Interactions between earthquakes and volcano activity
Nicolas Lemarchand and Jean-Robert Grasso

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[1] Using the 1973–2005 worldwide catalogues for $M \geq 4.8$ seismicity and VEI $\geq 0$ volcano eruptions, we find a significant, when tested against catalogue randomizations, increase of eruption onsets on the earthquake day. This result emerges from stacking time series of daily eruption rates relatively to earthquake time, $t_0$, over the whole seismicity catalogue. It is stronger for earthquake-volcano pairs for which the volcano is within ten rupture size from the epicenter. These results show that $M \geq 4.8 - $ VEI $\geq 0$ earthquake-volcano pairs are as important for interaction processes as the $M \geq 7$ and VEI $\geq 2$–5 pairs previously reported to interact. The clustering in time for earthquake-eruption pairs is not bounded to $t_0$. It remains above the background noise 6–10 days before and after $t_0$, and follows a power law distribution. These patterns, which are similar to the direct and inverse Omori’s laws observed for tectonic earthquakes, are the first evidence for the volcano eruptions to be contemporary of a stochastic brittle damage in the earth crust. The clustering we observe in time and space (i) rejects the earthquake triggering as the single process that drive earthquake-volcano interactions; (ii) supports a regional tectonic coupling in the earth crust. The clustering we observe in time and space (i) rejects the earthquake triggering as the single process that drive earthquake-volcano interactions; (ii) supports a regional tectonic coupling in the earth crust. The clustering we observe in time and space (i) rejects the earthquake triggering as the single process that drive earthquake-volcano interactions; (ii) supports a regional tectonic coupling in the earth crust; (iii) is larger for earthquake-volcano interactions than the one observed by Lindé and Sacks [1998]. Conditioning the time on the earthquake occurrence time, $t_0$, we stack time series of eruptions relative to the earthquake date, for different distance and magnitude ranges (Figure 1). A peak in the daily eruption rate emerges on the earthquake day for volcano eruptions within 30 km from the earthquake (Figure 1b) whereas no signal is observed for earthquake-eruption pairs without distance sorting (Figure 1a) in the period 1973–2005. By exploring space and time interrelations between earthquakes and volcano eruptions, we recover similar patterns closer to the volcanoes than the one observed by Lindé and Sacks [1998] for larger earthquakes, i.e., a peak in eruption rate at $t_0$ for $M \geq 8$ earthquakes in the 0–250 and 500–750 km distance ranges from the volcanoes, and within 200 km distance for $M \sim 7$–7.9 earthquakes. Within 30 km of a volcano are the regular volcano tectonic (VT) earthquakes that are classically related to brittle failure induced by magma transport within the volcano edifice (for a review, see, e.g., McNutt [2002]). The VT patterns as resolved by local seismic monitoring are small $M \leq 3$–4 local earthquakes within the volcano edifice, $\Delta \leq 5$–10 km [e.g., McNutt, 2002; Scott, 1989]. These events are not included in our global $M \geq 4.8$ catalogue. Thus the peak value we observe for earthquake-volcano interaction for $\Delta \leq 30$ km is not due simply to regular VT events. Additionally the highest signal to noise ratio for the peak value of eruption rate is not observed on the volcano edifice per see, including a 15 km epicenter location accuracy (Figures 1 and 2).

[5] To explore further the possible coupling between earthquakes and volcano eruptions we select earthquake-eruption pairs by using the $\Delta/L$ ratio. This ratio normalizes the earthquake-eruption distance, $\Delta$, by the earthquake rupture length, $L$. This technique is derived from the one used to analyse earthquake-eruption interactions [e.g., Felzer and Brodsky, 2006].

[6] The best signal to noise ratio is obtained for $\Delta/L \leq 10$ (Figure 1d), i.e the earthquake-eruption interactions are
statistically higher when the volcanoes are at distances less than ten fault lengths from the earthquakes. For \( \Delta/L \leq 10 \) and all distance ranges, the signal to noise ratio is better than when selecting event pairs with distance criterion only (Figures 1b, 1c, and 2). The \( t_0 \) peak value of eruption rate is related to 29 earthquake-eruption pairs for which the magnitude ranges from a \( M = 4.8 \) earthquake on the volcano to \( M = 7.8 \) at a 160 km distance (Table S1).\(^1\) All these observations are given with a 99% confidence interval. Significant \( t_0 \) peak values of eruption rate are not observed when randomizing the initial catalogue either on eruption dates or volcano coordinates. By randomizing the volcano locations and the eruption times we always accept the peak value in the observed eruption rates at \( t_0 \) above the 99.7% confidence level.

3. Patterns of Earthquake-Eruption Interactions

3.1. Earthquake-Eruption Pairs at Intermediate Distance: Extending the VT Event Definition

\[^7\] The 29 earthquakes, that interact with eruptions at \( t_0, \Delta/L \leq 10 \), range in magnitudes and distances from a \( M = 4.8 \) earthquake on the volcano to \( M = 7.8 \) at a 160 km distance (Table S1). These patterns do not fit the long recognized volcanic earthquake definition [e.g., \( M \leq 3–4, \leq 5–10 \) km McNutt, 2002; Scott, 1989]. There is no quantitative definition of a VT event in the volcanology community, and also there is a lack of quantitative definition of earthquake foreshocks and aftershocks. Bak et al. [2002] successfully collapse all the California seismicity onto the same curve, demonstrating that there is neither a maximum distance nor a maximum time to define an earthquake as an aftershock, both limits being bounded by the catalog length. The same way, our results suggest that VT events, when defined as earthquakes which interact with a volcano eruption, can be as large as 4.8–7 in magnitude and up to \( \Delta/L \leq 10 \).

3.2. Sizes of Events for Interacting Earthquake-Eruption Pairs

\[^8\] The distribution in sizes of the 29 earthquake-eruption pairs that cluster at \( t_0 \), with \( \Delta/L \leq 10 \), significantly differ from the size distributions for both the earthquake catalogue and the eruption catalogue (Figure 3). The probability for a given VEI on the earthquake day is larger.
than the one expected from the Smithsonian catalogue (Figure 3a).

To compare the magnitude distribution of earthquakes which interact with a volcano eruption at $t_0$ with those of the USGS catalogue, we remove the effects on the probability distribution of sizes that are induced by conditioning the earthquake catalogue with $\Delta/L \leq 10$. Accordingly, we built synthetic earthquake catalogues by randomly picking one master point in seismic areas, and by selecting, in the USGS earthquake catalogue, 30 events which follow $\Delta/L \leq 10$ relatively to this master point. The synthetic distributions significantly differ from the initial Gutenberg-Richter distribution (Figure 3b). The distribution of sizes for the 29 earthquakes involved in the $\Delta/L \leq 10$ earthquake-eruption clustering at $t_0$, is significantly below the synthetic distributions we constructed using $\Delta/L \leq 10$ (Figure 3b). Selecting earthquake by (i) a $t_0$ coupling with eruption onsets and (ii) $\Delta/L \leq 10$, increases the relative contribution of the smaller earthquake size to earthquake-eruption interactions.

3.3. Time Clustering of Earthquake-Volcano Interactions: Inverse and Direct Omori-Like Patterns

Our analysis shows that the clustering of earthquake-eruption pairs exists a few days before and after $t_0$. Although the associate rate values are smaller than the $t_0$ peak value, they remain above the noise level and fit an increase of rate toward $t_0$ whether before or after the $t_0$ reference event (Figure 4). These power law increases of event rates toward $t_0$ are reminiscent of the direct and inverse Omori’s laws found before and after tectonic earthquakes, respectively. These two Omori’s laws are observed when stacking time series of seismicity before and after earthquakes in the same way we stacked eruption rates before and after earthquakes [Omori, 1896; Jones and Molnar, 1979; Helmsatter and Sornette, 2003].

The small number of events above the noise level imposes rough estimates of the exponent values for the Omori’s law in the 0.8–0.5 range.

To compare the after- and fore-shock rates triggered by earthquake-earthquake and earthquake-eruption interactions, respectively, we use the same criteria for both data sets. Accordingly, using the USGS catalogue for stacked...
Aftershocks of large earthquakes, ebook at the data within the one day accuracy we are bound to use for eruption onset timing, are negative evidences for any specific so-called dynamic triggering process at $t_0$.

4. Discussion and Conclusion

When applying the same techniques and using the same robustness criteria that demonstrate that $M \geq 7$ earthquakes are efficient in triggering both volcano eruptions and each other [e.g., Linde and Sacks, 1998; Manga and Brodsky, 2006], we show that using $M \geq 4.8$, $VEI \geq 0$ and $\Delta/L \leq 10$, 1973–2005, 0.3% of eruptions interact with earthquakes. This rate is the same than the 0.4% value found for $M \geq 8$, $VEI \geq 2$ and $\Delta \leq 800$ km, i.e., $\Delta/L \leq 5$, 1500–2005 by Manga and Brodsky [2006]. The pioneering work of Linde and Sacks [1998], which focussed on large $M \geq 7$ earthquake-large $VEI \geq 2$ eruption interactions is only one aspect of the possible earthquake-volcano interactions, with the Linde and Sacks results corresponding to $\Delta/L \leq 5$ interactions. By normalizing the earthquake-volcano distances by the earthquake size, we highlight the importance of smaller earthquakes in volcano-earthquake interactions, as proposed for earthquake-earthquake interactions by Helmstetter [2003]. The $\Delta/L \leq 10$ threshold for which the interactions are strongest selects distances within the distance range observed for earthquake interactions [Felzer and Brodsky, 2006], i.e., the so-called aftershocks patterns. Such normalisation by earthquake sizes supports a frequency independent process for earthquake-eruption interactions as suggested for earthquake-earthquake interactions [e.g., Manga and Brodsky, 2006].

Manga and Brodsky [2006] estimated that a 0.4% fraction of triggered eruptions is too large to be expected from the interplay between seismic loading rate and rate of increase of magma overpressure rate with time. This argues for the peak activity of eruptions we observe on the earthquake day not to be driven solely by earthquake triggering. Furthermore, any eruption rate triggered by earthquakes through either static stress changes or dynamic wave skaking is expected to spread a few days-months after the seismic loading, not before it; that is, we expect a non symmetrical eruption rate around $r = 0$. This is not in agreement with our observation (Figure 4). The time clustering of earthquake-eruption pairs is spread significantly 6–10 days around $t_0$.

On the other hand, these results can be read as purely volcano driven seismicity that extends beyond the regular definition of VT events, these later being bounded to be relatively small $M \leq 3–4$ events, in the immediate $\Delta \leq 5–10$ km vicinity of a volcano. Because the best signal to noise ratio is roughly constant up to $\Delta/L = 10$, our results are not driven solely by earthquakes very close, i.e., $\Delta/L \leq 1–3$, to the volcanoes. Thus the minimum interaction distance we
have to consider is in the 20–50 km range (see Table S1), i.e., the order of magnitude of crustal thicknesses.

[16] For seismicity rate after and before eruptions, we resolved a power law increase in event rates toward \( t_0 \). Although the limited number of earthquake-volcano pairs does not permit determination of accurate values for each of the power law exponents, the 0.5–0.8 values we observed after-before eruptions are significantly smaller than the 0.9–1 values before and after tectonic earthquake-earthquake pairs in the same catalogue. These changes in seismicity rate after and before eruptions are positive evidence for (i) the coupling between earthquakes and volcano eruptions to induce similar seismicity patterns to those observed for earthquake-earthquake interactions; (ii) the slower the loading rate change, i.e., the day-month scale of the volcano processes relative to the second-minute earthquake dynamics, the slower the induced seismicity rate changes. While most of these deterministic properties can be reproduced by many physical processes, including but not restricted to the rate and state friction law models [e.g., Dieterich, 1994], the statistical properties we observe for stacked time series of earthquake-eruption pairs remain unexplained.

[17] Because of the reciprocity in time between eruption onsets and earthquake occurrences, the clustering patterns we resolve in Figure 4 question the causality that drives the earthquake-eruption interactions. First we can reject earthquake triggering as the single process involved in increasing the eruption rate. Earthquake triggering cannot induce eruption before the earthquake, as tested by randomizing eruption time. Second, the observed coupling between earthquakes and eruption onsets can possibly be driven by volcano dynamics alone with power law seismicity rates after and before eruptions. This interpretation requires extending the regular VT earthquake definition to larger magnitude and larger distance, i.e., \( M \geq 4.8 \) and \( \Delta / L \leq 10 \). The last possible process is a common external driving that correlates the regional eruptions and seismicity. As resolved by the signal to noise ratio from the catalogues, the scale of such a crustal-mantle coupling has a minimum size of a few hundred km and a minimum time scale of a few tens of days. One possible candidate for such a regional stress field increase can be a deep magma storage area at the base of the lithosphere.

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References