Earth and Planetary Science Letters 277 (2009) 399–407

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

journal homepage: www.elsevier.com/locate/epsl

The influence of great earthquakes on volcanic eruption rate along the Chilean subduction zone

Sebastian F.L. Watt *, David M. Pyle, Tamsin A. Mather

Department of Earth Sciences, University of Oxford, Parks Road, Oxford OX1 3PR, UK

ARTICLE INFO

Article history:
Received 14 May 2008
Received in revised form 23 September 2008
Accepted 1 November 2008
Available online 13 December 2008

Editor: C.P. Jaupart

Keywords:
seismic triggering
volcanic eruption
Chile
Andean southern volcanic zone
great earthquake
eruption rates

ABSTRACT

Seismic activity has been postulated as a trigger of volcanic eruption on a range of timescales, but demonstrating the occurrence of triggered eruptions on timescales beyond a few days has proven difficult using global datasets. Here, we use the historic earthquake and eruption records of Chile and the Andean southern volcanic zone to investigate eruption rates following large earthquakes. We show a significant increase in eruption rate following earthquakes of $M_W>8$, notably in 1906 and 1960, with similar occurrences further back in the record. Eruption rates are enhanced above background levels for ~12 months following the 1906 and 1960 earthquakes, with the onset of 3–4 eruptions estimated to have been seismically influenced in each instance. Eruption locations suggest that these effects occur from the near-field to distances of ~500 km or more beyond the limits of the earthquake rupture zone. This suggests that both dynamic and static stresses associated with large earthquakes are important in eruption-triggering processes and have the potential to initiate volcanic eruption in arc settings over timescales of several months.

1. Introduction

Seismic activity can elicit a variety of dynamic crustal responses, including seismic (Hill et al., 1993), hydrological (e.g., Brodsky et al., 2003; Montgomery and Manga, 2003) and magmatic (Linde and Sacks, 1998; Hill et al., 2002; Manga and Brodsky, 2006; Harris and Ripepe, 2007; Walter and Amelung, 2007) effects, in some cases at distances far in excess of the source-fault dimensions. The precise mechanisms of seismic volcanic eruption triggering, particularly at distances far from the rupture zone, remain enigmatic. Static stress changes in regions close to the fault rupture may explain processes leading to eruption (e.g., Walter and Amelung, 2007), but the magnitude of these stresses decays rapidly with distance. At greater distances, dynamic stresses associated with the passage of seismic waves, are often cited as a possible eruption trigger (Linde and Sacks, 1998; Manga and Brodsky, 2006). These explanations focus on seismic wave interactions with crustal fluids, causing fluid movement, disruption or bubble growth through a variety of possible mechanisms leading to magmatic overpressure. The timescale of such responses is poorly understood. Where studies have identified examples of seismically-triggered phenomena they have focussed on occurrences immediately following an earthquake (e.g., Linde and Sacks, 1998), where the suggestion of a causal link is straightforward. One such case is the eruption of Cordón Caulle volcano, Chile, 38 h after the M_W 9.5 Chilean earthquake in 1960 (Barrientos, 1994; Lara et al., 2004). However, without an agreed causal mechanism or careful analysis of event records this and similar examples, might be dismissed as coincidences. To investigate whether a relationship exists between non-volcanic earthquakes and volcanic eruptions requires a systematic examination of event records. Linde and Sacks (1998) used global earthquake and eruption datasets and found that within two days of a large earthquake more explosive eruptions occur than expected, up to a distance of 750 km. Manga and Brodsky (2006) quantified this relationship, and showed that 0.4% of explosive volcanic eruptions occur within a few days of large distant earthquakes; a much larger proportion than expected (0.01–0.1%) if there were no causal relationship.

These prior studies did not find evidence for triggered response times longer than about five days, but this does not necessarily mean they do not occur. We might expect eruptions triggered after a longer delay to be less frequent than those triggered within days of an earthquake, with potential variability between arc settings. Thus, such events may not be detectable above the natural background variability of eruption rate in global data sets, as studied by Linde and Sacks (1998). Hydrological observations of wells have shown persistent pressure responses following distant earthquakes, but in some cases the response is delayed by several weeks, and is of a magnitude that cannot be explained by static stress changes alone (Brodsky et al., 2003; Montgomery and Manga, 2003). Similarly, interaction with crustal magma bodies may not be manifested as surface activity (eruption) for timescales of months or more. Mechanisms involving

* Corresponding author. Tel.: +44 1865 272000; fax: +44 1865 272072.
E-mail address: Sebastian.Watt@earth.ox.ac.uk (S.F.L. Watt).

0012-821X/$ – see front matter © 2008 Elsevier B.V. All rights reserved.
doi:10.1016/j.epsl.2008.11.005

near-field static stress changes have been highlighted by Walter and Amelung (2007), who note that following four $M_W \geq 9$ subduction zone earthquakes a number of adjacent arc volcanoes erupted in the following three years, while Marzocchi (2002) suggests that there is a relationship between earthquakes and the largest volcanic eruptions operating on a timescale of up to 35 yr. Thus, while individual instances of delayed triggered eruptions have been proposed, demonstrating whether a general relationship between earthquakes and eruptions exists on timescales longer than a few days has proven difficult. Examination of data from individual seismically-active volcanic regions, rather than a global data set, may help elucidate such relationships by investigating the effects of large earthquakes in a single setting. Here, we investigate the relationship between large earthquakes and volcanic eruptions using the historical records from one arc, the Andean southern volcanic zone (SVZ), and the earthquake history of the adjacent subduction zone.

2. The Chilean subduction zone and SVZ

The SVZ is an ideal area to study relationships between volcanism and earthquakes. For around 3200 km of the Chilean coastline the Nazca plate is subducted beneath the western edge of South America along a straight north–south margin. Subduction is oblique and proceeds at ~8.4 cm yr^{-1} (DeMets et al., 1990). The SVZ runs along the southern 1400 km of this margin, from 33.4–45.9°S, and contains over 60 volcanoes considered to have been active in the Holocene (Siebert and Simkin, 2008-). The plate boundary between these latitudes is characterised by extremely large thrust earthquakes, and includes the largest event recorded in modern times, the M_W 9.5 earthquake of 22nd May 1960 (Barrientos and Ward, 1990; Cisternas et al., 2005). It is also a region where a large proportion of the potential rapidly-triggered eruptions identified by Linde and Sacks (1998) and Manga and Brodsky (2006) are located.

2.1. Earthquake records

The dates of large earthquakes along the Chilean subduction zone are relatively well documented from the 16th century onwards (Lomnitz, 1970; Kelleher, 1972; Comte et al., 1986), although there are some uncertainties regarding magnitude and fault rupture length due to sparsely distributed damage reports. In Table 1, we have compiled a record of all events with estimated moment magnitudes, M_W, >7.5. There is general agreement in the literature as to which were the largest events ($M_W \geq$~8), hereafter termed 'great earthquakes'. These main ruptures dominate subduction zone convergence and regional crustal deformation (Klotz et al., 2001). We focus on events of this magnitude, since their rupture lengths (>150 km) greatly exceed the typical spacing of local arc volcanoes (~25 km) and they thus present the most likely candidates for producing an eruption response. The great earthquake record shows a remarkably well-defined cyclicity (Fig. 1), with main ruptures generally propagating southwards from the epicentre, and stepping southward along the subduction zone in a temporally clustered pattern, as adjacent locked segments are seismically loaded (e.g., Li and Kisslinger, 1984). This cyclic pattern has been noted numerous times (e.g., Kelleher, 1972; Lomnitz, 1985; Nishenko, 1985; Comte et al., 1986), though the magnitude of events is not necessarily consistent, with some smaller 'great earthquakes' only partially releasing seismic strain, and themselves being intercyclic to much larger events (e.g., Cisternas et al., 2005; Moernaut et al., 2007).

Table 1
Chilean main ruptures ($M_W \geq$~8) between 32° and 46°S

Date	Epicentre	Estimated rupture length (km)	Estimated M_W[a]
8 Feb 1570	36.7°S 73.0°W	280	8–8.5
17 Mar 1575[b]	32.5°S 71.5°W	110	? 7.5
16 Dec 1575	39.8°S 72.8°W	800	>8.5
13 May 1647	32.9°S 71.3°W	380	8.5
15 Mar 1657	36.7°S 73.0°W	390	8
8 Jul 1730	33.1°S 72.0°W	560	>8.75
24 Dec 1737	39.8°S 73.0°W	530	7.5–8
25 May 1751	36.7°S 73.0°W	440	>8.5
19 Nov 1822	33.1°S 71.8°W	220	8.3
20 Feb 1835	36.6°S 73.0°W	440	8–8.3
7 Nov 1837	40.0°S 73.0°W	630	>8
17 Aug 1906	33.0°S 72.0°W	330	8.3
1 Dec 1928	35.0°S 72.0°W	140	8.2
25 Jan 1939	36.3°S 72.3°W	190	8.0
22 May 1960	39.5°S 74.5°W	940	9.5
4 Mar 1985	33.2°S 71.9°W	170	8.0

[a] Magnitude estimated from data in Lomnitz (1970, 1985), Kelleher (1972), Comte et al. (1986), Nishenko (1985), Cisternas et al. (2005), Okal (2005).
[b] Considered a great earthquake by Comte et al. (1986), the event fits well with the seismic cycle, though contemporaneous reports provide very little information on damage.

Three main subduction zone segments are defined by great earthquake ruptures (Fig. 1) spanning the SVZ, which may relate to physical features on the subducting plate and variable coupling at the margin (e.g., Herron, 1981; von Huene et al., 1997; Hackney et al., 2006). In each of these segments repeat times show a high degree of consistency, in spite of a degree of variability in rupture length and magnitude (cf. Comte et al., 1986).

2.2. Eruption records

We compiled a record of all historic SVZ volcanic eruptions, using Siebert and Simkin (2008-), rejecting all events where there is uncertainty over authenticity of the record or the year of eruption. Of 325 eruptions listed since 1558, 63 are rejected for these reasons, leaving 262 eruptions from 25 volcanoes. While Linde and Sacks (1998) rejected small or non-explosive events (Volcanic Explosivity Index (VEI) <2) we do not take this approach, primarily because the VEI is uncertain for most events beyond the recent past, and to reject all eruptions of uncertain magnitude would make our catalogue too small for any meaningful study. Furthermore, we are interested not only in explosive events, but in any activity that may contribute to arc-wide fluctuations in rates of volcanism. Thus some volcanoes, such as Villarrica, which have gone through periods of persistent but low-level activity may appear to dominate the eruption record. However, since persistently-active volcanoes are more likely to have magma-filled plumbing systems than dormant volcanoes, they should be ideal candidates for showing responses to external forcing as suggested, for example, by the seasonal response to earth-surface deformation observed in small and near-continuous eruptions of Sakura-jima, Japan (Mason et al., 2004). Thus, while explosive eruptions at long-quiescent volcanoes may provide the most compelling individual cases of potentially earthquake-triggered eruptions (Marzocchi, 2002; Manga and Brodsky, 2006), we would not wish to reject potential evidence from smaller eruptions of arc-scale magmatic disturbance following great earthquakes.

Our measured parameter is eruption occurrence, and not magnitude. While this is an imperfect measurement of the rate of volcanism, it is the only viable parameter given the variable detail on eruption scale in historic records. We measure eruptions by their onset date, which form point events in time. Eruption length is not accounted for, and significant changes in behaviour during eruption are therefore not included in our record; in any case, such information is sparsely available. Of the 262 dated eruptions in our catalogue, the month of onset is known for 182 (69% of the total), and the day for 146 (56%). Each eruption is recorded by a decimal onset date, and shown by date and latitude in Fig. 1. For eruptions where the exact onset date is unknown we take the midpoint of the smallest time interval for which the event can be dated, following Bebbington and Lai (1996). To avoid clustering artefacts, if two or more events are given the same date by

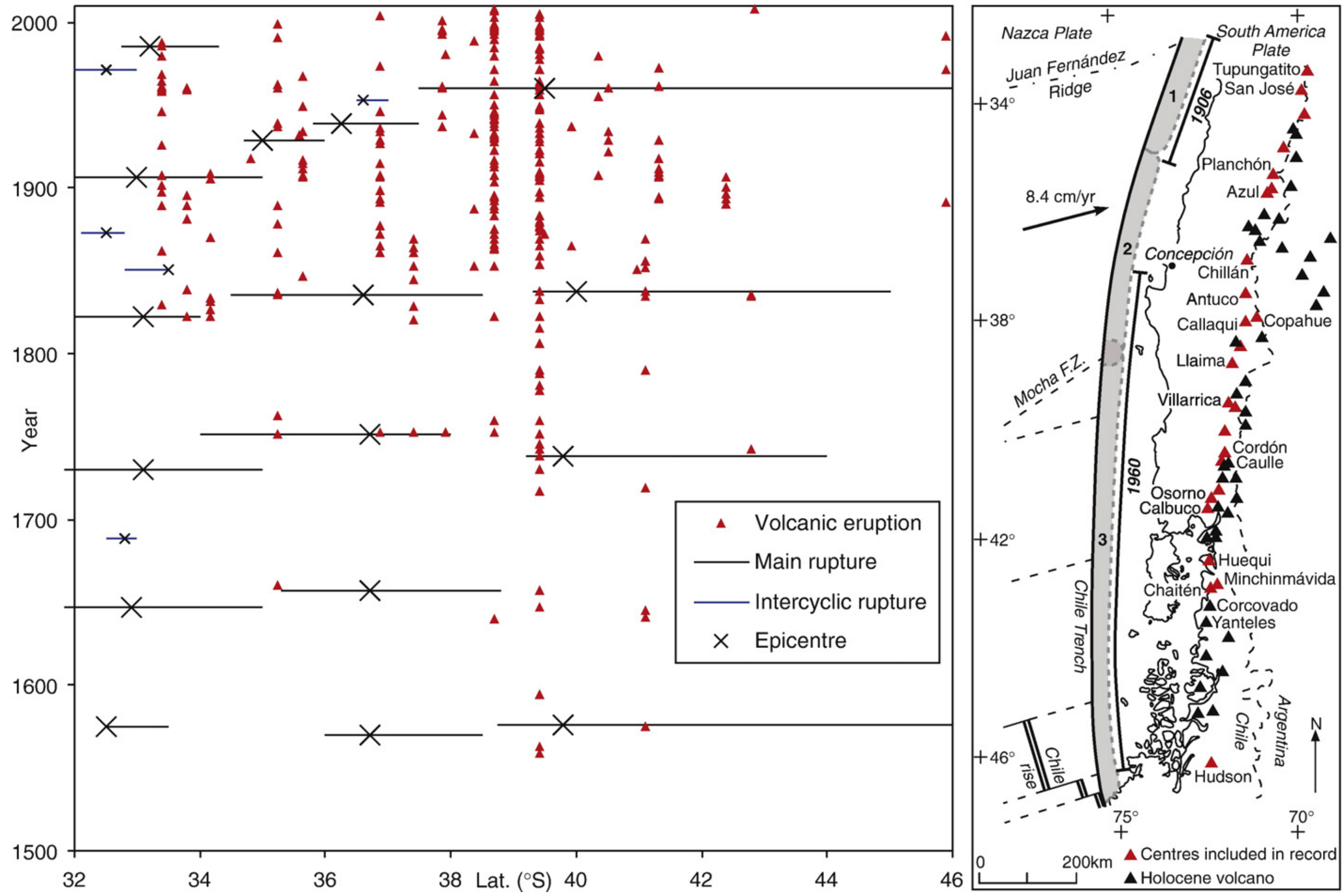

Fig. 1. Historically recorded large earthquakes (M_W>7.5) in central and southern Chile, showing main ruptures (M_W>8) and intercyclic events, with epicentres and approximate rupture lengths. Earthquake occurrence shows a cyclic temporal clustering and broad southward stepping pattern. Reliably recorded historic volcanic eruptions are depicted, showing that a few centres dominate the record, and that eruptions in the far south of the region appear to be underrepresented. The map depicts the regional tectonics, with volcano locations. Named volcanoes are discussed in the text. Schematic rupture zone segments are shown with approximate limits, corresponding to: Central Chile (1); Central Valley (2); and Southern Nazca (3) rupture zones (e.g., Lomnitz, 1970, 1985). The cause of the boundary between (1) and (2) is unclear, while the (2)–(3) boundary may be related to subduction of the Mocha fracture zone (e.g., Herron, 1981). Subduction of the Juan Fernández seamounts may be related to the onset of the volcanic gap north of the SVZ (e.g., von Huene et al., 1997).

this process (e.g., two events occurring in the same year at unknown dates) we spread the data evenly across the relevant time period. If this coincided with the year of a great earthquake the eruptions were removed from the analysis, since it was unclear if they preceded or post-dated the earthquake. It is clear that a few volcanoes dominate the eruption total, and that many centres are characterised by eruption clusters on decadal timescales.

Assuming a long-term constant rate of volcanism over the studied time-period, a cumulative plot offers a simple assessment of the period for which the record is approximately complete (Fig. 2). This suggests that from around 1850 onwards only a small proportion of eruptions have gone unrecorded, although the dating of recorded events improves only slightly, with the onset day known for 59% and the month for 73% of the 206 eruptions since 1852. Even in the period since 1970 the onset day is still recorded for <90% of eruptions. The long-term rate defined by Fig. 2 is 1.32 eruptions per year. Prior to ~1850 many eruptions presumably went unrecorded.

2.3. Record bias and interpretation

Historical records of earthquakes and eruptions must be examined with caution. By their nature both are notable phenomena, but especially so when endangering human life or property, and thus their recording is often sporadic. In particular, following great earthquakes, there may be a heightened awareness of other geological phenomena, including volcanic eruption (e.g., Simkin, 1993). Confusion and anecdotal evidence can lead to bias in the record, if not the misreporting of events. When looking for relationships between these processes, it is particularly important that doubtful reports are not included. Furthermore, to focus on any one individual eruption following a great earthquake as evidence of a causal relationship is misleading. Such coincidences would be expected (in limited numbers) for any independent stochastic processes, which is why

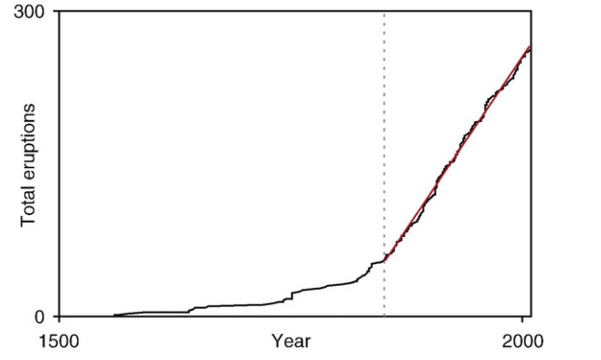

Fig. 2. Cumulative plot showing the number of recorded volcanic eruptions over time in the SVZ. Assuming an approximately constant long-term rate of volcanism, the record can be assumed to be approximately complete after ~1850.

we here consider eruption rates rather than listing individual events. These difficulties are well illustrated by the following examples.

Perhaps the most cited reports suggestive of earthquake-triggered volcanism are those of Darwin (1840); (Linde and Sacks, 1998; Manga and Brodsky, (2006); Walter and Amelung, 2007), following the 20th February 1835 Concepcion earthquake: “A few days after the earthquake, several volcanos within the Cordilleras, to the north of Concepcion, though previously quiescent, were in great activity… During the remainder of the year, the whole of the volcanic chain, from Osorno to Yantales…exhibited, at times, unusual activity.” Much of Darwin's record is based on secondary sources, and some eruptions, such as that of Antuco (Darwin, 1835) he later declares doubtful (Darwin, 1840). To the north of Concepcion, erupting volcanoes are described by Caldcleugh (1836), though his sources are unclear and reports are not corroborated elsewhere. Reports suggest activity in the Descabezado–Azul system and at Tupungatito. These are unlisted or declared uncertain by Siebert and Simkin (2008-) and so are excluded from our records, while only the year is known for the activity at Planchón. Of the listed volcanoes to the south, eruption reports at Corcovado and Yanteles (based on the disappearance of snow or the appearance of dark patches) and concomitant eruptions at Corcovado and Osorno later in the year, are doubtful or unlisted by Siebert and Simkin (2008-). Osorno and Minchinmávida were erupting before the earthquake, and are only stated as exhibiting renewed vigour. Minchinmávida is the only volcano with a reliable listed eruption record (Siebert and Simkin, 2008-) corresponding to the earthquake date. However, this is not a clear candidate for an earthquake-triggered event, since the volcano had started erupting three months before the earthquake, in November 1834, after nearly a century of quiescence: the 1835 record may simply be a result of heightened awareness of geological activity following the earthquake. In spite of this, the example of Yanteles and Minchinmávida are listed by Linde and Sacks (1998) as an earthquake-triggered eruption pair. This is not to say that Darwin's accounts are incorrect, but based on the information given in reports from the time, most of the listed eruptions cannot be included in our catalogue. Indeed, to quote FitzRoy (1839), following the same earthquake: “As to the state of neighbouring volcanoes, so various were the accounts of their action, both after and before the earthquake, that I had no means of ascertaining the full truth”. It is of interest to note, however, that the four erupting volcanoes south of Concepcion listed by Darwin (1840) are at latitudes well beyond the limit of the rupture zone, Minchinmávida being ~500 km distant.

The above discussion illustrates the difficulties in interpreting the historic record, but also demonstrates that individual eruptions can only ever be identified as *potential* examples of triggered events. For example, Tupungatito, listed by Walter and Amelung (2007) as one of four volcanoes that erupted within a year of the May 22nd 1960

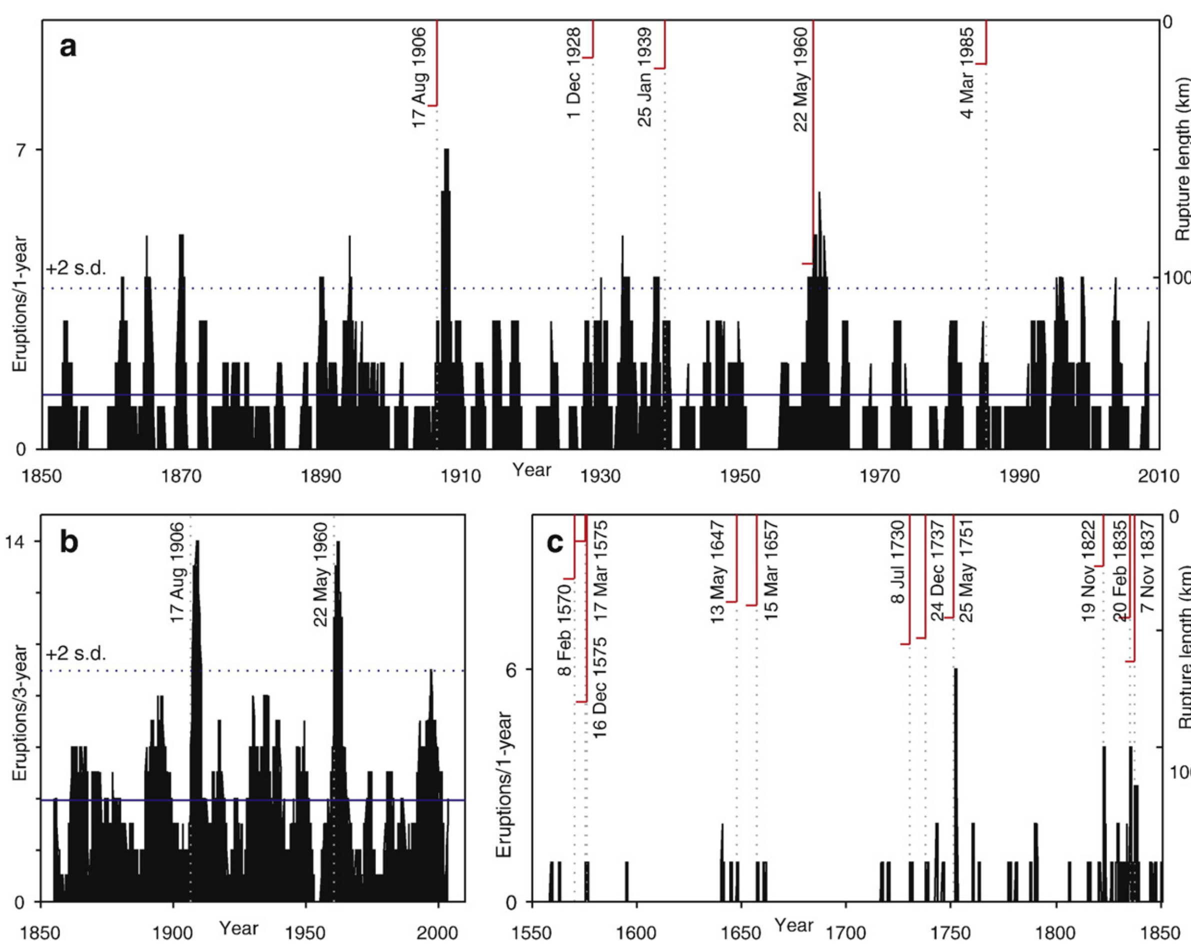

Fig. 3. Eruption rate plots, showing great earthquake timing and rupture length. Sharp eruption rate increases occur immediately following the 1906 and 1960 earthquakes, with rates above the background range lasting for ~1 yr. Using these calculated rates the mean and +2 standard deviation lines are shown to assess rate variability, and show the degree of deviation from the background of the peaks in 1906 and 1960. a: Rate since 1850 using a 12-month rectangular window passed over the eruption record at 0.05 year increments, such that only eruptions prior to the time point contribute to the eruption rate. b: Using the method described for a, with a 3-year window at 0.1 year increments, giving a smoother long-term eruption rate. c: Using the method described for a, with a 12-month window at 0.1 year increments, for the record prior to 1850. While data are likely to be highly incomplete, eruptions are almost always recorded in the year following great earthquakes, most notably in 1751.

Table 2
Potentially triggered eruptions, following the 1906 and 1960 earthquakes

Earthquake	Date	Latitude		
		Epicentre	Rupture N end	Rupture S end
	17 Aug 1906	33.0	32.0	35.0
	22 May 1960	39.5	37.5	46.0
Volcano[a]			Arc parallel distance (km)	
	Date	Latitude	From epicentre	From rupture zone[b]
Tupungatito	15 Feb 1907	33.4	0	–
Carrán-los Venados	9 Apr 1907	40.4	790	560
Calbuco[c]	22 Apr 1907	41.3	900	680
Villarrica	5 May 1907	39.4	690	460
Azul[c]	28 Jul 1907	35.7	260	30
Nevados de Chillán	1907	36.9	400	170
Llaima	1907	38.7	600	380
Huequi[d]	1906	42.4	1020	790
Cordón Caulle	24 May 1960	40.5	70	–
Planchón-Peteroa	10 Jul 1960	35.2	530	300
Tupungatito[e]	15 Jul 1960	33.4	730	510
Calbuco	1 Feb 1961	41.3	160	–
Copahue[f]	1961	37.9	230	–
Villarrica	1961	39.4	50	–
San José[d]	1960	33.8	690	460

[a] Volcanoes assumed to lie 200 km E of tabulated rupture line, which is taken to be trench-parallel.
[b] For volcanoes that lie outside the tabulated rupture zone limits.
[c] Also erupted in 1906, at unknown date (eruption excluded from analysis).
[d] Erupted at unknown date, so unclear if post-dated earthquake, and excluded from analysis (Fig. 3a).
[e] Also erupted on 5 May 1961.
[f] Also erupted in 1960, at unknown date (eruption excluded from analysis).

Chilean earthquake, certainly represents a candidate triggered eruption. However, in spite of the volcano's relatively infrequent activity, any stronger assertion is difficult to argue, given that it entered a new eruptive phase in 1958 after 12 yr of quiescence, and erupted twice in 1959 before erupting again two months after the 1960 earthquake.

By removing events from the record for which there is uncertainty over occurrence or timing, and by only examining in detail the record for which the long-term eruption rate is constant (Fig. 2), we aim to avoid erroneous reports and record bias. Furthermore, we focus on changes in arc-scale eruption rate, rather than attempting to explicitly identify individual triggered eruptions, which would not be justified by the data available.

3. Event analysis

From Fig. 2 it is assumed that the eruption record is complete since ~1850. The eruption rate since this time has been compared with the timing of great earthquakes (Fig. 1). We require a measure of short-term eruption rate on the arc scale based on eruption incidence, and not accounting for magnitude, to investigate whether great earthquakes lead to an eruption rate increase on a timescale greater than a few days. In order to verify, or otherwise, whether an increase in eruption rate occurs *after* great earthquakes, our eruption rate measurement in any time increment must be based upon the timing of recent prior eruptions, rather than future events. In this way, any abrupt increase in eruption rate will be shown at its real time position, rather than artificially early. To achieve this we pass a rectangular window across the full eruption records, with a minimum span of 12 months. Due to the poor quality of some data, in terms of eruption date, we did not choose a more complex or narrower window. This method has the advantage of producing data that are readily interpreted as the number of eruptions in the window-width period prior to that time point. This is not necessarily equal to the number of active volcanoes in that time period, since the figure may include two or more eruptions from a single volcano. Wider windows smooth the record and show longer-frequency variations, but artificially spread these variations along the time axis, though the onsets of rate-increase appear at their real time. The filter was run with various widths, and results are presented in Fig. 3 for three cases: 1850–present with a 12-month window (Fig. 3a); 1850–present with a 3-year window (Fig. 3b), and analysis of the incomplete pre-1850 eruption records with a 12-month window (Fig. 3c).

4. Discussion

4.1. Eruption rates

Fig. 3a shows that the eruption rate in the SVZ has fluctuated widely. However, two outlying maxima occur where the eruption rate is significantly increased above the norm. These two periods commence in late-1906 and mid-1960, and the rate of activity exceeds the background for ~12 months. The filter in Fig. 3b gives a smoothed longer-term eruption rate pattern which reveals the two peaks in eruption rate very clearly. These two periods correspond to times immediately following the two largest Chilean earthquakes since 1850: in August 1906 in the northern part of the study area, and in May 1960 in the southern half of the study area.

Following the 1906 earthquake and before the end of 1907 at least seven volcanoes erupted in the SVZ, in a possible maximum of ten eruptions (Table 2). Similarly, at least six, and possibly seven, volcanoes erupted by the end of 1961 following the 1960 earthquake, in up to nine eruptions. In the rest of the record, the eruption of five volcanoes in any 12-month period occurs three times, in 1863–4, 1869 and 1893, while four erupting volcanoes in a year occurs several times.

The distribution of eruption occurrence on an arc-scale may be modelled as a Poisson process, where it is expected that the occurrence of eruption at any one centre is independent of the timing of eruption at another volcano in the arc. This is confirmed when the time intervals between eruptions are examined (Fig. 4), using the sample mean of ~1.32 eruptions per year, which shows a good fit to the expected exponential distribution. Thus, if the arc-scale eruption rate is Poissonian, the probability of six eruptions occurring in a one-year period is 2.0×10^{-3}, indicating that such a rate would be expected once every 500 yr. In fact, this rate occurs twice in our study period, such that the probability of this rate occurring in any year is ~0.013, and hence these ‘eruption clusters’ do not fit well with a Poissonian model. Over the same period five great earthquakes occur, such that

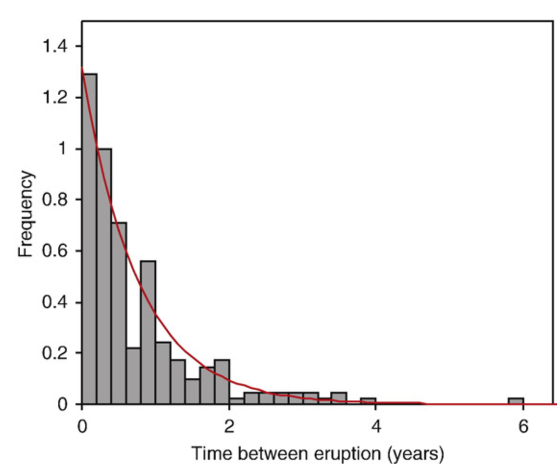

Fig. 4. Histogram showing the distribution of times between volcanic eruptions for the SVZ, with a 0.2 year bin width. The data fit closely the exponential curve shown, as expected if the arc eruption data are Poissonian. The peak in the histogram in the 0.8–1 year bin is an artefact from occasions in the record where only the year of eruption is recorded for consecutive events in consecutive years.

the probability of a great earthquake in any one year is ~0.032. Thus, from the product of these, the probability of a cluster and great earthquake coinciding is 4×10^{-4}. This occurs twice during our record, on both occasions where an eruption cluster is observed, the dates of which correspond to the two largest earthquakes in the record. Since the highest eruption rates occur more frequently than predicted by a Poissonian model, the eruption record may in fact be better modelled by a negative binomial distribution, suited to contagious events. Hence, we suggest that the highest eruption rates represent a departure from the typical Poissonian background eruption rate, presumed to be governed by internal and continuous-external factors at any individual system, and that eruption occurrence is not an independent variable in such cases, but is governed by factors relating to great earthquakes, which have the ability to affect a significant portion of the volcanoes in the arc. Following these great earthquakes, even if the background rate of volcanism were above average at the time, eruptions at 3–4 volcanoes may be plausibly assigned the status of seismically triggered events. Interestingly, the probability of five eruptions in a one-year period is 8.9×10^{-3}, but between 1863 and 1893 five volcanoes erupt in a yearly period on three occasions (in addition five eruptions occur at four volcanoes in 1932–33 and 1959) suggesting that there may be other underlying processes leading to enhanced eruption rates at certain times. It must be emphasised that there is no statistical basis by which any one of the eruptions following the 1906 or 1960 earthquakes may be asserted to be earthquake-triggered, but that we have identified an extremely small probability that the observed eruption rates occurred by chance, and hence suggest that a number of eruptions following the earthquakes are likely to have been triggered by the event.

Fig. 3 begs the question of why the 1906 and 1960 earthquakes show such a clear response in eruption rate, while the other main ruptures during this period do not. There is evidence of an above-average eruption rate after the 1928 M_W 8.2 earthquake, but no such evidence following the 1939 and 1985 (M_W 8.0) earthquakes. Thus, there is a lack of correlation between eruption rate and large earthquake occurrence for the 1939 and 1985 earthquakes. This observation has been made in studies of rapid triggering (e.g., Manga and Brodsky, 2006), highlighting cases such as the 2004 Sumatra earthquake. While immediate responses were not seen following this earthquake, subsequent eruptions have been postulated as triggered events (Walter and Amelung, 2007), but the time gap may make such association appear speculative. From our arc-scale approach, it is clear that not all great earthquakes elicit a similar eruptive response. The greater size of the 1906 and 1960 earthquakes is not an entirely satisfactory explanation; a smaller event would still be expected to show some response, albeit less pronounced. However, the mechanisms and timescale of these responses are not yet well understood.

The 1960 earthquake was much larger in terms of rupture length (940 km) and seismic moment (M_W 9.5) than the 1906 event (330 km and M_W 8.3), and yet the post-seismic eruption rate increases are of similar magnitude. If the system disturbance is ultimately related to stress changes associated with the earthquake, some relationship with earthquake magnitude might be expected. However, an additional factor will be the state of a particular volcano, in terms of how near the magmatic system is to the critical threshold at which dyke propagation may occur, which can be simply described by competing load and strength functions (cf. Jupp et al., 2004). The greater the stress changes associated with an earthquake, the higher the probability of conditions at a volcano exceeding the threshold for eruption. In other words, triggered eruptions occur at volcanoes that were already likely to erupt in the near future, had the earthquake not occurred, but the occurrence of the earthquake results in systems already primed for eruption crossing an eruption threshold, and thus may produce several concomitant eruptions. At any one time, the number of volcanoes in the arc approaching this threshold is likely to vary randomly. Such variation is likely to be complex, and related both to internal factors at any individual volcanic system, and external factors, such as the time since the previous large earthquake. The 1906 earthquake occurred after a relatively long gap in great earthquakes. This lapse in time may have allowed more volcanic centres to approach a critical state, undisturbed by large seismic events. Our analysis, revealing a variable magnitude of response to great earthquakes, suggests that the number of volcanoes 'primed' for eruption may be highly variable. On an arc scale, therefore, the likelihood of an observable response following a large earthquake is a function both of the state of the arc's volcanoes, as well as the scale of the earthquake, and this may explain why some large earthquakes do not register an increased eruption rate.

It has been proposed by other workers that the converse relationship, an eruptive trigger to earthquakes, may exist on a range of timescales (Acharya, 1982; Lemarchand and Grasso, 2007). We do not observe a clear relationship between great earthquakes and the immediately preceding arc-scale eruption rate, although prior to the 1960 earthquake the eruption rate was higher than average, with several eruptions in the SVZ and Austral volcanic zone, further south. However, this pattern is not repeated before other large earthquakes, and the eruption rate in 1959 does not constitute a statistical outlier from the record, unlike the post-earthquake records. However, it does reflect the inherent variability in eruption frequency and the fluctuating background rate of volcanism in the SVZ. Other periods of increased eruption rate (Fig. 3b) occur in the 1890s, 1930s and 1990s. This may simply be natural variation, but does hint at a periodicity, potentially related to changing crustal stress conditions through the seismic cycle (e.g., Klotz et al., 2001).

Ideally, we would incorporate eruption magnitude into our analysis. However, the major limitation to this approach is record quality and resolution, with accuracy biased towards recent events, and estimates of the past event magnitudes based on sparse data with poorly quantified uncertainties. As a preliminary analysis to investigate whether a magnitude response to great earthquakes exists, we consider all events with some estimate, however uncertain, of magnitude. The smallest events are under-represented in the record, and we thus consider only the 170 events of VEI≥2. This record includes five eruptions of VEI≥4 (including that of Chaitén in May 2008), which show no clear correspondence to great earthquake dates. Nor do the patterns of VEI 2 and 3 events suggest any variability in eruption scale that can be related to great earthquake occurrence. This is not to say such a relationship does not exist, but a larger dataset with a better quantified and more precise measure of eruption scale would be necessary to test this hypothesis.

4.2. Eruption locations and seismic stresses

The locations of potentially triggered eruptions can provide constraints on the stresses that disturb magmatic systems following great earthquakes. Static stress changes decay more rapidly with distance from the fault zone than dynamic stresses: the magnitude of static stress change after an M8 earthquake is of the order of 10^{-1} MPa at 100 km, decreasing to 10^{-4} MPa at 1000 km, comparing with 3 to 0.06 MPa for dynamic stress changes in the same locations (cf. Manga and Brodsky, 2006). The two are not directly comparable, due to the transience of dynamic stresses, but the significantly greater magnitude of dynamic stresses at large distances makes them a more plausible far-field trigger. If the mechanisms involved in eruption triggering are solely due to static stress change, candidate triggered eruptions would be located predominantly in the arc section parallel to the rupture zone.

From Fig. 1 the subduction zone can be split into three segments, defined by portions of the plate boundary which fail in each seismic cycle, and spanning the latitudes 31.5–34.8°S, 34.8–38.4°S and 37.8–46°S. We analysed data for each of these segments individually, but found no strong pattern of eruption rate variation following earthquakes within the same segments. It is only on the arc-scale that the relationship between earthquakes and volcanism is seen, and this response is not limited to the arc-section adjacent to the rupture zone. This is clear from Table 2, where, in both 1906 and 1960, several of the candidate triggered eruptions occurred at locations far beyond the rupture zone. In 1906, seven of eight potentially triggered volcanoes lie outside the rupture zone, at distances of up to 790 km. In 1960, three of the potentially triggered volcanoes lie outside the rupture zone, up to a maximum distance of 450 km. This concurs with Linde and Sacks' (1998) conclusion, that seismic triggering mechanisms are capable of acting at distances of several hundred kilometres from a great earthquake rupture zone.

The strain modelling of Walter and Amelung (2007) shows that following the 1960 earthquake, the whole of the SVZ experienced volumetric expansion, with strain experienced at candidate triggered volcanoes ranging from >25 μ at Cordón Caulle, Calbuco and Villarica, to 10–15 μ at Copahue, and approaching 0 μ at Planchón, San José and Tupungatito. From the modelling, all of these systems would have experienced magma chamber dilatation, but if static changes were the sole initiator of eruptive processes, a non-zero strain threshold for the observation of such effects would be expected. Static stress changes may be a plausible mechanism for eruptions such as that of Cordón Caulle, but is less so at locations experiencing very low strain, which would also have experienced much larger dynamic stresses. This is reiterated by the 1906 triggered eruptions, half of which lie at >450 km beyond the rupture zone (Table 2). This smaller (though poorly constrained) earthquake is unlikely to have generated significant static strains at such distances (cf. Walter and Amelung, 2007). The prevalence of candidate triggered eruptions at distances where static stresses were minor indicates a likely important role for dynamic stresses in triggering eruptions, and suggests that both dynamic and static stresses can initiate magmatic changes that may not be manifested as eruptions for periods of several months.

The reliable eruption record only covers a period of five main ruptures. Further back in time data become sparse. While reporting bias following large earthquakes is more likely in this period, the pattern seen in Fig. 3c is still striking. Virtually all of the great earthquakes are followed by eruptions in the following year. For example, six well-dated eruptions occurred in the nine months following the 1751 Concepcion earthquake, at Planchón, Villarrica, Llaima, Callaqui, Chillán and Antuco. Four of these volcanoes lie outside the rupture zone, at distances of up to 160 km, although in this case the rupture zone is poorly defined. In contrast, there is no clear response to the 1835 earthquake in Fig. 3c (two eruptions occur in the three months prior to the earthquake, contributing to the peak). As discussed in Section 2.3., the contemporary reports of Caldcleugh (1836) and Darwin (1840) strongly suggest a significant volcanic response, but the record quality is insufficient for the reliable identification of specific eruptions, and hence no response is identifiable from our statistical analysis. What is also notable about this earthquake, and not cited in recent work, is a possible triggered eruption at Robinson Crusoe (Darwin, 1840; Siebert and Simkin, 2008), in the Juan Fernández Islands. This is the only well-reported historical eruption at this Pacific Ocean island group, 635 km from Concepcion, and occurred on the same day as the earthquake. If genuine, it suggests that perturbations to volcanic systems may occur quite remote from the rupture zone and in a wholly different tectonic setting, presumably due to dynamic stresses.

4.3. Triggering mechanisms

We have shown that both static and dynamic stresses, arising from great earthquakes, are likely triggers of volcanic eruption in the SVZ, over periods of several months. However, mechanisms are required by which these stresses precipitate magmatic processes that ultimately generate eruption. Several hypotheses for such processes have been proposed, and while our data are not suitable to test these, it is useful to discuss potential mechanisms to account for our observed relationships. Whether the effects of great earthquakes on volcanic systems persist for longer periods cannot be deduced from these records, since after ~12 months eruption rates fall to within the background range. The period of response we identify is on the order of 1 yr, though this figure is difficult to constrain since it is not possible to identify specific seismically triggered eruptions.

Magmatic overpressure is likely to be a key factor determining the apparent time lag before eruption (e.g. Tait et al., 1989; McLeod and Tait, 1999). Eruptions within days of a large earthquake will only occur at magma bodies that were already near a critical eruptive overpressure. Volcanoes displaying a slightly longer response may be slightly further below this tipping point, but stress changes must still be significant enough to produce permanent pressure changes that set the system on the path to eruption. In more general terms, the triggering event must be sufficient to initiate failure by exceeding a threshold, but the failure event (eruption) might then occur after an extended incubation period (cf. Jupp et al., 2004).

Various models have been proposed by which stresses effect physical changes and movement within the magmatic system. Static stress changes have been invoked for rapidly triggered events within the rupture zone, notably for the 1960 eruption of Cordón Caulle (Barrientos, 1994). This eruption, within 48 h of the May 22nd earthquake, was considered to be triggered as a result of strain arising from extension beneath the volcano (Barrientos, 1994), which was in a mature stage of its eruptive cycle. Walter and Amelung (2007) proposed that volumetric strain expansion of magma would trigger eruption by magmatic decompression and gas exsolution. The same strain changes may enhance the unclamping of fracture systems, and allow dyke formation. This model provides a plausible explanation for responses on the timescale we observed, while the viscous relaxation model of Marzocchi (2002) is likely to act on longer timescales. However, it remains unclear why some volcanoes show a near-immediate response to static stress changes, while others may take several months for triggering effects to result in eruption. Magma composition and rheology, storage depth and state of volatile saturation, as well as system overpressure, may all play a role (e.g., Woods and Pyle, 1997; McLeod and Tait, 1999; Jellinek and DePaolo, 2003). Particular mechanisms may also operate on different timescales, but the physical differences between each system, in terms of plumbing and crustal structure, may ultimately be the most important control in influencing the timing at which eruption occurs following the initial triggering process. Thus the time lag may not inform directly on potential mechanisms. Such time lags have been observed before, and have been difficult to account for. For example, coupling relationships between eruptions of Vesuvius and Apennine earthquakes has been suggested to operate over a timescale of several years (Nostro et al., 1998), due to static stress changes, and in both directions, with eruptions both preceding and following earthquakes.

Dynamic stresses, due to their transience, require a mechanism by which they are rapidly converted into permanent pressure changes (Manga and Brodsky, 2006). While such a response must be rapid, nevertheless it may not result in eruption for several months. Several mechanisms proposed for triggering through the passage of seismic waves involve bubbles, which, through nucleation and growth, are of primary importance in initiating eruption. One such process involves advective overpressure, by which bubbles are dislodged by seismic waves and, through rising, transmit larger pressures across a decreasing pressure gradient. However, in most real magmas, such processes produce insufficiently large pressure changes to initiate eruption (Pyle and Pyle, 1995). A second proposed mechanism involves rectified diffusion, whereby bubble oscillation during the passage of seismic waves results in a net addition of mass into the bubble, increasing bubble volume and magmatic pressure. Again, modelling of this process with realistic parameters suggests the resulting pressure increase is insignificant in most cases (Ichihara and Brodsky, 2006). While both these processes increase pressure, if the change is too minor to initiate pre-eruption processes then the excess pressure is likely to dissipate over a short timescale due to gas loss (e.g., Manga and Brodsky, 2006), and hence, in most cases, these are not considered to be important eruption triggering mechanisms. Finally, if the magma is close to volatile saturation, the pressure change associated with dynamic stresses may be sufficient to generate rapid bubble nucleation, leading to eruption (Manga and Brodsky, 2006).

Magmatic overturn, whereby a crystal mush at the roof of a magma chamber may be dislodged as inter-crystal yield strength decreases due to dynamic stresses, may also initiate eruption processes, by promoting vesiculation of rising magma and convection. Physical models suggest this process to be plausible (Manga and Brodsky, 2006; Davis et al., 2007), and one that may operate on the timescales of interest.

Sumita and Manga (2008) show that candidate rapidly-triggered eruptions, as well as hydrological and other seismic responses, fall within a general distance–magnitude bound, corresponding closely to the liquefaction limit for hydrological system responses. The correlation suggests a common mechanism influencing both magmatic and hydrological systems, and they suggest that liquefaction may play a role in eruption-triggering. For the 1960 earthquake, all of the candidate triggered eruptions lie inside the bounds shown by Sumita and Manga (2008), and liquefaction may provide a plausible triggering process for these events. However, for the 1906 earthquake the same magnitude–distance bounds suggest that triggered effects should not occur beyond 800 km, while three of our candidate triggered eruptions lie at or beyond this limit. Thus, results based on previously postulated triggered responses suggest that these eruptions are too distal to be earthquake-triggered.

While it is not possible to differentiate between the discussed mechanisms on the basis of our record alone, distinguishing between the postulated models may be attainable through petrological and geochemical analysis of triggered-eruption products. For example, chemical zonation in phenocrysts may reveal the short timescale (<1 month to <1 yr) pre-eruption changes in the magma system that would occur as a consequence of inter-mixing and re-equilibration of disturbed crystalline mush and melt (e.g., Morgan et al., 2004, Martin et al., 2008).

5. Conclusions

Volcanoes in the Andean SVZ respond to earthquakes of $M>8$ in the adjacent subduction zone through an increased rate of eruption. Using earthquake and eruption records since 1850 we have investigated temporal relationships between earthquakes and volcanic eruption. The mean eruption rate in the arc is 1.32 eruptions per year. The highest eruption rates in the record show a deviation from the long-term Poissonian background trend, and occur in the 12–18 month periods following the 1906 and 1960 earthquakes, with up to eight volcanoes erupting in 1906–7. The probability of this event coinciding with the date of a great earthquake is 4×10^{-4}, a situation which occurs in the two periods of highest eruption rates. We thus suggest that the arc-scale volcanic eruption rate is not independent during the most elevated periods, but is governed by the occurrence of great earthquakes in the adjacent subduction zone, which have the ability to trigger multiple eruptions over a period of several months. Based on the average eruption rate during the remainder of the record, we estimate that at least 3–4 eruptions were seismically triggered following each earthquake. Not all large earthquakes elicit similar responses in eruption rate, and the relationship between the two is not simply based on earthquake magnitude, but an interplay between magnitude and the number of arc volcanoes in a pre-eruptive state at any one time, itself highly variable and subject to multiple internal and external factors. The locations of candidate triggered eruptions suggest that triggering occurs both in the near-field, where static stress changes are likely to be important, and also at distances up to ≥500 km beyond the rupture zone. Here, dynamic stresses associated with the passage of seismic waves are likely to be the primary cause of magma body disturbance. While individual triggered eruptions cannot be identified, we have shown that seismic eruption triggering following large earthquakes, with delays of several months, is a significant process in volcanic arcs.

Acknowledgements

The authors acknowledge the Natural Environment Research Council UK and the Royal Society for financial support, and thank two anonymous reviewers for comments that improved the manuscript.

References

Acharya, H., 1982. Volcanic activity and large earthquakes. J. Volcanol. Geotherm. Res. 13, 373–378.
Barrientos, S.E., 1994. Large thrust earthquakes and volcanic eruptions. Pure Appl. Geophys. 142, 225–237.
Barrientos, S.E., Ward, S.N., 1990. The 1960 Chile earthquake: inversion for slip distribution from surface deformation. Geophys. J. Int. 103, 589–598.
Bebbington, M.S., Lai, C.D., 1996. On homogeneous models for volcanic eruptions. Math. Geol. 28, 585–600.
Brodsky, E.E., Roeloffs, E., Woodcock, D., Gall, I., Manga, M., 2003. A mechanism for sustained ground water pressure changes induced by distant earthquakes. J. Geophys. Res. 108. doi:10.1029/2002JB002321.
Caldcleugh, A., 1836. An Account of the great earthquake experienced in Chile on the 20th of February, 1835. Philos. Trans. R. Soc. Lond. 126, 21–26.
Cisternas, M., Atwater, B.F., Torrejon, T., Sawai, Y., Machuca, G., et al., 2005. Predecessors of the giant 1960 Chile earthquake. Nature 437, 404–407.
Comte, D., Eisenberg, A., Lorca, E., Pardo, M., Ponce, L., et al., 1986. The central Chile earthquake of 3 March 1985: a repeat of previous great earthquakes in the region? Science 233, 449–453.
Darwin, C.R., 1835. Darwin's Beagle Diary (1831–1836), p. 546. http://darwin-online.org.uk/.
Darwin, C.R., 1840. On the connexion of certain volcanic phenomena in South America; and on the formation of mountain chains and volcanos, as the effect of the same power by which continents are elevated. Trans. Geol. Soc. Lond. 5, 601–631.
Davis, M., Koenders, M.A., Petford, N., 2007. Vibro-agitation of chambered magma. J. Volcanol. Geotherm. Res. 167, 24–36.
DeMets, C., Gordon, R.G., Argus, D.F., Stein, S., 1990. Current plate motions. Geophys. J. Int. 101, 425–478.
FitzRoy, R., 1839. Narrative of the surveying voyages of His Majesty's Ships Adventure and Beagle between the Years 1826 and 1836, vol. 2. Henry Colburn, London, p. 418.
Hackney, R.I., Echtler, H.P., Franz, G., Götze, H.-J., Lucassen, F., et al., 2006. The segmented overriding plate and coupling at the South-Central Chilean margin (36–42°S). In: Oncken, O., Chong, G., Franz, G., Giese, P., Götze, H.-J., et al. (Eds.), The Andes — Active Subduction Orogeny. Springer-Verlag, Heidelberg, pp. 355–374.
Harris, A.J.L., Ripepe, M., 2007. Regional earthquake as a trigger for enhanced volcanic activity: evidence from MODIS thermal data. Geophys. Res. Lett. 34, L02304. doi:10.1029/2006GL028251.
Herron, E.M., 1981. Chile margin near lat 38°S: evidence for a genetic relationship between continental and marine geologic features or a case of curious coincidences? Nazca Plate: crustal formation and Andean convergence. GSA Memoir 154, 755–760.
Hill, D.P., Reasenberg, P.A., Michael, A., Arabaz, W., Beroza, G.C., et al., 1993. Seismicity in the western United States remotely triggered by the M 7.4 Landers, California, earthquake of June 28, 1992. Science 260, 1617–1623.
Hill, D.P., Pollitz, F., Newhall, C., 2002. Earthquake–volcano interactions. Phys. Today 55, 41–47.
Ichihara, M., Brodsky, E.E., 2006. A limit on the effect of rectified diffusion in volcanic systems. Geophys. Res. Lett. 33. doi:10.1029/2005GL024753.
Jellinek, A.M., DePaolo, D.J., 2003. A model for the origin of large silicic magma chambers: precursors of caldera-forming eruptions. Bull. Volcanol. 65, 363–381.
Jupp, T.E., Pyle, D.M., Mason, B.G., Dade, W.B., 2004. A statistical model for the timing of earthquakes and volcanic eruptions influenced by periodic processes. J. Geophys. Res. 109, B02206. doi:10.1029/2003JB002584.
Kelleher, J.A., 1972. Rupture zones of large South American earthquakes and some predictions. J. Geophys. Res. 77, 2087–2103.
Klotz, J., Khazaradze, G., Angermann, D., Reigber, C., Perdomo, R., Cifuentes, O., 2001. Earthquake cycle dominates contemporary crustal deformation in Central and Southern Andes. Earth Planet. Sci. Lett. 193, 437–446.
Lara, L.E., Naranjo, J.A., Moreno, H., 2004. Rhyodacitic fissure eruption in Southern Andes (Cordón Caulle, 40.5°S) after the 1960 (Mw:9.5) Chilean earthquake: a structural interpretation. J. Volcanol. Geotherm. Res. 138, 127–138.
Lemarchand, N., Grasso, J.-R., 2007. Interactions between earthquakes and volcano activity. Geophys. Res. Lett. 34, L24303. doi:10.1029/2007GL031438.

Li, V.C., Kisslinger, C., 1984. Stress transfer and nonlinear stress accumulation at subduction-type plate boundaries: application to the Aleutians. Pure Appl. Geophys. 122, 812–830.

Linde, A.T., Sacks, I.S., 1998. Triggering of volcanic eruptions. Nature 395, 888–890.

Lomnitz, C., 1970. Major earthquakes and tsunamis in Chile during the period 1535 to 1955. Geol. Rundsch. Z. Allg. Geol. 59, 938–960.

Lomnitz, C., 1985. Tectonic feedback and the earthquake cycle. Pure Appl. Geophys. 123, 667–682.

Manga, M., Brodsky, E.E., 2006. Seismic triggering of eruptions in the far field: volcanoes and geysers. Annu. Rev. Earth Planet. Sci. 34, 263–291.

Martin, V.M., Morgan, D.J., Jerram, D.A., Caddick, M.J., Prior, D.J., Davidson, J.P., 2008. Bang! Month-scale eruption triggering at Santorini volcano. Science 321, 1178-1178.

Marzocchi, W., 2002. Remote seismic influence on large explosive eruptions. J. Geophys. Res. 107 (B1), 2018. doi:10.1029/2001JB000307.

Mason, B.G., Pyle, D.M., Dade, W.B., Jupp, T., 2004. Seasonality of volcanic eruptions. J. Geophys. Res. 109, B04206. doi:10.1029/2002JB002293.

McLeod, P., Tait, S., 1999. The growth of dykes from magma chambers. J. Volcanol. Geotherm. Res. 92, 231–246.

Moernaut, J., De Batist, M., Charlet, F., Heirman, K., Chapron, E., et al., 2007. Giant earthquakes in South-Central Chile revealed by Holocene mass-wasting events in Lake Puyehue. Sediment. Geol. 195, 239–256.

Montgomery, D.R., Manga, M., 2003. Streamflow and water well responses to earthquakes. Science 300, 2047–2049.

Morgan, D.J., Blake, S., Rogers, N.W., DeVivo, B., Rolandi, G., Macdonald, R., Hawkesworth, C.J., 2004. Time scales of crystal residence and magma chamber volume from modeling of diffusion profiles in phenocrysts: Vesuvius 1944. Earth Planet. Sci. Lett. 222, 933–946.

Nishenko, S.P., 1985. Seismic potential for large and great interplate earthquakes along the Chilean and southern Peruvian margins of South America: a quantitative reappraisal. J. Geophys. Res. 90, 3589–3616.

Nostro, C., Stein, R.S., Cocco, M., Belardinelli, M.E., Marzocchi, W., 1998. Two-way coupling between Vesuvius eruptions and southern Apennine earthquakes, Italy, by elastic stress transfer. J. Geophys. Res. 103, 24487–24504.

Okal, E.A., 2005. A re-evaluation of the great Aleutian and Chilean earthquakes of 1906 August 17. Geophys. J. Int. 161, 268–282.

Pyle, D.M., Pyle, D.L., 1995. Bubble migration and the initiation of volcanic eruptions. J. Volcanol. Geotherm. Res. 67, 227–232.

Siebert, L., Simkin, T., 2008-. Volcanoes of the world: an illustrated catalog of Holocene volcanoes and their eruptions. Smithsonian Institution, Global Volcanism Program Digital Information Series, GVP-3, http://www.volcano.si.edu/world/.

Simkin, T., 1993. Terrestrial volcanism in space and time. Annu. Rev. Earth Planet. Sci. 21, 427–452.

Sumita, I., Manga, M., 2008. Suspension rheology under oscillatory shear and its geophysical implications. Earth Planet. Sci. Lett. 269, 467–476.

Tait, S., Jaupart, C., Vergniolle, S., 1989. Pressure, gas content and eruption periodicity of a shallow, crystallizing magma chamber. Earth Planet. Sci. Lett. 92, 107–123.

von Huene, R., Corvalán, J., Flueh, E.R., Hinz, K., Korstgard, J., et al., 1997. Tectonic control of the subducting Juan Fernández Ridge on the Andean margin near Valparaiso, Chile. Tectonics 16, 474–488.

Walter, T.R., Amelung, F., 2007. Volcanic eruptions following M≥9 megathrust earthquakes: Implications for the Sumatra-Andaman volcanoes. Geology 35, 539–542.

Woods, A.W., Pyle, D.M., 1997. The control of chamber geometry on triggering volcanic eruptions. Earth Planet. Sci. Lett. 151, 155–166.