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Stress and mass changes at a “wet” volcano: Example during the 2011–2012 volcanic unrest at Kawah Ijen volcano (Indonesia)

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Abstract Since 2010, Kawah Ijen volcano has been equipped with seismometers, and its extremely acid volcanic lake has been monitored using temperature and leveling sensors, providing unprecedented time resolution of multiparametric data for an acidic volcanic lake. The nature of stress and mass changes of the volcano is studied by combining seismic analyses and volcanic lake measurements that were made during the strongest unrest ever recorded by the seismic network at Kawah Ijen. The distal VT earthquake swarm that occurred in May 2011 was the precursor of volcanic unrest in October 2011 that caused an increase in shallow earthquakes. The proximal VT earthquakes opened pathways for fluids to ascend by increasing the permeability of the rock matrix. The following months were characterized by two periods of strong heat and mass discharge into the lake and by the initiation of monochromatic tremor (MT) activity when steam/gases interacted with shallow portions of the aquifer. Significant seismic velocity variations, concurrent with water level rises in which water contained a large amount of steam/gas, were associated with the crises, that caused an although the unrest did not affect the shallow hydrothermal system at a large scale. Whereas shallow VT earthquakes likely reflect a magmatic intrusion, MT and relative seismic velocity changes are clearly associated with shallow hydrothermal processes. These results will facilitate the forecast of future crises.

1. Introduction

Kawah Ijen (2386 m) is a stratovolcano located within Ijen caldera, situated at the easternmost part of Java Island (Figure 1a). The visible superficial manifestations of the hydrothermal system associated with Kawah Ijen consist of a crater lake, arguably the largest natural reservoir of hot acidic waters on Earth ($V = 27 \times 10^6$ m$^3$, $T > 30^\circ$C, and pH $< 0.3$), and a few thermal discharges and crater fumaroles [Delmelle et al., 2000] which produce significant amounts of sulfur. Specific dynamics and seismicity are triggered at “wet” volcanoes by the hydrothermal activity [Syahbana et al., 2014]. The most recent magmatic eruption of the volcano took place in 1817 [Junghuhn, 1853], but phreatic and geyser-like activity frequently occurred since that time [Caudron et al., 2015a]. Recently, the volcanic lake has shown signs of instability that were coupled with an increased seismicity [Caudron et al., 2015a]. The daily production of several tons of sulfur has led to the development of mining activities in the crater. More than 200 persons/day mine the sulfur close to the lake and are thus vulnerable to phreatic eruptions. Even a small eruption could expel an important amount of acidic water from the lake and trigger destructive acidic lahars which threaten more than 10,000 people living in the Kawah Ijen caldera. Although periodic geochemical/geophysical campaigns were carried out in the past, to date no effort was undertaken for continuously monitoring the volcano hydrothermal activity (from a research perspective).

Since June 2010, the volcano and its surroundings have been equipped with nine seismometers and several sensors immersed in lake waters. Although finding sensors capable of surviving in such an extreme environment has been challenging, they significantly improved the monitoring capabilities and data set. A swarm of distal volcano-tectonic (VT) earthquakes occurred to the SW of Kawah Ijen in May 2011. It was the first
expression of a magmatic intrusion. Unusual shallow seismic activity began in October 2011 and culminated in mid-December 2011 and was followed by another episode in March 2012. These significant changes in volcanic activity, hereafter termed unrest, affected the shallow portions of the volcanic system and led the authorities to increase the volcanic alert level to 3 (on a scale of 1 to 4). The earthquake swarms were the strongest ever recorded since the volcano has been instrumentally monitored in 1987 and showed different patterns for the parameters investigated hereafter. Unusual seismicity, heat discharge, and mass changes were recorded by our sensors with an exceptional temporal resolution of data for an acidic volcanic lake.

Recently, Sens-Schönfelder and Wegler [2006] proposed the use of the repetitive waveforms of seismic noise cross correlations to track for subsurface volcanic edifice seismic velocity changes. Although this technique successfully forecast eruptions at the basaltic Piton de la Fournaise volcano (La Réunion, France) [Brenguier et al., 2008; Duputel et al., 2009; Obermann et al., 2013; Sens-Schönfelder et al., 2014; Rivet et al., 2014] which has a weakly developed hydrothermal system [Zecevic et al., 2013], other studies at different volcanoes required less straightforward interpretations of volcanic phenomena [Mordret et al., 2010; Anggono et al., 2012]. Several mechanisms produce seismic velocity variation at volcanoes when environmental and external influences are discarded [e.g., Sens-Schönfelder and Wegler, 2006]: stress changes due to volcanic pressure sources [Brenguier et al., 2008; Mordret et al., 2010], caldera formation [Anggono et al., 2012], or topographic changes [Anggono et al., 2012]. Other phenomena linked to the dynamics might explain seismic velocity variations, especially for a well-developed system such as Kawah Ijen.

In this paper, we investigate the nature of stress and mass changes that occurred at the volcano by combining seismic analyses and volcanic lake measurements. The biases and the limits of the seismic velocity variation...
retrieval in a system such as Kawah Ijen will be first assessed and discussed. The mechanisms that could trigger the observed seismicity are then discussed, followed by an investigation of the enthalpies and thermal input required at the lake bottom to heat the large volcanic lake. Finally, an interpretation of the unrest by combining all the parameters will be proposed.

2. Network and Sensors

Before this study, only one short-period (vertical component) seismometer had been operating since 1987 to continuously monitor Kawah Ijen volcano (CVGHM instrument (Center for Volcanology and Geological Hazards Mitigation, Indonesia)). The Royal Observatory of Belgium and the U.S. AID-U.S. Geological Survey Volcano Disaster Assistance Program installed three broadband (three components Trillium 120P sensors, Taurus data loggers, and Nanometrics) and four short-period seismometers (three components L22D and one component L4 sensors, Mark Product), respectively, in the vicinity of the volcano (Figure 1). Seismic data are acquired at 100 Hz and transmitted to the Kawah Ijen Observatory, located at ~15 km to the south of the crater (Tamansari in Figure 1). We also used and analyzed seismic data from a broadband station (GE-JAGI from GEOfon network, http://geofon.gfz-potsdam.de/) located ~45 km to the south of the volcano (Figure 1a).

A temperature sensor was immersed in lake waters close to the western shore (Figure 1b, DAM), at a depth of ~5 m (iButton, accuracy of 0.5°C, resolution of 0.625°C). The reliability of the measurements was verified by other probes (Troll 500 device, In-Situ Inc., 0.1°C of accuracy and 0.1°C of resolution for temperature recordings and 0.1% of accuracy 0.005% of resolution for pressure measurements) that were installed at the same position, before the time of our study. The instruments measure the lake water temperature at least every hour.

The lake level is monitored weekly by CVGHM using a leveled wooden stick. Based on the historical record, the lake level generally varies by ~4 m over a year [Caudron et al., 2015a]. The volume of the lake was estimated from echo sounding surveys carried out in May 2010 and July 2011 using a Kongsberg-Simrad ES60 single-beam sonar equipped with a dual-frequency (50 and 200 kHz) transducer.

3. Methodologies and Results

3.1. Ambient Noise Cross Correlation and Velocity Variations

In this section, we first introduce the procedure applied to process the seismic data. We then present the seismic velocity variations and the cross-correlation coefficients evolution during the period of study.

We applied the MSNoise program which is an open source python package for monitoring seismic velocity changes using ambient seismic noise. It computes the cross-correlation function (CCF) of ambient noise time series for individual pairs of sensors (vertical components only in this study) then measures velocity variations of different arrivals (direct or coda waves) between these individual CCFs and a defined reference. A detailed methodological description and a link to the program can be found in Lecocq et al. [2014].

A cross-correlation function (CCF, Figure S1 in the supporting information) contains negative and positive lag times, named acausal and causal parts, respectively. The causal and acausal sides correspond to the waves traveling from the first station to the second and from the second to the first, respectively [Bensen et al., 2007]. A closer inspection of the CCF (Figure S1) reveals a strong asymmetry between both parts that reflects an inhomogeneous distribution of sources [Stehly et al., 2006]. One can take advantage of this asymmetry to assess the microseismic source region by, for example, taking pairs at sufficiently long distances such as KWUI-POSI (first CCF (thick red line) in Figure S1, interdistance of 10.65 km) versus POSI-PSG (third CCF (light green curve) in Figure S1, interdistance of 8.85 km) or POSI-TRWI (second CCF (light black curve) in Figure S1, interdistance of 9.55 km). The maximum is shifted toward the acausal part of the CCF for KWUI-POSI, whereas the contrary is observed for the two other pairs, clearly indicating that the background noise propagates northward across the network. This result agrees well with the location of seismic noise sources and significant wave height studied by Stehly et al. [2006]. It is worth noting that the spatial instability of the source distribution is not a limitation for noise-based correlation monitoring as long as part of the noise spatial distribution is stable [Hadziioannou et al., 2009].

The plot of the stacked CCFs (117 days, 0.5–1.0 Hz, Figure S1) as a function of their distances shows that the wavelets with the maximum amplitudes propagate with a group velocity of around 892 m s$^{-1}$ (Figure S1, $r^2 = 0.99$). The linear regression is performed by using the maxima of the CCF envelopes.
Brenguier et al. [2011] has shown two main drawbacks of the technique. The first consideration is the determination of the appropriate length of seismic noise to converge to stable CCFs which depends on the nature of the noise, the distance between the sensors, and the intrinsic attenuation and scattering properties of the medium. Many studies [Brenguier et al., 2008; Duputel et al., 2009; Mordret et al., 2010; Clarke et al., 2011] used a 10 day stacking procedure to recover reliable velocity variations. Our tests revealed that a stack of the 19 preceding daily CCFs and the day considered (hereafter referred to as a 20 day stack) to compute the current CCF is providing a satisfying signal-to-noise ratio. In that case, coherence values are systemically above 0.6.

The second consideration explained by Brenguier et al. [2011] concerns the nonstationarity of ambient seismic noise. For a reliable analysis, relatively stable sources are expected in the frequency band of interest. The data were band passed between 0.1–1 Hz (Figure S2) and 0.5–1 Hz (Figure S3) where sufficient microseismic energy can be radiated and to avoid perturbations induced by the continuous volcanic tremor whose energy is mostly distributed between 2 and 10 Hz and periods of monochromatic tremor (dominant frequencies of 1–2 Hz; see below). Indeed, we clearly detect tremor periods in the direct arrivals of CCFs filtered between 0.5 and 2 Hz (causal side of the CCFs for KWUI-POSI, Figure S2) in November–December 2011 and January–March 2012. Two methods were used to estimate the velocity variations; the stretching technique which stretches and compresses the reference CCF, in the time domain, in order to obtain the best coherency with the current CCF [Sens-Schönfelder and Wegler, 2006]; and the moving window cross-spectral analysis [Poupinet et al., 1984], hereafter referred as doublet which measures the traveltime difference between two waveforms in each time window by fitting the phase differences in the frequency domain [Zhan et al., 2013].

The choice of the CCF analyzing window is important and should reflect strongly scattered waves. The analysis for KWUI-POSI was performed in the 17–57 s windows of lag time to take advantage of sufficiently scattered waves. For the doublet, Clarke et al. [2011] decreased the errors by increasing the number of sliding windows and the number of points used to transform the windowed cross correlations into the Fourier domain. Moreover, a choice of broadly overlapping windows improves the stability of the velocity measurements. We took into account the aforementioned recommendations and used a window length of 12 s shifted by 4 s for each computation of the time delays in the frequency domain.

Some tests were also performed using the raw and instrumentally corrected data. The results of our comparisons revealed that the effect of instrumental correction is obvious in the CCF, whereas it does not affect the calculated velocity variations.

The retrieved velocities strongly fluctuate depending on the station pair. Among the pairs that extensively sample the edifice, KWUI-POSI presents the best data availability. KWUI-POSI interferogram includes very low amplitudes in the causal part compared to the acausal one for the direct waves (e.g., Figure S2), most likely due to the anisotropic distribution of the sources [Lesage et al., 2014]. However, clear arrivals are observed in the scattered portions of the CCFs, both for the causal and acausal parts. We computed the velocity variations in the causal and acausal parts of the CCF and in the acausal only. These tests revealed negligible differences, and the results presented in this study derive from computations performed using both sides.

Results derived from the stretching technique are much more stable than using the doublet technique (Figure 2b) and will be used throughout this study. Background velocity variations \( \frac{dv}{v} \) do not exceed \( \pm 0.1\% \) for KWUI-POSI pair. However, fluctuations were more larger from October 2011 to April 2012 (Figure 2b). This period is characterized by two episodes of strong unrest, which are subdivided following CVGHM alert levels (1A-2A-3A for unrest A and 2B-3B for unrest B (Table 1)). This terminology will be used to separate the important periods of observations throughout the document. Two significant drops \( dv/v > 0.2\% \) are observed. The first occurred at the beginning of the unrest, i.e., period 2A-3A, and was followed by an increase until the end of period 3A. The second seismic velocity decrease started by the end of period 3A and culminated at the beginning of period 3B. Velocity variations then got back to their background precrisis values. No other variation exceeded 0.1\% until the end of the period of study for this pair of stations. Before the beginning of October 2011, the cross-correlation coefficient (CCC), in the 17–57 s windows, was above 0.9 (Figure 2a). Interestingly, velocity drops do not systematically coincide with an associated CCC decrease (Figure 2a). After the crisis, the CCCs progressively returned to the previous values and remained above 0.9.

### 3.2. Volcano Seismic Events

The crises were also characterized by an unusual numbers of volcano-tectonic (VT) earthquakes and monochromatic tremor (MT) that occurred at different periods. Whereas MTs are not recorded during normal...
volcanic activity, only a few VTs are typically observed in normal conditions (2/month). Those events are very important to understand the nature of the crises since they are not related to the same processes: i.e., VTs map stress concentration in the crust providing information on the state of solid [Kumagai et al., 2002], while MTs are related to pressure disturbances accompanying fluid mass transport [Chouet, 1996]. We describe and analyze these events using specific processing tools in the following section.

The earthquake swarm that occurred on 21 May 2011 likely constituted the first step in the generation of the successive crises. Several distal VT earthquakes having magnitude range of 1.3 to 3.3 were located nearby the caldera (Figure 1). Unfortunately, the seismic network was not fully operational (only POS, PSG, IJEN, RAUN, and JAGI were recording). The events recorded on 21 May are representative of a seismic swarm, rather than a classical main shock-aftershock sequence, and were felt at Sempol and Tamansari (Figure 1). These events are the first stress change associated with an intrusion of new magma or fluids into the volcanic system [White and McCausland, 2015]. At this time, no change in shallow seismic activity, i.e., low-frequency events or tremor,
occurred at Kawah Ijen and Raung volcanoes. Moreover, no significant change was observed in the Kawah Ijen lake system. This type of seismic pattern has been observed at more than 60 volcanoes worldwide [White and McCausland, 2015; Pozgay et al., 2005; Surono et al., 2012] and reflect the driving force (intrusion) of the later unrest.

Renewed seismic unrest started in October 2011 with several VT events recorded every day (Figure 2d). VTs are characterized by sharp, mostly impulsive onsets of \( P \) and \( S \) waves, with typically broad spectra extending up to 15 Hz [Lahr et al., 1994] (Figures 3a and 3b). VT activity peaked at the transition between periods 2A and 3A (Figure 2d), and some earthquakes were felt by the sulfur miners in the crater and on the volcano flanks (Sempol).

After reviewing all the VT waveforms for the period of interest, 172 earthquakes could be located. In total, 744 \( P \) phases and 376 \( S \) phases have been manually picked. In most cases, we did not use IJEN and RAUNG stations (Figure 1) which are not always synchronized with the rest of the network. Instead, we used the respectively colocated TRWI and MLLR stations which were recently installed. If TRWI and MLLR data were not available, IJEN and RAUNG measurements were still included in the location procedure, but with a lower weight (flagged as less reliable). In most cases, an \( S \) phase arrival time was measured at TRWI and at least at one broadband sensor (POS or PSG).

The location procedure has been carried out using a modified version of Hypo71 [Lee and Lahr, 1975] which takes into account the elevation of the stations. The next major problem was the lack of a reliable seismic velocity model for the Kawah Ijen. The only published 1-D model [Sulaeman, 2006] is presented in the Table S1. The inferred interface depths as well as the very low and variable \( V_p/V_s \) ratio remain doubtful. It is also unclear whether the layer 0 includes the terrains above sea level or if the 0 is relative to the highest altitude of the stations. Instead, we applied the beginning of the approach described in You et al. [2013]: all the events are measured manually and are located using a simple velocity model, layered 2 by 2 km and interpolated from the worldwide ak135 model (flat-Earth approximation) [Kennett et al., 1995] and with constant \( V_p/V_s = 1.73 \) ratio (14 layers in total, Table S2). We then slightly lowered the velocity for the layer above sea level (2.7 km/s for \( P \) wave) with the same \( V_p/V_s \) ratio.
Figure 4. Reliably located VTs during the studied period (0 km corresponds to the sea level). Red colors are events in the volcanic area, and blue circles lie outside. (a) Planar view. (b) Cross section (along the latitude). (c) Cross section (along the longitude).

Of the 172 events, 123 could be located with a vertical error less than 1.5 km. They are displayed in Figure 4, where events situated in Kawah Ijen volcano are colored in red and others in blue. Most epicenters are found close to the Kawah Ijen crater, and their focal depths are spread between the surface (\(\sim 2\) km above sea level (asl)) and 1 km below sea level.

Several features distinguish MTs from the broadband continuous tremor and other specific seismic events. MTs correspond to sustained tremor bursts (up to 50 min, Figure 3c), sometimes having extended coda, and mostly showing one dominant spectral peak (Figure 3d). Hence, these events are distinct from other seismic manifestations, and their detection is straightforward. An event starts when the amplitude exceeds by a factor of 2 the background amplitude and ends when it gets back to the background level for more than 10 s. The MT is then encoded as an event in the database along with its duration. MTs were recorded during the crisis, and only few were detected during period 1A (Figure 2c). A large number were observed during periods 3A and 2B (Figure 2c) and were mostly detected at IJEN station (Figure 5a). Their activity strongly decreased at the end of period 3A (Figure 2c). Several MTs were again recorded during period 2B. By contrast, few VTs were detected. This activity continued until the very beginning of period 3B (Figure 2c) then strongly diminished and finally vanished. Their dominant frequency oscillated between 1.3 and 1.6 Hz during sustained activity (Figures 5a and 5b) and gradually evolved from 0.9 to 1.6 Hz during period 2B. When their activity decreased (at the end of period 3A and beginning of 3B), they displayed a higher dominant frequency and shorter durations (Figures 5a, 5b, and 5e).

Seismic events such as MTs are commonly associated with eruptive activity and unrest [Chouet et al., 1994] and, hence, represent an important seismic signature from which the origin of the unrest can be better constrained. The dominant spectral peak of these events was mostly similar even if recorded at different seismic
Figure 5. Dominant frequency (F0) of monochromatic tremor (MT) retrieved at (a) IJEN (log scale) and (b) KWUI (log scale) stations and (c) the difference of the dominant frequencies. (d) Maximum amplitude (in mm, IJEN station). (e) Duration measured (in seconds, IJEN station). Light grey and dark grey shadings represent the periods of crises (alert 2 and alert 3, respectively).

stations (Figure 5c) and was sometimes accompanied by higher-frequency peaks at some sites. Overtones were detected when MT events were particularly energetic. The fact that the main peak dominated the spectra at different stations (more than 90% of the total spectral energy) indicates that one mode of vibration is preferentially excited and points to a source effect rather than path effect [Chouet et al., 1994; Chouet, 1996].

3.3. Volcanic Lake

After assessing their reliability and biases, this section presents the outstanding temperature and volume variations recorded during the crises. These are key parameters to constrain the mass changes that occurred in the shallow hydrothermal system.

Based on echo sounding profiling carried out in June 2010 and July 2011, it has been determined that the lake has a volume of $2.85 \times 10^7$ m$^3$ (for a lake surface of $3.95 \times 10^5$ m$^2$) and the maximum depth reaches 170 m [Caudron et al., 2015a]. After the typical seasonal decrease, the lake temperature started to rise in mid-October 2011 and remained stable for a week until the end of period 1A (Figure 6a). At the beginning of the first unrest (i.e., period 2A), it sharply increased again, the lake color suddenly turned to grey (Figure 7a), and a strong smell of sulfur was reported. After remaining stable, the temperature changed dynamically, as indicated by sharp drops and increases between the end of 2B and the beginning of 3B (black rectangle, Figure 6a). The temperatures of the lake water close to the device increased by 4° C in 2 h. It is unlikely that this would have occurred over the whole lake volume [Pasternack and Varekamp, 1997], as 71.7 GW would have been required over 2 h. CVGHM staff observed upwelling of bubbles close to the sensor location during period 3A. New fractures may have opened during the unrest which could explain the sudden heatings. They could also reflect an enhanced convection nearby the sensor location. Although these measurements obviously do not reflect the whole lake volume variations, a strong localized heating of lake waters over a longer time period is probable. Temperature measured by CVGHM staff on 17 March 2012 was similar (Figure 6a). A photograph taken on the same day shows that the whole lake surface was strongly evaporating (Figure 7b). Moreover, a sharp
Figure 6. (a) Lake temperature measured by different probes. (b) Volumes derived from CVGHM lake level measurements. (c) Velocity variations (in %) using different frequency bands for KWUI-POSI pair.

temperature drop would have been expected as soon as the batch of hot water reequilibrated with colder surrounding lake waters. However, this was not observed. The whole lake volume still appears to have been heated but probably over a longer time period.

To constrain the mass changes in the system, we related our level measurements to volume changes. After a calibration of the data that were acquired by CVGHM with our high-performance device (Troll 500 sensor, one measurement/30 min, recording between November 2010 and October 2011), we derived the fluctuations of the lake volumes from surface lake level measurements by using the volume determined from the bathymetrical map. We calculated the volume considering that the level fluctuations only affect the upper 10 m of the lake. Lake level started to increase before the first unrest (period 1A, Figure 6b). Within 2 weeks, the lake volume had increased by 460,000 m$^3$ and remained constant the week before the onset of the first unrest. The next measurement acquired at the transition between periods 3A and 2B indicates 1.25 million m$^3$ of volume increase within 3 months, which is extremely unusual. After 1 month without any important fluctuation, the volume finally increased by ~630,000 m$^3$ in 1 week (period 3B). By analyzing photographs taken on 29 September 2011 and 21 January 2012, we estimated the level increase to be 2.5 m, which is in agreement with the level measured by CVGHM staff. Considering the lake level measurements acquired since 2006 (in normal conditions, seasonal variation of 4 m is expected or a corresponding volume of 1.2 \times 10^6 m$^3$ per year), the estimated increase of volume between December 2011 and March 2012 is highly unusual. It suggests that this unrest was characterized by an input of about 2.3 million m$^3$ of hot water at the bottom of the lake with a more abrupt discharge before the onset of the second unrest period (period 2B, 0.63 million of m$^3$). These values could be underestimated, as some more water may have seeped through the rocks. Unfortunately, it could not be quantified with the data acquired during that period.

4. Discussion

Before interpreting the results, we first discuss the different nonvolcanic and volcanic phenomena that could explain the $dv/v$ observed. We then discuss the reliability of the VT event locations and investigate the
processes that may explain the peculiar MT seismicity in such an environment. We finally estimate the enthalpies and thermal input required to heat up the lake during each period of crisis and assess the reliability of these results.

4.1. External Influences on Velocity Variations
Several natural mechanisms and/or artifacts may produce the significant $dv/v$ observed and might perturb the retrieval of seismic velocity changes related to volcanic activity. We will first discuss them to ensure the interpretation is based on changes that may have been induced by volcanic processes.

Time clock problems can induce unwanted time delays [Sens-Schönfelder, 2008]. If this would have happened, an equal shift would be expected between positive and negative lag times [Sens-Schönfelder, 2008] which is not observed in any interferogram.

A water level increase causes a decrease in seismic velocity [Sens-Schönfelder and Wegler, 2006; Anggono et al., 2012]. Infiltration of rain water into the ground increases pore pressure which in turn decreases effective stress and causes seismic velocity decrease [Anggono et al., 2012]. Sens-Schönfelder and Wegler [2006], Wegler et al. [2006], and Meier et al. [2010] attributed seasonal variations of seismic velocity to water level changes due to precipitation. Richter et al. [2014] inferred an atmospheric origin from the velocity changes they observed using a model based on thermally induced stress, whereas Tsai [2011] attributed the wave speed variations to thermoelastic and hydrologic variations. Even if the velocity variations were observed during the rainy season at Kawah Ijen, they are not correlated to heavy rain periods recorded nearby (Figure S5). Hence, a different cause has to be found to explain these short-term variations.

Strong ground motion could also cause seismic velocity decrease in the volcanic edifice at shallow depths [Anggono et al., 2012; Battaglia et al., 2012; Brenguier et al., 2014; Lesage et al., 2014]. A strong earthquake hits Bali on 13 October 2011 ($M_w = 6.2$) (Figure 1a and Figure S5, vertical line), and no clear change is observed. The earthquake that occurred on 11 April 2012 ($M = 8.2$) in the Indian Ocean off Aceh (Sumatra, Indonesia) did not produce any drop, and no other significant earthquake in the vicinity of the volcano has been noted.

The few studies that retrieved velocity variations on volcanoes also evidenced spatial heterogeneities of those variations. When sufficient sensors are recording, a regional approach helps to locate the observed perturbations [Brenguier et al., 2008; Obermann et al., 2013; Sens-Schönfelder et al., 2014]; however, this is not the case at Kawah Ijen. Since the velocity variation is clearly localized, the velocity depends on the time lag, and its
relation with velocity changes is not simple [Lesage et al., 2014]. Hence, the measured velocity changes in this study are apparent relative velocity fluctuations. The problem of precisely estimating the lateral [Larose et al., 2010] and depth sensitivities [Obermann et al., 2013] is beyond the scope of this paper, but it certainly deserves attention.

4.2. Processes and Mechanisms of the Seismic Event Activity

As the results have shown, VTs and MTs present distinct characteristics. After briefly discussing the well-constrained VT seismicity, we investigate the processes and mechanisms responsible for MT events. We describe the most likely candidate, i.e., resonance in fluid-filled cracks, which is widely recorded in such environments and is supported by several observations in this study.

VTs are ordinary double-couple earthquakes occurring in the brittle rock within a volcanic edifice or the crust. They originate from shear failure caused by stress buildup and resulting slip on a fault plane [Chouet and Matoza, 2013]. Stresses may be induced by dike propagation as indicated by propagating hypocenters and fault plane solutions reflecting regional to local stresses. Dike inflation, as indicated by randomly distributed hypocenters and fault plane solutions with pressure axes rotated $\sim 90^\circ$ to regional maximum compression [Roman et al., 2006], may also stress the medium. Alternatively, intrusion of magma or fluids into the volcanic system, increases stresses at and around the volcano causing swarms of VT earthquakes on local faults [White and McCausland, 2015].

We use $V_p/V_s = 1.73$ to locate the VTs. This provides stable results for the event locations. A joint inversion of the event locations and the velocity model should be done in the future, using VELEST [Kissling et al., 1995] or similar approaches. Wadati diagrams for the 376 S-P measurements do not fit a straight line which could be an indication for different $V_p/V_s$ values with depth and/or location.

While the origin of VT seismicity is well constrained and reflects a stress buildup in the system, possibly driven by a magmatic intrusion, MTs represent a subset of low frequency (LF) seismicity [Chouet and Matoza, 2013]. The processes responsible for MT events may be related to the resonance of fluid channels encased in rocks, such as fluid-filled hydrofractures, dykes, and cylindrical conduits [Rust et al., 2008]. The presence of bubbles in magma and hydrothermal fluids lowers the speed of sound in these fluids, inducing a sharp contrast in velocity between the fluid and encasing solid [Chouet and Matoza, 2013]. This favors the entrapment of acoustic energy in the fluid volume source region [Chouet and Matoza, 2013]. Other processes and mechanisms are able to trigger harmonic/monochromatic tremor, but the resonance of a fluid-filled crack or cavity is the most likely candidate [e.g., Chouet, 1996; Neuberg, 2000; Chouet and Matoza, 2013], especially in the presence of such a large hydrothermal system.

At Kusatsu-Shirane volcano (Japan), Nakano et al. [2003] proposed a conceptual model in which magmatic heating causes a gradual buildup of steam pressure in a hydrothermal crack, which subsequently causes repeating discharges of steam. The rapid discharge of fluid may cause the collapse of the fluid-filled crack, which afterward excites resonance and leads to the LF coda. This model involves boiling of groundwater by rising magmatic gases, producing steam and raising the overall pressure of hydrothermal fluids in a distributed network of fractures. Impulsive discharges of steam (or more generally fluids) and collapse of these hydrothermal fractures give rise to impulsive excitation and resonance. Similar observations and source mechanisms have linked shallow LF events to magmatic-hydrothermal interactions at a variety of volcanoes (see Chouet and Matoza [2013] for an exhaustive list of references). Matoza and Chouet [2010] discussed the trigger mechanism of LF seismicity and the thermodynamic changes that may occur in a pressurized hydrothermal crack heated by a magmatic source. They proposed that more energetic phase changes could result in LF events. We note that if MT events originated in the shallower part of the hydrothermal system, phase changes should be rather restricted. However, a large amount of CO$_2$ was measured inside the lake. As more CO$_2$ is added to the system, the potential of the system to generate seismic energy increase and smaller pressure drops can lead to more violent phase changes [Matoza and Chouet, 2010].

Visual observations at Kawah Ijen, such as the sustained upwelling of bubbles, could provide additional constraints. Between 8 and 10 March 2012, the closed-circuit television (CCTV) installed by CVGHM captured a continuous upwelling of bubbles at the lake surface. At the same time, the seismometers recorded the most vigorous activity of MTs (Figure 2c). Together with the heightened heat flux, as suggested by the sharp increase in lake temperature (Figure 6a), bubble upwelling points to a higher contribution of liquid, steam, and/or gas into the lake system which in turn may trigger the resonance inside a crack or a network of cracks beneath
the volcano. The high-frequency onset of some MTs may be related to the opening of the valve sealing the crack (brittle failure). The few Q factors that could be derived (not shown here) range between 1 and 20 and may be attributed to bubbly magma, bubbly water, or steam [Kumagai and Chouet, 2000; Kumagai et al., 2002]. Although we could not locate MTs at Kawah Ijen due to the small number of stations distributed in the near field of the active volcanic crater, a contribution from bubbly magma is not likely, because there is no evidence for fresh magma entering the shallow layer of the volcano. A hydrothermal fluid such as a bubbly water or steam may thus account for the observed Q factors. Envelopes of MT events display irregular modulation of the amplitudes and relatively long duration. This might be attributed to an imperfect sealing of the crack which would explain how acoustic energy may be trapped for variable durations. The differences in waveforms could be explained by a less efficient valve mechanism in which an increased fluid and/or heat flux prevents the regular valve closure.

4.3. Lake Heating Episodes and Volume Changes

The December 2011 and March 2012 crises were characterized by an enhanced heat discharge based on the temperature patterns, which are different in each case. We estimate the enthalpies and thermal input required at the lake bottom to heat the lake during each period of crisis. We finally compare them to similar environments.

After the onset of the seasonal recharge (between November and the beginning of December 2011, period 1A, ~0.8°C/week, Figure 6a), the temperature started to rise abnormally in mid-December 2011 (period 2A, ~1.3°C/week, Figure 6a). This crisis implied a gradual heating of the lake, whereas for the second unrest (2B and 3B), stronger variations were observed (Figure 6a). By comparison, at the Ruapehu volcano (New Zealand), water temperature can rise up to 1°C/d and reach 60°C without leading to any significant eruption [Hurst and Dibble, 1981; Vandemeulebroeck et al., 2005]. While the increase in temperature between December 2011 and January 2012 is reasonable, this is not the case in March 2012. Our calculations indicate that 9°C heating within a few days is unlikely to affect the whole water mass. We note that the temperature pattern starts to be unstable during period 2B (Figure 6a) indicating that the lake was not well mixed and that heatings and coolings may only reflect local conditions. Although lake temperature could not have increased that much within 4 days, CVGHM measurements, visual observations, and long-term decaying pattern during period 3B suggest that the whole lake volume temperature could have increased from 40°C to 49°C over a more reasonable number of days (at least 10 to 15 days).

To heat up the whole lake volume by 9°C, an amount of energy around $1.2 \times 10^{15}$ J is required. The change of volume observed during period 2B is around $\sim 620,000 \text{ m}^3$ (Figure 6b) or, considering a density of 1083 kg m$^{-3}$, a consequent mass input between 250 and 1000 kg/s for this period ($\Delta V/\Delta t$ where $\Delta V$ is the volume change and $\Delta t$ the time period (7–28 days)). Assuming 23–28 days a reasonable time to heat the whole lake and increase its volume (Figure 6a, black rectangle), we evaluate an enthalpy for the fluid at the lake bottom close to 2000–2300 kJ/kg, i.e., a power input of 434–587 MW. This simple calculation assumes constant evaporation and seepage rates. Seepage is difficult to monitor, and the use of a constant rate does not have much impact on the energy estimate (a few percent of the final result based on Terada and Sudo [2012]). Evaporation fluctuates depending on local wind conditions. Generally, strong winds favor evaporation which, in turn, cools down the lake water temperatures. In that case, our enthalpies would be underestimated. Unfortunately, the meteorological station installed at the summit did not record at that time. However, the wind speed recordings of the previous year only revealed a weak influence of this parameter on lake water temperature at a timescale of a few hours. The main parameter controlling the enthalpy over a time range of a few days is most likely the heat flux provided by the shallow hydrothermal system. Nonetheless, it is still important to point out that the permeability increase enhances the lake water recycling into the hot rocks underneath the lake which, in consequence, could lead to an overestimation of the enthalpy.

We applied the same calculations for the more gradual temperature increase observed between periods 2A and 3A (Figure 6a), yielding to a lower enthalpy of 1100 kJ/kg. At the bottom of the lake (pressure of at least ~ 18 bars), the boiling temperature is estimated to be around 210°C. As explained by Delmelle [1995], it seems reasonable to assume a temperature close to 200°C for the fluid discharged at the bottom owing to the existence of liquid sulfur pools (temperature reported elsewhere around 177°C at Ruapehu (New Zealand) [Christenson, 1994] and 116°C at Kusatsu-Shirane (Japan) [Takano et al., 1994]). At those conditions, the enthalpy of pure liquid is around 800 kJ/kg and of pure steam 2800 kJ/kg.
Figure 8. Conceptual model of Kawah Ijen volcano seismicity that may have triggered the unrest (21 May).

At Aso volcano (Japan), Terada and Sudo [2012] evidenced an enthalpy fluctuating between 1840 and 3030 kJ/kg, using a numerical approach, whereas Hurst et al. [2012] estimated that an apparent enthalpy much higher than 3000 kJ/kg was required to avoid most of the negative apparent meltflow at Ruapehu (New Zealand). The calculated thermal flux is also reasonable. The calculated thermal fluxes of 400–600 MW (during periods of unrest) and 200 MW (during quiet periods, for dry season, following Barbier [2010]) are consistent with studies at other volcanoes. For example, the power input by Hurst and Dibble [1981] at Ruapehu ranges between 100 and 700 MW. Therefore, the enthalpy values obtained for periods 2A–3A (1100 kJ/kg) and period 2B (2300 kJ/kg) suggest an increase in gas/steam contribution to the lake heating between the two unrest periods.

5. Nature and Origin of the Crises

5.1. Distal Seismicity: Trigger of the Unrest (May 2011, Period 1A)

The earthquake swarm that occurred on 21 May 2011 (period 1A) may have constituted the earliest stage in the generation of the successive crises. This first important stress change in the area was likely induced by a deep magmatic intrusion that triggered earthquakes along distal faults. A swarm including several strong volcano-tectonic earthquakes (maximum magnitude of 3.3) hit the caldera and its vicinity. This type of seismic pattern has been observed at more than 60 volcanoes worldwide [e.g., White and McCausland, 2015; Pozgay et al., 2005; Surono et al., 2012] and might have driven the later unrest. We do not have independent geodetic parameter that could confirm this hypothesis, just an increased amount of partial pressure of gaseous CO₂ measured in the lake (3 times more) in July–August 2011 compared to May 2010, that could indicate an increase in magmatic degassing. This episode is illustrated in Figure 8.

5.2. Volcanic Unrest

5.2.1. Magmatic Intrusion (Period 1A to 2A)

Between October and December 2011 (period 1A), strong shallow VTs (~2 km asl) were recorded and may indicate the first stage of a renewed activity at a shallow level (Figure 9). Shallow proximal VTs reflect the brittle response of volcanic rock to change in strain rate associated with a magmatic intrusion, a dyke intrusion
or propagation [Roman et al., 2006], or the movement driven by fluids rising from the magma [Chouet and Matoza, 2013]. A magmatic/dyke intrusion might have taken place below sea level and triggered the shallow VTs.

5.2.2. Open System (Periods 3A, 2B, and 3B)

The strong VT earthquakes that occurred during periods 1A and 2A probably opened pathways for the fluids to ascend by increasing the permeability of the system (Figure 9). Concurrently, hot gases or fluids, released from the supposed intruding body, impacted the shallow hydrothermal system, by heating and pressurizing the system. This triggered significant MT events, increases in lake temperature and level, and significant velocity variations.

We measured velocity drops coincident with MT activity (Figure 2) and lake heatings (Figure 6). Our \( \frac{dv}{v} \) observations during periods 3A-2B could be consistent with observations of \( \frac{dv}{v} \) drops in other hydrothermal settings. For instance, Cros [2011] measured velocity drops due to boiling water and steam flows around Old Faithful geyser (Yellowstone, USA). In steam flood sites, dramatic seismic velocity decreases, up to 40%, were measured in the steam area compared to the liquid area [Lumley, 2001]. The injection of steam in the shallow system, indeed, generates not only fluid phase changes affecting the seismic velocities but also dilatancy by thermal expansion which can be the source of velocity changes. Although the dilatancy model is plausible, the saturation of the medium could also decrease the seismic surface wave velocities [Cros, 2011; Letort et al., 2012]. The majority of those velocity drops took place during the period of increased MT activity rather than VTs, lake heatings, and volume increases, which may indicate a hydrothermal origin and a system that is more “opened” to heat and mass transfers, where fluid fluxes easily reached the surface (Figure 10).

Drastic changes in lake water temperature have been reported at several volcanoes. However, very few were linked to fluctuations in seismicity. At Ruapehu (New Zealand), Sherburn et al. [1999] linked the 3 Hz tremor to periods when the temperature of Crater Lake water was above 30°C. At Aso (Japan), Terada and Sudo [2012] found that the heating of the hydrothermal system drives the volcanic tremor, whereas at Poas (Costa Rica), Martínez et al. [2000] related an increase in the flux of heat and gases discharged through the bottom of the crater to increases in summit seismic activity. However, those studies could not achieve a better temporal resolution than 1 month for lake temperature data.
Figure 10. Conceptual model of Kawah Ijen volcano seismicity for the unrest ranging between periods 3A and 3B.

If we refer to the terminology of Kumagai and Chouet [2000], enthalpies calculated for period 2A point to a liquid-gas mixture (foam) with 25% of gas volume fraction and between 60 and 75% for period 2B. The observations of bubbles at the lake surface and our enthalpy calculations indicate an increased steam/gas input to the lake bottom during the crises. The seismic velocity drops may also suggest the presence of a large amount of steam in the shallow system. Thermal expansion of water alone could only explain 10% of the volume variations observed during period 2B. As suggested by Dibble [1974] (in Vandemeulebroeck et al. [2005]), steam pushes out liquid water during the heating phase and is replaced by liquid during cooling. The interface between liquid water and two-phase layer is thus vertically displaced, as in the model of Vandemeulebroeck et al. [2005].

The period when MTs were detected is coincident with the heating phase and with our calculations displaying a contribution of steam during these periods. Indeed, a strong impedance contrast exists between liquid and steam zones [Vandemeulebroeck et al., 2005] and could explain the appearance of MT seismicity specifically within these time spans. The fact that higher amplitude and higher frequencies of MTs were observed at the end of the second crisis (beginning of period 3B) would support the hypothesis of an enhanced steam contribution (Figure 2c). We know that more gas/steam was observed during period 3B, based on the images captured by CVGHM CCTV and enthalpy calculations. These were exactly correlated to the increase in dominant frequencies of MT (Figures 5a and 5b, period 3B) to the last velocity drop (Figure 2b, period 3B) and to an increase in temperature (Figure 6a, period 3B). If we recall the crack model, a straightforward interpretation would be that the crack was progressively heated up and/or was fed with a higher gas volume fraction. Nonetheless, for a crack of constant geometry and size filled with bubbly liquids (gas volume fractions between 2 and 10%), an increase in the dominant frequency can be explained by a decrease of the gas volume fraction in the hydrothermal mixture; increasing the volatile content is known to decrease the acoustic velocity of a fluid and to increase its impedance contrast with the surrounding rock [De Angelis and McNutt, 2005]. Calculations by Kumagai and Chouet [2000] showed that dimensionless frequency is almost constant over the entire ranges of gas volume fraction and pressure. However, even though the dimensionless frequency in water foams is nearly constant up to a gas volume fraction of 60%, it increases rapidly if gas volume fraction exceeds 60%. According to Kumagai et al. [2002], it may indicate a drying process inside the crack due to a gradual heating of the hydrothermal system in response to a magmatic heat pulse beneath the volcano. This is in agreement with our observations. The high frequency at the end of the unrest could indicate a decrease of the crack size that finally collapse due to an overall pressure decrease in the system [Kumagai et al., 2002].
An alternative explanation arises from the recent crack/cavity’s model proposed by Jolly et al. [2012] who proposed that changes in spectral properties for a stationary earthquake source process are not only a result of temporal changes in resonance but also the near-source scattering. Reasonable changes in bubble content may produce dramatic velocity changes (2 to 300 m/s) [Kieffer, 1977], and attendant impedance contrasts, which is a viable LF resonance mechanism, but may also promote or inhibit scattering, depending on the strength of the velocity contrast. The frequency of excitation for any given cavity would be fixed by its geometry and the velocity of the fluid, but the level of excitation would depend on the bubble content. The enrichment of higher frequencies then reflects an increase in bubble volume fraction, which reduces the acoustic velocity and enhances resonance for each cavity in the hydrothermal system [Jolly et al., 2012]. This is in agreement with the observation of higher frequencies at the beginning of period 3B together with lower frequencies at other times.

In summary, the presence of a sufficient amount of steam/fluids appears as the essential condition to generate MTs in the Kawah Ijen hydrothermal system. MT tremor is thus a key feature of the volcanic activity at Kawah Ijen and must be closely monitored in the future.

MTs activity vanished when the velocity fluctuations became small and the lake temperature and volume decreased, which indicates a general drop of pressure in the shallow system. After the March 2012 crisis, the CCCs indeed recovered to the precrises values. It is known that a small dislocation of the scatterers in random directions causes a decrease in the CCC comparing seismograms before and after the perturbation [Snieder et al., 2002]. This was, for example, also observed before and after the mid-Niigata earthquake (Mw = 6.6) (Japan) by Wegler and Sens-Schönfelder [2007]. We reprocessed the data using a reference calculated only after the crises (May–September 2012) and obtained a similar result. Our results show that the whole process driving the crises is clearly reversible and ephemeral and did not alter the shallow hydrothermal system at a large scale, as the CCCs recovered to their background values after the crises.

6. Conclusions

We studied the strongest sequence of unrest ever recorded by seismometers at the wet Kawah Ijen volcano [Caudron et al., 2015a]. Temperature and pressure sensors immersed in the lake waters recorded this episode with an exceptional time resolution for a volcanic lake. We used the seismic ambient noise cross-correlation technique to retrieve the seismic velocity variations in the shallow portions of the volcanic edifice. Several processing parameters have been investigated to compute reliable seismic velocity variations. Comparing a stack of 20 days of the cross-correlation function to a reference has given reliable results.

In May 2011, a strong swarm of distal VTs was recorded. It is most likely the results of a deep magmatic intrusion, as almost no changes were observed at shallow depths beneath the volcano until 6 months later. The strong unrest commenced in October 2011 (period 1A) with increased shallow VTs and LF events which culminated in mid-December 2011 (period 2A) and was manifested at the surface by strong gas emissions and lake color changes. This was concomitant with enhanced heat and fluid discharges to the crater. We interpret our results as a buildup of stress inside the volcano, through a magmatic intrusion, which was followed by the entrance of hydrothermal fluids into the shallow hydrothermal system. Increased pore pressure gave rise to increasing fracture. By increasing the permeability of the system, VT earthquakes opened pathways for the fluids to ascend. This allowed the later initiation (period 3A) of MT events when steam/gases interacted with the shallow portions of the aquifer. Our calculations point to a higher contribution of steam (gas volume fraction above 60%) for the second unrest (periods 2B and 3B) leading to an increase of the MT frequency when bubbles were observed at the surface. This period was also characterized by significant seismic velocity variations and important heat and mass discharges to the lake. The stress buildup likely resulted from water level rises containing a large amount of bubbles (gas/steam) and/or rapid seal development.

This study identified seismic precursors of this sequence (an increased number of shallow VT earthquakes), specific seismic manifestations during the unrest (MTs and their dominant frequency increase and seismic velocity drops), and heat and mass flux increases associated with the crises. However, the major challenge still remains to understand why the unrest did not lead to an eruption and to identify precursory signs of a phreatic eruption. A small phreatic eruption would be devastating for the people working daily in the crater and the ones living near the voluminous acidic lake. Along with the seismic network and the development of other processing techniques, new sensors could greatly improve the monitoring (e.g., GPS antennas, gravimeters, hydrophone, CO2 probes, and self-potential [Caudron et al., 2015b]) and thus the identification of the potential precursory signs of phreatic eruptions.
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