Volcano-Tectonic Earthquakes: a new tool for estimating intrusive volumes and forecasting eruptions

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Abstract:

We present data on 136 high-frequency earthquakes and swarms, termed volcano-tectonic (VT) seismicity, which preceded 111 eruptions at 83 volcanoes, plus data on VT swarms that preceded intrusions at 21 other volcanoes. We find that VT seismicity is usually the earliest reported seismic precursor for eruptions at volcanoes that have been dormant for decades or more, and precedes eruptions of all magma types from basaltic to rhyolitic and all explosivities from VEI 0 to ultraplinian VEI 6 at such previously long-dormant volcanoes. Because large eruptions occur most commonly during resumption of activity at long-dormant volcanoes, VT seismicity is an important precursor for the Earth's most dangerous eruptions. VT seismicity precedes all explosive eruptions of $\text{VEI} \geq 5$ and most (all?) VEI 4 eruptions in our data set. Surprisingly we find that the VT seismicity originates at distal locations on tectonic fault structures at distances of a few to tens of kilometers laterally from the site of the eventual eruption, and rarely if ever starts beneath the eruption site itself. The distal VT swarms generally occur at depths almost equal to the horizontal distance of the swarm from the summit out to about 15 km distance, beyond which hypocenter depths level out. We summarize several important characteristics of this distal VT seismicity including: swarm-like nature, onset days to years prior to the beginning of magmatic eruptions, peaking of activity at the time of the initial eruption whether phreatic or magmatic, and large non-double couple component to focal mechanisms. Most importantly we show that the intruded magma volume can be simply estimated from the cumulative seismic moment of the VT seismicity from:

$\text{Log}_{10} V = 0.71 \text{ Log}_{10} \Sigma \text{Moment} - 5.32$, with volume, V , in cubic meters and seismic moment in Newton meters. Because the cumulative seismic moment can be approximated from the size of just the few largest events, and is quite insensitive to precise locations, the intruded magma volume can be quickly and easily estimated with few short-period seismic stations.

Notable cases in which distal VT events preceded eruptions at long-dormant volcanoes include: Nevado del Ruiz (1984-1985), Pinatubo (1991), Unzen (1989-1995), Soufriere Hills (1995), Shishaldin (1989-1999), Tacana' (1985-1986), Pacaya (1980), Rabaul (1994), Cotopaxi (2001). Additional cases are recognized at frequently active volcanoes including Popocateptl (2001-2003) and Mauna Loa (1984). We present four case studies (Pinatubo, Soufriere Hills, Unzen, and Tacana') in which we demonstrate the above mentioned VT characteristics prior to eruptions. Using regional data recorded by NEIC, we recognized in near-real time that a huge distal VT swarm was occurring, deduced that a proportionately huge magmatic intrusion was taking place beneath the long dormant Sulu Range, New Britain Island, Papua New Guinea, that it was likely to lead to eruptive activity, and warned Rabaul Volcano Observatory days before a phreatic eruption occurred. This confirms the value of this technique for eruption forecasting. We also present a counter-example where we deduced that a VT swarm at Volcan Cosiguina, Nicaragua, indicated a small intrusion, insufficient to reach the surface and erupt. Finally, we discuss limitations of the method and propose a mechanism by which this distal VT seismicity is triggered by magmatic intrusion.

Keywords: Volcano Seismology, Eruption Forecasting, Seismic Moment, Volcano Hazards, Precursory Seismicity, Volcano-Tectonic Earthquakes, Non-double couple

1. Introduction:

The Volcano Disaster Assistance Program (VDAP) was formed by the US Geological Survey and the Office of Foreign Disaster Assistance following the disastrous eruption of Nevado del Ruiz, Colombia, in 1985. Since its inception, the VDAP team has helped build monitoring infrastructure in fourteen countries, responded to more than 25 volcanic crises worldwide, and had the opportunity to review tens of thousands of seismic records from those plus scores of additional eruptions and volcanic crises. One of the first phenomena the VDAP team noticed was that explosive eruptions are preceded by high-frequency volcano tectonic earthquakes that occur in swarm-like distributions and lack the typical main shock – aftershock distribution of tectonic earthquake sequences. Further, we recognized that this VT seismicity originates at foci well away from the eventual eruption vent at lateral distances ranging from one or two to more than 30 km. We call this “distal volcano tectonic” (distal VT or dVT) seismicity. To document and interpret this unusual seismicity, we have cataloged 75 cases where distal VT earthquake swarms have occurred prior to eruptions (Table 1), 37 more cases for which we infer that precursory seismicity was distal VT (Table 2), and 19 additional cases of distal VT seismicity during apparent magmatic intrusions (Table 3). Over the last 25 years, we have personally encountered more than 40 of these cases and have shown that distal VTs are related to magma intrusion and are invaluable for eruption forecasting (White and Power, 2001; White, 2006; White and Rowe, 2006).

In this paper we demonstrate the following characteristics of dVT earthquakes:

1. dVT's are often the earliest seismic precursor to eruptions at volcanoes that have been in repose for two decades or longer,
2. dVT's begin days to years prior to onset of eruptive activity,

3. dVT's occur at offset distances ranging from less than 2 km to 45 km laterally from the eventual eruption site,
4. the dVT swarms occur at depths generally very similar to the offset distance of the swarm from the eventual eruption site to distances of ~15 km, beyond which the depths level out at 15-20 km,
5. dVT earthquakes occur as swarms of events which generally increase in number and average energy over time where the largest events typically occur near or after the middle of the swarm and often include several events within $\frac{1}{2}$ magnitude unit of the largest event,
6. cumulative seismic moment of the dVT's is found empirically to be proportional to the intruded volume, in cases where that volume was determined independently.

We present four case studies to illustrate the above characteristics. We then present data from 75 eruptions at 61 volcanoes, each known to be have been preceded by dVT seismicity, the eruptions ranging in size and type from VEI 0 phreatic and/or passive basaltic extrusions to ultraplinian VEI 5 to 6 dacitic and rhyolitic eruptions. We discuss aspects of these data that are most useful for eruption forecasting and present three examples of how the VDAF team has used them in forecasting. Finally we suggest a mechanism driving the distal VT phenomenon which is compatible with the constraints presented in this paper.

2. Methods:

The data presented in this paper is in large part the result of examination of tens of thousands of paper seismograms and digital records of precursory seismicity conducted over the past 26 years during visits to volcano observatories around the world, including those in Mexico, Guatemala, El Salvador, Nicaragua, Costa Rica, Panama, Colombia, Ecuador, Peru, Chile, Montserrat, the Congo, Saudi Arabia, Indonesia, Papua New Guinea, Philippines, Japan, and Kamchatka, Russia (see Acknowledgements section for more details)

as well as the U.S. Geological Survey's Alaska, Cascades, Hawaii, and California Volcano Observatories, and the Emergency Management Office of the Commonwealth of the Northern Mariana Islands. Additional data also come from hypocenter searches of the National Earthquake Information Center (NEIC) or International Seismic Center (ISC) worldwide seismic data bases, from the Smithsonian Institute's Global Volcanism Program reports and from published literature. The data for the Unzen case study come from Dr. Kodo Umakoshi.

Seismicity was classified into typical event types including low frequency (LF), hybrid, high frequency volcano tectonic (VT), tremor, explosion, and other surface flow events. Our definitions for these event types are consistent with definitions used by others (e.g. Lahr et al., 1994; Power et al., 1996; Chouet, 2003; and McNutt, 2005). In the cases where there were just paper records, events were classified by signal characteristics using the same station at distances less than 5 km where possible. Where digital data were available, events were classified using signal characteristics and spectral characteristics (e.g. bandwidth and dominant frequencies).

Network quality is variable for several reasons. In some cases the volcano was not recognized as potentially active and therefore the local network was installed after the onset of precursory activity (e.g. Chaiten, Pinatubo, and Kasatochi). In other cases remote location or lack of sufficient infrastructure prohibited the installation of more than one or two seismic stations. In such cases, events were located based on felt reports, S-P times and relative move-out across seismic stations. In cases with networks of more than 3 stations, events were located by local observatory personnel using P- and S-wave phases on at least 4 stations using standard location algorithms. In at least two cases, as we show for Soufriere Hills and Mt. Pinatubo, high precision relocation study results demonstrate that the preliminary locations are sufficient to determine whether events are distal or proximal.

Examples in the tables were compiled from seismic data sets directly examined by the authors, from published papers or reports, and/or from the Smithsonian Volcanoes of the World database (Siebert et al., 2010). Earthquake magnitudes are from observatory

catalogs or the NEIC and ISC catalogues. For eruptions that pre-date reliable seismic reporting (before 1960), examples are from volcanoes with historical records of felt reports or where there were working seismometers (e.g. Sakurajima, 1914; Usu, 1910 and 1944; and Bezymianny, 1956). Volume changes are calculated from published data sources including modeled InSAR deformation (Sulu Range, Shishaldin, Akutan, and Peulik), deep well-injection records (e.g. Paradox Well, KTB Well, and Denver Well) which injections we propose trigger seismicity analogous to magma intrusion, modeled EDM deformation (Long Valley Caldera, Matsushiro, Yellowstone), and modeled magnetometer data (Poas).

3. Case Studies:

3.1 Case study 1: Mt. Pinatubo, Philippines, 1991 VEI 6:

The 1991 VEI 6 eruption of Mt. Pinatubo was the largest eruption worldwide during the last 100 years. Mt. Pinatubo was a relatively unknown, heavily forested 1745-m-high lava dome complex with no record of historical eruptions. Unrest near the volcano was first noticed when earthquakes began to be felt in Yamut and Tarao towns on 15 March, 1991 (Sabit et al., 1996), located about 5 to 6 km from the eruption vent on the west and northwest sides of Mt. Pinatubo. This seismicity increased in rate and intensity on the morning of April 2 and again shortly before the initial phreatic explosions about 1600 (local) that afternoon. We estimate that the largest felt events were about M4 to M4.5 based on the felt shaking reports. Local portable seismographs were installed three days later (Sabit et al., 1996) and the USGS and PHIVOLCS installed a permanent telemetered seismic network 5 weeks later (Harlow et al., 1996). The networks demonstrated that these earthquakes were occurring almost exclusively in that same area as the early felt reports, 5-6 km NW of the summit and at depths of 5-6 km deep, and that they were generally

decreasing in average rate of occurrence and average magnitude. An examination of the May seismograms shows that these were high-frequency dVT events. Figures 1a and 1b show the relationship in map view and cross-section of the network locations for the dVT seismicity (blue) and the proximal seismicity (red). High-precision relocation of the distal VT swarm events (Jones et al., 2001) shows that the relocated dVT events occurred in a small volume and apparently define a NE-striking NW-dipping fault. The dVT locations and the continuing felt reports from Yamut and Tarao suggest that the prior felt events of mid-March to April 2 were also dVT events. There were no reports of felt events anywhere near the summit prior to April 2. Some proximal seismicity, in the form of long-period tremor and a few long period events (Sabit et al., 1996), was also recorded beneath the summit area as early as 5 April and we infer that this proximal, low-frequency seismicity began on April 2 with the initial phreatic explosions. Hybrid events began to occur beneath the summit on May 26 and increased with time until a dacite dome began to extrude late on 7 June and continued as the dome extruded. Plinian eruptions began on 12 June and the cataclysmic ultraplinian eruption occurred on 15 June during which up to 30 km³ of tephra (about 5 km³ dense rock equivalent) was explosively ejected (Koyaguchi, 1996).

[Figure 1 near here]

3.2 Case study 2: Soufriere Hills volcano, Montserrat, W.I., 1995-present, VEI 3.

Soufriere Hills Volcano, Montserrat, British West Indies, began erupting in 1995 and eventually produced several VEI 3 magmatic eruptions; see the special issues of Geophysical Research Letters 25 (18 and 19) (1998) and Druitt and Kokelaar (2002). This dominantly andesitic volcano has a summit area consisting primarily of a series of lava domes within a

1-km-wide crater. A 17th-century eruption produced the Castle Peak lava dome but no subsequent eruptions were recorded on Montserrat until 1995 (Siebert et al., 2010). Volcanic seismicity during 1897-1900 and 1933-1937 was strong enough to damage buildings and a period of much less and weaker volcanic seismicity occurred during 1966-67 (Miller et al. 1997).

Although the regional network was very sparse prior to late July 1995, the recent eruptive period apparently began with pulses of increasingly intense swarms of VT seismicity in January 1992, in January 1993, in 1994, in early 1995, then finally in June 1995 (Figure 2). These VTs reached M3.6 and the few locatable events had epicenters (Figure 2a) several to many kilometers northeast and northwest of the summit (epicentral errors are several kilometers). On seismograms, almost all events appear VT with impulsive onsets. Beginning on 14 July 1995 and peaking on 15 July, a strong ($\sim$ M3.5) dVT swarm occurred and located just offshore to the east of Soufrière Hills volcano. On 18 July the first rather small phreatic eruption occurred at the summit. Further phreatic explosions from a second vent occurred on 28 July, the same day that the Seismic Research Institute installed a telemetered seismic network. On 5 August another dVT swarm began in a different location than the previous dVT swarm, 3-5 km ENE of the crater at 3-5 km depth. The dVT swarm started and increased in magnitude for 7 hours with largest events reaching M3.4, and then declined and was followed by increased LF tremor at the summit vents. On 12 August immediately after a phreatic explosion, another dVT swarm at a third location, 3-4 km WNW of the crater and 3-5 km deep, started and increased in magnitude for over 6 hours, with largest events reaching M4.0 before it declined. On 8 September another dVT swarm occurred at a fourth location, 4 km NNW of the crater and about 4 km deep, with largest events reaching M3.8. On 17 September yet another dVT swarm occurred at a new location, about 1 km to the east of the crater and 1-2 km deep with the largest events reaching M3.8. In each of these swarms, the largest events occurred after the middle of the swarm and there were at least 3 events within one-half of a magnitude unit of the largest event of the swarm. Focal

mechanisms and relocations for the largest dVT's show double-couple mechanisms and indicate slip on pre-existing local faults. After September 17, the dVT's died off. Each of the dVT swarms correlated with activity from a different one of the four summit area phreatic vents and the swarms died off as each of the vents appeared to boil dry (Gardner and White, 2002). Hybrid-type seismicity then dominated as a dome emerged beginning on Sept. 25, 1995 (Gardner and White 2002). During the next several years seismicity was almost entirely of hybrid-type as dome lobes continued to be extruded and destroyed during collapse and/or vulcanian explosions (Baptie et al, 2002).

[Figure 2 near here]

3.3 Case study 3: Unzen volcano (Mt. Fugen), Japan, 1991-1995, VEI 1:

The massive Unzen volcanic complex comprises much of the Shimabara Peninsula, with historical dacitic eruptions restricted to the summit and flanks of Fugen-dake, the middle of three large stratovolcanoes with complex structures. The last previous eruptive activity was during the 1790's and formed a lava dome at the summit.

The largest overall dVT energy release occurred in 1984. According to Umakoshi et al. (2001), the first strong dVT seismicity occurred beginning in May 1984 beneath the western part of the Unzen Graben, and epicenters expanded westward into Tachibana Bay with time (Figure 3). The swarm peaked on 6 August 1984 with three main shocks with magnitudes M4.9, M5.3 and M4.8 (magnitudes from USGS-National Earthquake Information Center (NEIC) plus another event of similar magnitude, M5.1, on 19 October within Tachibana Bay (see Figure 3). This dVT seismicity occurred at least 12-15 km west of Unzen and continued

sporadically at decreasing levels over 5 years until November 1989, when there was a significant resurgence in dVT swarms (largest magnitude 3.7) under Tachibana Bay (Umakoshi et al., 2001). The VT seismicity returned in 1989 and through May 1991 located 6-20 km west of the summit. Beginning in April, 1990 dVT seismicity became gradually stronger within the Shimabara Peninsula especially under the western flank of the volcano and on 7 July 1990 a dVT earthquake of M4.9 occurred coincident with the onset of tremor (Nakada et al, 1999). Distal VT seismicity ramped up and peaked at the time of the 1st phreatic explosion in November 1990 then died off rapidly. It then ramped up again and peaked at the time of the 1st phreatomagmatic eruption in February 1991, then decayed. Activity ramped up yet again and peaked in May, 1991 at the time of initial dome extrusion, after which the dVT seismicity died off for good, replaced largely by low-frequency (LF) earthquakes. Dacite dome extrusion began in May 1991 and lasted approximately 4 years. Once the dome appeared, the dVT earthquakes decreased dramatically and were replaced mainly by LF and a few VT events beneath the summit.

[Figure 3 near here]

3.4 Case Study 4: Volcano Tacaná', Guatemala and Mexico, 1986, VEI 1 phreatic eruption:

Another case study from the VDAP archives provides an example from a very sparse network of only 3 vertical component short-period seismic stations. One of the earliest volcano crises for VDAP was associated with Tacaná', a 4060-m-high composite stratovolcano on the border of Guatemala and Mexico with a summit dominated by a series of andesitic lava domes and having viscous lava flow complexes on the flanks. The last eruption previous to 1986 was in 1949-50. Figure 4a shows the distal VT swarm, which began 15 December 1985, was located 10-15 km ESE of Tacaná'. The swarm peaked 19

December with a M4.8 earthquake and then decreased greatly through January. Another pulse of distal VT seismicity then ramped up quickly in the same location peaking on 3 February with a M3.7 earthquake in the same area. Earthquakes continued through 7 February with rumbling heard by local residents at the base of the volcano. Low amplitude harmonic tremor began to appear on 8 February and strengthened the next day when a minor phreatic eruption occurred. More pulses of dVT energy release again occurred during 18-25 February and 2-10 March with events felt and heard by local residents. A final, large pulse of distal VT seismicity ramped up sharply on 7 May with events averaging 1 per minute accompanied by thunder-like noises for 23 hours. Inspected seismograms showed that these events were also from the same area as the previous VTs. On 8 May near noon (local time), a moderate phreatic explosion opened a 20-m-diameter vent. After the phreatic explosion all seismicity was very shallow, low frequency and located within 3 km of the vent. The seismicity decreased slowly through mid-June and was accompanied by the emission of a white steam plume which also decreased gradually from an initial height of 500 m.

[Figure 4 near here]

4. Synthesis

The above four case studies illustrate that distal VT earthquakes are usually the earliest known precursor to eruptions at long dormant volcanoes. The seismicity began at least 7 weeks to 6 years before phreatic and 12 weeks before magmatic eruptions at locations 5 to >15 km away laterally from the eruption vent at from one to several distinct locations. In each case the swarm locations were disconnected spatially from the LF/hybrid seismicity beneath the crater. The dVT s occurred in pulses of swarm-like seismicity that peaked in event rate and average

magnitude about the time of the initial phreatic activity, of subsequent phreatic explosions and, at Soufriere Hills and Unzen, also just prior to the initial magmatic eruptions.

Although the above cases are all for eruptions at intermediate composition volcanoes, this pattern of precursory dVT's also occurs before eruptions of *mafic composition*, such as at Shishaldin, Alaska, and at Mauna Loa, Hawaii, and for silicic composition such as at Chaiten, Chile. At Shishaldin, a seismic network installed only months before the eruption recorded dVT seismicity including a M5.2 earthquake located 10–15 km west of Shishaldin Volcano on 4 March 1999 and a smaller swarm with magnitudes $M \leq 4.4$ at the same location on 13 April 1999, after which the volcano began to erupt, eventually producing a VEI 3 sub-Plinian basaltic-andesite eruption with a column to 16 km height (Moran et al., 2002).

After 25 years of quiescence, Mauna Loa, the world's largest volcano, produced voluminous but very passive VEI 0 eruptions in 1975 and 1984. The 1975 eruption was preceded by 16 months of precursory dVT seismicity at locations 5-8 km WNW, 3-5 km SW, 5-9 km NNW and 9-15 km WNW of the summit caldera, and at depths similar to those distances from the caldera center. Interestingly, dVT swarms again occurred prior to the 1984 eruption as noted by Lockwood et al. (1987): "The overall distribution of earthquakes in the 16 months before the eruption is remarkably similar to the distribution of earthquakes in the 16 months before the July 5, 1975 eruption."

In the case of the 2008 rhyolitic eruption of Chaiten, Chile, the onset was poorly recorded owing to no seismic stations within 250 km. A temporary seismic network deployed in the area two years earlier recorded significant seismicity at Chaiten, possibly indicating that VT seismicity may have begun two or more years prior to the eruption (Lange et al. 2007). However precursory VT seismicity recorded from more than 250 km away began ramping

up 2 days prior to the 1st explosion, and increased to a peak (M5.3) two hours afterward. The two cataclysmic eruptions occurred 4 and 6 days later.

5. dVT Database and Interpretations

Where monitoring is adequate with at least 2 well-operating seismic stations (short-period is sufficient) within 10 km of the vent, the dVT seismicity almost always occurs as a swarm: it ramps up in both the event rate (as in Figure 5) and the average magnitude of events, thereby lacking the classic main shock-aftershock sequence characteristic of tectonic seismicity. This swarm-like character is well illustrated in Figure 5, which shows an example of seismicity leading to the Kasatochi eruption of 2008 (Ruppert et al., 2011). In hindsight, minor VT seismicity had begun a couple of weeks earlier, and then ramped up to the eruption in less than two days. Such behavior is typical of precursory dVT seismicity, for which the event rate and average event magnitude tend to increase with time and multiple events often have magnitudes within one-half magnitude unit of the largest event. This contrasts with tectonic seismicity, for which the largest event occurs at or very near the onset, is more than 1 magnitude unit larger than the next largest event, and the seismicity quickly dies down in event rate and magnitude with time.

[Figure 5 near here]

Table 1 provides examples of 98 dVT swarms for 74 eruptions at 57 volcanoes where we have documented evidence for the pattern of precursory dVT seismicity described above. Table 2 presents data for an additional 38 eruptions at 26 volcanoes where we infer, from descriptions of the precursory seismicity, that those eruptions were very likely also

preceded by this pattern of dVT seismicity. In the following analysis we shall focus mainly on the better documented data in Table 1.

[Table 1 near here]

[Table 2 near here]

Columns 2-4 of Table 1 show the onset date of dVT seismicity, the date of initial phreatic, or phreatomagmatic activity, and the date of initial magmatic eruptions where those dates are known. These data show that dVT seismicity generally preceded the magmatic eruptions by less than a day to several years, with the great majority beginning weeks to a few months before the eruption. All of the five highly explosive ($VEI \geq 5$) eruptions in Table 1 occurred at volcanoes that had not erupted in at least 20 years. Consequently, we believe that dVT seismicity can provide a useful warning for such highly explosive eruptions. Including data from both Tables 1 and 2, we see that the VEI 5 magmatic eruptions at Bezymianny in 1955, Mount St. Helens in 1980, El Chichon' in 1982 and the VEI 6 eruption of Mount Pinatubo 1991 were all preceded by VT seismicity by a minimum of 10 weeks (see Tables 1 and 2 for references). A search of the ISC catalog, even though its reliability is restricted to earthquakes with magnitudes greater than 5 since 1955, and of the Global Volcanism Program database, shows that there have been at least 27 of 35 eruptions of VEI 4 and greater (Siebert et al., 2010) that were preceded by significant seismicity, including all eruptions of VEI 5 and VEI 6. At the volcanoes that hadn't erupted during the previous 20 years, the eruptions for at least 11 of 14 volcanoes, and probably all of them, were preceded by significant VT seismicity; the remaining three volcanoes had no local seismic stations to detect possible dVT seismicity.

It appears that an eruption at a volcano dormant for more than 25 years (see column 6, Table 1) is always preceded by significant dVT seismicity. However, dVT seismicity sometimes precedes eruptions at more frequently active volcanoes. For instance there are at least eight eruptions of VEI 4 since 1955 at volcanoes with less than 20 years eruptive quiescence that were preceded by significant dVT seismicity. More often though, the more frequently active a volcano is, the briefer and less energetic are the precursory dVT swarm(s).

Column 9, Table 1, shows the known depth (where known) and offset and azimuth of the center of the hypocentral clusters of the dVT swarms with respect to the eventual eruption sites. Offset values for dVT swarms range from 1-2 to 45 kilometers from the eventual eruption site, with 60% of 94 fairly well constrained values ranging from 2 to 8 kilometers away and 84% from 2-15 km away. The depths of dVT swarms are remarkably similar to the distances of the swarms from the eventual eruption site out to about 15 kilometers, beyond which the depths level out at roughly 15 kilometers (Figure 6). From this we infer that the dVT seismicity originates near the base of the brittle crust which is known to be thin beneath volcanoes but descends to roughly 15 to 20 kilometers deep elsewhere in arcs (Ito, 1993).

[Figure 6 near here]

The locations and focal mechanisms of dVT's where available often indicate that they originate on local tectonic faults where such faults have been extensively mapped. For example, prior to the VEI 3 eruption of Guagua Pichincha, Ecuador in 1999, dVT seismicity occurred for more than a year on a pre-existing splay of the Romeral Fault Zone about 15 km away from the volcano (Roman, De Angelis et al. 2008). The dVT's waxed and waned in

close relationship to the phreatic activity at the volcano, peaking at the time of the first phreatic explosion and at times of subsequent phreatic explosions, yet the dVT hypocenters never approached to the volcano as the time of the eruption approached. The 2008 eruption of Nevado del Huila, Colombia was preceded by distal VT seismicity NNE and SSW of the summit along the Cauca-Romeral Fault Zone (Cardona, Santacoloma et al. 2009), (Chicangana, 2005). The general lack of migration of dVT events toward vents, the association with nearby fault zones, and lack of geographic connection to proximal seismicity and eruption vent are indicative of triggering of seismicity by aqueous fluid, as will be discussed in a later section.

Column 7, Table 1, shows the magnitude of the largest earthquake in each dVT swarm. We see that eruptions of VEI >4 are almost always preceded by a dVT of at least magnitude 3.9 and rarely as large as M6.7-7 (at Sakurajima, 1914). The largest dVT prior to less explosive eruptions may range anywhere from less than M~3 to the M6.7 prior to the Mauna Loa eruption of 1984 and possibly even as large as the M7.5-M8 near Mauna Loa in 1868.

6. Cumulative dVT moment as proxy for intruded magma volume:

Along with the early eruption warnings that distal VT's can provide, another useful characteristic of distal VT's is that the cumulative seismic moment ΣM_0 is proportional to the intruded magma volume. Column 9 of Table 1 shows the estimated ΣM_0 for precursory VT swarms. McGarr (1976) presented evidence that the ΣM_0 is proportional to a cumulative volume change such as associated with forcibly injecting water into deep well holes. We find that, where intruded magma volume at volcanoes can be estimated from deformation data, those data fit the McGarr relationship. Table 3 and Figure 7 show the ΣM_0 and the

volume change for deep well injections (McGarr, 2014) and for magmatic intrusions. We find that the intruded magma volume can be simply estimated from the Σ Magnitude and ΣM_0 of VT seismicity as follows: We use magnitudes published by ISC or NEIC where available (generally for larger events only), and otherwise use magnitudes published by individual observatories, often calculated from event durations. We assume that the magnitudes are within a factor of two of moment magnitudes. We convert the values of the largest events to seismic moment with $\text{Log}_{10}\Sigma\text{Moment} = 16.1 + (3/2) \Sigma\text{Magnitude}$ from Hanks and Kanamori (1978). Then we sum these moments to get ΣM_0 and the corresponding cumulative magnitude for a single VT earthquake having the cumulative energy for all VT events in the swarm combined.

Then we relate this to intruded magma volume from Figure 7 using the best-fit line:

$\text{Log}_{10} V = 0.71 \log \Sigma\text{Moment}_0 - 5.32$, with the volume change V in cubic meters and seismic moment in Newton meters.

[Figure 7 near here]

[Table 3 near here]

Because ΣM_0 is so heavily dominated by the few largest events, it can be closely approximated by simply summing the moment of those few events within $\frac{1}{2}$ magnitude unit of the largest event. That value is generally about double the value of the moment for the largest event as found by comparing formulas 7 and 11 of McGarr, (2014) for $b \sim 1$ of the Gutenberg-Richter relationship, as is generally the case for VT and well injection seismicity. Secondly, magnitude and moment calculations sometimes include a factor for the station-to-source distance, but for the near-source stations of volcano networks, that factor is negligible. Also note that our formula relies on the ΣM_0 for distal VTs only; however, as mentioned above, distal VT seismicity swamps proximal VT seismicity by an order of magnitude or more. In fact for closed systems with only minor or no fumarolic activity, the

distal VT seismicity prior to onset of phreatic explosions dwarfs all precursory proximal seismicity, both VT and LF, so it is not necessary to separate distal from proximal VT seismicity, nor VT from LF seismicity. A sufficient approximation of intruded volume can be estimated by simply summing the seismic moment for all seismicity prior to phreatic onset. After phreatic onset, one simply sums all VT seismicity

For the reasons just stated, the Σ Moment for VTs can be estimated from the magnitude of the few largest events recorded by just a single local well-calibrated short-period vertical component seismograph regardless of the location of the VTs, assuming S-P intervals ≤ 3 sec, i.e. the VT events are within roughly 20 km of the station. Consequently and fortunately, it is possible to approximate intruded volume at volcanoes with only minimal seismic networks, the situation at many of the world's long-dormant volcanoes. However, it is more precise to estimate the magnitude and to identify dVT swarms using data from multiple stations at different azimuths.

Because the Σ Moment is proportional to cumulative intruded volume, we may infer that small cumulative moments result from small intrusions that are likely to stall before reaching the surface. Table 4 lists 21 examples where dVT swarms occurred near or beneath the flanks of volcanoes without an immediately subsequent eruption but the swarms were accompanied by inflation, magmatic gas emissions and/or high temperatures, indicating that they were caused by magmatic intrusions.

[Table 4 near here]

7. Application of these dVT observations to eruption forecasting:

7.1 Case study 5: Forecast of volcanic activity at Sulu Range, New Britain Island, Papua New Guinea:

On the assumption that some minimum magma volume is required to rise from a magmatic storage region at 5 km or more depth to the surface for long dormant volcanoes, we have used this method of estimating the intruded volume to forecast several eruptions. Table 1 shows that for 28 of 30 eruptions at volcanoes dormant more than 20 years, dVT magnitudes rose to at least M3.7 prior to magmatic eruption and Table 2 lists another 42 of 44 apparently similar cases. We used the dVT to magma volume relation to issue an unusual eruption forecast, unusual in that the forecast was made based on teleseismic data recorded by NEIC. The intrusion volume estimated from the Σ Moment at the Sulu Range was very large, later verified by InSAR (Wicks et al., 2007), and although no magmatic eruption occurred, the intrusion caused phreatic explosions from three locations. The Sulu Range consists of a cluster of partially overlapping small basaltic to rhyolitic stratovolcanoes and lava domes in north-central New Britain. No historical eruptions are known from the Sulu Range although some of the cones display a relatively undissected morphology.

On 6 July 2006, several earthquakes were first reported felt by residents in the Sulu range. Distal VT seismicity began to be recorded by NEIC the same day, with four earthquakes of M4.2- M4.5. On 9 July, additional and larger quakes of M4.8-M5.2 began to occur and by late 11 July, 15 quakes of $M \geq 4.5$ -M5.3 had occurred. With this dVT swarm characteristic of increasing numbers and magnitudes seen from the NEIC data, we alerted the Rabaul Volcano Observatory that a significant intrusion was underway. The moment centroid solutions for these events locate in the Sulu Range and west-northwest to the coast. Booming noises were heard at the Kaiamu and Silanga areas during felt earthquakes and vegetation was reported to having died off. Explosive emissions of mud and steam occurred from the Silanga Hot Springs, Mato Hot Springs, and Bakama Hot Springs. On 13

to 15 July, a temporary seismograph station recorded continuous strings of high frequency earthquakes (including many that were felt) at rates as high as one event every 2 to 3 minutes. Moment centroids of the six largest events, of M5.1-M5.3, locate a few km offshore to the west-northwest. On July 19, a M6.4 event occurred further to the west-northwest about 28 km from the Sulu range. This largest event of the dVT sequence was well recorded and the focal mechanism has a large non-double couple component of 39% and 61% double-couple component. This was followed a few hours later by a magnitude M5.9 quake 10 km even more distant to the west-northwest (along the dashed purple arrow in Figure 8a), after which the seismicity dropped off rapidly. By 24 July, 45 dVT's of $M \geq 4.5$ had occurred. Overall, the seismic moment release occurred in increasingly energetic pulses that peaked on 6, 9, 11, 16 and 19 July. The double couple component of the focal mechanisms show strike-slip with one plane striking west-northwest, strongly indicating that these events occurred along a west-northwest-striking strike-slip fault.

The seismic moment is dominated by the largest event, the M6.4, which together with combined energy of the thousands of smaller events, brings the total energy to the equivalent of one event of M6.5, the seismic moment of which correlates an intruded volume of nearly 200 million m^3 . Subsequently, an InSAR study by Wicks et al., (2007) demonstrated that approximately 1 m deformation occurred at and just west-northwest of the Sulu Range between the dates of 6 June and 22 July 2006, during which this seismic swarm occurred. Wicks estimated that the deformation correlates with an intruded volume of about 600 million m^3 , on the same order of magnitude as the value we estimated from the VT's recorded by NEIC stations thousands of kilometers away.

[Figure 8 near here]

7.2 Case study 6: Forecast of no eruption at Volcan Cosiguina, Nicaragua:

In 2002 two earthquake swarms with the characteristics of dVT swarms occurred 4-6 km northwest of Cosiguina volcano, the northwesternmost of Nicaraguan volcanoes. The swarms consisted of increasingly large events, with events increasing up to M3.5 during the first pulse of 4-6 September 2002, and increasing to M3.9 during the second pulse on 9-10 September (V. Tenorio, INETER, written communication, 2002). Six events had magnitudes within $\frac{1}{2}$ magnitude unit of the largest event of M3.9. During the swarm there were several smaller pulses of VT events spaced at 5.4 to 7.3 hour intervals. This pulsing of seismic energy over time would be highly unusual for tectonic seismicity but is quite common for distal VT swarms and very similar to the 7-10 hour intervals at Soufriere Hills. The cumulative seismic moment for the entire swarm correlates with a small intruded volume of 0.25 to 0.5 million m³ of magma. Given the lack of recent eruptions and very low level of unrest at Cosiguina, and our experience at similar volcanoes, we believed that it was unlikely that such a small intruded volume would reach the surface. Consequently, VDAP and INETER issued a forecast of no eruption unless there were additional dVT swarms. Since then, nothing noteworthy has occurred there.

8. Discussion and suggested mechanism for distal VT earthquakes:

The compendium of field data suggests that, in addition to intrusions of magma, dVTs are commonly associated with: (1) an alignment along regional faults (Jones et al., 2001; Chicangana, 2005; Roman et al., 2008; Cardona et al., 2009; Legrand et al., 2009), (2) a lack of systematic migrations in dVT hypocenters with time and (3) non-double couple movements that involve local tension, as well as shear (e.g. Long Valley, 1980 and 1997

(Ekstrom and Dziewonski, 1985; Julian and Sipkin, 1985; Kanamori et al., 1993; Foulger et al., 2004); Miyake-jima, 2000 (Minson et al., 2007); Sulu Range, 2006 (Section 7.1); Kasatochi, 2008 (Ruppert et al., 2011)). These features are consistent with the opening and slippage along faults following magma intrusion. The lack of migrating hypocenters suggests that the locations of dVTs do not follow a lateral migration of magma; instead, dVTs appear to reflect movements along faults following the intrusion of magma beneath the future location of eruption.

Although changes in the local stress field may trigger fault slip in the vicinity of an intrusion, it is unlikely that they can induce movements tens of kilometers away. Distant connections, however, could result from the propagation of high pore-fluid pressures along faults (Kanamori et al., 1993). Such a mechanism could also explain why dVT swarms tend to peak and die off at the time of the first significant phreatic activity. Whereas phreatic events are triggered by a build-up of high pore-fluid pressures, their occurrence allows the hydrothermal system to depressurize and to halt further phreatic activity until it has again become re-pressurized.

Based on theoretical stress-distributions at volcanoes, Hill (1977) argued that long-distance connections among aquifers could be established by meshes of parallel cracks around and beneath volcanoes, and evidence for such structures has subsequently been found at exhumed volcanic systems (Sibson, 1996). In addition, compared with the flow of magma through dykes or of aqueous fluids through permeable crust, it is physically easier to transmit a pressure pulse through an aquifer to distances of up 15-40 kilometers in the required time frame of hours. We therefore propose that dVT seismicity occurs when aquifers are over-pressurized by the intrusion of magma. Each pulse of intruding magma transmits pressure pulses through aquifers that intersect tectonically pre-stressed faults, thereby reducing the effective normal stress on those pre-stressed regional faults.

The proposed mechanism explains why the dVT phenomenon is ubiquitous at long-dormant volcanoes but uncommon at frequently active volcanoes. Thus, local faults that slipped significantly during dVT swarms before a recent eruption need time to become tectonically

re-stressed, and aquifers which boiled off during a recent eruption are no longer confined and so need time to rehydrate and re-pressurize. The mechanism can also account for the following features of dVT seismicity:

- (1) The occurrence of a large non-double-couple component, as well as double-couple slippage, to the focal mechanisms.
- (2) The swarm-like nature of dVT seismicity and how it ramps up in numbers and magnitudes over weeks on a particular stretch of a fault: as magma slowly intrudes it would gradually over-pressurize an aquifer and so increasingly lubricate the local fault, allowing more and larger patches to slip.
- (3) The occurrence of several main shocks produced by the continued intrusion of magma.
- (4) The decay of dVTs on faults 15 to 40 kilometers away following depressurization of aquifers after initial phreatic explosions.

Although the cumulative dVT moment tells us about the volume of intruded magma, it is clear from Table 1 that this intruded volume does not tell us how explosive the eventual eruption will be. The eruptions of Mount Pinatubo and Mount St. Helens were preceded by dVT swarms consistent with only modest intruded volumes of 0.5-1 million m³ yet they were highly explosive in contrast the inferred intruded volumes at Long Valley and the Sulu Range were about 1000 times larger but did not lead to magmatic eruptions at all. Other factors such as magma viscosity, volatile content, caldera vs. stratovolcano, and ascent time are also important controls on explosivity. For instance, we do see indications where moderately small dVT swarms persisted over several years, such as at Soufriere Hills, Montserrat during 1992-1995, and at Nevado del Huila, Colombia during 2000-2008, apparently implying that magma ascent was very slow; thus the magma was mostly degassed by the time it reached the surface, and produced fairly passive extrusive eruptions. Similarly, at Unzen, by far the largest dVT swarm, with events of M4.9, M5.3,

M4.8 and M5.1, occurred in 1984 over six years prior to extrusion, indicating that the bulk of the intrusion had been degassing for many years before finally reaching the surface during 1991 and producing a relatively passive eruption rather than a much more explosive eruption such as at Mount Pinatubo in 1991 or Mount St. Helens in 1980.

9. Conclusions

Based on our analysis of data for 111 eruptions at 83 volcanoes described here and summarized in Tables 1 and 2, we find that distal VT seismicity is generally the earliest recognized seismic precursor to eruptions at long dormant volcanoes, by which we mean volcanoes without an eruption for at least 25 years. This appears especially true for dVT activity preceding the most explosive eruptions of VEI 4 and greater. For long dormant volcanoes, we find that dVT seismicity precedes eruptions of all magma types from basaltic to rhyolitic and all explosivities from VEI 0 to ultraplinian VEI 6. We find that this distal VT seismicity appears to originate on tectonic fault structures offset anywhere from a few km to more than 30 km laterally from the site of the eventual eruption. The distal VT swarms generally occur at depths comparable to the horizontal distance of the swarm from the summit out to about 15 km distance, beyond which depths level out. We note several important characteristics of this distal VT seismicity including its swarm-like nature, beginning days to years prior to the onset of magmatic eruptions, generally peaking at the time of the initial phreatic or phreatomagmatic eruption and having a large volumetric opening component to focal mechanisms. Most importantly we find that the intruded magma volume can be simply estimated from the cumulative seismic moment of the VT seismicity using the empirical relation: **$\text{Log}_{10}dV = 0.71 \log \Sigma M_0 - 5.32$** , with volume V in cubic meters and moment M_0 in Newton meters. Because the cumulative seismic moment can be approximated from the size of the few largest events, and is insensitive to precise locations, the intruded magma volume can be quickly and easily estimated with but a few

short-period seismic stations. We also document cases in which this knowledge has been used in forecasting an eruption (and the lack of an eruption). We propose a mechanism to account for the dVT phenomenon in which pulses of magma intruding over time transmit pressure pulses through aquifers that intersect tectonically pre-stressed faults, lubricating and opening such faults and thus generating swarms of dVT events with both double-couple and large, at least temporary, opening components in their focal mechanisms.

10. Acknowledgments

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worldwide seismic databases, from the Smithsonian Institute's Global Volcanism Program reports and from published literature. The data for the Unzen case study come from Dr. Kodo Umakoshi. The authors acknowledge long-term support to the Volcano Disaster Assistance Program from the

USAID-Office of Foreign Disaster Assistance, without which this work would not have been possible. Most of all we wish to recognize and thank the hardworking members of the VDAP team who have helped install seismometers and computers on volcanoes all over the world on at least 50 volcanoes since 1985 from which most of the data for this paper have come. We especially want to thank John Power and Chris Newhall for the early discussions that got this work started, Tom Murray for inventing the RSAM system which made much of this analysis much easier, Andy Lockhart for installing most of the seismic equipment and Jeff Marso for installing most of the computer equipment at so many of the foreign observatories. And we also wish to thank the VDAP Chief Scientist - Dr. John Pallister for his leadership and his excellent review and Dave Hill and two anonymous reviewers for their excellent reviews and suggestions that have significantly improved the manuscript.

TABLE 1 Eruptions with known precursory distal volcano-tectonic (VT) earthquake swarms

Volcano Name	Date of 1 st known dVT events	Date of 1st Phreatic (P) or Phreato-magmatic (PM) activity	Date of Magmatic Eruption onset	V E I	Years since prior Eruption	Maximum VT Magnitude	Log10 cumulative Moment nm	D e p t h km	D i s t. km	Azi-muth	References
Anatahan	≤20030506	20030510 PM	20030510	3	>400	M4.1	15.4	3 3	2.5 3	N SW	This study, (Pozgay, 2005)
Augustine	200504	20051220 P	20060111	3	20	M3.7	15.0	14 1	25 1	E E	AVO (Fisher, 2010; Power, 1988)
Augustine	19850705	19860111 P	19860228	4?	10	M4.3	15.7	2 1	1 1	E?	AVO (Power, 2010)
Augustine	19750502	19760122 P	19760123	4	5	M3.5	??	2 1	1 1	SE?	AVO (Power, 2010)
Cerro Hudson	20111025	20111026 P	20111027	3	20	M5.0	16.6	9 9	9 9	W	(Hermosilla Pineda 2012) SERNAGEOMIN
Cerro Negro (Nicaragua)	19950524	19950529 P	19951119	2	4	M3.4	15.0	Z? 8	Z? 5-10	E NW	This study, INETER
Cerro Negro (Nicaragua)	19990223	19990508 P	19990805	2	4	M5	17.2	10 8	12 8	E NW	This study, INETER (La Femina, 2004)
Chaiten	≤20080501	20080502 PM	20080506	5	~400	M5.3	17.2	6 6	5 5	NE NW	This study, (Basualto, 2008; Wicks, 2011) SERNAGEOMIN
Chokai	197312	19740301 P	none	1	140	M3.0?		Z? 0-10	0-10	W	(Nakamura, 1975; (Ui, et al. 1977)
Colima	19910214	10010216 P	19910301	1	3	M3.5	14.5	2 6	6 6	NNW	This study, UCO (Nunez Cornu, 1994;
Colima	19940704	19940721 P	19940721	1	2	M3.4	14.5	15 15	10-15 10-15	W?	This study, UCO (Jimenez, 1995; Zamora-Camacho, 2007)
Colima	19970620	19980706 P	19981120	3	4	M4.0	15.1	3 2	2 2	W	This study., UCO (Zamora-Camacho, 2007)
Cotopaxi	20010601	20020122 P	none	1	60	M3.9	15.9	7 15	10 30	N W	This study, IGEP
El Chichon	198003	19820329 PM	19820403	5	132	M4.2	15.7	5 5	8 8	SE	This study, (Espindola, 2006; Jiménez, 1999; Havskov, 1983)
Fuego, Guate.	19820217	19820223 P	none	1	1	M3.6	M3.7	7-10 7-10	8 8	S	This study, INSIVUMEH
Galeras	19930401	19930404 P	19930607	2	0.5	M4.5	16.4?	6 6	3 3	N	This study, INGEOMINAS (Gomez M, 1997; Jimenez, 2009)
Galeras	20051020	none?	20060113	2	1	M4.7	16.5	6 6	4 4	NW	This study, INGEOMINAS (Jimenez, 2009)
Galeras	20090509	20090509 P	20090607	2	0.1	M4.0	15.1	10 3 4	6 6 5	NE SSE SW	This study, INGEOMINAS
G. Pichincha	19810812	~19810820 P	none	1	100	M<4	<15.1	4-5 4-5	5-7 5-7	SE	This study, IGEP
G. Pichincha	198806	~19881007 P	none	1	5	M3.5	14.7	5 5	5-7 5-7	SE	This study, IGEP
G. Pichincha	~19980701	19980807 P	19990728	3	5	M4	16	14 14	15 15	NE	This study, IGEP (Villagomez, 2003)
Irazu	199410	19941208 P	none	2	17	M3.4	14.2	4 4	3 3	NNW	This study, OVSICORI
Izu-Tobu	19890630	19890711? P	19890713	1	Inf-nite	M5.5	17.5	Z? 4	<40 3	ESE N	(Okada, 1991; Kasahara, 1991; Murase, 2010)
Kanlaon	20020205	20021128 P	none	1	6	M<4?				N	This study, PHIVOLCS
Karymsky	199505	????	19960102	3	26	M6.6	19.0	10? 10?	5 5	S?	(Pavlov, 2003)
Kasatochi	<20080720	2008080714 P	2008080720	4	109+	M5.8	18.0	10 10	10 10	W	AVO (Ruppert, 2011)
Kick'em Jenny	200110	???	20011204	1?	10?	M3	13.6	Z? 2?	3-5 3-5	NW	(Lindsay, 2005)
Kizimen	200904	20101016 P	20101213	3	82	M5.2	17.0	5 5	5 5	E	(Ji, 2013)
Krafla	19770120	???	19770427	1	1	M4.5		Z? 2?	10-40 10-40	N	(Brandsdottir1979) GVP
Kuju	??	19951011 P	none	1	257	M3.0		0-6 0-6	5 5	NW	(Sudo 1998)
Loihi	19960716	???	19960522	1?	???	M5.1	17.5	Z? 2?	4.5 4.5	SW	(Caplan-Auerbach, 2001) HVO
Mauna Loa	18680321	???	18680327	2	2	M7.75	~20.8	"deep" 20	20	S	(Wood, 1914) HVO
Mauna Loa	197401	None	19750705	1	25	M5.5	17.5	6 6 4	6 6 4	WNW NNW SW	(Lockwood, 1987) HVO

Mauna Loa	198104	19840318 P	19840325	1	9	M6.5v	19.1	6 6 WNW 6 6 NNW 4 4 SW	(Lockwood, 1987; Baher, 2003; Koyanagi, 1987; Lockwood, 1987) HVO
Miyake-Jima	19620504	19620824 P	19620824	2	22	M5.9	~18.4	2.5 2 NW	(Minakami, 1964)
Miyake-Jima	20000626	20000627 P	20000829	2	17	M6.4	19.0	15 27 WNW 6? 45 NNW	(Nakada, 2005; Uhira, 2005)
Mount Pago	2002070?	20020803 P	20020805	3	69	M4.3	15.9	5 5 N	This study, RVO
Mount Pinatubo	19910315	19910402 P	19910612	6	~540	M~4.5	~15.9	5 4 NW	This study, PHIVOLCS (Sabit, 1996; Harlow, 1996)
Mt. Spurr	199108	19920611 P	19920627	4	41	M1.7	12.6	4 4 N	AVO
Mt. St Helens	19800315	19800327 P	19800518	5	123	M3.9	15.7	? 4-7 SSE	This study, (Endo, 1981) CVO
Nevado Huila	<200702	20070219 P	200811	2	453	M4.6	16.1	2 2 NNE 7 8 SSW	This study, INGEOMINAS
Nevado Ruiz	198411	19841222 P	19851113	3	68	M4.5	16.0	5 5.5 W 2.5 2 ESE 7 4.5 NNW	This study, INGEOMINAS
On-take	197608	19791028 P	none	1	>1000	4.3	M4.6	4 7-15 ??	(Newhall 1988)
Pacaya	197903	19790810 P	198005	2	3	M5.0	16.9	15 30 SE	This study, INSIVUMEH
Planchon		20100906 P	none?	2	12	M4.5	16.0	13.5 15 SE	L Franco: SERNAGEOMIN
Poas	19900325	19900425? P	none	1	3	M3.1	M3.3	5 5 NE 15 5 SW	This study, OVSICORI
Popocatepetl	199006	19941221 P	19990329&	3	47	M3.2	14.2	5 6 SE	This study, CENAPRED; (De la Cruz-Reyna, 2008; Valdes-Gonzalez, 1997)
Puyehue	201104	201104 P	20110604	3	51	M5.1	16.9	4-6 4-10 ??	L Franco: SERNAGEOMIN
Rabaul	1971	19950918? P	19950919	4?	51	M5.0	17.0	10 25 NNE	This study, Itikarai (MPH thesis)
Raoul	20060312	20060317 P	none	1?	52	M4.8	16.5	"near island"	(Christenson,2006;Cole,2006)
Reventador	20021006	???	20021103	4	26	M4.1	15.4	3 3 SSW?	This study, IGEP
Ruang	20020924	2002092501 P	2002092511	4	53	M3.7	15.1	15 15 NE	This study, PVMBG
Ruapehu	19941104	19950111 P	19950923	3	5	M4.8	16.7	13 17 W	(Bryan, 1999;Hurst, 1999)
Sabancaya	199005	1990 0529 P	19900605	3	2	M4.5		10? 10 NW	ISC
Sabancaya	20130122	20121205 P	None yet	1	22	M6.0	18.3	6 10 NW 10 10 ESE	ISCIGP, INGEMMET
Sakurajima	19140109	19140112 PM	19140112	4	54	M7.0	19.7	10 10 SW Z? 7 NE	(Abe, 1981)
Sakurajima	19550805	19550829 P	19551013	3	5	M~4.5	16.1	6-9 7 SW	(Hidayati, 2007)
Shishaldin	19990225	19990209 P	19990413-7?	3	4	M5.2, M4.4	17.0 15.7	5 13 W	(Moran, 2002)
Soufriere Guadelupe	197506	19760608 P	19760812	2	21	M4.2	15.5	3 1-2 NW?	(Feuillard, 1983)
Soufriere Hills	1992	19950718 P	19950925	3	~365	M4.0	15.4	3 3 NE 4 4 WNW 4 4 NNW	This study, MVO, (Aspinall, 1998; Miller, 2006; Roman, 2008)
Sulu Range	20060706	20060710 P	none	1	>>100	M6.4	18.7	20 20 NW	This study, RVO
Tacana'	198512	19860310 P	19860508	1	37	M4.8	16.6	~10 10 SE	This study, INSIVUMEH
Telica	199903	199905 P	19990810	2	10	M4.0	15.4	4 6 SE	This study, INETER
Tiatia	19730618	≤19730714 P	19730714	4	161	M4.3	15.8	Z? 25 S	ISC
Tolbachik	19750628	???	19750706	4+	5	M4.9	16.6	Z? 20 SW	(Tokarev, 1978)
Tungurahua	199809	19990813 P	19991025	3?	81	M4.0	15.2	10 10 NE, 6 5 SW	This study, IGEP
Turrialba	199605	20100105 P	None	1	130	M3.9	15.5	7 6.5 SW	This study, OVSICORI
Ulawun	20000609	20000927 P	20000928	4	1	M3.9	15.1	5-10 5-10 ??	This study, RVO
Unzen	198405	19901117 P	19910520	3	204	M5.7 M4.0	17.8	15 16 W 12 12 W 7 6 W	(Nakada, 1999; Umakoshi, 2001)
Usu	19431228	19440623 P	19441115	2	34	M5.0	16.9	3 5.5 S	(Minakami, 1951)
Usu	197300806	19770807 P	19780824	3	34	M3.7	15.1	1 1 S	(Okada, 2003; Suzuki, 1980)
Usu	20000328	2000033113 P	20000401	2	22	M4.8	16.6	4 4 S	(Onizawa, 2003)
Yake-dake	19620520	19620617 P	none	2	24	M4.8	16.6	Z? 6 ??	(Newhall 1988)

Column 2: Date of 1st known dVT events; $\leq$ indicates 1st known events were either felt events or events recorded by distant stations so small dVT events likely started before this date.

Column 3: Date of initial explosion, whether phreatic (P) or phreatomagmatic (PM), or of a major upswing on fumarolic/phreatic activity in a few cases.

Column 5: VEI explosivity is taken from “Volcanoes of the World” where available, otherwise as estimated from eruption column height.

Column 7: Magnitude of Largest VT, taken from ISC or NEIC catalog where available, otherwise from local coda magnitude. Where two magnitudes are shown, they are the largest for each of two pulses of distal VT seismicity.

Column 8: Log10 of Cumulative Seismic Moment is computed by summing the moments of the 10 largest events where known; otherwise is estimated as twice the moment of the single largest VT.

Column 9: Average Depth and Distance from summit and Azimuth to center of distal VT swarm;

Column 10: References:

“This study” indicates authors personally reviewed seismograms at each observatory, or more recently, over the internet.

For publications, we show only 1st author’s last name and date of publication.

Observatories personally visited by these authors to review seismograms:

AVO = Alaska Volcanoes Observatory, US Geological Survey (Alaska, USA)

CVO = Cascades Volcanoes Observatory, US Geological Survey (Washington, USA)

HVO = Hawaiian Volcanoes Observatory, US Geological Survey (Hawaii, USA)

IGEP = Instituto Geofísico Escuela Politécnica (Quito, Ecuador)

IGP = Instituto Geológico de Perú (Arequipa, Perú)

INGEMMET = Instituto Geológico Minero y Metalúrgico (Arequipa, Perú)

INETER = Instituto Nicaragüense de Estudios Territoriales (Managua, Nicaragua),

INGEOMINAS (now SGC = Servicio Geológico Colombiano (Manizales, Popayan and Pasto, Colombia)

INSIVUMEH = Instituto Sismológico, Vulcanológico, Meteorológico, y Hidrológico (Guatemala City, Guatemala),

KVERT = Kamchatka Volcanic Eruption Response Team: Institute of Volcanology and Seismology FEB (Russia RAS)

MVO = Montserrat Volcano Observatory (Montserrat, British West Indies)

OVSICORI = Observatorio Vulcanológico y Sismológico (Heredia, Costa Rica)

PHIVOLCS = Philippine Institute of Volcanology and seismology (Quezon City, Philippines)

PVMBG = Pusat Vulkanologi & Mitigasi Bencana Geologi (Bandung, Indonesia)

RVO = Rabaul Volcano Observatory (Rabaul, Papua New Guinea)

SERNAGEOMIN = Servicio Nacional de Geología y Minería (Temuco, Chile)

SNET (now MARN = Ministerio de Medio Ambiente y Recursos Naturales (San Salvador, el Salvador)

UCO = University of Colima (Colima, Mexico)

TABLE 2 Eruptions with inferred precursory distal VT earthquake swarms

Volcano	Start Date of Magmatic Eruption	VEI	Maximum Magnitude	References #
Alaid	19810427	4	M4.6?	GVP, ISC
Arenal	196806	3	<4.5	(Barquero, 1992)
Askja	18750218	5	M6-7	(Thorarinsson, 1962)
Augustine	19860228	4?	M4.3	AVO (Power, 2010)
Beerenberg	19700918	3	M5.7	(Zobin, 2001)
Beerenberg	19850106	2	M5.0	(Zobin, 2001)
Bezmyanny	~19551116	5	M~5.5	(Gorshkov, 1959)
Cerro Hudson	19910814	5+	M5.0	(Naranjo S., 1991) SERNAGEOMIN
Colo Una-Una	19830720	4	M5.8	(Conte, 1999 ;Katili, 1990)
Colo Una-Una	18980502	3+	M5?	(Katili, 1990)
Curacoa	19730712	3	M5.9	(Okada, 1983)(Latter, 1976)
Deception Is	19671204	3	M~5	(Lorca, 1976)
Fernandina	19680611	4	M5.4?	(Filson, 1973;Kaufman, 1980)
Fernandina	20050513	2	M5.0	GVP, ISC
Fuji	17071216	5	~M4?	(Tsuya, 1955;Omori, 1911)
Grimsvotn	19960930	3	M5.6	(Zobin, 2001)
Hibok-Hibok	19511204	3	??	(Macdonald, 1956)
Huaynaputina	16000220	6	M7	(Adams, 2001;De Silva, 1998)
Kliuchevskoi	19511119	2	???	(Gorshkov, 1960)
Kelut	19900210	4	M<4	GVP
Kelut	20071004	2	≤M3.7	ISC
Kliuchevskoi	1966	3	M4?	(Tokarev, 1966)
Lamington	19510121	4	??	(Taylor, 1991)
Loguimay	19881225	3	M5.2	(Zobin, 2001)
Momotombo	1905	2	M5.0	(Leeds, 1974).
Novarupta	19120606	6	<M4.5, M7	(Abe, 1992)
Pagan	19810515	4	~M4.5	(Banks, 1984)
Santa Maria	19021024	6	M5+?	(Williams, 1983) Rose
Sheveluch	19641112	4+	M5.5	(Zobin, 1970, 2001)
Souf. St. Vincent	18120430	4	M~4?	(Anderson, 1903)
Souf. St. Vincent	19020505	4	M~4	(Anderson, 1903)
Tolbachik	20121126	3	M3.5?	KVERT
Ukinrek Maars	19770403	3	M4.5	(Kienle, 1980) AVO
Ulawun	19801006	3	strong	GVP
Unzen	17920210	2	M5-5.5	(Omori, 1928).
Usu	16630816	5	M>4	(Omori, 1911)
Usu	18220312	4	M>4	(Omori, 1911)
Usu	19100725	2	M5.5	(Omori, 1911)

References give last name of 1st author only, and date of publication; ALL CAPS indicate institutions detailed in Table 1

Table 3 Cumulative Magnitude and Seismic Moment versus 1) intruded magma volume where known, 2) extruded magma volume for basaltic eruptions, and 3) and injected H₂O volume at deep wells.

Date	Location	Maximum Magnitude	Cumulative Magnitude	Cumulative Moment (N-m)	Intruded Volume millions of m ³	Extruded Volume* millions of m ³	Log10 of Volume in m ³	References#
	<u>VOLCANOES</u>							
1868	Mauna Loa	7.8	7.80	6.31E+20		3200	9.50	Tilling (1987), Klein (2008)
1977	Ukinrek Maars	4.5	4.70	1.41E+16		2.5	6.40	Kienle (1980)
1999	Shishaldin	5.1	5.20	7.94E+16		14	7.10	Moran (2012)
1996	Akutan	5.1	6.00	2.70E+18	34		7.50	Lu (2005)
1996-98	Mt. Peulik	5.2	5.90	7.17E+17	50		7.70	Lu (2005)
1978-80*	Long Valley	6.2	6.40	5.01E+18	70		7.90	Langbein (1995)
1981-99	Long Valley	6.4	6.70	7.08E+18	180		8.40	Langbein (1995)
1966	Matsushiro	5.3	6.30	3.55E+18	125		8.00	Ohtake (1976)
1989	Poas	3.2	3.40	1.58E+14	0.04		4.70	Rymer (2000)
1984	Mauna Loa	6.7	6.75	1.68E+19		220	8.35	Lockwood (1987)
1975	Mauna Loa	5.3	5.70	4.56E+17		30	7.48	Lockwood (1987)
2010	Kizimen	5.0	5.40	1.58E+17	25		7.40	Ji (2013)
2000	Usu	5.0	5.30	1.00E+17	40		7.60	Jousset (2003)
1984	Tori Shima	5.5	5.60	4.00E+17	40		7.60	Kanamori (1993)
	<u>DEEP WELLS</u>				Injected volume 10e6 m³			
2000	KTB well	0.7	1.50	2.24E+11	0.0002		2.30	McGarr (2014)
1970	Matsushiro	2.6	2.80	2.00E+13	0.0029		3.45	Ohtake (1974)
1991-2005	Paradox well	4.3	4.50	7.10E+15	0.4500		5.67	McGarr (2014)
1967-68	Denver, CO	4.9	5.1	5.10E+16	0.6250		5.93	McGarr (2014)
2010-11	Bowland, UK	2.3	2.50	3.20E+12	0.0042		3.61	McGarr (2014)
2011	Garvin, OK	3.0	3.3	1.12E+14	0.0089		3.95	McGarr (2014)
2000-06	Souls, France	2.9	3.20	7.94E+13	0.0389		4.60	McGarr (2014)
2006	Basel, Switzerland	3.4	3.70	4.47E+14	0.0115		4.10	McGarr (2014)
1987	Ashtabula, OH	3.6	3.90	8.91E+14	0.0617		4.80	McGarr (2014)
2001	Ashtabula, OH	3.9	4.20	2.51E+15	0.3400		5.55	McGarr (2014)
2001	Raton Bas, CO	4.4	4.70	4.50E+15	0.3650		5.60	McGarr (2014)
2005	Raton Bas, CO	5.1	5.30	1.12E+17	2.9200		6.50	McGarr (2014)
2011	Raton Bas, CO	5.3	5.60	3.16E+17	4.9500		6.70	McGarr (2014)
2010-11	Guy, AR	4.7	5.0	4.00E+16	0.6290		5.70	McGarr (2014)
1985-86	Painesville, OH	4.8	5.1	5.62E+16	1.1900		6.10	McGarr (2014)
2010-11	Prague, OK	5.7	6.0	1.26E+18	12.0000		7.10	McGarr (2014)

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* Note: for Mauna Loa and other basaltic volcanoes such as Ukinrek Maars and Shishaldin, all of the intruded volume is generally considered to eventually erupt.

TABLE 4 Intrusions with known precursory distal VT earthquake swarms

Volcano	Date of VT swarm	Depth (km), Distance, Azimuth	Maximum Magnitude	Cumulative Magnitude	References #	Evidence of magma intrusion
Akutan	199603	5, 6 ESE	5.1	6	AVO, (Lu, 2000)	InSAR inflation
Apoyeque	20120906	5, 6 SE	3.8	4.3	This study, INETER	Episodic seismic swarms
Baru	1985, 2006	6? 6 E	4.2	4.4	This study, Univ. Panama	Episodic seismic swarms
Chanbaishan	2002-03	3.5 3 SW	2.8	2.9	(Wu 2007)	Inflation, geochemical
Cosiguina	200209	5, 4-6 NW	3.9	4.3	This study, INETER	Episodic seismic swarms
Deception Is	1998-9	2, 3 NNE	3.4	4.6	(Ibanez 2003)	SO ₂
Dutton	198807-8	1-10, 1-10SE	3.8	4.0	AVO	Episodic seismic swarms
Great Sitkin	20020518	? 6 SSE	4.6	4.7	AVO	Tremor
Hakone	199008	? 5 E	5.1	5.2?	GVP	Episodic seismic swarms
Iliamna	199609-12	5, 5 S	3.2	3.4?	AVO (Roman 2004)	InSAR inflation
Koryaksky	200803-04	5, 5 SW	3.3	3.5	GVP	SO ₂
Long Valley	1978-80	5, 5 S	6.2	6.5	CalVO (Hill 2006)	Inflation, CO ₂
Puerto Aysen	200701-04	? ? ?	6.2	6.3	SERNAGEOMIN, (Legrand 2011)	InSAR inflation
Machin	200081108	6, 6 SE	4.6	4.7	This study, INGEOMINAS	Inflation, springs appear and/or turn white
Mammoth Mtn	198905	8, 6 S	3.4	3.7	CalVO, (Hill 1990; Hill 2005)	Inflation, CO ₂ , Trees die
Mt Minakame, Matsushiro,	1966-67	2 2.5 SE	5.3	6.3	(Ohtake 1976; Stuart 1975)	H ₂ O, deformation
Mehetia	19800306	13 10 SE	4.3	4.6	GVP	High b-value
Momotombo	20000609	5, 5 SE,E	4.1	4.3	GVP, INETER	High temperatures
Mount Peulik	1996-98	15?, 30 NW	5.2	5.4	AVO, Lu 2002)	InSAR inflation
Taal	1991-92	2-5, 3 E	3	3.2?	(Newhall 1988)	Phreatic explosions, Cracks
Yellowstone	198506	10, 20 NW	4.9	5.1?	(Waite 2002)	Inflation/deflation

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Figures

Figure 1: Seismicity prior to the 1991 VEI 6 eruption of Mount Pinatubo, the largest eruption of the last 100 years. A: Instrumentally located epicenters from May 15 to Cataclysmic eruption onset June 15, 1991. Blue circles are distal VT events; red circles are proximal events, mainly hybrid events; black circles are relocated events within those two areas (Jones et al., 2001); Black squares are towns where early felt events were reported. B: Seismicity along NW-SE transect versus time. C: Cross-section along NW-SE transect. Blue shading between the two gray lines indicates area where early felt events were reported 4-8 km west and northwest of the summit. D: Log10 of the daily cumulative Seismic Moment Cumulative and the daily cumulative magnitude versus time. Blue line is for the distal VT moment, dashed where inferred from felt seismicity reports; fuchsia line is for low frequency seismicity and tremor; red line indicates hybrid-type low frequency seismicity. Blue arrow indicates initial phreatic explosion. Smaller orange-red arrow indicates the start and growth of dome extrusion; large orange-red arrow indicates cataclysmic eruption.

Figure 2: Seismicity prior to and during onset of the 1995-2003 eruption of Soufriere Hills volcano, Montserrat, British West Indies. A: epicenters for 1 Oct 1994 to 1 Jan 1996. Pale blue circles are distal VT's located by the sparse regional seismograph network of the Seismic Research Unit; the rest are epicenters determined using the on-island seismic network. Dark blue circles show the best 10% of preliminary distal VT locations; green circles show distal event relocations for many of the same events (Roman et al., 2007); red circles show proximal preliminary epicenter locations while orange circles show high-precision relocations for most of those same events (Rowe et al., 2004). The 4 circle sizes, smallest to largest, indicate $0 < M < 1$, $1 \leq M < 2$, $2 \leq M < 3$, and $3 \leq M < 4$ respectively. B: Seismicity versus time along W-E transect. C: Cross-section seismicity along W-E transect (East at top). D: Cumulative magnitude and Log10 of cumulative seismic moment versus time. Blue dots indicate distal VT's, red triangles indicate proximal events; red line is 3-day running average of the proximal events. Blue arrows indicate times of phreatic eruptions; orange-red arrows indicate onsets of dome extrusions.

Figure 3: Seismicity prior to and during the 1991-1995 eruption of Mount Unzen, Japan. Blue circles are VT events from the ISC catalog prior to Dec 1984, and from Dec 1984 onward from Dr. Kodo Umakoshi and Dr. Hiroshi Shimizu (personal communication; red circles are proximal events, mainly hybrid events, from Dr. Kodo Umakoshi and Dr. Hiroshi Shimizu (personal communication). Circles indicate magnitude, from smallest to largest: $M < 1$, $1 \leq M < 2$, $2 \leq M < 3$, $3 \leq M < 4$, $4 \leq M < 5$ and $5 \leq M \leq 5.7$; B: Seismicity along SW-NE transect versus time. C: Cross-section along SW-NE transect; symbols same as in B. D: Daily cumulative magnitude and Log10 of the daily cumulative seismic moment versus time. Large blue circle indicates largest event, M 5.7; red circles indicate proximal events; red line is weekly running average of proximal seismicity. 1st blue arrow indicates 1st phreatic explosion, 2nd blue arrow

indicates time of resumed phreatic eruptions; orange-red bar indicates duration of dome extrusions.

Figure 4: Seismicity prior to and during the 1986 phreatic eruption of Tacana', on the Guatemala-Mexico border. A: Blue circles indicate distal VT seismicity; red circles indicate proximal seismicity. B: Daily cumulative magnitude and Log10 of the daily cumulative seismic moment versus time. Blue line indicates distal VT seismicity; red line indicates proximal seismicity; dashed blue arrow indicates possible small phreatic eruption; solid lined blue arrow indicates phreatic explosion. Green bar indicates period of small amplitude tremor.

Figure 5: Kasatochi seismograms for short-period seismic station GSMY for 48 hours starting at 00:00 Alaskan Daylight time August 6, 2008. Each line is 30 minutes long. Inset shows how the total seismic energy released per unit time gradually ramped up, with the magnitude of the largest event increasing at least 10 times up to the largest event which occurred late in the sequence. Although this ramp up was relatively rapid compared to most eruptions in Table 1, ramp up is typical of DVT swarms and is the most obvious difference from tectonic seismicity in which the largest event occurs at or very near the beginning of a sequence.

Figure 6: Average depth and distance of each distal VT swarm from summit or eruption site (top left corner) for each eventual eruption. Dots indicate the approximate center of each distal VT swarm. Note that swarms appear to sweep a curve, generally occurring at greater depths farther from the summit. Note that the DVT swarms seem to occur within the lower portion of the brittle crust.

Figure 7: Daily cumulative magnitude and Log10 of the cumulative seismic moment versus Log10 of intruded magma volume for eruptions, where the volume can be estimated from deformation or other non-seismic data, and Log10 of injected H2O volume for deep wells.

Figure 8: Seismicity associated with the large 2006 intrusion and phreatic eruptions at the Sulu Range, New Britain Island, Papua New Guinea. A: Beach balls show the Harvard Moment Tensor solutions for the epicenters of $M \geq 5.0$ distal VT events. Diagonal yellow line indicates inferred fault on which these $M \geq 5.0$ events occurred. Colors indicate the ALOS L-Band Interferogram spanning June 6 to July 22, 2006 showing about 1 m of peak to peak deformation near the Sulu Range (from Wicks et al., 2011). Black circles show epicenters of mainly the onshore events located by a temporary seismograph network installed by Geoscience Australia in late July 2006. B: Log10 of the seismic moment. Blue line indicates VT seismicity, dashed where inferred. Blue arrow indicates time of phreatic explosions

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