Distinguishing high surf from volcanic long-period earthquakes

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Abstract Repeating long-period (LP) earthquakes are observed at active volcanoes worldwide and are typically attributed to unsteady pressure fluctuations associated with fluid migration through the volcanic plumbing system. Nonvolcanic sources of LP signals include ice movement and glacial outburst floods, and the waveform characteristics and frequency content of these events often make them difficult to distinguish from volcanic LP events. We analyze seismic and infrasound data from an LP swarm recorded at Pagan volcano on 12–14 October 2013 and compare the results to ocean wave data from a nearby buoy. We demonstrate that although the events show strong similarity to volcanic LP signals, the events are not volcanic but due to intense surf generated by a passing typhoon. Seismo-acoustic methods allow for rapid distinction of volcanic LP signals from those generated by large surf and other sources, a critical task for volcano monitoring.

1. Introduction

Long-period (LP) seismicity is widely observed at active volcanoes, and LP events are discrete transients characterized by energy in the 0.5–5 Hz band. Numerous source models exist to explain LP signals at volcanoes. These often invoke pressure fluctuations that result from the unsteady movement of bubbly magma [e.g., Chouet, 1996; Kumagai et al, 2002] or steam [Waite et al., 2008] through volcanic conduits and sometimes extend to include the venting of pressurized fluids into the atmosphere and generation of infrasound [Matoza et al., 2009]. Periodic strain changes induced by the oceanic microseism have also been cited as a trigger for hydrothermal LP events [Stich et al., 2011]. Locating and interpreting LP seismicity at volcanoes is complicated by the lack of sharp first arrivals and the potential for waveform modification by complex subsurface structures and topography [Bean et al., 2008]. A sudden increase in either the number or amplitude of LP events is often a sign of volcanic unrest, and LP signals are frequently seen as swarms or groups of repeating events concentrated in time. Identification of LP swarms is important in volcano monitoring because they often precede significant eruptions [Pinatubo Volcano Observatory, 1991; Chouet et al., 1994; Neuberg et al., 2000; Varley et al., 2010]. Not all LP seismicity has a volcanic origin. LP events at Cotopaxi volcano, Ecuador, have been attributed to the resonance of water-filled cracks in glaciers that mantle the volcano [Métaxian et al., 2003]. Ice movement has also been cited as the source of LP swarms observed at Katla volcano, Iceland [Jónsdóttir et al., 2009], as well as at a number of glaciers not located near volcanoes [e.g., Ekström et al., 2006; O’Neel and Pfeffer, 2007; West et al., 2010]. In 1992, a glacial outburst flood on Mount Spurr, Alaska, generated seismic signals having both LP and tremor characteristics that were initially thought to be an eruption precursor [Nye et al., 1995]. Infrasound recordings near coastlines show that breaking ocean waves are also capable of producing repeating events whose energy spectra overlap the LP band and have similar durations [Aucan et al., 2006; Garcés et al., 2003; Le Pichon et al., 2004]. For volcano monitoring and research purposes, it is critical to be able to quickly distinguish whether an increase in LP activity is due to volcanic unrest or to nonvolcanic sources such as wave action or a surging glacier.

We present data from a new seismo-acoustic monitoring network on Pagan volcano, Northern Mariana Islands (Figure 1) that identify breaking ocean waves as a previously undocumented source of swarms of LP seismic signals. The initial nonvolcanic LP swarm was recorded on 12–14 October 2013 in association with Typhoon Wipha passing 770 km west of Pagan. Analysis of data from the seismic network, infrasound arrays and ocean buoy data shows that these breaking wave-generated swarms have several distinct characteristics.
Figure 1
that allow them to be quickly distinguished from volcanic LP activity, despite the remarkable similarity in waveforms and energy content.

2. Instrumentation

The volcano monitoring network at Pagan was installed by the U.S. Geological Survey (USGS) in June 2013 and consists of seven Guralp 6TD intermediate band (0.033–50 Hz), three-component seismometers, two 6-element infrasound arrays (average interelement spacing ~75 m) equipped with VDP-10 differential pressure transducers built by the Cascades Volcano Observatory (0.0125–25 Hz with a nominal response of 10 mV/Pa) and two webcams (Figure 1a). The seismic and infrasound data are sampled at 50 Hz and digitized and time-stamped on site. Webcam images are taken every 30 min. All data are sent via satellite uplink to the Alaska Volcano Observatory (AVO) in real time.

3. Observed Activity and Signal Processing

Although only several months of seismo-acoustic data exist thus far for Pagan, volcanic LP activity dominates the records. During quiet weather times, 10–50 LP signals are generated per hour in association with continuous degassing from the summit vent (Figures 1b–1d). The seismo-acoustic signals recorded during the 12–14 October 2013 LP swarm have energy peaked between 1 and 3 Hz, a duration of ~6–10 s, and frequently came in clusters or sets of 4–14 events with interevent times of 12–16 s (Figure 2). The typical Pagan LP events have similar frequency contents as LP signals from the 12–14 October swarm, but occur less frequently (~30 versus 200/h) and are longer in duration (20–30 s versus 5–10 s). The seismic and infrasound amplitudes of both LP types vary substantially from event to event, but the 12–14 October swarm generally produced seismo-acoustic amplitudes 10–15 times higher than the average LP events at station PGBF (~0.1 versus 1 Pa and ~8 versus 100 μm/s).

Analysis of infrasound data reveals other important differences between the typical Pagan LP events and those generated during the 12–14 October swarm. The time between the seismic and infrasonic first arrivals at station PGBF for typical LP events is ~9 s (Figure 1d), in agreement with the 3050 m distance from the vent to sensors assuming average wave speeds in the solid earth and atmosphere. However, the time between the first arrivals for the swarm LP events decreased to ~1 s (Figure 2c). While it is possible that the seismic wave speed decreased for these events and/or the sound speed in air increased, the magnitude of the change is more suggestive of a source closer to PGBF than the vent.

Further evidence of a change in source location away from the vent for the LP swarm events comes from infrasound array processing. We calculate the azimuth to the strongest infrasonic signal recorded at PGBF in 2 s windows with 50% overlap by the semblance method [Neidell and Taner, 1971] using the average of the correlation coefficients of all possible channel pairs [e.g., Almendros and Chouet, 2003] over a range of possible slowness values from ~3 to 3 s/km. A typical Pagan LP event from 1 October 2013 indicates a stable source azimuth (~53°) in the direction of the vent (Figure 1e). Several swarm LP events from 13 October 2013 show a source azimuth that is initially northwest of PGBF (~330°) that switches to southwest of PGBF (~220°) after several seconds (Figure 2d). Additionally, between swarm LP events, the source azimuth switches back to the vent direction for a few seconds. This suggests that the volcanic source did not shut off during the swarm, but was overpowered by the sources west of PGBF.

In order to better understand the short-term observations and constrain the source of the 12–14 October LP swarm, we analyze seismo-acoustic and ocean buoy data for the entire month of October. The relative...
seismic and infrasound amplitudes across the local network provide a metric to compare activity levels through time and obtain some information about source locations [e.g., Battaglia and Aki, 2003]. In order to compare the amount of LP energy from the swarm across the network to preswarm and postswarm values, we calculate the root-mean-square (RMS) spectral energy in 20 min, nonoverlapping windows for the entire month of October 2013. We sum the 1–3 Hz portion of the spectra and smooth the spectral energy using the running RMS value in 2 h windows (Figure 3a). Six of the permanent seismic stations were operational during October 2013, and they provide a good spread of station distances from the active vent. We assume that most of the energy generated in the 1–3 Hz band comes from the volcano because it is above the microseism, the majority of the stations are located away from trees and other sources of wind noise and no anthropogenic noise sources exist on the essentially uninhabited island. Figure 3a shows that this assumption is valid during most of the month, with energy levels generally scaling with distance from the vent. However, station PGBF shows significantly greater LP energy than PGWW during the 12–14 October LP swarm, despite being farther from the crater. These dates are in good agreement with the timing of the passing of Typhoon Wipha ~770 km west of Pagan. Wipha was a large storm that produced gale strength winds (63–87 km/h) over 800 km from its center (Japan Meteorological Agency http://www.jma.go.jp/en/ typh/). LP energy levels at PGBF subsequently surpassed those at PGWW several times after the 12–14 October swarm (Figure 3a), significantly on 20–21 October correlating with the passing of Typhoon Francisco to the west of Pagan and 24 October corresponding to the passage of Typhoon Lekima to the northeast of Pagan.

Figure 2. Example waveforms recorded during the 12–14 October 2013 LP swarm at station PGBF as in Figure 1. (a) One-hour waveforms of infrasound and seismic data starting at 22:20 UTC on 13 October. The repeating LP signals are less clear in the infrasound trace because of significant high-frequency wind noise. (b) Waveform and spectrogram of 13 min of seismic data indicating that LP events arrive in clusters and have energy peaked at 1.5–2 Hz. (c) One minute of seismo-acoustic data showing the regular spacing of four consecutive events at ~14 s. Arrows indicate approximate first arrivals as in Figure 1c. The infrasound arrives ~1 s after the seismicity, suggesting a source significantly closer to PGBF than the typical volcanic LP (Figure 1c). (d) Azimuth determined as in Figure 1e. Each LP first shows energy originating from ~330° that switches to ~220° midway through the event. Between the LP events, the most energetic source is from the azimuth of the volcanic vent (~53°; red line).
Figure 3. (a) The 1–3 Hz spectral energy for the vertical component (Z) of each seismic station and infrasound sensors (i) at PGNW and PGBF. Typically volcanic signals dominate the LP band, reflected in highest LP energy at the station closest to the vent (PGWW). As Typhoon Wipha passed, LP energy at the station closest to the west beaches (PGBF) surpassed levels at PGWW. Three other passing storms can also be seen in the data, as well as teleseismic energy from a M7.1 east of Japan. (b) Significant wave height and wave direction from ocean buoy 52211 located 300 km south of Pagan. Open ocean wave heights of 7 m were recorded in association with Typhoon Wipha and correlate with the peak in seismic LP energy recorded during the LP swarm at Pagan. (c) Azimuth determined as in Figures 1e and 2d, except that window length is 10 min. The continuous degassing from the volcano is typically the most energetic source, but large breaking waves on the west side of the island overwhelmed the volcano during 12–14 October. The shift in wave direction recorded at the buoy from the southwest to the northwest is reflected in the change in infrasound source locations during 12–14 October at PGBF.
We compare the wave height and direction from the ocean buoy nearest Pagan with the trends in LP energy. No direct observations of wave height are available at Pagan, and the closest buoy recording wave information is 300 km south of Pagan and 9 km northwest of the island of Saipan (National Oceanic and Atmospheric Administration National Data Buoy Center http://www.ndbc.noaa.gov/station_page.php?station=52211). This buoy recorded an increase in significant wave height starting on 11 October that peaked on 13 October at 7 m (Figure 3b), indicating very high and energetic breaking waves along the coastline. The peak in wave buoy height occurs about 1 day prior to the peak in the 1–3 Hz seismic energy (Figure 3a) because the storm tracked from the southwest to the northwest and passed due west of Saipan about 1 day before passing west of Pagan.

Long-term changes in the locations of the most energetic LP infrasound during October 2013 also support the hypothesis that the 12–14 October LP swarm is not volcanic in origin. Following the methodology outlined above, we calculate the azimuth to the strongest infrasound signal recorded at each array in the 1–3 Hz band for successive, nonoverlapping, 10 min windows. Since the installation in June 2013, the dominant direction of the LP infrasonic source is remarkably consistent for both arrays and can be attributed to robust, continuous degassing from the summit vent (Figure 3c). However, during the 12–14 October swarm, both arrays indicate a westward shift in the most energetic LP source, although the source azimuths do not always intersect. The source azimuth from the PGBF array from 12–13 October varies from 213–330°, indicative of multiple or nonstationary sources away from the active vent (Figure 3c). The azimuth to the source then moves north on 14 October to 339°. The southwest to northwest shift in infrasound source locations matches the change in wave direction sensed by the ocean buoy (Figure 3b) and agrees with the storm track of Typhoon Wipha. The PGNW array shows a similar shift in azimuth during the course of the storm but the change in angle is subtle (~213 to 228°), possibly because the array is farther from the western beaches where the waves were breaking. Other major storms in October also generated LP energy that overpowers the volcanic source, but never as consistently as during the 12–14 October swarm and for periods of less than a day (Figure 3a).

We test whether the occurrence interval of LP signals recorded on Pagan shows a similar period to those from storm-generated swell, which typically range from 10 to 30 s [Barber and Ursell, 1948]. Here we determine the periodicity of our LP signals by calculating the spectra of the 1–3 Hz filtered waveform envelope. Aucan et al. [2006] performed similar analyses on infrasound and nearshore sea-surface elevation data during high surf in Hawaii and found that incoming waves produced a 0.07 Hz (14.3 s) peak in the spectra. Figure 4 shows envelope spectra from the vertical seismic channel and colocated infrasound microphone at station PGBF. The spectra were calculated in 1 h windows, summed and individually normalized. The first period covers quiet weather during 15–31 July 2013 when typical volcanic LP events were observed at a rate of 30–50/h. The second period covers quiet weather prior to the LP swarm on 6–9 October 2013. The third period spans the LP swarm on 12–14 October and has a clear spectral peak at 0.07 Hz (14.3 s) that is not seen in either the July or early October spectra, indicating that volcanic LP activity is not modulated or triggered by ocean waves or the microseism. Instead, the LP signals generated by ocean waves breaking on the western shore of Pagan are modulated by the period of the large swell.

Figure 4. Spectra of the envelope of 1–3 Hz filtered waveforms from the vertical seismic (z) and colocated infrasound (i) channels at station PGBF during a quiet weather period when volcanic LP events dominated (15–31 July 2013), 4 days leading up to the swarm (6–9 October 2013) and during the 12–14 October LP swarm. The peak at 14 s (0.07 Hz) during 12–14 October is due to the regular spacing of ocean waves that broke along the western shore of Pagan and produced the 1–3 Hz LP signals.
4. Discussion

Our results show that high surf is capable of generating significant seismic energy in the same frequency bands associated with volcanic unrest and can generate repeating events that share characteristics of volcanic LP swarms that often precede eruptions. This result is based on joint analysis of seismic, infrasound and buoy data and highlights the advantage of multiparameter observations for analyzing LP signals. Seismic analysis alone, for example, may have produced a different result. Seismic first arrivals from the 12–14 October LP signals show energy arriving first at the station closest to the summit (PGWW) followed by PGBF (Figure S1 of the supporting information). This is inconsistent with the distribution of seismic energy seen in the LP band (Figure 3a). However, since accurate locations of LP earthquakes are difficult, one interpretation would be that the swarm was deeper and was located away from the summit and that the anomalous energy ratios are due in part to site effects. Particle motion analysis of the LP signals indicate that the events recorded at PGBF are dominated by Rayleigh waves, while the same events at PGWW are dominated by body waves (S) (Figure S1 of the supporting information). Although PGWW is laterally farther from the western beaches and ~200 m higher in elevation than PGBF, it appears that the LP energy generated by the strong surf propagated to PGBF through a low-velocity, near-surface layer, while taking a deeper and faster path to PGWW.

These complexities reinforce how the spatially varying velocity structure of volcanic edifices can complicate interpretation of LP seismic data. To our knowledge, the only other reference to similar LP signals recorded at a volcano comes from Stich et al. (2011). From a single seismometer, Stich et al. (2011) observed LP seismic signals with interevent times of 10–20 s at Deception Island volcano and attributed the swarms to dynamic triggering of a hydrothermal LP source through a volumetric strain change induced by the microseism. Although we have not reanalyzed the Deception Island data, the waveform characteristics, event spacing and grouping of LP signals into sets of ~4–12 are remarkably similar to what we recorded at Pagan and what is expected from surf [Aucan et al., 2006; Barber and Ursell, 1948]. The possibility of surf as the origin of LP signals with interevent times of 10–20 s should be considered and tested in addition to other source models. Our interpretation of surf-generated LP signals greatly benefited from having multiple seismic stations, infrasound arrays and buoy data.

We have identified high surf as a source for LP seismic swarms, adding to a list that includes volcanic activity, ice movement and glacial outburst floods. Large surf-generated LP events are capable of overpowering volcanic LP seismic-acoustic signals in cases where the sensors are relatively close to the shore, and the waveform similarity and frequency content can lead to confusion about the source of the LP events. We distinguish surf-generated from volcanic-generated LP signals through changes in the time between seismic and infrasonic first arrivals, spatial comparisons of the RMS seismic energy, changes in source locations of infrasonic LP energy and identification of the periodic event occurrence in the spectra of the waveform envelopes. This study highlights the utility of multiparameter observations for characterizing volcanic and nonvolcanic LP sources at remotely monitored volcanoes.

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