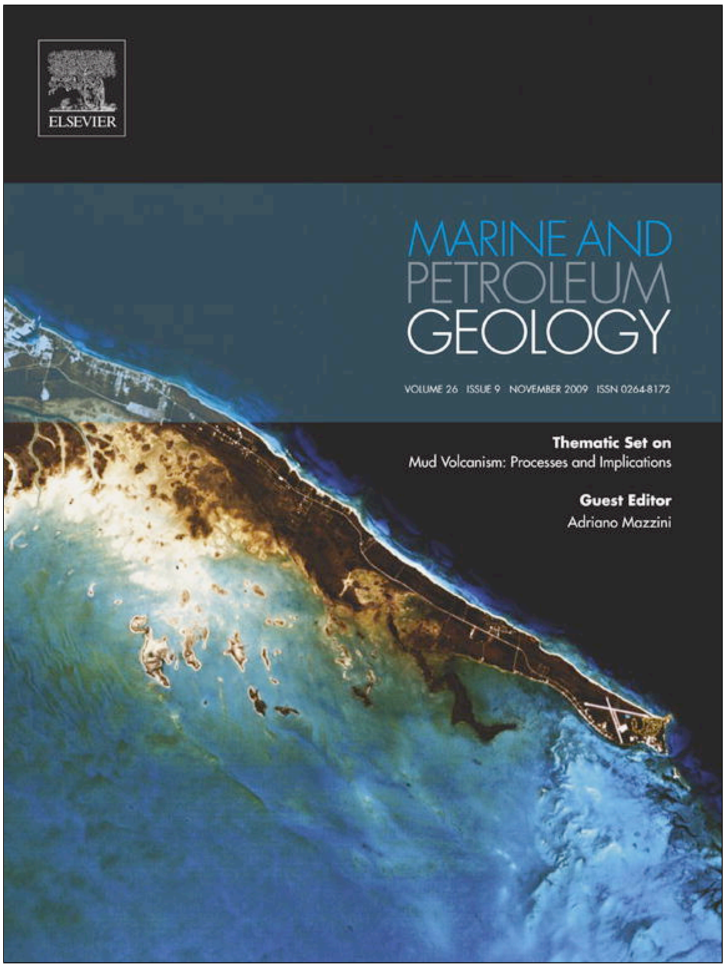
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Thematic Set on
Mud Volcanism: Processes and Implications
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Earthquake triggering of mud volcanoes

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ABSTRACT

Mud volcanoes sometimes erupt within days after nearby earthquakes. The number of such nearly coincident events is larger than would be expected by chance and the eruptions are thus assumed to be triggered by earthquakes. Here we compile observations of the response of mud volcanoes and other geologic systems (earthquakes, volcanoes, liquefaction, ground water, and geysers) to earthquakes. The compilation shows a clear magnitude–distance threshold for triggering, suggesting that these seemingly disparate phenomena may share similar underlying triggering mechanisms. The compilation also shows that pre-existing geysers and already-erupting volcanoes and mud volcanoes are much more sensitive to earthquakes than quiescent systems.

Several changes produced by earthquakes have been proposed as triggering mechanisms, including liquefaction and loss of strength, increased hydraulic permeability or removing hydraulic barriers, and bubble nucleation and growth. We present new measurements of the response of erupted mud samples to oscillatory shear at seismic frequencies and amplitudes, and find that loss of strength occurs at strain amplitudes greater than 10^{-3}. This is much larger than the peak dynamic strains associated with earthquakes that may have triggered eruptions or influenced already-erupting mud volcanoes. Therefore, we do not favor loss of strength as a general triggering mechanism. Mechanisms involving bubbles require significant supersaturation or incompressible mud, and neither condition is likely to be relevant. We analyze the response of the Niikappu group of mud volcanoes in Japan to several earthquakes. We find that this system is insensitive to earthquakes if an eruption has occurred within the previous couple of years, and that static strain magnitudes are very small and not correlated with triggering suggesting that triggering likely results from dynamic strain. Moreover, triggering may be frequency-dependent with longer period seismic waves being more effective at triggering. Available data are insufficient, however, to determine whether triggering characteristics at Niikappu are representative of triggered eruptions in general. Nor can we determine the exact mechanism by which dynamic (long-period) strains induce eruption, but given the apparent failure of all mechanisms except increasing permeability and breaching barriers we favor these. More observations and longer records are needed. In particular, gas measurements and broadband seismic data can be collected remotely and continuously, and provide key information about processes that occur during and immediately after the arrival of seismic waves.

1. Introduction

It has long been recognized that there is a spatial connection between the location of mud volcanoes and faults, leading to the proposal that "fault activity may be the key player governing the rate of mud extrusion" (Kopf, 2008). Indeed, the location of mud volcanoes is sometimes used to infer the presence of active fault systems (e.g. Lynch and Hudnut, 2008). Mud volcanoes have also been observed to erupt shortly after nearby earthquakes (e.g. Chigira and Tanaka, 1997). Mellors et al. (2007) analyzed the temporal and spatial relationships between earthquakes and mud volcano eruptions, and showed that there is a statistically significant number of eruptions triggered by earthquakes. However, this analysis, dominated by eruptions in Azerbaijan, also shows that most eruptions are not triggered – at least directly – by earthquakes. Nevertheless, understanding the causes of triggered eruptions provides the opportunity to learn about processes within mud volcanoes that are inaccessible to direct measurement or monitoring. Mud volcanoes are more than natural curiosities; they play an important role in subduction zone fluid cycling (Kopf et al., 2001) and contribute to the global atmospheric methane budget (Etiope and Milkov, 2004; Kopf, 2003).

Establishing a causal link between any particular earthquake–eruption pair is difficult. This uncertainty is highlighted by recent

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debate over whether the May 29 2006 eruption of a mud volcano in Sidoarjo, Indonesia was triggered by an earthquake (Manga, 2007; Mazzini et al., 2007), or by drilling activities (Davies et al., 2007, 2008). This question has important legal and financial implications. Whatever the cause, the eruption has had significant human and environmental impacts, with tens of thousands of people displaced and an estimated cost of tens of trillions of Indonesian rupiahs (billions USD). A better understanding of when and how earthquakes can trigger mud volcano eruptions is essential to address this controversy.

In this paper we review observations of mud volcano eruptions, and the response of ongoing eruptions to earthquakes. Because there are not very many documented examples of triggered mud volcano eruptions, we also discuss several geological systems unrelated to mud volcanoes, all of which respond to earthquakes. These analogous systems may share underlying physics that govern the response of geological fluids to earthquakes. Others have previously reviewed the effects of earthquakes on stream flow (Montgomery and Manga, 2003), ground water (Manga and Wang, 2007), volcanoes and geysers (Hill et al., 2002; Manga and Brodsky, 2006), and seismicity (Freed, 2005; Hill and Prejean, 2007; King, 2007). Given that this paper is presented as a review, there is inevitable overlap in content and conclusions with previous summaries (e.g., Mellors et al., 2007; Manga and Brodsky, 2006). The present discussion differs in that it adds to previous data sets, examines the response of already-erupting systems to earthquakes, and uses this albeit limited data set to test hypotheses and draw some new conclusions. In particular we show that 1) the magnitude and sign of static strain is not clearly correlated with triggered eruptions supporting mechanisms that result from dynamic strain; 2) we confirm that a repose time is needed for triggering; 3) we show that there may be a frequency dependence of dynamic triggering; 4) we finally document, using new lab measurements, that the strain amplitudes needed to reduce the strength of mud are much larger than those that trigger eruptions and that affect already-erupting mud volcanoes.

2. Mud volcanoes

Mud volcano eruptions are driven by a combination of high fluid pressures, gas exsolution, and the buoyant force of a low-density source layer; they are inhibited by impermeable structures between the mud source and the surface. Earthquakes can thus influence eruptions by increasing fluid pressure, causing the growth or nucleation of gas bubbles, or creating a hydraulic connection between the source and the surface by breaching the seal of overpressured reservoirs. In the absence of earthquakes, eruptions may be precipitated by changes in tectonic stress, changes in overburden (e.g. due to landslides), and/or changes in fluid pressure due to internal diagenetic processes such as clay mineral dewatering or methanogenesis.

Mud volcanoes occur in rapidly deposited sequences of under-consolidated sediments (Kopf, 2002). They have been found in many tectonic settings, but are more commonly found in compressive settings, where the tectonic stress assists the development of high pore pressures.

Mud source beds typically occur at depths of ∼1–3 km (Kopf, 2002; Davies and Stewart, 2005; Mazzini et al., 2007) but greater depths are possible (Kopf, 2008). Surface structures resemble those of their magmatic counterparts, featuring cones, vents, flows and calderas. While surface vents may have small diameters, seismic imaging suggests that conduits at depth may have diameters up to a few kilometer (e.g. Loseth et al., 2003; Talukder et al., 2003).

Gas emissions from onshore mud volcanoes are composed primarily of methane (70–99%) and CO_2 (Kopf, 2002). Submarine mud volcanoes emit methane and CO_2 along with other hydrocarbons, sulfur, and noble gases. The gas source can be significantly deeper than the mud source, e.g., ∼10–12 km in the Caucasus (Cooper, 2001).

3. Earthquakes as triggers

The energy delivered by an earthquake produces two different types of deformation: permanent deformation and a corresponding change in the stress state of the crust, and transient deformation associated with the passage of seismic waves. Hereafter, we refer to these strains as static and dynamic strain, respectively. The strain amplitude depends on distance from the fault rupture (see Table 1), moment release, and mechanical properties. Details of the rupture process such as directivity and regional differences in geology will affect the attenuation of dynamic strain with increasing distance.

Earthquakes in the brittle portion of the crust also place stress on the more plastic layers underneath; as stresses in these deeper layers relax, the stress diffuses outwards. This delayed, quasi-static stress transfer is similar to static stress transfer, except that it occurs over a period of years to decades (Freed and Lin, 2002; Marzocchi et al., 2002). This may play a significant role in triggering both volcanic eruptions and earthquakes (e.g. Marzocchi et al., 2002; Marzocchi, 2002; Pollitz and Sacks, 1997). Mellors et al. (2007) found, however, only very weak correlations between earthquakes and possible delayed triggered eruption of mud volcanoes.

4. Triggered eruption of mud volcanoes

Conceptually, an earthquake-triggered eruption is one in which the passage of seismic waves or coseismic change in crustal stress initiate a process leading to eruption. In practice, it is difficult to distinguish a true trigger from a mere coincidence. To prevent overinterpretation of statistical noise, many authors have required that events be proximate in time and/or space before they are considered examples of triggering (Linde and Sacks, 1998). Mellors et al. (2007), for example, analyzed only those eruptions occurring within ∼100 km of earthquakes. Others have required events at any distance to occur within a short time after the passage of seismic waves. For example, Moran et al. (2004) used a 1 h limit and Kane et al. (2007) a 2 day window to examine triggered earthquakes. Selva et al. (2004) used more intricate space–time constraints corresponding to the evolution of stress in a model system.

Mellors et al. (2007), performing an analysis similar to that done by Linde and Sacks (1998) for magmatic volcanoes, examined the historical record of mud volcano eruptions and found that significantly more mud volcanoes occur on the same day as earthquakes than would be expected by chance. We similarly restrict ourselves

Table 1
Approximate magnitudes of commonly induced stresses (Christiansen et al., 2005; Hill and Prejean, 2007; King et al., 1994; Saar and Manga, 2003). Dynamic stresses for the Landers earthquake are given for the region to the north, where rupture directivity produced maximum amplitudes.

	Stress (all in kPa)		
Earth tides		1	
Barometric pressure variations		2–4	
Ground water recharge		$1–10^2$	
1992 Landers earthquake – M_w 7.3	50 km	250 km	1000 km
Static stresses	10^2	1	10^{-2}
Dynamic stresses	10^4	10^3	10^2

to events that take place within a few days of the triggering earthquake. While we don't impose any additional a priori constraints on the spatial scale of triggering, because we are compiling results from other authors who do use limited spatial scales we may overlook some examples of extremely long-distance triggered events.

Fig. 1 shows the magnitude and hypocentral distance of earthquakes that triggered mud volcano eruptions. The examples plotted come from the compilation in Mellors et al. (2007), with additional events (1968, 1993) at Niikappu, Japan (see Table 2 and Chigira and Tanaka, 1997), Andaman Island after the 2004 Sumatra earthquake (Manga and Brodsky, 2006), three events in Makran (Delisle, 2005; Byrne et al., 1992), Baluchistan (Snead, 1964; Ambraseys and Bilham, 2003), Kandewari, Pakistan (Deville, submitted for publication), and two eruptions in Mongolia (Rukavickova and Hanzl, 2008). We also include eruptions triggered within 2 days compiled by Bonini (submitted for publication) from a variety of sources (Martinelli et al., 1989; Trabucco, 1895; De Brignoli di Brunhoff, 1836; Guidoboni, 1989; D'Alessandro et al., 1996; Silvestri, 1878; La Via, 1828). Table 2 summarizes this data. While the number of examples is modest, 29, there is a distance threshold over which triggering appears to occur and this distance threshold increases with increasing magnitude. Mellors et al. (2007) found that a distance threshold equivalent to ground shaking with Mercalli intensity about 6 was necessary, though not sufficient, to trigger eruptions. For reference, in this figure we also include an empirically derived distance threshold for the occurrence of liquefaction

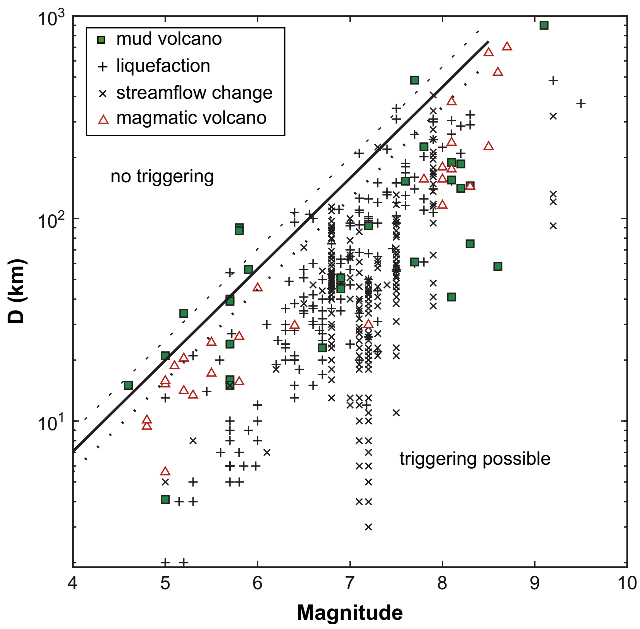

Fig. 1. Relationship between earthquake magnitude and the distance over which a variety of responses have been documented. Stream flow responses are indicated by a black ×, liquefaction by a black +, volcanoes by a red triangle, and mud volcanoes by a green square. Data from compilations in Lemarchand and Grasso (2007), Manga and Brodsky (2006), Mellors et al. (2007), Montgomery and Manga (2003), Wang et al. (2006) with additional mud volcanoes eruptions reported in Delisle (2005), Snead (1964), Ambraseys and Bilham (2003), Deville (submitted for publication), Bonini (submitted for publication) and Rukavickova and Hanzl (2008). The solid and dotted lines are the empirical liquefaction limit and error, respectively, determined by Wang et al. (2006). The two mud volcano outliers are Gobi Altav, Mongolia (Rukavickova and Hanzl, 2008) located 75 km from the a M 5.8 earthquake, and Kadewari, Pakistan (Deville, submitted for publication) located 482 km away from the epicenter of the 26 January 2001 Bhuj earthquake. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

(Wang et al., 2006), which proves also to provide a good (though empirical) estimate for the threshold distance for triggered mud volcano eruptions. The sloping lines in Fig. 1 show this empirical threshold.

Because of the limited number of triggered mud volcano eruptions, we include in Fig. 1 several other kinds of triggered events: magmatic volcanoes, liquefaction, and changes in stream flow. These are discussed in Section 5. Fig. 1 shows that all these triggered events can be described by a similar distance threshold.

4.1. Niikappu

Several mud volcanoes lie along the crest of an anticline near the city of Niikappu, Japan, and have been observed to erupt in response to earthquakes (Chigira and Tanaka, 1997). Fig. 2 shows the magnitude and hypocentral distance of historical earthquakes that occurred near these mud volcanoes (earthquakes from the National Earthquake Information Center catalog, hereafter NEIC, and Dunbar et al. (1992), both available at http://neic.usgs.gov/neis/epic/epic_global.html). In Fig. 2 we normalize the distance D from the epicenter by the square root of the fault rupture area, $\sqrt{A}$ (a characteristic length scale for earthquakes), according to an empirical relationship derived by Wells and Coppersmith (1994). We chose the square root of the rupture area instead of the rupture length because the aspect ratio of an earthquake rupture is magnitude-dependent. We emphasize that, while this normalization is physically meaningful in the near-field; in the intermediate- and far-field where triggered eruptions occur, there is an additional (but modest) distance dependence for dynamic strain (and will be seen later in Fig. 5). Nevertheless, the liquefaction threshold, shown in Fig. 1, has a nearly constant normalized distance of about 6. The normalized distance $D/\sqrt{A}$ provides a convenient measure of the distance of the triggered event relative to the fault size. In subsequent figures we will continue to normalize distance in this manner noting that earthquake magnitude influences the values of both the x and y axes. The history of triggered eruptions at Niikappu is consistent with the liquefaction limit of Wang et al. (2006) in that triggered events fall below the threshold distance. Significant earthquakes shown in Fig. 2 are also listed in Table 3.

The gray circles in Fig. 2 denote earthquakes that occurred within two years of an eruption, and the black circles show all other earthquakes. With one possible exception, triggered eruptions occur for all earthquakes within distances that might be expected to induce triggering based on the global compilation in Fig. 1, provided the earthquakes occur more than 2 years after previous eruptions. The exception is the November 6 1958 Kurile Islands earthquake, which is listed in Dunbar et al. (1992) with a surface wave magnitude of 8.7 but is given by others (e.g. Lander and Lockridge, 1989) as a magnitude 8.1–8.3 event. If we use instead a magnitude of 8.1, rather than 8.7, then the normalized distance $D/\sqrt{A}$ doubles and the 1958 Kurile earthquake is no longer an outlier. The pattern of response observed at Niikappu supports the finding of Mellors et al. (2007), based on an analysis of mud volcanoes in Azerbaijan, that a period of 1–2 years is required for the system to return to a critical state between eruptions.

Because the Niikappu mud volcano has responded to multiple earthquakes, we can use its response (or lack of response) to test different hypotheses about the triggering mechanism. We do so in Section 6 when we discuss proposed triggering mechanisms.

4.2. Lusi

On May 29th, 2006, a mud volcano began to erupt near the town of Sidoarjo in eastern Java. The ongoing eruption, now known as "Lusi", covers an area of 6.5 km^2 and has displaced tens of

Table 2
Triggered mud volcanoes.

Location	Date	Magnitude	Epicentral distance (km)	Reference
Niikappu, Japan	4-Mar-1952	8.6	58	Chigira and Tanaka (1997)
Niikappu, Japan	16-May-1968	8.2	186	Chigira and Tanaka (1997)
Niikappu, Japan	21-Mar-1982	6.7	23	Chigira and Tanaka (1997)
Niikappu, Japan	15-Jan-1993	7.6	153	Chigira and Tanaka (1997)
Niikappu, Japan	28-Dec-1994	7.8	226	Chigira and Tanaka (1997)
Niikappu, Japan	25-Sep-2003	8.3	145	Manga and Brodsky (2006)
Andaman Islands	26-Dec-2004	9.1	900	Manga and Brodsky (2006)
Kalamaddyn, Azerbaijan	28-Jan-1872	5.7	24	Mellors et al. (2007)
Shikhzairli, Azerbaijan	28-Jan-1872	5.7	40	Mellors et al. (2007)
Shikhzairli, Azerbaijan	13-Feb-1902	6.9	45	Mellors et al. (2007)
Bozakhtarma, Azerbaijan	13-Feb-1902	6.9	51	Mellors et al. (2007)
Ormara, Makran	28-Nov-1945	8.1	41	Delisle (2005)
Hingol, Makran	28-Nov-1945	8.1	189	Delisle (2005)
Gwadar, Makran	28-Nov-1945	8.1	155	Delisle (2005)
Thok, Baluchistan	30-May-1935	7.7	61	Snead (1964)
Livanoca, South Caspian	8-July-1895	8.2	141	Mellors et al. (2007)
Baciu, Romania	4-Mar-1977	7.2	92	Mellors et al. (2007)
Marazy, Azerbaijan	24-Sept-1848	4.6	15	Mellors et al. (2007)
Gobi Altay, Mongolia	4-Dec-1957	8.3	75	Rukavickova and Hanzl (2008)
Gobi Altay, Mongolia	15-June-2006	5.8	90	Rukavickova and Hanzl (2008)
Kandewari, Pakistan	26-Jan-2001	7.7	482	Deville (submitted for publication)
Regnano, Italy	11-Oct-1915	5.0	21	Martinelli et al. (1989) in Bonini (submitted for publication)
Portico di Romagna, Italy	4-Sept-1895	5.0	4.1	Trabucco (1895) in Bonini (submitted for publication)
Montegibbio, Italy	5-Apr-1781	5.8	87	De Brignoli di Brunhoff (1836) in Bonini (submitted for publication)
Montegibbio, Italy	91 BCE	5.7	16	Guidoboni (1989) in Bonini (submitted for publication)
Nirano, Italy	91 BCE	5.7	15	Guidoboni (1989) in Bonini (submitted for publication)
Paterno, Italy	13-Dec-1990	5.7	39	D'Alessandro et al. (1996) in Bonini (submitted for publication)
Paterno, Italy	4-Oct-1978	5.2	34	Silvestri (1978) in Bonini (submitted for publication)
Caltanizetta, Italy	5-Mar-1828	5.9	56	La Via (1828) in Bonini (submitted for publication)

thousands of people (Cyranoski, 2007). Prior to the eruption, Indonesian energy company PT Lapindo Brantas had been drilling an exploration well, Banjar Panji-1 (BJP-1), located about 150–200 m from the initial eruption location.

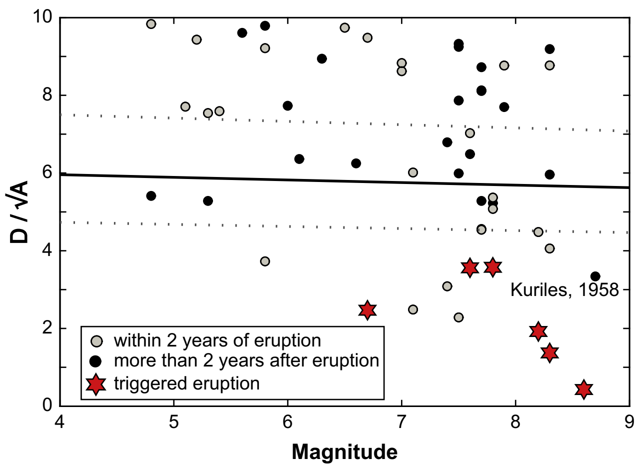

Fig. 2. Eruptions of Niikappu mud volcanoes in response to earthquakes. Red stars are earthquakes that triggered an eruption, black circles are earthquakes that did not trigger an eruption, and gray circles are earthquakes that did not trigger an eruption but occurred less than two years after a previous eruption (triggered eruptions from Chigira and Tanaka, 1997 and Mellors et al., 2007). Hypocentral distances D are normalized by the square root of the rupture area $\sqrt{A}$ (Wells and Coppersmith, 1994) so that the liquefaction limit (an empirical threshold in Fig. 1) becomes an approximately horizontal line. $D/\sqrt{A}$ is a measure of the number of fault dimensions separating the response from the hypocenter. The one outlier black circle corresponds to the 1958 Kurile earthquake whose magnitude is controversial. If a smaller magnitude of 8.1 is used (e.g., Lander and Lockridge, 1989) then it is no longer an outlier. The nearly horizontal line shows the empirical liquefaction threshold and uncertainty from Wang et al. (2006) (see also Fig. 1). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

The cause of the eruption is disputed. Two days before the eruption began, the M_W 6.3 Yogyakarta earthquake occurred 250 km to the southeast. Mazzini et al. (2007) argue that the eruption was triggered indirectly by the earthquake by reactivating a nearby fault. As noted by Manga (2007), however, there were previous larger, closer earthquakes that did not trigger an eruption, and this event falls well outside of the empirically observed relationship between magnitude and distance for hydrological responses to earthquakes. Davies et al. (2008) show that between dozens and hundreds of other earthquakes produced greater dynamic strains at the eruption site and did not trigger an eruption. This conclusion holds for all measures of dynamic strain, including peak ground velocity, peak ground acceleration, and the Arias Intensity that also accounts for the duration of shaking. Davies et al. (2008) also showed that there was no unusual build up of seismic shaking before the Yogyakarta earthquake. If the eruption was

Table 3
Significant historical earthquakes near Niikappu mud volcanoes. Earthquakes listed above the line triggered an eruption, earthquakes below the line did not. Not listed are earthquakes that occurred within two years of an eruption that did not trigger new activity. Data from Chigira and Tanaka (1997), Mellors et al. (2007), and the NEIC earthquake catalog. For static strain, + denotes volumetric expansion, and − volumetric contraction. Sources for focal mechanisms or slip models for the static strain calculations are given in the text.

Date	Magnitude	Depth (km)	Epicentral distance (km)	Static strain
4-Mar-1952	8.6	25	58	$+10^{-6}$
16-May-1968	8.2	9	186	$+10^{-9}$
21-Mar-1982	6.7	44	23	$+10^{-6}$
15-Jan-1993	7.6	102	153	$+10^{-10}$
28-Dec-1994	7.8	27	226	$+10^{-6}$
25-Sep-2003	8.3	27	145	$+10^{-6}$
6-Nov-1958	8.7	32	545	$+10^{-8}$
13-Oct-1963	8.3	47	639	-10^{-8}
17-Jun-1973	7.7	48	299	$+10^{-8}$

triggered by the earthquake, it would represent an unusually sensitive response, or a mechanism not shared by other mud volcanoes and other triggered hydrologic effects. In contrast, Davies et al. (2008) argue that pressure of drilling fluids within the BJP-1 well during the days following the earthquake was sufficient to initiate hydrofracturing and trigger an eruption.

Lusi is an unusual mud volcano in several respects. First, the eruption did not occur at a pre-existing edifice. The extent to which it occurred through pre-existing fractures or zones of weakness, rather than creating fresh hydrofractures, is unknown. Second, the mud erupts at relatively high temperatures, near 100 °C (Mazzini et al., 2007). Finally, most mud volcano eruptions are short-lived episodes, lasting only hours to a few days (Kopf, 2003; Mazzini et al., 2007), while the Lusi eruption has been ongoing for more than two years.

4.3. *Response of ongoing eruptions to earthquakes*

Fig. 3 shows the magnitude and normalized distances $D/\sqrt{A}$ of earthquakes that have produced a response in ongoing mud (Mazzini et al., 2007) and magmatic (Harris and Ripepe, 2007; Walter et al., 2007) eruptions, geysers (Hutchinson, 1985; Silver and Valette-Silver, 1992; Husen et al., 2004), and oil wells (Beresnev and Johnson, 1994). These ongoing eruptions are influenced by earthquakes at greater distances than the events shown in Fig. 1 – they fall well beyond the empirical liquefaction limit shown with the solid line. Changes to ongoing eruptions and well production can be accomplished by altering existing flow paths, rather than opening new ones. As discussed in Section 5.4, geyser behavior is also very sensitive to the structure of the existing conduit. The relative insensitivity of new eruptions (and other effects discussed in Section 5) to earthquakes may therefore be related to the additional energy required to create or re-open closed flow paths, or to cause rheological transitions such as liquefaction.

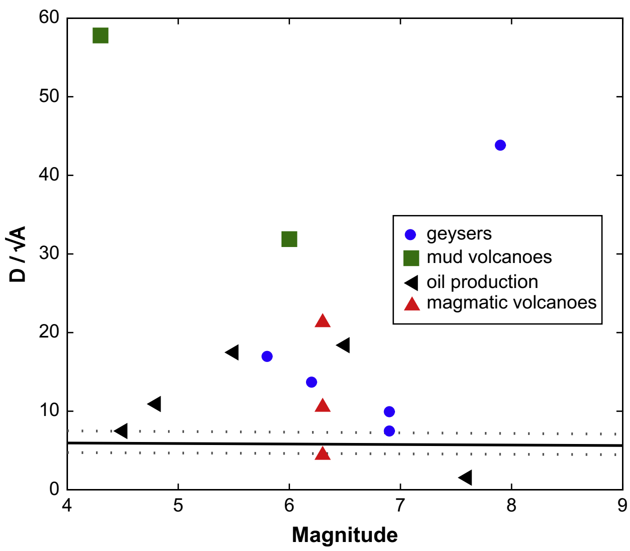

Fig. 3. Earthquakes which have triggered responses to ongoing magmatic (red triangles) and mud volcano eruptions (green squares), changes in geyser eruption frequency (blue circles), and changes in oil production (black triangles). Hypocentral distance is again normalized by the square root of the rupture area so that it is clear that these are triggered at distances far greater than those shown in Fig. 1. Data compiled from Hutchinson (1985), Silver and Valette-Silver (1992), Beresnev and Johnson (1994), Husen et al. (2004), Harris and Ripepe (2007), Mazzini et al. (2007), Walter et al. (2007). The nearly horizontal line shows the empirical liquefaction threshold and uncertainty from Wang et al. (2006) (see also Fig. 1). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Fig. 4 shows changes in eruption rate during the first six months of activity at the Lusi mud volcano. It also shows the distance to nearby earthquakes, again normalized to fault rupture area, to provide an estimate of relative triggering potential (note that an inverted scale for $D/\sqrt{A}$ is used compared to previous figures). Mazzini et al. (2007) noted that significant increases in eruption rate followed a couple of days after earthquakes on September 6 and 8. An earthquake on September 22 was also followed by increased flow. However, an equally significant earthquake on August 2 did not have any apparent effect on the eruption rate. Additionally, it is difficult to tell whether the earthquakes occurring on July 17 had an effect, because daily records were not kept during June and July 2006 (Mazzini et al., 2007). Finally, increases in flow rate observed at the beginning of August 2006 and during October and November 2006 were not associated with large earthquakes. However, these increases may also be artifacts that originate from the difficulty in accurately measuring discharge and the limited measurement interval of 4–5 days (Mazzini, personal communication).

Although the earthquakes of September 6 and 22 may have triggered changes in the ongoing eruption, the precise nature of the relationship between earthquakes and eruption dynamics is unclear. At times, Lusi has exhibited quasi-regular pulsations in eruption volume (Mazzini et al., 2007). During September 2006, these occurred roughly every 30 min, while during February 2007 they were every 1.5 h (Mazzini et al., 2007). Inconsistent and infrequent observations make it difficult to correlate these pulsations with earthquakes or other observable changes. If the pulsations are caused by dynamics similar to what drives geysers, however, we would expect their frequency to be very sensitive to passing seismic waves or other changes in subsurface conditions (see Section 5.4).

Automatic gas monitoring techniques permit sustained, high-frequency monitoring programs, although data interpretation is complicated by the response of mud pools to local weather conditions and the shifting of bubble release sites within the monitored pool. For example, radon levels have been considered indicative of flow conditions at depth (Martinelli et al., 1995). Intriguingly, Martinelli et al. (1995) found that radon levels increased significantly before two small local earthquakes. Albarello et al. (2003), however, found that radon measurements do not respond to the small strains caused by earth tides.

Yang et al. (2004, 2006) present the highest-frequency monitoring of gas emissions at mud volcanoes we are aware of, with concentrations of several gas species measured automatically every 2 min during 2001–2002. Several peaks in the observed CO_2/CH_4 ratio are roughly coincident with local earthquakes (Yang et al., 2006, Fig. 9). However, not all peaks are associated with earthquakes, and Yang et al. (2006) do not establish the statistical significance of the association. Such tentative associations are intriguing, but no study has yet shown a convincing relationship between gaseous emissions and earthquakes.

5. Related triggered events

Because of the limited number of observed earthquake-triggered mud volcano eruptions, it is useful to consider possible similarities with other earthquake-triggered events. Fig. 1 showing the magnitude and hypocentral distance of earthquakes that triggered mud volcano eruptions also shows liquefaction, changes in stream flow, and triggered magmatic eruptions.

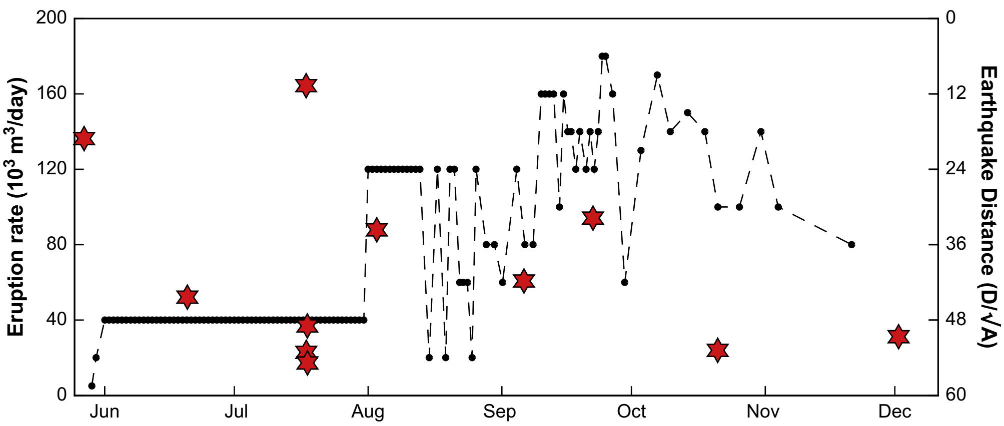

Fig. 4. Eruption rate at Lusi (Mazzini et al., 2007) and nearby earthquakes (red stars; NEIC catalog), May–November 2006. Monitoring during the months of June and July was not daily, and is not as reliable as the rest of the record; after September 26 observations were recorded every 4 days (Mazzini et al., 2007). Note that the vertical scale that characterizes the earthquake distance (normalized by earthquake size characterized by the square root of rupture are) is inverted compared to the previous figures. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

5.1. Liquefaction

Shaking a saturated, unconsolidated sediment can disrupt the grain structure and transfer pressure from grain–grain contacts to the pore fluid. When the pore pressure exceeds the overburden, the material behaves in a liquid-like manner, rather than as a solid; the process is known as liquefaction. This type of undrained consolidation by cyclic deformation is one way to induce liquefaction and the most thoroughly studied mechanism in the geotechnical literature. The loss of strength in liquefied soil is a cause of significant damage during earthquakes. Additionally, when liquefied material is ejected to the surface it can form small-scale volcano-like edifices.

Historical observations of earthquake-induced liquefaction are limited to shallow, saturated, unconsolidated sediments (Ambraseys, 1988). The liquefaction observations compiled in Fig. 1 are surface observations of this type of shallow liquefaction. However, some studies have suggested that preserved sand pipes were produced by liquefaction at greater depths (e.g. Kenkmann, 2003). Determining the conditions under which liquefaction can occur is of tremendous importance to seismic hazard evaluation.

The occurrence of liquefaction depends on both the duration and amplitude of shaking, as well as inherent soil properties. Wang and Chia (2008) showed that at the greatest distances where liquefaction occurs, the energy density in the seismic waves is more than 10^2 times smaller than the laboratory-determined energy density required to cause liquefaction in even sensitive soils (Green and Mitchell, 2004), and is also much greater than the energy density needed to initiate undrained consolidation. Wang and Chia (2008) thus conclude that most of the observations of liquefaction in Fig. 1, from between about one fault length up to the liquefaction threshold, are the result of permeability changes and/or breached barriers that redistribute pore pressure.

Liquefaction can begin with the arrival of the strongest ground acceleration (e.g. Youd et al., 2004), but pore pressure continues to rise even after the strongest ground motion has passed (e.g. Holzer et al., 1989). Liquefaction may also occur shortly after the main shock due to the development of Kelvin–Helmholtz instabilities (Heifetz et al., 2005). Most interestingly, liquefaction has also been observed minutes (Holzer et al., 2004) to one day (Ishihara, 1984) after earthquakes. Wang (2007) proposed that long delays arise from the creation of a hydraulic connection to pressurized aquifers (see Sections 6.4 and 6.5).

Fig. 1 includes an empirical magnitude–distance relationship for liquefaction (Wang et al., 2006). This relationship was derived using a compilation of historical observations of liquefaction. Wang et al. (2006) interpret this threshold as corresponding to a constant seismic energy density of $\sim$0.004–0.1 J/m^3. Other magnitude–distance relationships have been derived for the occurrence of liquefaction. In particular, Trifunac and Todorovska (2004) develop an empirical relationship based on the kinetic energy of seismic waves integrated over the duration of the earthquake, with a threshold of 1.5×10^{-3} J s. Trifunac and Todorovska (2004) also present a brief review of other liquefaction thresholds.

5.2. Volcanoes

Anecdotal observations of volcanic eruptions occurring shortly after large earthquakes are not uncommon (e.g. Darwin, 1840). Earthquake and eruption catalogs provide statistical support for the occurrence of long-distance triggering; both Linde and Sacks (1998) and Lemarchand and Grasso (2007) found that volcanic eruptions occur more frequently in the few days following large earthquakes. Reviews of proposed mechanisms for short-term dynamic triggering of volcanic eruptions are given by Hill et al. (2002) and Manga and Brodsky (2006). Static stress also plays a significant role in near-field interactions between tectonic earthquakes and volcanic eruptions (e.g. Feuillet et al., 2006; Walter and Amelung, 2006; Walter, 2007; Walter and Amelung, 2007).

Volcanic activity also generates seismicity (see McNutt, 2005, for a review), and it can be difficult to separate earthquakes that triggered an eruption from earthquake–eruption pairs associated with a common underlying process. Lemarchand and Grasso (2007) found that volcanic eruptions are more common immediately preceding large earthquakes not otherwise associated with the volcano, as well as immediately following, indicating that volcano–seismic coupling may take place at larger scales than previously thought.

Earthquakes can also influence ongoing volcanic eruptions. Harris and Ripepe (2007) observed increased thermal emissions from Indonesian volcanoes Merapi and Semeru beginning 3 days after the M6.3 Yogyakarta earthquake. These already-erupting volcanoes appear to be more sensitive to seismicity than dormant volcanoes (see Fig. 3).

The response of volcanoes to earthquakes happens on time-scales of days (Linde and Sacks, 1998) to decades (Marzocchi, 2002). Current eruption catalog data are not precise enough to find correlations between earthquakes and eruptions on timescales shorter than a day.

5.3. Changes to springs, streams, and ground water levels

Manga and Wang (2007) review the postseismic response of springs, streams, and ground water levels. In some cases, response patterns can be tied to the sign of volumetric stress change (e.g. Muir-Wood and King, 1993). For example, Jonsson et al. (2003) document a pattern of ground water level changes after an earthquake in Iceland that mimics the volumetric strain. In other cases, for example at Sespe Creek in Southern California (Manga et al., 2003), the observed spatial distribution of the responses can only be explained by a process that does not depend on the sign of static stress change. Changes caused by static and dynamic stresses are not mutually exclusive and either or both may be operating in any given case.

Changes to springs and streams can appear within a few minutes to a few days of an earthquake, and typically decay over a period of weeks to a few months (Montgomery and Manga, 2003).

5.4. Geysers

Many geysers exhibit strong eruption periodicity which is very sensitive to small changes in subsurface conditions. Geysers respond to events with static strains less than 10^{-7} and dynamic strains less than 10^{-6} (e.g. Hutchinson, 1985; Silver and Valette-Silver, 1992). While geysers in Yellowstone are not sensitive to strains less than 10^{-8}–10^{-9} (Rojstaczer et al., 2003), a lower limit has not been established for other systems. The observed response occurs very quickly after earthquakes, and decays over a period of days to months (Husen et al., 2004).

There are two classes of models that explain the response of geysers to seismic disturbance. One invokes bubble nucleation, the other changes in permeability.

Steinberg et al. (1981, 1982a,b,c) developed a conceptual model of the geyser process that requires supercritical temperatures inside a chamber. Steinberg et al. (1982b) attribute reductions in geyser period to increases in background seismicity. Laboratory model geysers based on this scheme (Shteinberg, 1999) are quite sensitive to vibration and will erupt in response to stomping or shouting. Additionally, slight changes in the available nucleation sites for bubbles will manifest as changes in the sensitivity of eruptions to earthquakes. The eruption timing of this model can also be altered by changing the rate of heating.

Ingebritsen and Rojstaczer (1993), by contrast, modeled geysers as a vertical, highly permeable fracture zone. They found that geyser periodicity is strongly dependent on the permeability contrast between the geyser chamber and the surrounding rock, and the length of the fracture zone. Small changes to permeability (as discussed in Section 6.4) can produce significant changes in eruption frequency.

Periodic eruptive behavior can also arise from the formation of gas slugs (Taylor bubbles) in long, narrow channels. Lu et al. (2005) present a review of geyser-like behavior in low-temperature two-phase flows (i.e., CO_2 and water) in wells and engineering applications. Although the effect of vibration on the eruption frequency of these systems has not been studied in detail, high seismic frequency vibration ($\sim$60 Hz) has been found to increase the amount of gas–liquid interaction in narrow capillaries (Vandu et al., 2004). If vibration affects mass transfer in the system it might also affect eruption dynamics and periodicity.

5.5. Triggered earthquakes

Earthquake triggering by other earthquakes is by far the most abundantly documented triggered phenomenon. We do not include triggered earthquakes in the compilation shown in Fig. 1 as they are not directly connected to fluid-flow changes seen at Earth's surface, even though fluids may play a role in triggering. Triggered earthquakes differ from the other phenomena shown in Fig. 1 in that their distance threshold occurs at much greater distances. Magnitude 9 earthquakes, such as the 2004 Sumatra–Andaman earthquake, appear to be able to trigger earthquakes globally (e.g. West et al., 2005; Velasco et al., 2008).

Conventionally, near-field aftershocks are thought to be a response to static stress changes. However, recent work has shown that dynamic stresses may be important aftershock triggers even in the near-field (Felzer and Brodsky, 2006; Pollitz and Johnston, 2006). Hill and Prejean (2007) and King (2007) provide more complete reviews of the current understanding of earthquake-triggered earthquakes by dynamic and static stresses, respectively.

Earthquakes triggered by distant events ($\gg$1 fault length) occur most commonly, though not exclusively, in active hydrothermal areas (Husker and Brodsky, 2004; Hill and Prejean, 2007) but they have been documented in a wide range of tectonic environments (Velasco et al., 2008). They are typically coincident with the arrival of surface waves (Hill and Prejean, 2007). Some earthquake sequences, however, occur during the days following an earthquake, on timescales consistent with fluid migration (e.g. Miller et al., 2004). West et al. (2005) additionally documented triggering during the phase of maximum extensional stress, though it is not yet clear whether this is the case in general. The phase correlations found by West et al. (2005) are best explained by simple shear failure. Hill (2008) notes that simple Coulomb failure models predict that triggering should occur with the arrival of the Love wave, and triggering by both Love and Rayleigh waves seems to be common (e.g. Velasco et al., 2008).

While fluids are often invoked to explain triggering, see review by Hill and Prejean (2007), triggered earthquakes might also be caused by nonlinear rate-and-state or granular time-dependent frictional processes (Gomberg, 2001; Johnson and Jia, 2005).

6. Proposed mechanisms

A wide range of mechanisms have been invoked to explain triggered phenomena. Here we list these, identify (when possible) what features of triggered eruptions they predict, and then test the proposed mechanism with available observations. In the sections that follow, we first identify features of the stresses that cause triggering. We then use these features to distinguish between proposed mechanisms.

It is important to reemphasize that most eruptions are not triggered by earthquakes. For example, Mellors et al. (2007) calculated that of the 299 eruptions recorded in Azerbaijan in the past 2 centuries, only 5 have been associated with earthquakes. The chance that a small additional fluid pressure will be sufficient to trigger an eruption can be estimated using an argument similar to that presented by Manga and Brodsky (2006) for magmatic volcanoes. Assuming that overpressure develops at a constant rate in the absence of earthquakes, the fraction of triggered eruptions should be approximately equal to $\Delta P_{eq} T_e / \Delta P_c T_{eq}$, where ΔP_{eq} and T_{eq} are the pressure generated during an earthquake and the earthquake recurrence times, respectively, ΔP_c is the critical overpressure at which the mud volcano will erupt, and T_e is the eruption recurrence time. In the Azerbaijan eruption catalog, there are about 3.2 eruptions per year (Mellors et al., 2007). Earthquakes large enough to produce shaking of intensity IV or greater occur in this region approximately once per year (Mellors et al., 2007). To produce the observed fraction of triggered eruptions, the additional overpressure generated by these earthquakes must be ~5% of the critical pressure.

If we assume that ΔP_c is approximately equal to the tensile strength of rocks, $\mathcal{O}$ (1 MPa) (e.g. Cho et al., 2003), proposed triggering mechanisms should generate overpressures of ~50 kPa. Alternatively, we can use the fluid pressures required to initiate mud loss from boreholes, commonly determined during drilling using a leak-off test. A leak-off test near the site of the Lusi mud volcano found a strength of ~20 MPa (Davies et al., 2008). Using this value means that triggering earthquakes should create overpressures of ~1 MPa.

6.1. Static or quasi-static stress transfer

In order to trigger an eruption, proposed mechanisms must either appeal to static stresses (Section 3), or convert transient stresses from ground shaking into a semi-permanent effect. They can occur by altering the hydraulic characteristics of the material surrounding the mud source (Sections 6.3 and 6.4) or by altering the mud itself (Sections 6.5 and 6.6).

Static stress transfer is often discounted as a potential trigger because it is of smaller magnitude and decays much faster with distance than dynamic stresses (e.g. Manga and Brodsky, 2006). Some triggered eruptions occur within one fault length of the earthquake (normalized distances less than about 1, see for example Fig. 2) where static stress changes might be large enough to trigger eruptions (see Table 1). However, triggered eruptions occur to distances about 7 fault lengths (see Fig. 1), where only dynamic stresses are significant.

If static stress changes play a dominant role in triggering eruptions, we expect to see a strong correlation between the occurrence of triggered eruptions, stress magnitude, and the sign (compression or dilation) of the static stress lobe in which they occur. Following the approach in Manga et al. (2003) we can test this hypothesis by examining the response of a single system to multiple earthquakes. In this case we use the Niikappu mud volcano. We also restrict our analysis to large earthquakes that occurred more than 3 years after eruptions so that our conclusions will not be influenced by the repose time mud volcanoes might require to be triggered again (see Section 4.1). We calculated the volumetric strain using Coulomb 3.1.09 (Lin and Stein, 2004; Toda et al., 2005). Calculated signs and orders of magnitude of the volumetric strain are listed in Table 2, where + denotes dilatation and − denotes contraction. Focal mechanisms and slip models come from a numbers of sources: for the 1982, 1993, and 1994 earthquakes we use the Global Centroid-Moment-Tensor (CMT) catalog (globalcmt.org); for the 1952 earthquake we used the slip distribution in Table 1 of Hirata et al. (2003); for the 1958, 1963, 1968, 1973 and 2003 we used, respectively, Fukao and Furumoto (1979), Kanamori (1970), Kanamori (1971), Nakanishi et al. (2004), and Yagi (2004). For this last earthquake, if we use instead the Global CMT focal mechanism, we obtain a volumetric strain of -10^{-6}.

There are two features of the strains in Table 2 that lead us to disfavor static stress change as a dominant control on triggering. First, the pore pressure changes and stresses caused by these strains are very small. For a linear poroelastic material, the pore pressure change for undrained conditions is given by $\Delta p = -BK_u\epsilon_{kk}/3$ where B is the Skempton coefficient (about 1 for mud), K_u is the undrained bulk modulus (about 6 GPa for mud), and ϵ_{kk} is the volumetric strain (Wang, 2000). The strains listed in Table 2 imply stresses and pore pressure changes less than about 3 kPa (sometimes less than 1 Pa) and hence smaller than those expected to influence eruptions and smaller than or comparable to those from tides and barometric pressure changes. Second there is no clear magnitude threshold for triggering. Moreover, pore pressure changes are decreases, the opposite sign of pressure changes normally expected to induce eruption (though we note that Walter and Amelung, 2007 also found a correlation between volcanic eruption and dilatation). Identifying the consequence of the sign of volumetric strains is difficult in this setting, and indeed in most volcanic arc settings, because the location of the thrust earthquakes along the subducted plate interface causes dilatation in almost all cases at the site of the Niikappu mud volcanoes. For the 1963 and 2003 earthquakes, the mud volcano, however, is located close to a nodal plane and hence the sign of the strain will be especially sensitive to details of the slip model.

6.2. Frequency dependence of observations

Large-magnitude earthquakes last longer than small ones, and release more energy at low frequencies. Thus, the magnitude dependence of observed effects can help to distinguish among mechanisms involving dynamic stress by constraining the importance of the frequency and duration of shaking.

We estimated peak ground velocity (hereafter PGV) and strain at Niikappu using a recently published attenuation relationship for PGV in Japan (Kanno et al., 2006) and an assumed shear wave velocity of 2500 m/s. The approximate peak transient strain is then given by:

$$\varepsilon \sim \mathrm{PGV}/V_s \tag{1}$$

where ϵ is strain amplitude and V_s is the shear wave velocity. This peak strain estimate has a large associated uncertainty, because it does not account for the earthquake focal mechanism or the effect of directivity, and approximates rupture location by the hypocenter.

The data shown in Fig. 5 suggests that mud volcanoes are more sensitive to longer period waves. Fig. 5 shows that strain amplitudes associated with triggered events are smaller than those from smaller earthquakes that did not trigger eruptions (compare the black circles for $6 < M < 7$ with the triggered eruptions for $M > 7.5$).

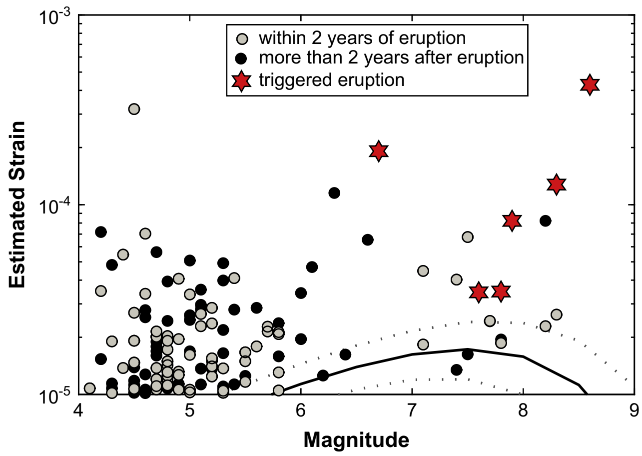

Fig. 5. Estimated peak transient strain from historical earthquakes near Niikappu mud volcanoes. Red stars are earthquakes that triggered eruptions, black circles are earthquakes that did not trigger eruptions, and gray circles are earthquakes that did not trigger eruptions, but occurred within two years of a previous eruption. The solid and dotted lines are the empirical liquefaction limit and error, respectively, determined by Wang et al. (2006). Strain was estimated using the peak ground velocity attenuation relationship of Kanno et al. (2006), and an assumed shear wave velocity of 2500 m/s. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

We leave this as a tentative conclusion because the estimated strain has significant uncertainty and we have a limited number of triggered events.

Other triggered phenomena show a frequency dependence. Common parameterizations of the intensity of ground motion during an earthquake, including peak ground acceleration (PGA) and peak ground velocity (PGV), are measures of different frequency components of seismic energy. Wong and Wang (2007) examined the regional distribution of PGA, PGV, and other measures of shaking after the 1999 M_w 7.6 Chi-Chi, Taiwan, earthquake, and found that PGV, which is sensitive to low frequencies, was the best predictor of liquefaction and changes in stream flow. This may indicate that the low frequency component of seismic energy is more effective at causing these effects. Youd and Carter (2005) and Holzer and Youd (2007) also document an association of liquefaction with low frequency ground motion. However, it may also be due to frequency-dependent attenuation in areas where liquefaction occurs. Liquefaction has been observed to amplify long-period ground motions and attenuate short periods (Youd et al., 2004). We address the frequency dependence of liquefaction in more detail in Section 6.5.

The frequency sensitivity of triggered earthquakes is unknown, and published results are contradictory. Brodsky and Prejean (2005) found that long-period (>30 s) waves are more effective at triggering earthquakes in the magmatically active Long Valley Caldera. However, Kane et al. (2007) found that if there is a frequency dependence for triggered earthquakes in Southern California, the most effective frequencies must be >2 Hz.

Seismic or ultrasonic waves are sometimes used to increase production at oil wells (Beresnev and Johnson, 1994). Beresnev et al. (2005) found that low frequencies (frequencies of 10–60 Hz were studied) are more effective at removing small droplets of oil that can block pore throats.

6.3. Dilatancy

Strong ground shaking can also disrupt the arrangement of larger grains in unconsolidated soils, increasing the material's pore volume as the material dilates. Bower and Heaton (1978) found that this effect can explain water level declines observed in Alaska after the 1964 Great Alaska earthquake. However, dilatancy is only observed in the near-field (e.g. Wang et al., 2001), and should always produce water level declines and a decrease in pore pressure. If dilatancy occurs in the source region for mud volcanoes, then there should be a zone around large earthquakes where eruptions are inhibited, rather than triggered.

The distance over which dilatancy occurs is just a few percent of the fault length, assume for argument sake 5%. Assuming triggering is possible over an area extending to 5 fault lengths, there is 10^4 times more area in which dilatancy does not occur yet triggering is possible. The number of triggered eruptions is far too small to identify such a small region with suppressed triggering. However, the strong association of faults and mud volcanoes implies that mud volcanoes have no difficulty forming in regions where dilatancy can occur, though whether the mud volcanoes along these faults can be triggered is not known and the compilation in Fig. 1 contains no examples.

6.4. Increasing permeability and opening fractures

Changes in permeability, or the breaching of barriers to permit fluid flow, have been invoked to explain all the triggered phenomena described in Section 5, including changes in stream flow (e.g. Rojstaczer et al., 1995; Wang et al., 2004a,b), liquefaction (e.g. Wang, 2007), and geyser eruption interval (e.g. Ingebritsen and Rojstaczer, 1993).

Seismic waves can increase permeability by flushing particles from fractures (Brodsky et al., 2003) or opening fissures (King et al., 1999). Elkhoury et al. (2006) documented this permeability increase at two wells in Southern California and found that the post-earthquake permeability increase scales with the peak ground velocity. Brodsky et al. (2003) calculated fluid velocities induced within an aquifer by passing seismic waves, and found that they are sufficient to flush small (micron-sized) particles from fractures. The removal of clay particle clogs from pore throats can also explain increased permeabilities seen at high seismic frequencies (25–75 Hz) by Roberts (2005).

In order for an increase in permeability to drive an eruption, it must create or enhance a connection to a pressurized water source. High pore pressures will then diffuse across the former barrier. The resulting elevated pore pressures in regions of lower clamping stress can be sufficient to continue fracture propagation on their own, even after the seismic waves have passed (Wang et al., 2004a,b). The timescale of the observed response will depend on the distance from the breached barrier and the hydraulic diffusivity of the intervening media, and will be reflected in time delay of the triggered eruption.

Hill et al. (2002) and Miller et al. (2004) consider scenarios where the high-pressure reservoir is located at near-lithostatic pressures – pockets of brine and steam below the brittle-plastic transition (Hill et al., 2002) or CO_2 (Miller et al., 2004). From aftershock patterns in the Northern Apennines, Miller et al. (2004) infer a high-pressure (10–20 MPa) pulse of fluid moving at about 1 km/day from a source at about 5 km depth. Such a mechanism could explain triggering delays of a few days.

Husen and Kissling (2001) proposed that subduction zone earthquakes increase permeability along the décollement and allow the episodic injection of fluids into the overlying wedge. This scenario is supported by narrow heat flow anomalies that can be associated with a burst of fluid (Grevemeyer et al., 2006) and observations of extensional fracturing in the outer portions of the fault damage zone (Vannucchi and Leoni, 2007). Husen and Kissling (2001) also found a 50 day lag between the Antofagasta earthquake and the appearance of increased fluid pressure in the lower crust, too slow to have a triggering effect at the surface within a few days of the earthquake. Brown et al. (2005), however, found that increased fluid flow occurred $\sim$20 days before the beginning of episodes of seismic tremor, possibly indicating a shared underlying cause.

We know of no test of this class of mechanisms using the existing data. However, by analogy to other situations in which dynamic strains changes permeability, we expect that mud volcano systems might experience similar changes. The short interval between the earthquake and eruption for the examples shown in Fig. 1 requires that the permeability changes or breached barriers lie close to, or within, the mud volcano source.

6.5. Liquefaction

Liquefaction can act to promote mud volcano eruption in two ways: by increasing pore fluid pressures, and by fluidizing the mud prior to eruption (thereby reducing the resistance to motion). In laboratory experiments and field observations, liquefaction is associated with excess pore water pressures of a few to several 10 s of kPa (e.g. Kostadinov and Towhata, 2002; Youd et al., 2004).

The sensitivity of liquefaction to different seismic frequencies is a topic of considerable debate, and the answer has important implications for seismic hazard evaluation. Laboratory experiments on saturated, unconsolidated soils (e.g. Yoshimi and Oh-Oka, 2005) and other dense particle suspensions (e.g. Sumita and Manga, 2008) have generally not shown frequency-dependent behavior. However, most laboratory setups are too small to capture any resonant effects in the soil column, and experiments with large soil columns are not performed at seismic amplitudes. Numerical models of high amplitude ground motion in full soil columns point to the possibility of frequency-dependent behavior. For example, Popescu (2002) and Popescu et al. (2006) found that low frequency inputs produce greater pore pressure rises, while Ghosh and Madabhushi (2003) found that frequencies of 1–2 Hz were most effective. As discussed in Section 6.2, field observations appear to favor low frequencies being more effective at causing liquefaction, and mud volcanoes also appear to be more sensitive to low frequency waves.

Producing or sustaining liquefaction much beyond the duration of ground shaking requires an alternate source of energy input into the system. This might come from a connection to a previously isolated pressurized aquifer, as suggested by Wang (2007). Such mechanisms are essentially the same as those discussed in the previous section on permeability changes. In the rest of this section, we thus only consider the possible role of liquefaction induced by cyclic shear deformation.

In order to better quantify the changes in mud strength expected from seismic waves, we measured the rheology of several muds obtained from the Lusi mud volcano using a cone-plate rheometer (Haake RheoScope1), the same instrument and procedure used by Sumita and Manga (2008). Rheological measurements have the potential to identify the relevant frequency components and thresholds for changes (Ancey and Jorrot, 2001). The samples were collected and made available to us by Adriano Mazzini and can be cross-referenced with Mazzini et al. (2007). The water content of the mud was 43% by mass, or 63% by volume for a solid density of 2.2 g/cm^3 – lower than volume fractions of the first erupted mud, ~70%, and greater than that of later erupted mud, ~30% (Mazzini et al., 2007). The rheometer deforms an unconfined sample by applying a cyclic stress $\tau(t)$ necessary to produce a specified time-varying strain:

$$\gamma(t) = \gamma_0 \sin(\omega t) \quad (2)$$

where $\omega = 2\pi f$ is the angular frequency of oscillation and γ_0 is strain amplitude. Because a viscoelastic material may exhibit a phase lag between stress and strain, the applied stress corresponding to the specified strain can be written:

$$\tau(t) = G'\gamma_0 \sin(\omega t) + G''\gamma_0 \cos(\omega t) \quad (3)$$

where G' is the storage modulus and G'' is the loss modulus. For each measurement, we pre-sheared the sample through a strain of 10^{-3} at a frequency of 1 Hz for 60 s. For each measurement point, the sample was deformed through 10 cycles and then measurements were averaged over the subsequent 10 cycles. To prevent sample dessication (and associated volume decrease and rheological changes), we surrounded the sample with an oil bath that did not come into contact with the rheometer cone. To assess sample contamination and dessication, we performed the pre-shear test after making measurements and compared the post-measurement storage and loss moduli to the pre-measurement values; the greatest change we observed was a 13% reduction. The results are shown in Fig. 6, and confirm that strength reduction occurs at a critical strain of between $10^{-2}$and 10^{-3}. Our results, like some others, do not show any significant dependence on the frequency of deformation.

The strain at which the mud begins to lose strength is similar to, though a bit larger than, the critical strains found in previous studies on sands for the onset of pore pressure increase (e.g. Dobry et al., 1982; Vucetic, 1994). We were not able to make measurements on samples with much larger water contents representative of early erupted mud, nor much lower water contents representative of mud in the source. We believe that the critical strain amplitude for weakening shown in Fig. 6, however, is likely to be representative because in experiments with an analogue system, the critical strain amplitude at which suspensions lost strength depended only very weakly, if at all, on the fluid concentration (Sumita and Manga, 2008). This strain is quite a bit larger than the peak dynamic strains associated with triggered eruptions at Niikappu (Fig. 5) and we conclude that liquefaction through shear deformation and undrained consolidation cannot be responsible for many or even most triggered eruptions.

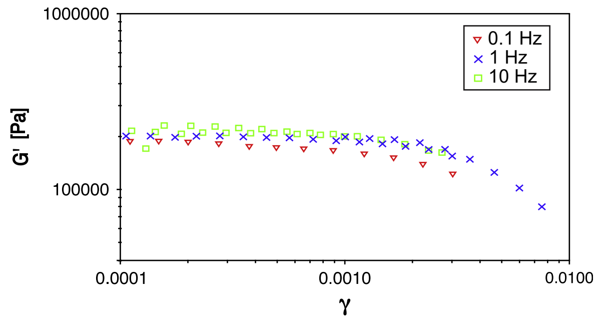

Fig. 6. Storage modulus of a representative mud sample (JV07-05) as a function of oscillatory strain amplitude at 0.1 Hz (red triangles), 1 Hz (blue crosses), and 10 Hz (green squares). We observed a decrease in storage modulus between strain amplitudes of 10^{-3} and 10^{-2} for all samples tested, independent of frequency. Mass fraction of water is 43%. The staggered appearance of the blue crosses and green squares is due to a small amount of hysteresis between two measurement runs with overlapping strain amplitudes. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

We can estimate the maximum strain produced by earthquakes that may have affected the ongoing eruption at Lusi from the predicted peak ground velocity, using the same procedure as in Section 4.1. Assuming a shear wave velocity of 2500 m/s, the dynamic strain produced at Lusi by these earthquakes is on the order of 10^{-6}–10^{-7}, very far from the critical strain required for strength reduction to occur. Therefore, the strength reduction of mud is unlikely to be a key contributor to either triggering or changes in eruption rate at the Lusi mud volcano.

6.6. Bubbles

Mechanisms involving bubble formation or growth have been proposed as ways by which earthquakes might trigger magmatic eruptions. Mud volcanoes typically also include gas phases, composed primarily of methane and CO_2, which may play a role in driving eruptions. The buoyant force from bubbles can act as a positive feedback to processes which reduce the external pressure on a system. Bubbles may also drive other rheological changes. Pralle et al. (2003), for example, found that injecting gas bubbles into a suspension of quartz powder promotes liquefaction.

For a new bubble to nucleate, the supersaturation pressure must be high enough to overcome surface tension. The rate of homogeneous bubble nucleation is very sensitive to the supersaturation pressure, which for many gases in water can be several MPa (e.g. Finkelstein and Tamir, 1985), much smaller than the nucleation barrier in magmas (e.g. Mangan and Sisson, 2000). The larger the critical pressure for bubble nucleation, the greater the potential for rapid bubble growth. The Steinberg models discussed in Section 5.5 appeal to this mechanism, namely for pressure variations caused by seismic waves to induce nucleation in a superheated (analogous to supersaturated) fluid. This model is consistent with laboratory geysers, but the degree of supersaturation in natural field settings should be much lower because particulate matter will aid nucleation and limit the degree of supersaturation.

Davis et al. (2007) presented a numerical model of the behavior of a dense (particle fraction >0.5) suspension of crystals in magma, where shaking on short timescales causes particle–particle pressures to increase and fluid pressures to decrease. In highly viscous magma the calculated pressure drop is enough to form bubbles, and the effect is strongest at high frequencies. This relationship is opposite to that suggested by the data in Fig. 5, namely that longer period waves are more effective at triggering. The effect is also strongly viscosity-dependent, and has not been observed in lab experiments with dense particle suspensions (see Stickel and Powell, 2005, for a review).

Bubbles will contract when pressurized and expand when depressurized. When a bubble expands, volatiles diffuse into it; when it contracts, volatiles diffuse back into the fluid. When bubbles are subjected to oscillatory strain with a period shorter than the diffusive timescale, they experience an asymmetric flux of volatiles across the bubble-fluid boundary. Because the bubble has a larger surface area during the expansive phase, there is a net flux of volatiles into the bubble (Hsieh and Plesset, 1961). This process is known as "rectified diffusion".

Sturtevant et al. (1996) found that rectified diffusion in a water–CO_2 system can generate significant ($\sim$10 kPa) overpressures in geothermal systems, although the pressure generated declines steeply at depths less than a few hundred meters (Sturtevant et al., 1996, Fig. 7). Accounting for the pressure dependence of water solubility in magma significantly limits the amount of pressure that can be generated in volcanoes (Ichihara and Brodsky, 2006). However, unlike magmatic systems, dissolved gases in water typically do obey Henry's Law, so this caveat is less restrictive for mud volcanoes.

The overpressure developed by rectified diffusion scales roughly linearly with the duration of shaking (Sturtevant et al., 1996). Therefore, if this is an important cause of triggered eruptions, we would expect to find that for a given peak transient strain amplitude, large-magnitude earthquakes trigger more eruptions than small ones, consistent with the observation in Fig. 5. Nevertheless, given that high supersaturation is unlikely and the theoretical results in Ichihara and Brodsky (2006), rectified diffusion is unlikely to play a role in triggering.

Bubbles that form at the bottom of a chamber are under higher pressure than bubbles that form at the top. If bubbles rise in an incompressible medium, they carry this overpressure with them, and pressure within the fluid will increase (Sahagian and Proussevitch, 1992). This process has been termed "advective overpressure". However, most sediments are highly compressible so, as pointed out in the quantitative analyses by Bagdassarov (1994) and Pyle and Pyle (1995), this mechanism is unlikely to be reasonable on physical grounds. Additionally, as bubbles rise within the mud volcano, they will lose gas to the solution unless the entirety of the system is already saturated or oversaturated. Advective overpressure requires a time delay related to the rise time of bubbles, and has no expected frequency or duration dependence (Linde et al., 1994).

6.7. *Gas hydrates*

Subaqueous mud volcanoes commonly occur in association with gas hydrates (Milkov, 2000). Gas hydrates are a high-pressure, low-temperature mineral phase in which gas molecules, typically methane, are surrounded by a cage of hydrogen-bonded water. Their presence is limited to deep-water sediments and polar environments where they occur in association with permafrost (Kvenvolden and Rogers, 2005).

The majority of (observed) triggered mud volcano eruptions are from subaerial mud volcanoes, where gas hydrates do not occur. However, gas hydrates may play a role in subaqueous eruptive processes. Mud volcanoes on the bottom of Lake Baikal, for example, are likely related to the dissolution of methane hydrates by hydrothermal fluids (Rensbergen et al., 2002). Obzhirov et al. (2004) attributed an increase in methane emissions and "flares" of high bubble concentrations to seismic activity. Similarly, Mau et al. (2007) found a link between seismic activity and methane emissions at seep sites off the coast of Costa Rica. We speculate that pulses of warm fluid released by earthquakes may dissolve gas hydrates and produce such a link. If this is the case, we would expect delays between increased methane concentrations and earthquakes to be consistent with the migration timescale of hydrothermal fluids from depth, but this test only works if we are lucky enough, or persistent enough, to observe the beginning of these episodes.

7. Concluding remarks

Several of the mechanisms discussed in this review may play a role in triggering eruptions. They are not for the most part mutually exclusive, and several may act simultaneously during any given instance of triggering. Moreover, different mud volcanoes may not be affected to the same extent by the same mechanisms. Determining which mechanisms predominate for specific eruptions, or in general, is likely to be quite difficult.

Despite the limited number of triggered eruptions, we can draw some tentative inferences about triggering mechanisms, though the limited number of eruptions prevents a statistical analysis of these inferences. First, dynamic rather than static strains are likely to dominate triggering because their magnitude is much larger. We find no evidence of a correlation of static stress changes and triggered eruption at the Niikappu mud volcano. Second, there appears to be a frequency dependence to triggering, with longer period waves being more effective. Third, the strain amplitudes required to initiate undrained consolidation, or liquefaction by cyclic deformation, are far too large to explain the great distances over which triggered eruptions occur. We do not favor mechanisms that directly involve nucleating bubbles or growing bubbles because previous theoretical analyses have shown that these processes are physically implausible or require very special (high supersaturation) conditions. This leaves processes that breach hydraulic barriers or increase permeability in some other manner. This class of mechanisms is not unique to mud volcanoes: it has also been proposed to explain many other triggered phenomena including changes in stream flow (e.g. Rojstaczer et al., 1995; Wang et al., 2004a,b), liquefaction at distances greater than about 1 fault length (e.g. Wang, 2007), triggered earthquakes (e.g. Brodsky et al., 2003), changes in spring temperatures (e.g. Mogi et al., 1989), and changes in water level in wells (e.g. Elkhoury et al., 2006).

More detailed and complete observations of triggered eruptions are needed to make progress and to perform meaningful statistical tests of proposed mechanisms. Information about the time delay between earthquakes and eruptions, and the frequency and duration of ground shaking that produce a response, should be particularly useful. Current mud volcano eruption catalogs list eruptions to the nearest day, but do not contain complete information about the time at which the eruption began or how eruptive behavior changes over time. In fact, defining a meaningful start time for eruptions is difficult, as the climactic eruption is probably preceded by various signs of unrest. Automatic monitoring of gas emissions and the installation of broadband seismometers near active mud volcanoes are two promising monitoring techniques.

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