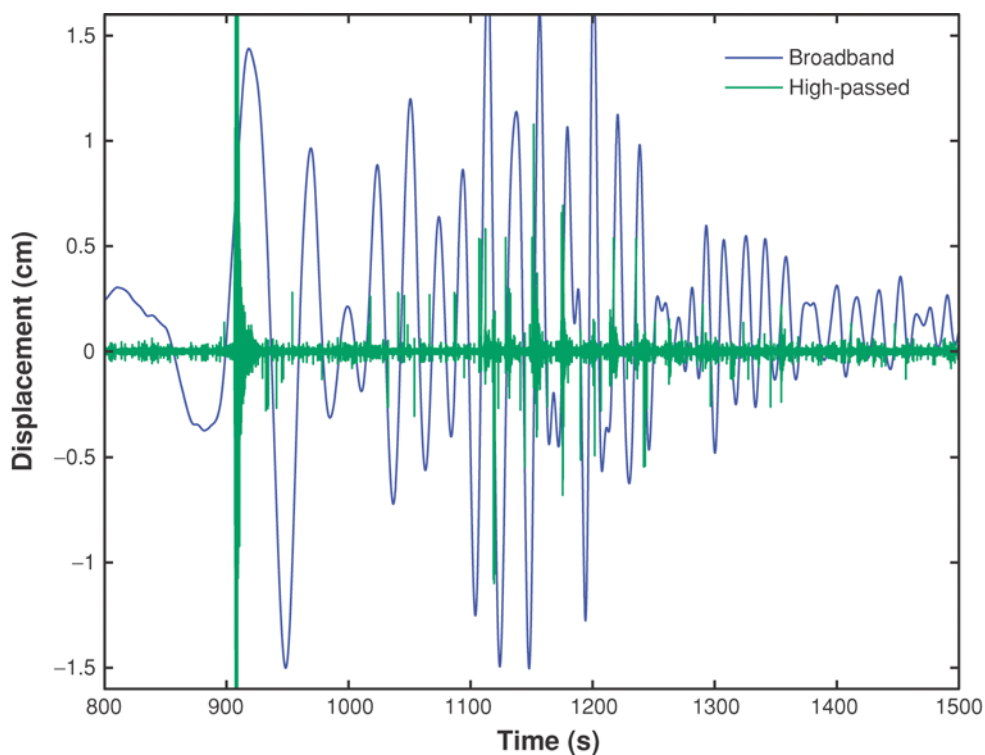


**Figure 5**

Broadband seismogram from Bozeman, Montana, showing surface waves from the  $M_w$  7.9 2002 Denali Fault earthquake (*top panel*) and high-passed records revealing local triggered earthquake (*middle and bottom panels*). Arrows indicate local events. Note that the local earthquakes begin at the time of the surface wave arrivals. Most of these local earthquakes are in Yellowstone.

occur in geothermal settings, although triggered earthquakes have been reported in other settings (e.g., Hough et al. 2003, Tibi et al. 2003).

As noted previously, at large distances the dynamic stresses from the originating mainshock are much larger than the static stress changes, and hence dynamic strains are usually invoked to explain distant triggered earthquakes. Close inspection of triggered earthquakes shows that swarms commence with the arrival of the surface waves. **Figure 5** shows an example of events triggered at Yellowstone National Park by the 2002 magnitude 7.9 Denali earthquake more than 3000 km away. Swarms persist through the wavetrain and, in some cases, continue for days (Hill et al. 1993). More recent evidence from Wrangell, Alaska, shows that triggered events occurred during the maximum horizontal extension of the surface waves (West et al. 2005). A similar observation of triggered earthquakes correlating with a particular phase of the Rayleigh waves can be made for earthquakes triggered at the Coso geothermal field from the Denali earthquake (**Figure 6**). The tight relationship between a particular



**Figure 6**

Triggering of local earthquakes from the Denali Fault earthquake at Coso geothermal field, as recorded on the broadband seismometer JRC. As in **Figure 5**, the local events are shown on a high-passed record. The high-passed record is amplified to appear on the same scale as the original broadband record. The local earthquakes appear in packets that correspond to the Rayleigh wave shaking.

part of the wavefield and the triggering further reinforces the conclusion that the seismic waves are directly responsible for the triggered earthquakes.

Proposed mechanisms for long-range triggering include frictional instabilities (e.g., Gomberg et al. 1997), bubble pressurization by rectified diffusion (e.g., Sturtevant et al. 1996) or advection (e.g., Linde et al. 1994), subcritical crack growth (e.g., Brodsky et al. 2000), and fracture unclogging (e.g., Brodsky et al. 2003). The mechanisms involving bubbles are those described in the previous section on bubbles and suffer from the same shortcomings.

Brodsky & Prejean (2005) examined triggered seismicity in Long Valley Caldera and concluded that any mechanism that relies strictly on cumulative strain energy is incompatible with the observations. Moreover, they found that long-period seismic waves are more effective at triggering earthquakes than short-period waves of similar amplitudes as recorded at the surface of the Earth. This conclusion contradicts the results of an earlier study (Gomberg & Davis 1994) that assumed that earthquakes

triggered by tides and seismic waves had the same frequency-amplitude dependence. The increased effectiveness of long-period waves at Long Valley is primarily due to the increased penetration depth of long-period surface waves into the seismogenic zone. Further enhancement of the long-period energy may be due to the triggering mechanism. For instance, if the seismic waves push water in and out of fault zones, the water pressure oscillations will be affected more by the long-period waves than the short-period waves.

A third conclusion of Brodsky & Prejean (2005) is that the amplitude of the observed waves is insufficient to trigger either frictional instabilities or stress corrosion unless the ambient pore pressure is extremely high ( $>99\%$  lithostatic). Although pockets of such high pore pressure may exist in Long Valley Caldera, elevated pore pressures are unlikely at another triggered site, The Geysers in northern California. At The Geysers, fluid pressures are observed to be nearly vaporstatic (Moore et al. 2000).

Brodsky et al. (2003) proposed that unblocking preexisting fractures by fluid flows driven by dynamic strain may allow for a redistribution of fluid pressure and hence trigger earthquakes. Because this process involves fluid flow, there is both an amplitude and frequency dependence of triggering on dynamic strain, consistent with the observations. Even in this mechanism, it is not clear quantitatively how the very small stresses,  $\sim 10^{-2}$  MPa, generated by the seismic waves can produce sufficient flow to unplug fractures. Perhaps the unblocking mechanism fails less readily than friction instabilities or subcritical crack growth because the critical stresses necessary to unplug fractures are less well understood. At Long Valley, observed changes in water level in deep wells (Roeloffs et al. 2003) and measured surface deformation (e.g., Johnston et al. 1995) appear to be compatible with upward flow of water from a breached hydrothermal system accompanying triggered earthquakes.

**4.2.2. Surface triggers.** Some eruptions are initiated when a surface load is removed, or a cap rock fails, and subsurface magma experiences rapid decompression. Earthquakes can trigger landslides, and landslides above stored magma can cause rapid decompression. If the subsurface magma has a high enough vesicularity it can erupt explosively (e.g., Spieler et al. 2004, Namiki & Manga 2005).

Although we are unaware of any eruption triggered by large, distant earthquakes in this manner, eruptions of juvenile magma have been initiated by landslides, for example, in 1980 at Mount St. Helens (Lipman & Mullineaux 1981) and 1964 at Shiveluch (Belousov 1995).

**4.2.3. Fatigue and subcritical crack growth.** Dobran (2001) notes that repeated or fluctuating stresses can cause material failure for stresses well below the usual failure stress. He describes a scenario of fatigue where cyclic strains cause subcritical cracks to grow to critical size. As discussed in Triggered Seismicity, above, the same mechanism has also been suggested as a way to trigger earthquakes from seismic waves.

In both cases, it is difficult to generate large crack growths from the small stresses of the seismic waves. Brodsky & Prejan (2005) calculate that the pore pressure must be within 99.9% lithostatic to reduce the effective pressure enough to allow the seismic



waves to have a significant effect. This strong condition does not seem plausible for very common occurrences, such as triggered seismicity in geothermal areas, but it may be applicable for more rare events, such as triggered eruptions. If pore pressure in the hydrothermal system increased steadily with time, then the statistics of triggering eruptions via subcritical crack growth can be calculated from an equation similar to Equation 3. Earthquakes can trigger eruptions via this mechanism 0.01% of the time, which would account for about 3% of the triggered eruptions.

**4.2.4. Viscoelastic relaxation.** Viscoelastic relaxation of earthquake-induced stresses has been invoked for delayed earthquake-earthquake triggering over intermediate to large distances (e.g., Pollitz et al. 1998, Freed & Lin 2001). Triggering by viscoelastic relaxation results in a time lag of years to decades. Marzocchi (2002) and Marzocchi et al. (2004) have argued that triggering of eruptions to distances of  $10^3$  km can also occur through viscoelastic relaxation of stresses. The magnitude of these stresses should be smaller than the static stresses listed in **Table 2**. Hill et al. (2002) suggest that viscoelastic relaxation might contribute to the increase in eruption frequency in the Cascade volcanic arc during the early 1800s following the 1700 AD magnitude 9 megathrust earthquake.

Because the stress diffusion caused by viscoelastic relaxation results in a nonlinear spatial and temporal evolution of stresses caused by the earthquake, quantifying the relationship between earthquakes and eruptions from observations will undoubtedly remain challenging.

## 5. MUD VOLCANOES

Mud volcanoes range from small centimeter-sized structures to large edifices up to few hundred meters high and several kilometers across. Mud volcanoes erupt water, fine sediment, fragments of country rock, and sometimes hydrocarbons and gas (methane and carbon dioxide). They are found most often in areas where high sedimentation rates and fine-grained materials allow high pore pressures to develop (Kopf 2002), and regional compression sometimes provides the driving force for eruption (e.g., Deville et al. 2003). Hereafter, we only consider large mud volcanoes that erupt from depths of at least several hundred meters.

Mud volcanoes, like magmatic volcanoes, may also erupt in response to earthquakes (e.g., Panahi 2005), although the number and quality of observations make it especially difficult to quantify and verify the correlation. Some specific examples include the reactivation of mud volcanoes in the Andaman Islands following the December 26, 2004 magnitude 9.3 earthquake whose epicenter was 1100 km away (although within 60 km of the rupture); the Niikappu Mud Volcano, Hokkaido, Japan, pictured in **Figure 7**, erupted within hours of the magnitude 7.9 Tokachi-oki earthquake 160 km away, and responded to at least four other large, distant earthquakes (Chigira & Tanaka 1997; N. Makoto, personal communication).

A necessary condition to create mud volcanoes is the liquefaction or fluidization of erupted materials (Pralle et al. 2003). Liquefaction can be induced when shear strains, e.g., those generated by earthquakes or anisotropic consolidation, cause

**Figure 7**

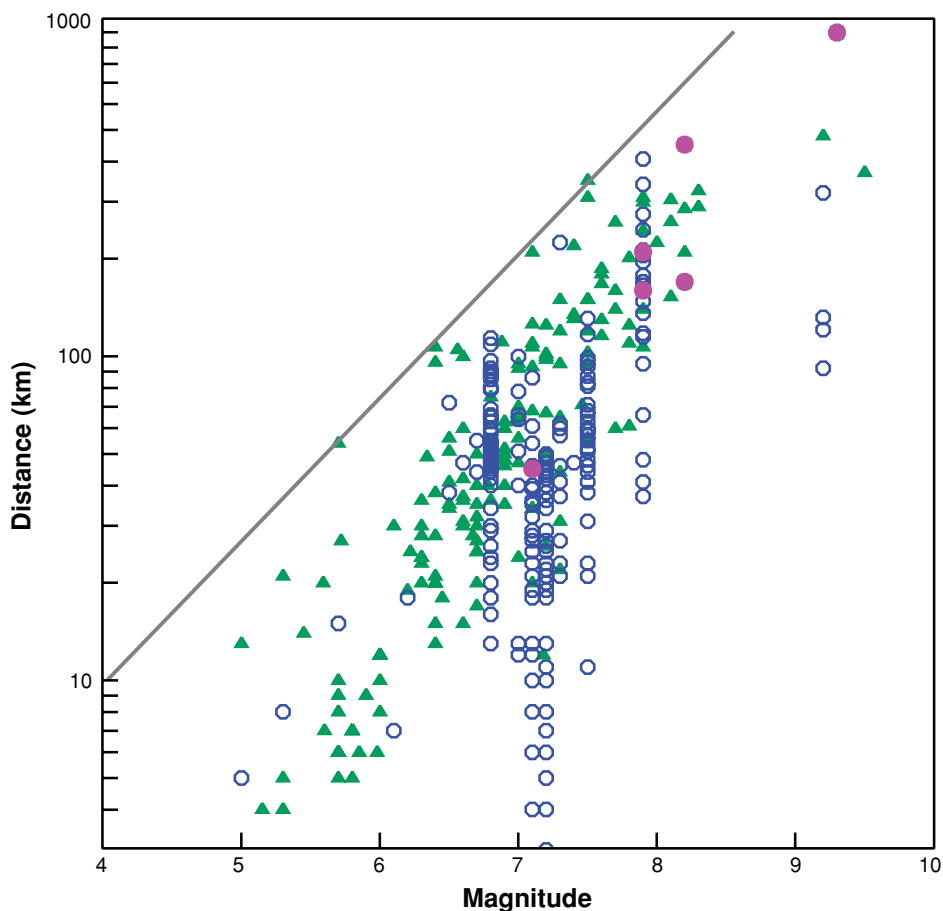
Photo of the Niikappu Mud Volcano, Hokkaido, Japan. Image provided by Nakamukae Makoto.



unconsolidated materials to consolidate through a rearrangement of grains. The pressure in the fluid between grains must then increase in response to the volume decrease. If undrained consolidation proceeds sufficiently far that the effective stress becomes zero, the material loses strength and can behave in a liquid-like manner. Fluidization is caused by fluid flowing upward at a sufficient rate to support the weight of grains. Fluidization of subsurface sediment might arise if strains caused by the earthquake rupture hydraulic seals that had isolated overpressured fluids.

Liquefaction occurs commonly after earthquakes in shallow soils. In fact, small mud volcanoes are one manifestation of earthquake liquefaction. **Figure 8** shows the distance over which liquefaction occurs following earthquakes. The compilation shown includes observations reported in several previous compilations (Kuribayashi & Tatsuoka 1975, Ambraseys 1988, Galli 2000, Wang et al. 2005). The solid curve in **Figure 8** is an empirically determined relationship for the maximum distance  $R_{\max}$  over which liquefaction has been found,  $M = -5.0 + 2.26 \log R_{\max}$ , where  $R_{\max}$  is in meters (Wang et al. 2005). It should be noted that **Figure 8** includes only two parameters that influence liquefaction: distance and earthquake magnitude. Properties of the earthquake source, elastic properties of the region through which the waves travel, and most importantly, the liquefaction susceptibility also influence the occurrence of liquefaction. For this reason, the solid line is best viewed as an estimate of the maximum possible distance at which liquefaction could be anticipated.

Also included in **Figure 8** are observations of increases in streamflow that occurred following earthquakes (compilation from Manga 2001, Montgomery & Manga 2003, Wang et al. 2004). Increases in streamflow require changes in either hydraulic head (Muir-Wood & King 1992, Manga 2001) or changes in permeability (Rojstaczer et al. 1995). Although the origin of streamflow increases may be controversial (Montgomery & Manga 2003), all explanations involve significant hydrological changes in the subsurface that could influence mud volcanism.



**Figure 8**

Relationship between earthquake magnitude and distance over which liquefaction has been reported (*solid green triangles*). Also shown is the distance over which significant changes in streamflow have been reported (*open blue circles*). The latter require changes in either subsurface hydraulic properties or fluid pressure. The solid purple circles show the distance from the epicenter of large mud volcanoes that originate from depths greater than hundreds of meters and erupted within two days of the earthquake; shown are Andaman Island (responding to magnitude 9.3 Sumatra earthquake) and Niikappu Mud Volcano, Hokkaido, Japan (responding to events with magnitudes 7.1, 7.9, 7.9, 8.2, and 8.2 in 1982, 1968, 2003, 1952, and 1994, respectively). The gray line is an empirical upper bound for these observations from Wang et al. (2005):  $M = -5.0 + 2.26 \log R_{\max}$  where  $R_{\max}$  is in meters.

Assuming the empirical scaling shown in **Figure 8** applies to liquefaction at depth, mud volcanoes might be expected to respond to magnitude 6–7 earthquakes up to a few hundred kilometers away and perhaps over 1000 km away for a magnitude greater than 9 earthquake (although we emphasize that there are far too few observations to constrain the maximum distance for liquefaction for large events). The reports

of triggered mud volcanism cited previously fall within the bounds suggested by **Figure 8**.

Liquefaction is usually thought to be a shallow phenomenon, which is confined to the upper few to tens of meters. This is because the increase in fluid pressure needed to reach lithostatic pressure usually increases with depth. However, sedimentary basins with high sedimentation rates and fine sediments (with low permeability) often have high fluid pressures. Indeed, it is these settings in which mud volcanoes seem to occur. In addition, the presence of gas bubbles, whose presence may be inferred from the eruption of gas (primarily methane and carbon dioxide), enhances local deformation, and hence pore pressure changes (Pralle et al. 2003). Yassir (2003) performed cyclic loading experiments on low-permeability clays consolidated to a mean stress of 30 MPa (equivalent to a depth of about 2 km). Cyclic loading/unloading increased the pore pressure by 15 MPa, although liquefaction did not occur in this particular experiment.

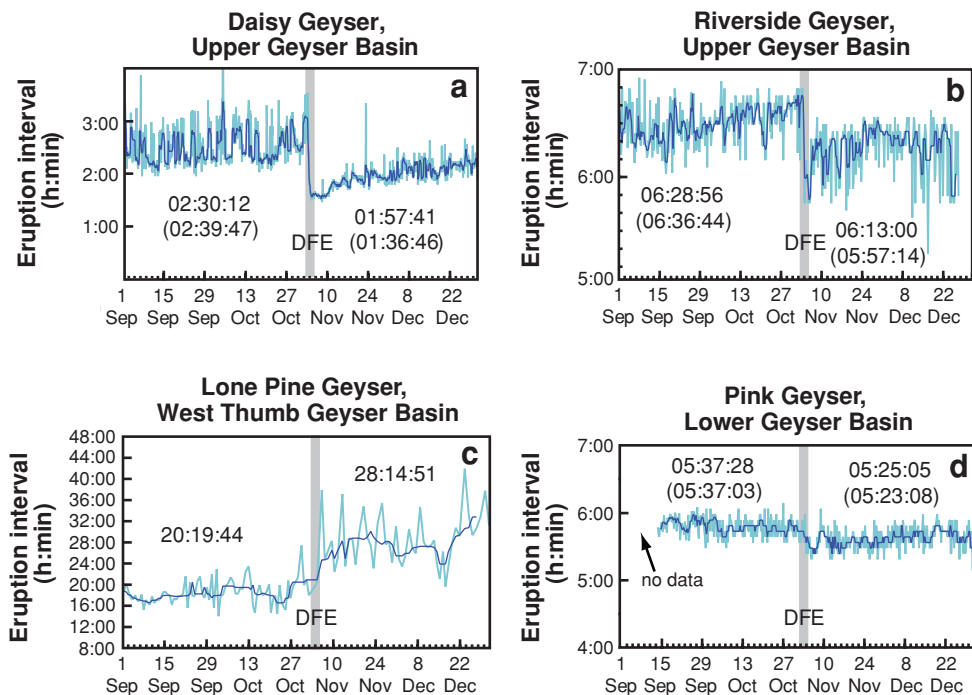
Liquefying or fluidizing unconsolidated sediment does not require earthquakes; mineral dehydration, gas expansion, tectonic stresses, inflow of externally derived fluids, and even ocean waves are examples of processes that can in some cases promote liquefaction of subsurface sediments (Maltman & Bolton 2003). The fact the mud volcanoes are widespread indicates that, like magmatic volcanoes, earthquakes only enhance other processes that produce high fluid pressures.

We conclude that that liquefaction by dynamic strain (cyclic deformation) may be a plausible mechanism for triggering the eruption of mud volcanoes. The relationship between earthquake magnitude and distance for shallow occurrence of liquefaction is not inconsistent with the small number of reported examples of mud volcanoes that erupted immediately after earthquakes. We should note, however, that the quantitative relationship between dynamic strain and occurrence of liquefaction in deep sedimentary basins could be quite different from that shown in **Figure 8** for the shallow subsurface because the pore pressures are likely to already be high, and both a free gas and dissolved gas, which promote eruption, may be present. A second possible explanation is the breaching of sealed, pressurized reservoirs—a mechanism similar to that proposed for triggered earthquakes—to induce flow and fluidization.

## 6. GEYSERS

Geysers are periodically erupting springs driven by steam and/or noncondensable gas. Geysers are few in number, reflecting the special conditions required to generate sufficient hot water and the right combination of conduit geometry and permeability (Ingebritsen & Rojstaczer 1993, 1996). Natural geysers, at least on Earth, appear to be confined to the discharge regions of hydrothermal systems.

Geysers show clear changes in eruption frequency following distant earthquakes that generate coseismic static strains smaller than  $10^{-7}$  or dynamic strains less than  $10^{-6}$  (e.g., Hutchinson 1985, Silver & Vallette-Silver 1992). **Figure 9** shows the response of geysers in Yellowstone National Park to the magnitude 7.9 Denali earthquake located >3000 km away (Husen et al. 2004). At Daisy geyser, for example, the eruption interval decreased by about a factor of two and then slowly increased to the



**Figure 9**

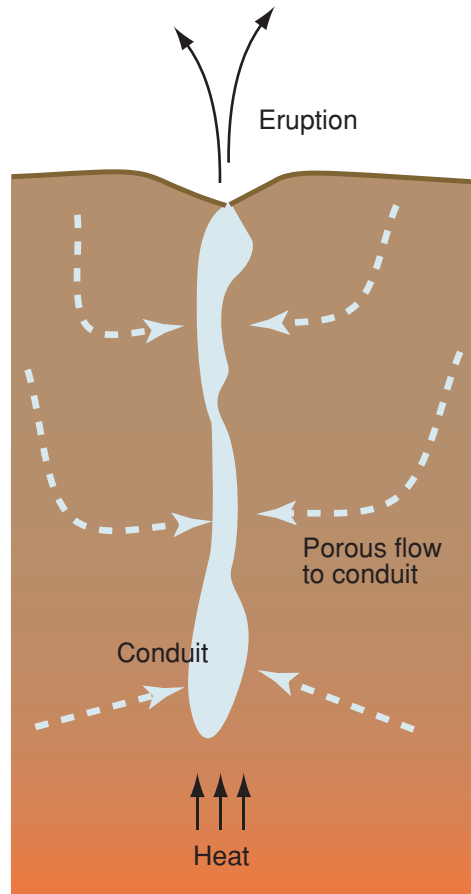
Eruption interval (time between two successive eruptions) in hour:minute format as a function of time for four geysers in Yellowstone National Park: (a) Daisy geyser, (b) Riverside geyser, (c) Lone Pine geyser, and (d) Pink geyser. Light blue lines show raw data, that is, the actual measured eruption intervals. The dark blue lines are smoother data obtained by averaging over several measurements. Median eruption intervals are also shown in hour:minute:second format. Median eruption intervals are computed over a several week period; intervals in parentheses are based on a few days. Time of the 2002 magnitude 7.9 Denali earthquake is shown by the vertical line. Figure provided by Stephane Husen and made of data presented in Husen et al. (2004).

pre-earthquake frequency over a period of a few months. Following this earthquake, similar to previous earthquakes, the eruption frequency of some geysers increased, while for others it decreased.

The response of geysers to barometric pressure changes (e.g., White 1967), solid Earth tides (e.g., Rinehart 1972), and hypothetical preseismic strains (Silver & Vallette-Silver 1992) is by comparison weaker. Indeed, the very existence of some of these apparent correlations is controversial. Rojstaczer et al. (2003) analyzed eruption records in Yellowstone and found that geysers were insensitive to Earth tides and diurnal barometric pressure changes and concluded that geysers were not sensitive to strains less than about  $10^{-8}$ – $10^{-7}$ . One of these records was from Daisy geyser (Figure 9). Given that Daisy and other Yellowstone geysers did in fact respond to earthquakes that caused equally small static strain, we conclude that it is the dynamic strains that are primarily responsible for the observed responses.

**Figure 10**

Schematic illustration of a geyser showing the subsurface conduit and pathways through which water moves.



**Figure 10** shows schematically one model for how a geyser works and the pathways through which water moves in a geyser system. Water is recharged at the surface. Heat is provided from below. If the conduit cannot discharge water fast enough, the heat added from below generates a confined vapor phase, which in turn leads to an eruption. Ejection of liquid water and vapor from the conduit lowers pressure in the conduit, promotes boiling, and increases the amount of vapor present. Eventually the fluid pressure in the conduit becomes low enough to allow recharge from the surrounding rock matrix (Hutchinson et al. 1997), and the eruption cycle begins again. Geysers are thus similar to volcanoes in that they generate gas-driven, repeated eruptions. They differ, however, in the physical processes through which the erupting fluid enters the conduit and is stored in the subsurface.

In general, geyser eruptions begin with the discharge of water at temperatures less than the boiling point, followed by liquid-dominated fountaining, and finally a steam-dominated discharge that tapers off to a quiescent period. Ingebritsen & Rojstaczer (1993, 1996) show, using a numerical model of the flow shown in **Figure 9**,

that this eruption sequence can be reproduced for particular combinations of basal heat flow, conduit geometry, and permeability of the matrix and conduit. They infer a large ( $\geq 10^3$ ) permeability contrast between the geyser conduit and adjacent matrix. Ingebritsen & Rojstaczer (1993) conclude that much of the observed temporal variability in eruption frequency might be explained by changes in matrix permeability, which in turn govern the recharge of the conduit.

The high sensitivity and permanent (over a timescale much longer than eruption interval) response of geysers to small seismic strains would seem to require reopening, unblocking, or creation of fractures to create large enough permeability changes to influence eruption frequency. Geyser frequency is highly sensitive to the permeability of the geyser conduit. However, because the conduit is already very permeable, small strains in the conduit itself seem unlikely to have a significant effect on geyser eruptions (Ingebritsen et al. 2005).

In summary, the response of geysers to distant earthquakes is most easily explained by changes in permeability, and the sensitivity of geysers to earthquakes compared with Earth tides and changes in barometric pressure suggests that dynamic strains cause the response.

## 7. CONCLUDING REMARKS

Determining whether earthquakes trigger volcanic eruptions is of value beyond satisfying the curious looking for connections between catastrophic geological phenomena. If a connection exists, triggered eruptions provide a probe of magmatic and hydrothermal processes that might otherwise be difficult to study. These connections also provide insight into the physical processes that initiate eruptions, an understanding that is critical for recognizing and interpreting precursors to volcanic eruptions.

Several of the mechanisms discussed in this review, either individually or simultaneously, are plausible. Determining which are important in a given eruption is probably very challenging, and as far as we know, none have been convincingly identified for any of the triggered eruptions listed in **Table 1**. The best opportunity for identifying mechanisms is integrated monitoring of volcanoes. A combination of deformation, seismicity, and gas flux measurements may be able to identify the triggering processes that lead to eruption.

Hill et al. (2002) begin their review with the question commonly asked of volcanologists and seismologists when eruptions occur closely in time and space to large earthquakes: "Is this eruption related to that earthquake?" Hill et al. (2002) answer this question, "Well, maybe." We concur with Schminke (2004, p. 55) who suggests that the answer to the question whether earthquakes can trigger eruptions should now be, "In general yes, in this particular instance maybe—but we do not know exactly how."

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