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Detection of Pre-eruptive Seismic Velocity Variations at an Andesitic Volcano Using Ambient Noise Correlation on 3-Component Stations: Ubinas Volcano, Peru, 2014

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Abstract

Volcano monitoring and eruption forecasting are based on the observation and joined interpretation of several precursory phenomena. It is thus important to detect new types of precursor and to study their relationship with forthcoming eruptions. In the last years, variations of seismic velocity have been observed in some volcanoes, mainly basaltic, before eruptions. In this paper, we look for velocity variations and waveform decorrelations before the 2014 eruptive sequence of the andesitic Ubinas volcano in Peru. We compute velocity changes by using seismic ambient noise cross-correlation (between pairs of stations) and cross-components correlation (between vertical and horizontal components of single stations), as well as coda wave interferometry of seismic multiplets. With these different approaches, we show that the major explosions that occurred from 13 to 19 April were preceded by a clear velocity decrease and waveform decorrelation. The amplitude of velocity change is generally larger on single-station cross-components correlation than on two-stations cross-correlation in all the frequency ranges tested (between 0.1 and 8 Hz). We highlight an apparent anisotropy of velocity change in single-station cross-components correlation, with larger amplitudes when correlating vertical and tangential components

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than using vertical and radial components with respect to the crater. The $M_w = 8.1$ Iquique earthquake on 1 April 2014 produced also a marked co-seismic velocity drop detected in a high frequency range (3-5 Hz) in both single-station cross-components correlation and cross-correlations. We locate in the horizontal plane and in depth the velocity perturbation and the structural change related with decorrelation. During the main phase of eruptive activity, the velocity decrease at low frequency (0.1-1 Hz) appears to affect the whole edifice mainly at depth of about 1 to 3 km below the surface. The structural perturbation is more concentrated on the south flank of the volcano, a zone that corresponds to an ancient collapse. We suggest that the observed velocity variations are due to the dilatation of the edifice and to microfracturation induced by magma pressurization. The structural change may be locally enhanced by a possible zone of material weakness in the southern sector. The co-seismic velocity perturbation is located mostly in the southeast flank, at depth smaller than 0.5 to 1 km, and may be related to the presence of the hydrothermal system of the volcano.

Keywords: Seismic velocity variation, coda wave interferometry, eruption precursor, volcano monitoring, Ubinas volcano, volcano seismology

1. Introduction

The early detection of volcanic unrest before a forthcoming eruption is one of the primary goals of volcano observatories. For that purpose, it is necessary to identify and analyze a maximum number of precursory phenomena that are the basis of eruption forecasting. The most widely used precursors are the increase of the seismic activity, the ground deformation (inflation) and the variation of gas flow and composition (Scarpa and Gasparini, 1996). However, these phenomena are not always observed before eruptions (Kato et al., 2015; Romero et al., 2016) and it is important to investigate other types of precursors.

The variations of seismic velocity in volcanic structures appear to be a promising precursory phenomena that have been recently observed on some volcanoes. Seismic velocity changes caused by large earthquakes or volcanic ac-

13 tivity have been detected first by using repeating earthquakes (Poupinet et al.,
14 1984; Ratdomopurbo and Poupinet, 1995) or active seismic sources (Nishimura
15 et al., 2005). However, this approach is not easily applicable for continuous
16 monitoring because it depends on the occurrence of repeating earthquakes with
17 similar sources. Several authors have demonstrated that the Earth’s impulse
18 response (Green’s function) between two passive receivers can be retrieved by
19 cross-correlating ambient seismic noise (Weaver and Lobkis, 2001; Shapiro and
20 Campillo, 2004); Ambient Noise is a permanent vibration of the ground surface
21 due to human activity or natural processes such as the interaction of the oceanic
22 swell with the solid Earth. These virtual records provide the opportunity for
23 imaging and monitoring a medium without using active seismic sources (Shapiro
24 et al., 2005; Wegler et al., 2009; Larose et al., 2015). Furthermore, Larose et al.,
25 (2010), Obermann et al., (2013) and Planès et al., (2014) introduced a method
26 to locate, in the horizontal plane, spatially heterogeneous velocity or structure
27 perturbations.

28 Brenguier et al., (2008) calculated velocity variations by comparing ambient
29 noise cross-correlation functions obtained at different periods and evidenced ve-
30 locity decreases prior to eruptions at Piton de la Fournaise volcano, La Réunion
31 Island. More recently De Plaen et al., (2016) also estimated velocity reductions
32 before eruptions of Piton de la Fournaise by using auto-correlation and correla-
33 tion between components of a single station. Haney et al. (2015); Bennington
34 et al. (2015; 2018) also show the application of ambient noise cross-components
35 correlation of single stations for study seasonal and magmatic velocity variations
36 at Okmok and Veniaminof volcanoes.

37 Temporal changes in the subsurface velocity structure during volcanic ac-
38 tivity are likely controlled by several factors, such as stress, deformation, and
39 migration of magmatic or hydrothermal fluids (e.g. Sens-Schönfelder et al.,
40 2014; Donaldson et al., 2017). Up to now, velocity variations preceding volcanic
41 eruptions have been detected only at a limited number of volcanoes such as
42 Piton de la Fournaise (Brenguier et al., 2008; Duputel et al., 2009; Clarke et
43 al., 2013; Obermann et al., 2013; Rivet et al., 2014; 2015; Sens-Schönfelder et

44 al., 2014; De Plaen et al., 2016), Ruapehu (Mordret et al., 2010), Miyakejima
45 (Anggono et al., 2012), Etna (Cannata, 2012; De Plaen et al., 2019), Okmok
46 (Bennington et al., 2015), Mt St Helens (Hotovec-Ellis et al., 2015), Merapi
47 (Budi-Santoso and Lesage, 2016), or Kilauea (Donaldson et al., 2017), Hakone
48 (Yukutake et al., 2016), Veniaminof (Bennington et al., 2018) or White Island
49 (Yates et al., 2019). However, at some open systems such as Volcán de Colima,
50 the velocity variations were almost undetectable (Lesage et al., 2014), even be-
51 fore large eruptive events (Lesage et al., 2018). It is thus important to keep
52 evaluating and documenting this approach on an increasing number of cases.

53 In this study, we investigate the velocity changes associated with the 2014
54 eruptive cycle of Ubinas volcano, Peru, and with the $M_w = 8.1$ Iquique earth-
55 quake. We use ambient Noise Correlation Function (NCF) calculated between
56 station pairs, Noise Cross-Correlation Function (NCCF) and Noise Single-station
57 Cross-components correlation Functions (NSCF) calculated between different
58 components of the same stations. Additionally, we analyze similar events (mul-
59 tiplets) to detect pre-eruptive velocity changes in the structure of Ubinas by
60 using also Coda Wave Interferometry (Snieder, 2006). We explore the velocity
61 variations in several frequency bands in order to discriminate possible veloc-
62 ity perturbations at different depths and to identify the best spectral intervals
63 in terms of forecasting. We localize in the horizontal plane the velocity varia-
64 tions and the structure changes associated with the main eruption, using the
65 sensitivity of multiply scattered waves to weak changes in the medium (Ober-
66 mann et al., 2013). Then we interpret the estimated velocity changes in relation
67 with other observations, such as seismicity, plume elevation, satellite thermal
68 anomalies and SO_2 flux, associated with the volcanic activity.

69 **2. Geological setting**

70 *2.1. Tectonic context*

71 Southern Peru is an active tectonic zone where the Nazca plate is subducted
72 beneath the South American plate, with a convergence rate of about 62 mm/year

73 (Villegas et al., 2016). This subduction is accompanied by a high level of seis-
74 micity (Chlieh et al., 2011) and volcanism along an active continental margin
75 (Figure 1a). Active stratovolcanoes are located on the Central Volcanic Zone
76 (CVZ), a segment associated to a steeply dipping ($25 - 30^\circ$) slab extending
77 from north of Chile to southern Peru. In this part of Peru, several volcanoes are
78 considered as potentially dangerous among others El Misti, Ubinas, Sabancaya,
79 Ticsani, Tutupaca, Coropuna, Yucamane, Huaynaputina. The region was also
80 the site of the largest explosive eruption in historical times within the Andes
81 (Huaynaputina volcano; AD 1600; VEI 6; Thouret et al., 1999). Several large
82 earthquakes ($M_w > 7.0$) occurred along the subduction zone in the past decades.
83 For instance, on 1 April 2014, a M_w 8.1 interplate thrust earthquake occurred
84 off-shore of Iquique city, Northern Chile (epicenter 19.572° S, 70.908° W, ~ 360
85 km south of Ubinas volcano, figure 1a). The global Centroid Moment Tensor
86 (gCMT) solution indicates an almost pure double-couple faulting geometry with
87 strike 357° , dip 18° , and rake 109° , at a centroid depth of 21.9 km and centroid
88 location south of the hypocenter, (Lay et al., 2014, Duputel et al., 2015). In
89 Arequipa and Ubinas Valley the movement was felt for more than a minute with
90 IV Modified Mercalli Intensity.

91 *2.2. Ubinas volcano*

92 Ubinas volcano (Figure 1) is considered as the most active volcano of Peru
93 (Thouret et al., 2005; Rivera et al., 2014) and is located at about ~ 65 km to the
94 East of Arequipa city. Two main periods were identified in its eruptive history.
95 The first one, “Ubinas I”, from middle Pleistocene to ~ 370 ka ago, is charac-
96 terized by andesite and dacite lava flows that built the lower part of the edifice.
97 The south flank of this edifice collapsed, resulting in a debris-avalanche deposit
98 on the SE of the summit. The second one, “Ubinas II”, (~ 370 ka to present)
99 comprises several stages. The summit cone was built by a series of andesite and
100 dacite lava flows and pyroclastic deposits. During the last stage, between 20
101 and 1 ka, the eruptive behavior has been dominantly explosive, and the summit
102 caldera was formed in association with a large-scale Plinian eruption, between

103 20 and 14 ka (Figure 3a). The last Plinian eruption occurred in 980 ± 60 year
104 BP. Since then, Ubinas displays a persistent phreatic and explosive activity. In
105 the past 550 years, 26 volcanic unrests have been reported (VEI 1 – 3), the
106 last two eruptions occurred in 2006-2009 and 2014-2017. An average of 6 to 7
107 eruptions per century was estimated (Rivera et al., 2014).

108

109 *2.3. Chronology of the 2013 – 2016 eruption*

110 The previous eruptive period of Ubinas Volcano was on 2006 – 2009. It
111 included episodes of magma extrusion and vulcanian explosions associated to
112 strong degassing and a large number of elevated plumes (Macedo et al., 2009;
113 Rivera et al., 2010, 2014; Traversa et al., 2011; Inza et al., 2014; Machacca, 2012;
114 Figure 2). In 2013, after four years of quiescence, a new eruptive activity started
115 with nine phreatic explosions and tremor activity between September 2 and 7. In
116 the following months, the activity came back to the normal background level. On
117 January 2014 a new increase in seismic activity started, followed on 1 February
118 by an increase of tremor activity that marked the beginning of magma extrusion
119 and degassing. On 20 February, 2014, the first harmonic tremor was recorded,
120 indicating lava extrusion in the crater (INGEMMET internal reports; Machacca
121 et al., SGP, 2014). The lava extrusion was confirmed during a field survey on 1
122 March by the staff of Instituto Geofísico del Perú who reported an incandescent
123 lava body at the bottom of the crater (Figure 3c; delimited by the red dashed
124 line). Thermal anomalies were also detected by the MIROVA hot-spot detection
125 system (Coppola et al., 2015). This activity increased significantly until the first
126 major explosion on 13 April.

127 Several large explosive events occurred between 13 and 19 April, 2014, and
128 ejected blocks of fresh basaltic andesite magma, $40 \times 40 \times 50$ cm in size, up
129 to 2.6 km from the active crater. The volcanic plume exceeded 5 km above
130 the crater level and a block of $5 \times 4 \times 2$ m has been found inside the summit
131 caldera, at 660 m from the active vent (INGEMMET internal reports). After the
132 major explosion on 19 April, the seismic energy and explosive activity gradually

133 declined, with weak plume emissions and sporadic series of explosions (Figure
134 2).

135 **3. Data and methods**

136 *3.1. Seismic network and processing*

137 During the study period, January to May, 2014, the Ubinas Network of IN-
138 GEMMET Volcano Observatory (OVI) was composed by seven seismic stations
139 (with codes UBN01 to UBN07) distributed around the volcano (Figure 1b). The
140 distances between stations ranged from 2 to 12 km. The stations were equipped
141 with three-components Guralp 6TD broadband seismometers with frequency
142 range 30s - 100 Hz and recorded with sampling frequency of 100 Hz and A/D
143 resolution of 24 bit. Station UBN06 began to record on July 29, 2013, UBN02
144 on 14 February 2014 for and UBN01, UBN03, UBN04, UBN05 and UBN07 on
145 20 to 25 March. During the eruption, UBN02 and UBN06 transmitted the data
146 via telemetry radio to the monitoring center of OVI in Arequipa-Peru.

147 The seismic activity at Ubinas volcano during the 2014 crisis presented sev-
148 eral types of earthquake including Volcano-Tectonic (VT) events, Long-Period
149 events (LP), Tremor, Hybrid events (HYB), Very-Long Period events (VLP)
150 and other types (Machacca et al., 2014; Figure 13a), classification was made
151 considering their waveforms, spectrum and spectrogram.

152 Real-time seismic amplitude measurement (RSAM; Endo and Murray, 1991)
153 was calculated after removing manually regional earthquakes from the records,
154 by filtering in the [0.1 – 1] and [1 – 20] Hz frequency band and using a 10 min
155 moving window.

156 Seismic energy has been calculated using the equation formulated by Johnson
157 and Aster (2005).

$$E(iT) = \frac{2\pi r^2 \rho c S^2}{A} \int_0^T y^2(t) dt \quad (1)$$

158 where r is the distance from source, ρ is the density, c is the P wave velocity,
159 A is the attenuation correction, S the seismic site response correction and $y(t)$

160 is the particle velocity. We assumed that the source was at shallow depth below
161 the crater, $\rho = 2600 \text{ kgm}^{-3}$, $c = 3000 \text{ ms}^{-1}$, A and S were fixed at 1.

162 *3.2. Other observations*

163 Several complementary methods were used to observe the volcanic activity at
164 Ubinas. Deformations of the edifice were monitored by using electronic distance
165 measurement (e.d.m.) on several lines as shown on figure 1b. Two video cameras
166 (Simple webcam and AXIS Q1765-LE), located at 6.5 and 25 km from the active
167 crater, which recorded one picture every 30 s. Plume elevation from mid-2013
168 to 30 June 2014 was estimated using the nearest camera which has a maximum
169 visibility on the crater. The thermal anomalies of Ubinas volcano were detected
170 by MIROVA (Middle InfraRed Observations of Volcanic Activity; Coppola et
171 al. 2015). The SO_2 flux was measured with a mobile scanning DOAS during
172 field surveys. Temperature in hot spring “Ubinas Termal” was measured every 5
173 minutes by a data-logger and a sensor with a precision of 0.01 °C and calibrated
174 before installation. This spring is located at 2.2 km from UBN06 seismic station
175 and at 5.8 km from the active crater (Figure 1b).

176 *3.3. Calculation of seismic ambient noise correlation functions*

177 We first prepared waveforms from each component and station in one hour
178 long segments, after removing the mean and trend. Instrumental corrections
179 were not necessary because the same type of sensor is used at all the stations,
180 and the instrument responses are stable over time. As a second step, we down
181 sampled the signal to 50Hz, then we applied spectral whitening and we filtered
182 the records in several frequency ranges (0.1 – 1 Hz, 0.3 – 1 Hz, 1 – 3 Hz, 3 – 5 Hz,
183 5 – 8 Hz) in order to study the velocity perturbations at different depths. Then,
184 in order to suppress high amplitude events, we performed amplitude normal-
185 ization in the time domain. We tested three methods — one-bit normalization
186 (Larose et al., 2004), division by the envelope (Budi-Santoso and Lesage, 2016),
187 and amplitude clipping (Bensen et al., 2007) — and we decided to use the clip-
188 ping method because it produced slightly smoother curves of velocity variation.

189 This method consists in clipping the amplitudes larger than 3 times the RMS
190 value of the record, thus reducing the weight of large events while keeping all
191 the information contained in the low-amplitude continuous signal.

192 After that we computed the noise correlation function (NCF) for each hour
193 and averaged them over 24 hours (Wapenaar et al., 2010). For simplicity, the
194 acronym ‘NCF’ will refer to both NCCF and NSCF. Daily NCFs calculated with
195 less than 6 hours of data are discarded because of their low signal-to-noise ratio
196 (SNR). We calculated NCCF for 21 pairs of vertical components and NSCF
197 between vertical and horizontal components of all stations. The NCCF are sen-
198 sitive to perturbations in the medium around the path between two stations and
199 NSCF are sensitive to the medium around one station. Figure 4 displays exam-
200 ples of daily NSCFs and NCCFs calculated over several months (correlograms).
201 In our case, the correlation functions obtained between pairs of stations and be-
202 tween components of single stations are approximately symmetrical and stable
203 over time. Thus we merged the causal and acausal sides of the NCFs and we
204 kept a section of their coda, starting after the Rayleigh waves and ending before
205 the SNR is too low (for example for delays of 10 to 60 s; black rectangles with
206 spaced triangles in figure 4). Finally, before compute velocity changes (dv/v),
207 we applied the Butterworth bandpass filters with the same previous frequency
208 bands and the Wiener filter once to the correlograms (2-D array) for increasing
209 the SNR and smoothing the NCFs (Hadziioannou et al., 2011, Moreau et al.,
210 2017).

211 *3.4. Estimation of velocity variations*

212 Any change in velocity or structure of the propagation medium generates
213 modifications of the Green’s functions such as variations of the travel times
214 of direct and coda waves and changes in the waveforms. The corresponding
215 ambient seismic noise correlations functions (NCFs) can thus be used to detect
216 perturbations in the medium by comparing the current NCF to a reference
217 NCF. In particular, two methods can be used to estimate the variations of
218 waves travel times: the Stretching Method (Lobkis and Weaver, 2003) and

219 the Moving Window Cross-Spectral (MWCS) method also known as doublet
 220 method (Poupinet et al., 1984; Clarke et al., 2011). We tested both methods,
 221 using as reference the average NCF over the whole study period and we obtained
 222 similar patterns of relative velocity changes dv/v (Figure 5). However in our
 223 case, the stretching method appears to produce more stable curves of velocity
 224 variation, i.e. with less fluctuations, and thus we decided to use this method in
 225 the following computations.

226 The stretching method operates in the time domain and consists in stretching
 227 or compressing the reference or the current NCF in order to maximize the cor-
 228 relation coefficient (CC) between the two functions in a selected delay window,
 229 generally in the coda. More precisely, a grid-search on the stretching coefficient
 230 is carried out and the value ε that gives the maximum of CC is assumed to
 231 correspond to the relative velocity variation of the medium: $\varepsilon = dt/t = -dv/v$.
 232 This relation stands for a homogeneous velocity change in the structure sam-
 233 pled by the seismic waves. In the general case, the perturbation is not uniform
 234 and dv/v should be considered as an ‘apparent’ velocity variation (AVV). The
 235 correlation coefficient is calculated as:

$$CC(\varepsilon) = \frac{\int_{t_1}^{t_2} NCF_{c,\varepsilon}(t)NCF_r(t)dt}{\sqrt{\int_{t_1}^{t_2} (NCF_{c,\varepsilon}(t))^2 dt} \sqrt{\int_{t_1}^{t_2} (NCF_r(t))^2 dt}} \quad (2)$$

236 where $NCF_r(t)$ and $NCF_c(t)$ denote respectively the reference and the current
 237 stretched noise correlation functions and t_1 and t_2 are the limits of the time
 238 window used. The subscript ε stands for the stretched version of the NCF.
 239 The uncertainty on dv/v is estimated using the theoretical formula proposed by
 240 Weaver et al. (2011):
 241

$$\sigma_d = \frac{\sqrt{1 - CC^2}}{2CC} \sqrt{\frac{6\sqrt{\frac{\pi}{2}}T}{\omega_c^2(t_2^3 - t_1^3)}} \quad (3)$$

242 where T is the inverse of the frequency bandwidth and ω_c is the central fre-
 243 quency. The same procedure is applied when using noise cross-correlation func-
 244

245 tions calculated between pairs of station and single-station cross-components.
246 Figure 6 displays an example of velocity change over four months, with the
247 corresponding CC and uncertainty obtained by the stretching method using as
248 reference the averaged NCF.

249 *3.5. Estimation of velocity variations without reference*

250 When the medium is affected by significant structural and velocity changes,
251 the estimation of velocity variations may depend on the choice of the reference
252 NCF (Sens-Schönfelder et al., 2014). In order to improve the precision and to
253 increase the robustness of the estimation of velocity variations, we calculated
254 the velocity variation without reference, following the formulation proposed by
255 Brenguier et al., (2014). In this method we calculate the velocity variation
256 between all the pairs of daily NCFs with the stretching method. Each day is
257 thus compared to each other days, making a full 2D matrix of relative velocity
258 variations. Then, we reconstruct the time series of velocity variation by using
259 a Bayesian least-squares inversion (Tarantola and Valette, 1982). Significant
260 differences are obtained in the velocity variations calculated using the stack of
261 all daily NCFs as a reference (figure 6). For example, the amplitude of the large
262 velocity decrease in April (-0.7 %) obtained with a fixed reference is smaller
263 than that calculated without reference (-1.0 %). Furthermore, on May the ve-
264 locity calculated with reference almost recovers its previous value ($\sim 0\%$), while
265 the estimation without reference stabilizes at about -0.3 %. Advantages of the
266 technique without reference have been discussed by Brenguier et al. (2014) and
267 Gómez-García et al. (2018). In this study, we decided to use the stretching
268 method without reference to estimate AVV, because several stations were in-
269 stalled at the end of March 2014, precluding the identification of stable periods
270 that could be used as references.

271 *3.6. Seismic multiplets*

272 Velocity variations can also be tracked by using families of seismic events
273 with similar waveforms called seismic multiplets (Ratdomopurbo and Poupinet,

1995; Cannata, 2012; Hotovec-Ellis et al., 2014; Budi-Santoso and Lesage, 2016).
For our period of study, multiplets were obtained using the records of LP events
at UBN02 station from February to April 2014. Families of similar events were
identified by waveform cross-correlation with the GISMO toolkit (Thompson
and Reyes, 2017). The traces were band-pass filtered in the range 0.5 to 5 Hz
and cross-correlated over a window from 2 s before to 3 s after the picked P-wave
arrival time. Then correlation coefficients were calculated between each pair of
events and clusters were defined for a correlation threshold of 0.8. Finally, those
multiplets signals are used to estimate velocity variations, similar to the steps
described in subsection 3.4.

3.7. Location of velocity and de-correlation perturbations in 2D

We used the apparent velocity variations and the amplitudes of decorrelation
estimated for various station pairs and delay windows in the coda of NCF
to locate the source of these perturbations in the horizontal plane. Decorrelation
(DC) corresponds to one minus the correlation coefficient (CC). For this
purpose, a relationship between the AVV and the physical local velocity pertur-
bations $dv/v(x_0)$ was defined. Then the spatial distribution of velocity change
was estimated as an inverse problem. Similar relationship between decorrela-
tions and scattering cross-section density can be found and used. We followed an
approach proposed by Larose et al. (2010), and Planès (2013) to carry out the
mapping of velocity variations and structural changes in the horizontal plane.
This procedure assumes that the coda of NCFs are mainly multiply scattered
surface waves (Pacheco and Snieder, 2005) and it uses sensitivity kernels based
on a solution of the radiative transfer equation (Shang and Gao, 1988; Sato,
1993; Paasschens, 1997; Planès, 2013). Analytical developments and details on
the inversion procedure are presented in Appendix A.

4. Analysis of apparent velocity variations

In the following we examine the temporal AVV estimated using NCCF be-
tween several station pairs, NSCF between different components of single sta-

303 tions and their dependence with frequency. We computed the NCFs in several
304 frequency bands (0.1 – 1 Hz, 0.3 – 1 Hz, 1 – 3 Hz, 3 – 5 Hz and 5 – 8 Hz) for
305 estimating AVV at different depths below the crater, and then stretched them
306 in several time windows in the coda (with delays 10 – 60 s, 8 – 60 s, 5 – 40 s, 4
307 – 35 s and 3 – 30 s, respectively).

308 *4.1. Influence of frequency band*

309 We computed the NCCF and corresponding velocity variations for the 21
310 possible pairs of station in five frequency bands. As an example, the behavior
311 of velocity variations calculated for pair [UBN05 – UBN06] strongly depends on
312 the frequency considered (Figure 7a). At high frequency [1 – 3 Hz, 3 – 5 Hz
313 and 5 – 8 Hz], a sharp velocity drop occurred during or just after the 8.1 Mw
314 Iquique earthquake. This sudden change is not observed at low frequencies [0.1
315 – 1 Hz and 0.3 – 1 Hz]. However, in the later spectral bands, a marked velocity
316 decrease is detected prior and during the eruptive activity of Ubinas volcano.
317 The velocities returned toward their previous values after the main explosions
318 with different characteristic times. The frequency dependence of the results will
319 be interpreted later (see section 6) with regard to the penetration depth and
320 dispersion of surface waves.

321 We further calculated NSCF between vertical and horizontal components of
322 single stations for several frequency bands and we estimated the corresponding
323 AVV. The results are quite similar to those obtained from NCCF (Figure 7b). A
324 sharp velocity drop, concomitant with the Iquique earthquake, is also observed
325 for the three highest frequency bands. The amplitude of this velocity drop is
326 generally larger on NSCF than on NCCF. This may result partly from differences
327 in the volumes of the medium sampled by the diffuse waves used by the two types
328 of correlation function. In the two low frequency bands, progressive decrease
329 of velocity down to -0.8 % is obtained beginning mid-March, about one month
330 before the major explosion of 19 April 2014.

331 *4.2. Velocity variations and decorrelations associated with the volcanic activity*

332 As demonstrated in the preceding section, while the velocity is strongly
333 perturbed by the large tectonic earthquake at high frequency, clear decreases of
334 the velocity are detected up to one month before the main eruptive activity of
335 Ubinas volcano. Here, we focus on estimations of velocity variation in the band
336 $[0.1 - 1 \text{ Hz}]$ using both NCCF and NSCF at several stations.

337 Figure 8a displays the velocity variations obtained for several pairs of sta-
338 tion, indicated in figure 1b. Here, we chose station pairs for which the records
339 are more continuous, because some stations as UBN01 and UBN03 have gaps in
340 the data availability. All the curves present similar behavior characterized by a
341 clear decrease before the eruption and velocity recovering after the main explo-
342 sions. The amplitude of velocity variations depends on the station pair used,
343 an observation that will be useful to locate the perturbations in the structure
344 (see section 5). Figure 8b displays the variations of the DC obtained with the
345 stretching method for the same station pairs as in figure 8a. Similarly, there
346 are clear differences between pairs in the variations of DC . The largest velocity
347 variation (AVV) and decorrelation (DC) prior the main eruption are obtained
348 for pair (UBN05-UBN06). They partially recover after the eruptive period.

349 In figure 8c, we display some AVVs calculated from NSCF, using component
350 pairs for which the AVVs are larger. A progressive velocity decrease is observed
351 for all the stations, starting on March 18 for the southern station UBN06 where
352 the largest variation (-0.8 %) is obtained. After one of the major explosions,
353 on 13 April 2014, the velocity began to recover without returning to its initial
354 values in the study period.

355 Overall, and considering various seismic station pairs and frequency bands,
356 we observe a clear pre-eruptive velocity decrease using ambient noise correlation
357 in all the analyzed station pairs in $0.1 - 1 \text{ Hz}$ frequency band.

358 *4.3. Anisotropy of apparent velocity variations*

359 The velocity changes obtained from NSCF generally have different ampli-
360 tude when calculated with Vertical-East (Z-E) and Vertical-North (Z-N) com-

361 ponents at the same station. For example, at UBN06 the velocity decrease in
362 Z–E components is larger than for Z–N components, and at UBN04 the velocity
363 decrease in Z–E components is smaller than for Z–N components. It appears
364 that velocity changes are stronger when they are calculated between the vertical
365 and the horizontal component closest to the tangential direction with respect
366 to the crater. With this observation in mind, we computed the noise corre-
367 lation function and velocity variations for the Vertical–Radial (Z–R) and the
368 Vertical–Transverse (Z–T) components with respect to the crater. The results
369 are displayed in figure 9 where it can be observed that the velocity changes are
370 larger when using the Z–T components than the Z–R components. In some
371 cases, no variations at all are detected using Z–R components. The error bars
372 calculated with equation 3 are smaller than the velocity variations obtained for
373 the Z–T components (Figure 9c). This confirms the reliability of the differences
374 in behavior between the two component pairs.

375 Cross-correlation between Z–T and Z–R components therefore suggest an
376 apparent anisotropy in the velocity change. In term of early warning, AVV for
377 Z–T components seems to be the more useful combination.

378 *4.4. Velocity variations calculated by using seismic multiplets*

379 We analyzed the 20 most populated families of LP events recorded at UBN02
380 during the volcanic unrest. Two events belong to the same family if the cor-
381 relation coeficient between them is larger or equal to 0.8. The corresponding
382 waveforms, their spectrum and stack, as well as the number of events in each
383 family are presented in figure 10. The spectra of most families have a dominant
384 peak at about 3 Hz. Some of them have spectral peaks close to 1 Hz. The
385 occurrence of those events is displayed on figure 11a. While some families (# 1,
386 2, 3, 6, 10, 16) occurred only before the eruptive crisis, other families (# 12, 20)
387 occurred after the main explosions. Most of the remaining families (# 4, 5, 7, 9,
388 11, 14, 18) were active during the eruptive cycle but they spanned over periods
389 too short to be used for velocity change detections. Finally, 5 families (# 6, 8,
390 12, 13, 19) could be used to estimate temporal velocity variations as they are

391 relatively well spread in time. However, none of the multiplets extended over
392 the whole study period.

393 Multiplet can be obtained by clustering repetitive earthquakes, and then
394 used also to estimate AVV. They can be used to check the validity of AVV from
395 noise correlations, or as additional independent inputs.

396 To estimate AVV with multiplets, the waveforms of each family were aligned
397 on the first arrivals. To improve those times, we cross-correlated all events
398 between -2 s before and to 3 s after their manually picked first arrival, then we
399 corrected the times adding the delay that maximize the CC. Finally, using the
400 first event as reference, the coda of the remaining events were stretched in the
401 delay window [7 – 30 s] in order to maximize the correlation coefficient. When
402 the resulting maximum of correlation was smaller than 0.5, the corresponding
403 value of velocity variation was discarded. Figure 11b present the comparison
404 between the velocity variations estimated with the five selected multiplets and
405 the velocity variations obtained from NSCF. Although the values obtained with
406 multiplets are somewhat scattered and extend over a short duration, they are
407 consistent with the trend obtained by ambient noise. From mid-March to 10
408 April, the results obtained with both approaches indicate a decreasing velocity,
409 while before this period and after the main explosions, they show stable or
410 slightly increasing velocity.

411 The AVV obtained from multiplets confirm those obtained from noise corre-
412 lations. As a matter of fact, due to their sparse distribution in time, they also
413 show lower time-resolved AVV than AVV observed from NCFs.

414 **5. Spatial localization of medium perturbations**

415 As shown earlier in the article, the estimated velocity variations present
416 different amplitudes, depending on the station pairs used and their position
417 with respect to the crater. For example, the velocity decreases associated to
418 station pairs for which the direct path crosses the crater are larger than the
419 others. Similar patterns are observed in the time series of decorrelation. These

420 observations were used to locate the perturbations of velocity and structural
421 changes in the medium.

422 In figures 8a and 8b, we define 4 periods: the first one (21 to 27 March
423 2014), just after the completion of the seismic network, is used as a reference
424 (*Ref*); P_1 (29 March to 01 April 2014) when the velocity began to decrease; P_2
425 (04 to 09 April 2014) when the velocity reached its minimum, prior to the main
426 explosion; and P_3 (20 to 24 April 2014) when the velocity came back to its initial
427 values after the main explosions. Therefore, the localization of velocity change
428 and decorrelation for the periods of interest (P_1 , P_2 and P_3) are calculated with
429 respect to the reference (*Ref*) period. Here only the 0.1 – 1 Hz band is used.
430 Additionally, we locate perturbations of velocity at high frequencies triggered
431 by the Iquique earthquake. In figure 7 the dv/v shape for 1–3, 3–5 and 5–8
432 Hz bands are quite similar, indicating a change occurring over a large range
433 of depth. We chose the intermediate frequency (3-5 Hz) to locate the velocity
434 variations in the horizontal plane, for two differently defined periods: *Ref2* (24
435 to 31 March 2014) as reference and P_{AEQ} (6 to 13 April 2014) when the velocity
436 dropped after the earthquake.

437 5.1. Results

438 Figure 12 displays the maps of the velocity perturbations and of the struc-
439 tural changes for the three periods. For period P_1 (Figure 12a & 12d), a weak
440 negative variation in velocity is observed in the whole edifice, especially close to
441 the crater, as well as a small structural change on the East flank of the volcano
442 near station UBN02. During period P_2 , a pre-eruptive velocity decrease with
443 larger amplitude extends again over the whole edifice. The zone of maximal
444 variation coincides with that of structural change on the south flank of the vol-
445 cano (Figure 12b & 12e). After the main explosion on 19 April (period P_3), the
446 velocity returns close to the level of period P_1 . However, a structural change
447 with lower amplitude remains on the Southern and Eastern flanks (Figure 12c &
448 12f). The restitution index (Vergely et al., 2010) is larger than one in the whole
449 edifice (Figure 12g) which indicates that the perturbations are well recovered.

450 On the other hand, the velocity perturbation due to the Iquique earthquake is
451 located mainly in the Southeast flank of the volcano (Figure 12h).

452 *5.2. Sensitivity of velocity variations to depth of perturbation*

453 In order to estimate the depth of the perturbations, we assumed that Rayleigh
454 waves are dominant in the coda and we computed the sensitivity of their veloc-
455 ity to velocity perturbation at depth. The corresponding sensitivity kernels are
456 the derivative of the Rayleigh wave phase velocity with respect to the S wave
457 velocity for various frequencies. We used a stratified velocity model (Figure
458 13a) obtained by solving the coupled hypocenter-velocity problem for 264 local
459 volcano-tectonic earthquakes recorded by 4 or more stations at Ubinas (Kissling,
460 1994). In this model, the flat surface lies at 4500 m o.s.l. As expected, the depth
461 of highest sensitivity increases as frequency decreases. Apparent velocity vari-
462 ations calculated in frequency ranges higher than 1 Hz are mainly sensitive to
463 perturbations at depths smaller than 0.5 to 1 km below the surface (Figure 13a).
464 This confirms that sharp velocity decreases in the shallow layers of volcanoes
465 can be induced by the shaking of large earthquakes (Figure 7). The effect can
466 result from the mechanical softening of the granular volcanic material due to
467 nonlinear elastic behavior (Johnson and Jia, 2005; Lesage et al., 2014).

468 The velocity changes at frequencies lower than 1 Hz are sensitive to per-
469 turbations in deeper structures of the volcano (1 - 3 km). For example, the
470 sensitivity kernel at a frequency of 0.3 Hz has large values up to 2 km below the
471 surface and coincides with the distribution in depth of the VT seismic activity
472 (Figure 13b). The corresponding apparent velocity are thus sensitive to pertur-
473 bations in the seismogenic zone of Ubinas (Figure 13c) similar to that observed
474 for events on 2009 (Inza et al., 2014).

475 **6. Discussion**

476 *6.1. Other observations*

477 Figure 14 displays measurements of seismicity, plume elevation, thermal
478 anomalies and SO_2 flux that can be compared with the estimated velocity vari-

479 ations.

480 The volcano-tectonic activity mainly occurred in February and ended in
481 the first days of March (Figure 14a). During the whole eruptive period, the
482 dominant seismicity was the LP activity which peaked at a rate of over 500
483 events per day. The volcanic system generated also important activity of tremor
484 with total duration over 10 h/day after the first explosions (Figure 14b). A
485 significant increase of RSAM is observed in the days before the main explosions,
486 especially in the high-frequency band (Figure 14d). It corresponds to the high
487 level of LP, hybrid and tremor activity. After the eruption the RSAM returned
488 to the background level. The curve of cumulative energy shows that energy was
489 released mostly between 29 March and 13 April 2014. After the main explosion
490 on 19 April, the energy release came back to a low level.

491 The plume elevation began to increase on 2 February 2014 and reached its
492 maximum during the main explosion of 19 April (Figure 14e) and some plumes
493 possibly exceeded 5 km. The coloration of the plume got darker after the first
494 explosions of April, indicating the presence of ash.

495 The increase of volcanic radiative power (VRP) prior the eruption indicates
496 the emplacement of magma in the crater. The maximum value was registered
497 on 4 April 2014. The VRP remained almost constant until the main explosion
498 on April 19 (Figure 14f).

499 The SO_2 flux measured with a mobile scanning DOAS during field surveys
500 presented a progressive increase beginning in February and up to the main
501 explosions (Figure 14g), a behavior similar to those of thermal anomaly and
502 plume elevation. There were not measurements between 25 March and 12 April.

503 A sudden decrease of about $0.35\text{ }^\circ\text{C}$ of the water temperature occurred im-
504 mediately after the 1 April 2014 Iquique earthquake (Figure 14h). This temper-
505 ature perturbation is highly correlated with the velocity drop observed during
506 the passing of seismic waves. It partially recovered during the study period and
507 after five months the temperature had returned to its initial value.

508 *6.2. Summary of main results and interpretation*

509 The main finding of the present work is the clear velocity decrease down to -
510 0.8 % obtained by NSCF, that occurred in the three weeks prior to the explosive
511 sequence of April 2014. This velocity decrease exceeded the seasonal fluctuation
512 of dv/v for Ubinas volcano that ranges between $\pm 0.2\%$ (Figure 15). Such a
513 seasonal change is consistent with observation in other volcanos such as Okmok
514 (Haney et al., 2015; Bennington et al., 2015), Mount St. Helens (Hotovec et
515 al., 2015), and Veniaminof (Bennington et al., 2018). The largest variations of
516 AVVs by NSCF were obtained at the farthest stations from the crater (UBN06
517 and UBN07). This result differs from those of most studies where the largest
518 AVVs are observed at the closest stations to the active crater (e.g. De Plaen et
519 al., 2016;2019; Takano et al., 2017). This could result, at least for UBN06, from
520 its position close to the structurally weakest part of the volcano (Figure 12).
521 The velocity decrease appeared also clearly when using NCCF on all the station
522 pairs, suggesting a global effect in the structure. The localization of the velocity
523 perturbations in the horizontal plane using the AVVs obtained by NCCF, sug-
524 gests that the sources of velocity variation were distributed in the whole edifice,
525 with a maximal amplitude close to the crater. However this spread distribution
526 may partly result from the poor spatial resolution of the localization method
527 which is based on the use of diffusive waves and thus depends on their mean
528 free path. Structural perturbations were also localized using measurements of
529 decorrelation. The maximum of perturbation was located on the southern flank
530 which is probably a weak zone resulting from an ancient flank collapse, and
531 where many rockfalls take place (Figure 3f).

532 The velocity variations related to the volcanic activity were obtained at fre-
533 quencies lower than 1 Hz. The corresponding perturbations were mainly located
534 at depths of 1 to 3 km below the surface, which corresponds to the seismogenic
535 zone. This suggests a possible relationship between velocity decrease and seismic
536 activity through damaging or pressurization of the medium (Lamb et al., 2017).
537 The velocity variations estimated by Coda Wave Interferometry using the fami-
538 lies of similar events are consistent with those detected from noise correlation in

539 the pre-eruptive period. At frequencies higher than 1 Hz, a sharp velocity drop
540 was observed in NCFs during the large tectonic earthquake of Iquique. This
541 perturbation affected layers shallower than 1000 m and was located mainly in
542 the southeast flank that corresponds to a zone of poorly consolidated material
543 where a hydrothermal system was detected by Self-Potential studies (Gonzales
544 et al., 2014; Byrdina et al., 2013). Numerous springs are located in this flank
545 including “Ubinas Termal”.

546 From the methodological point of view, our estimations of the apparent
547 velocity variations were carried out without reference correlation functions.
548 With this approach, a time series of AVV is estimated from the calculation
549 of $N(N - 1)/2$ velocity variations instead of N values when using a reference
550 NCF. It can thus produce more reliable results (Brenquier et al., 2014), es-
551 pecially when the studied time series is too short and does not include stable
552 periods that can be used as reference or when both the velocity and the structure
553 of the medium are perturbed and generate decorrelation of the NCFs. In this
554 case, and unlike the usual method which uses a unique reference NCF, values
555 of AVV can still be obtained for the daily NCFs that are well correlated with
556 only few of the other ones.

557 We showed that interesting information can be obtained by calculating cor-
558 relation functions between the components of single stations. From these NSCFs
559 we could estimate velocity variations with large amplitudes in the pre-eruptive
560 period. This confirms the results obtained for Piton de la Fournaise, La Réunion
561 Island by De Plaen et al. (2016), at Etna volcano by De Plaen et al. (2018), at
562 Whakaari volcano by Yates et al. (2019), at Veniaminof volcano by Bennington
563 et al. (2018) and provides a complementary tool for detecting precursory sig-
564 nals of impending eruptions. Moreover we demonstrated, in the case of Ubinas
565 volcano, that the velocity decrease was much larger when using combination of
566 vertical and tangential components, with respect to the direction of the crater,
567 than with the vertical and radial components. To our knowledge, this is the
568 first evidence of an anisotropic effect of velocity variation reported in a volcanic
569 context. It may result from an anisotropy of the fractures distribution, as sug-

570 gested by the predominance of radial faults in the shallow structure (Figure 12)
571 and/or by the largest compliance of some families of crack or by anisotropy of
572 velocity structure as observed by Mordret et al. 2015. It may also be related to
573 phenomena of temporal variations of seismic anisotropy due to stress changes
574 observed on some volcanoes (Gerst et al., 2004; Bianco et al., 2006). More
575 observations in other volcanic or geologic contexts and theoretical studies are
576 required to validate and interpret this phenomenon.

577 The 2014 eruptive crisis of Ubinas was well observed thanks to the use of
578 a variety of sensors and methods (Figure 14). In the following, we propose a
579 scenario for interpreting the set of observations.

- 580 • Mid-January 2014: First observation of hybrid seismic events and pro-
581 gressive increase of hybrid and LP activity. On 1 February, beginning of
582 spasmodic tremor activity and small phreatic explosions. These phenom-
583 ena may be induced by a deep magma intrusion, that interacts with the
584 hydrothermal system of the volcano (first km below the summit crater).
585 Hot gas with higher mobility interacts with the hydrothermal producing
586 the small explosions and the tremor.
- 587 • 10 February: New thermal anomaly detected by MIROVA, SO_2 flux in-
588 crease. 20 February; beginning of harmonic tremor activity. These obser-
589 vations indicate magma extrusion in the crater.
- 590 • 18 March: velocity decreases are detected on the stations in operation.
- 591 • 29 March: LP and tremor activity increases. Strong SO_2 flux is measured.
592 Velocity decrease and decorrelation are detected on all station pairs and
593 component pairs of single station at low frequency (0.3 – 1 Hz).
- 594 • 1 April: Mw 8.1 Iquique earthquake occurs at 360 km from Ubinas. Sharp
595 velocity drop is observed at high frequencies (1 – 3, 3 – 5 and 5 – 8
596 Hz) on all NCFs. It probably occurs mainly in the shallow layers of the
597 edifice. It could result from the temporal softening of poorly consolidated
598 granular material induced by ground shaking (Johnson and Jia, 2005;

599 Lesage et al., 2014). The passing of seismic waves can also increase the
600 rock permeability, due to the mobilization of colloidal particles, droplets
601 or bubbles trapped in pores (Elkhoury et al., 2006; Manga et al., 2012).
602 Meteoric water lying in this porous material could have been released and
603 mixed to warmer thermal water producing the rapid temperature decrease
604 at Termal Ubina spring. The common origin of the velocity perturbation
605 and of the spring water disturbance may explain the temporal correlation
606 between their occurrences. The ground shaking associated with the large
607 tectonic earthquake may have also contributed to trigger the forthcoming
608 eruption.

- 609 • Beginning of April: Strong increase of RSAM and seismic energy release.
610 Velocity variations reach their maximal absolute values. Tremor dura-
611 tion still increases. Largest thermal anomaly as revealed by MIROVA,
612 plume heights and ash contents. Estimated extrusion rate peaks at $1.4 \pm$
613 $0.02m^3s^{-1}$. All these observations can be related to the raising of magma
614 in the conduit and the progressive pressurization of the magmatic system.
615 The pressure source could produce compressional strain in the surround-
616 ing medium and extensional strain at shallower depth (Budi-Santoso and
617 Lesage, 2016; Donaldson et al., 2017) and thus could induce the aperture
618 of some crack families and rock damaging yielding to decreasing velocity.
- 619 • 13 - 19 April: Series of large explosions producing the highest and most
620 ash-laden plumes, maximal values of SO_2 flux, extrusion and destruction
621 of domes. RSAM and energy release decrease and seismic velocity begins
622 to recover its previous value. The large explosions opened the conduit
623 and depressurized the system. Magma extrusion continued as observed
624 by thermal anomaly.

625 7. Concluding remarks

626 Since the development of methods based on ambient noise correlation, only a
627 few numbers of cases presenting variations of seismic velocity before an eruption

628 of andesitic volcano have been identified. In the present study, we have obtained
629 a very clear example of velocity decrease up to three weeks before the onset of
630 an explosive sequence at Ubinas volcano. Our results suggest that this precursor
631 has similar behavior on andesitic and basaltic volcanoes. Estimations of velocity
632 changes from the analysis of seismic multiplets by coda wave interferometry
633 gave consistent results. Velocity variations with amplitude down to -0.8 % were
634 also detected by using seismic noise correlation between vertical and horizontal
635 components of single stations. Moreover, the amplitudes of variation are larger
636 when using vertical and tangential components, with respect to the direction to
637 the crater, than using vertical and radial components. This observation suggests
638 an anisotropic effect in the response of the seismic velocity to pressurization
639 of the volcanic structure. Similar studies on other volcanoes and geological
640 contexts are required to determine if this behavior is common or if it is a specific
641 feature of Ubinas volcano. Theoretical developments would also be necessary in
642 order to understand the origin of this anisotropy.

643 Our results have also showed that precursory velocity variations can be de-
644 tected with NSCF calculated at single stations even if they are not close to the
645 crater. This observation, if it is corroborated by other studies and if possible
646 local effects of weaker structure can be discarded, would be of great interest for
647 volcano observatories that operate sparse monitoring networks.

648 The velocity variations at Ubinas volcano in 2014 are a relatively complex
649 phenomenon because they were induced by at least two processes: the volcanic
650 activity and the large tectonic Iquique earthquake that occurred less than two
651 weeks before the main explosions. However we could separate the two effects
652 thanks to their sensitivity in different frequency ranges and the good temporal
653 resolution of ours NCCFs. While the seismic waves generated by the earthquake
654 produced a decrease of velocity in the shallow layers of the edifice (< 1 km),
655 the velocity perturbations related with the impending eruption were located at
656 larger depth (1 – 3 km) and distributed in the whole volcanic structure. The
657 question of the possible coupling between the large earthquake and the vol-
658 canic reactivation remains open. Decorrelations of the NCFs were also detected

659 prior to the eruption. They may result from structural changes centered on the
660 southern flank of the edifice, a weak zone due to an ancient flank collapse.

661 The 2014 eruptive crisis of Ubinas volcano was observed by different types
662 of sensors, such as seismometers, DOAS, camera, satellite IR detector. Thanks
663 to the analysis of this set of observations, we proposed an interpretative sce-
664 nario of the pre- and co-eruptive periods which relates the velocity variations at
665 low frequency to the progressive pressurization of the magmatic system before
666 the main explosions, followed by its depressurization after the opening of the
667 magmatic conduit.

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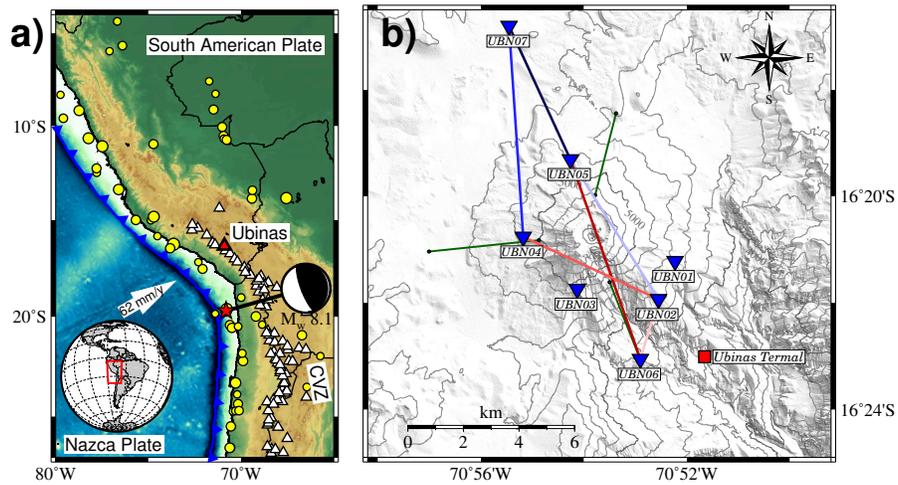


Figure 1: a) Map of the Central Volcanic Zone of southern Peru. White triangles show potentially active volcanoes. Red triangle indicates the location of Ubinas Volcano. Yellow circles are the epicenters of earthquake with magnitude larger than 7 Mw (1900-2018). The red star shows the location of the 2014 Iquique earthquake (8.1 Mw) with its focal mechanism. The blue line in the ocean is the location of the subduction trench. b) Map of Ubinas Seismic Network deployed during 2014. Inverted blue triangles: location of seismic stations. Red square: position of thermometer in spring water “Ubinas Termal”. The color lines correspond to the station pairs discussed in section 4.2. Electronic distance measurements were carried out along the green lines in 2014.

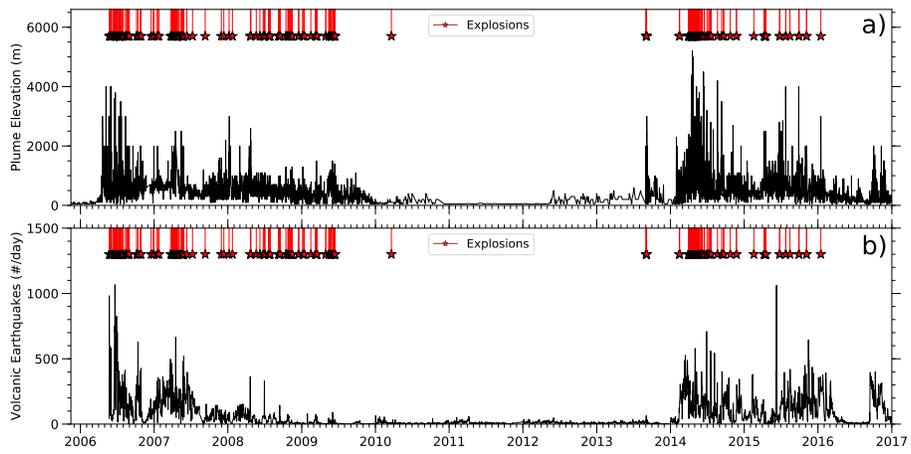


Figure 2: a) Plume elevation above crater level of Ubinas volcano recorded in the last 10 years. b) Seismic activity (sources: INGEMMET Report, 2014; IGP Report, 2013; Machacca, 2012).

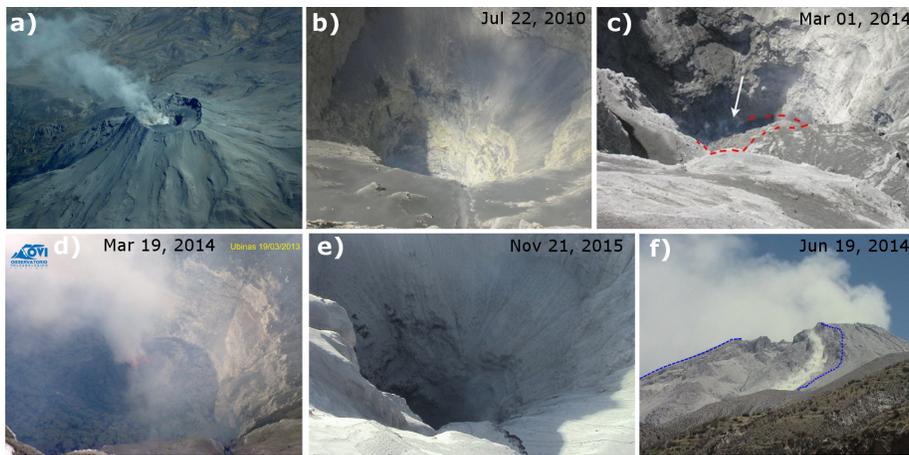


Figure 3: a) Aerial view of Ubinas volcano taken from the Northeast displaying the caldera, the active crater and fumaroles. Pictures of the active vent of Ubinas volcano on b) 2010, c) 1 March 2014, d) 19 March 2014, e) 13 June 2015. f) Rock fall during an eruption on the south flank; the old collapse scarp is indicated by blue dashed lines. Photos are from <http://lechaudrondevulcain.com> (a), R. Machacca (b), IGP Report, 2014[31] (c), J. Acosta (d), E. Alvarez (e,f).

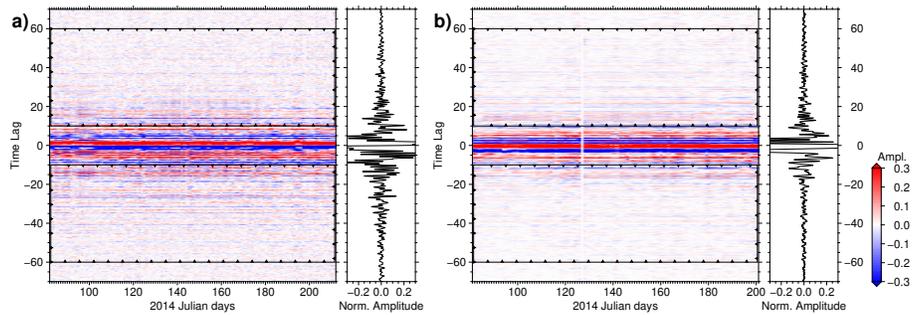


Figure 4: Daily cross-correlation functions calculated in the band $[0.1 - 1 \text{ Hz}]$. a) Noise Single-station Cross-components correlation Functions (NSCF) between Z and E components of UBN07 station. b) Noise Cross-correlation Functions (NCCF) of vertical components between UBN02 and UBN05. The respective waveforms correspond to the mean NCF. For this figure, we clipped the normalized amplitude from ± 0.3 for better visualization of the coda.

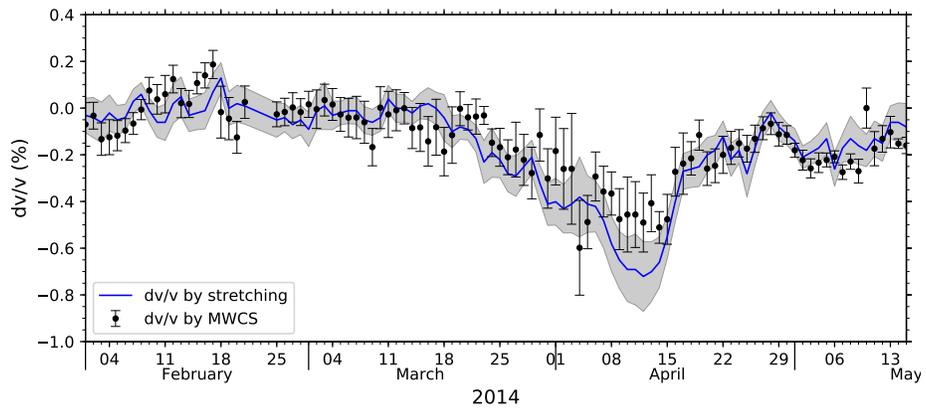


Figure 5: Comparison of the velocity changes dv/v calculated by the stretching and the MWCS methods with their respective uncertainties. The example corresponds to the single-station cross-components of Z and E components at UBN06 station [0.1-1Hz].

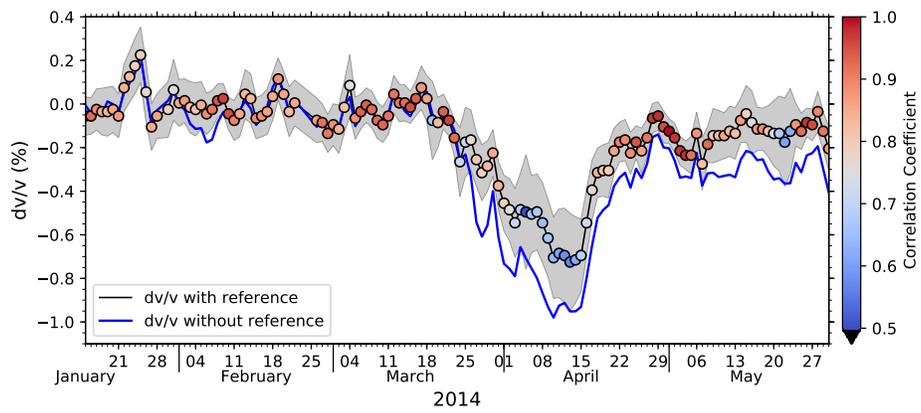


Figure 6: Comparison of the velocity changes dv/v calculated with respect to a reference NCF (black line with uncertainties represented by grey zone and correlation coefficients CC by color code) and calculated without reference (Blue line). The example corresponds to the single-station cross-components between Z and E components of station UBN06 [0.3 – 1 Hz].

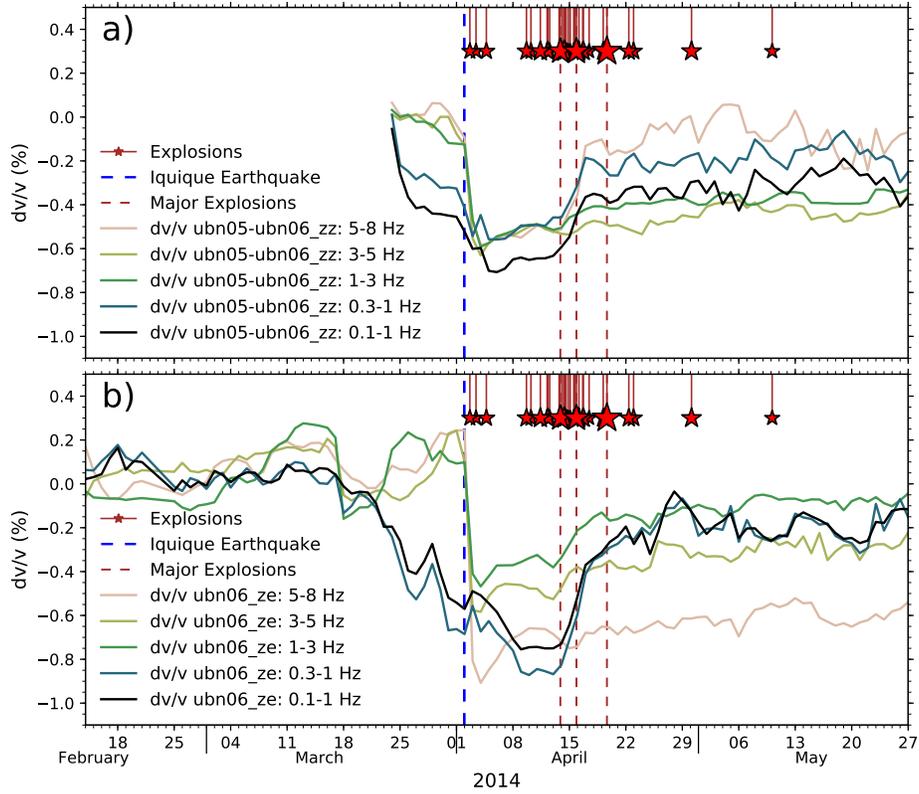


Figure 7: a) Apparent velocity variations calculated from cross-correlation functions between vertical components of UBN05 and UBN06 in five frequency bands (see legend). b) Apparent velocity variations calculated from single-station cross-components between the vertical and east components of UBN06. Vertical dashed blue and red lines indicate the occurrence of Iquique earthquake and of the major explosions, respectively. The star sizes correspond to the energy of main explosions.

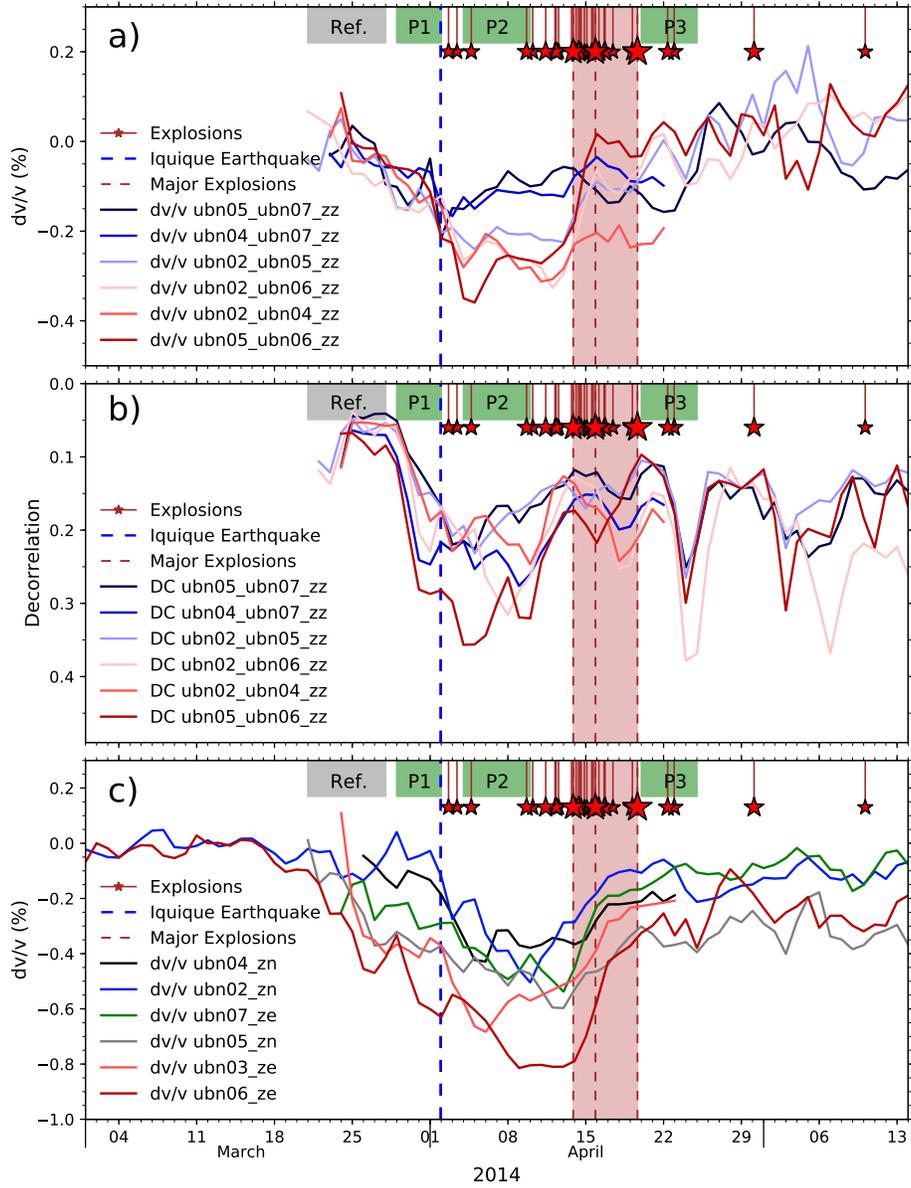


Figure 8: a) Apparent velocity variations AVV calculated from NCCF of vertical components for the pairs of station displayed in Figure 1b. Same colors are used to plot paths between stations and corresponding AVVs. b) Corresponding time series of the decorrelation (DC). c) AVVs obtained from NSCF between vertical and horizontal components at several seismic stations. In all cases the dv/v and DC are computed for frequency range $0.1 - 1$ Hz and delay windows $10 - 60$ s in the coda. The $Ref.$ and P_i boxes represent periods used for localization in 2D of velocity and decorrelation changes discussed in section 5. The red shaded zone corresponds to the main eruptive period. 44

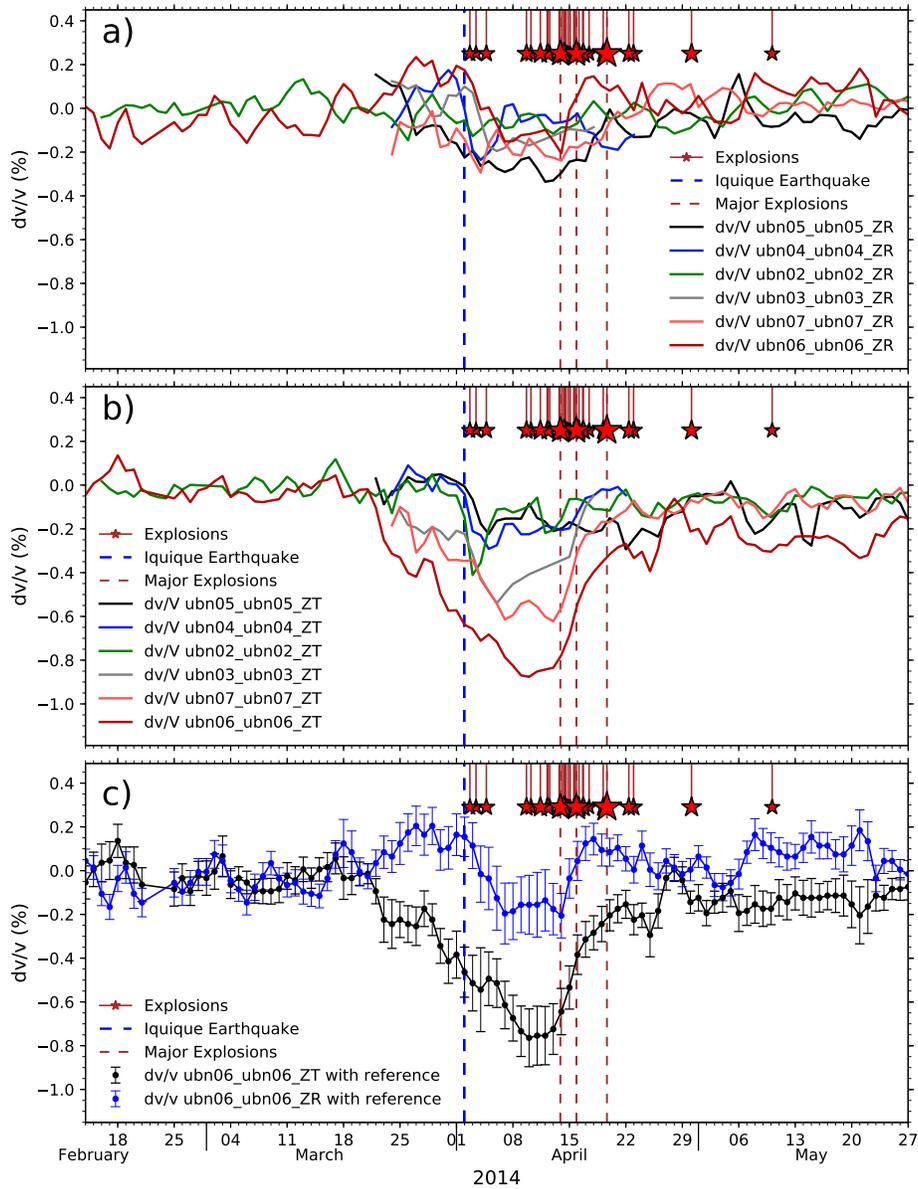


Figure 9: AVV calculated from single-station cross-components between a) Vertical and Radial components and b) Vertical and Transverse components. c) AVV obtained with reference for Z-T and Z-R components combination for station UBN06. Radial and Transverse components are with respect to the directions of the crater from each station. In all cases dv/v was computed in the range 0.1 – 1 Hz and delay window 10 – 60 s in coda.

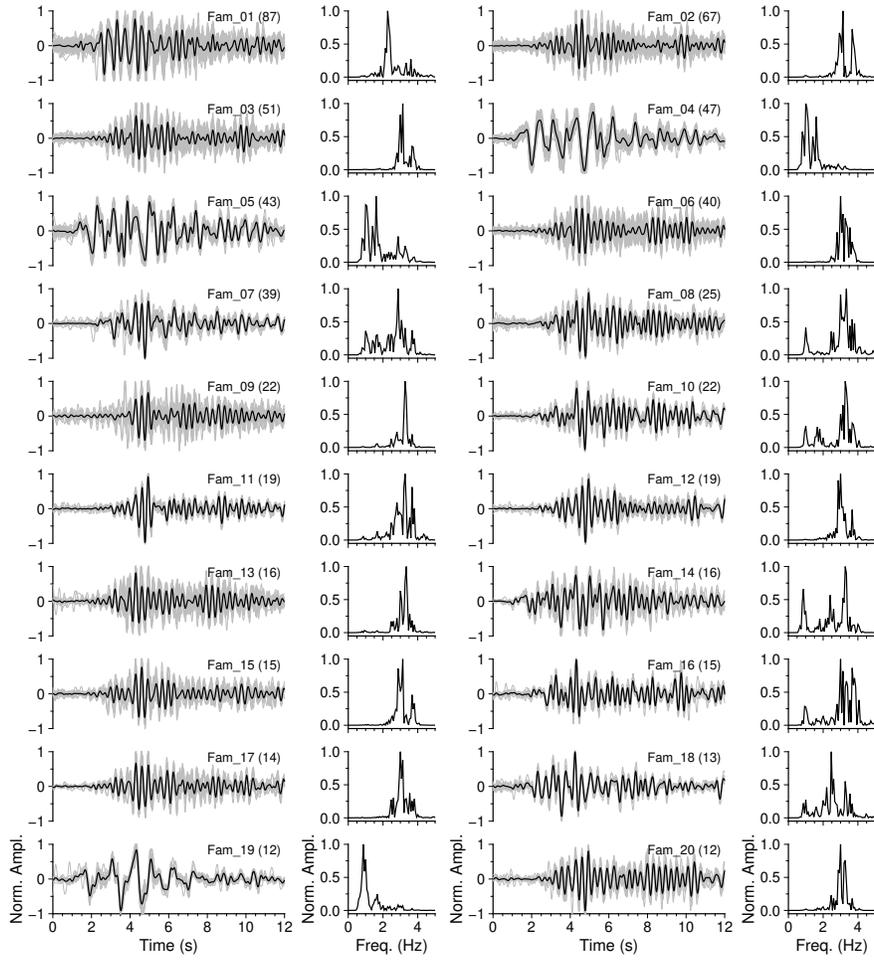


Figure 10: Twenty most populated seismic multiplets. For each cluster, the event waveforms recorded on the vertical component of UBN02 station and filtered between 0.5 and 5 Hz are plotted in grey together with their stack in black. The numbers of event in each family are indicated in parentheses. The Fourier spectra of the stacked traces are also displayed alongside.

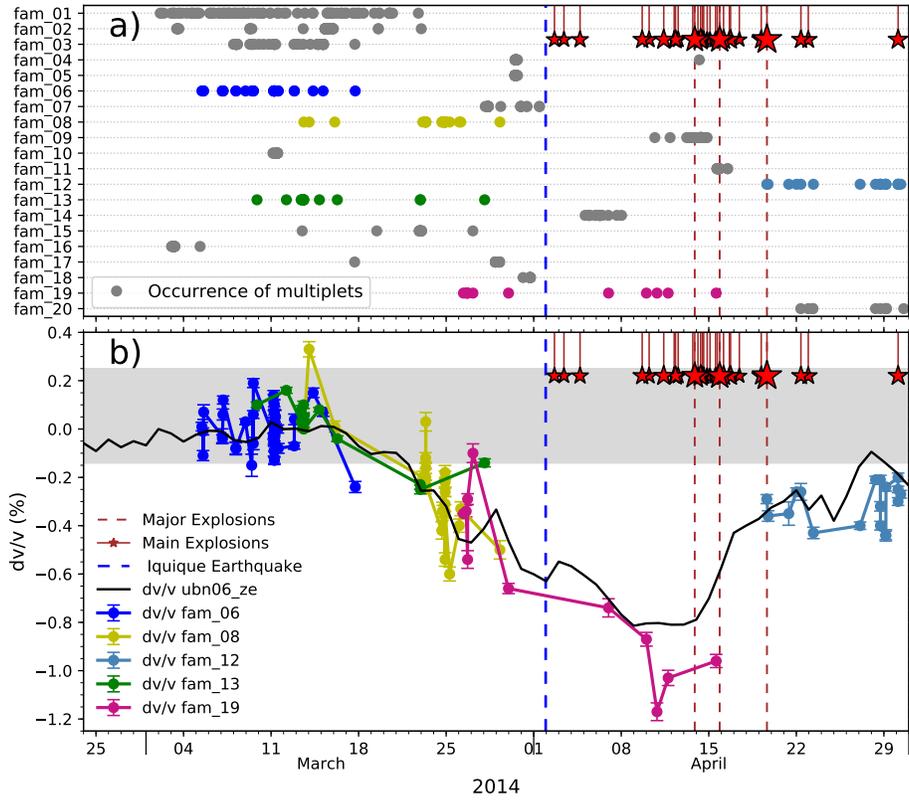


Figure 11: a) Occurrence of the LP events that belong to the main 20 families; in gray, families that have not been used to estimate dv/v . b) Velocity variations estimated by NSCF (UBN06 Z–E, black line) and multiplets with corresponding uncertainties. The width of the horizontal gray zone represents the amplitude of dv/v fluctuations obtained from NSCF before the eruption. The AVVs of families 12 and 19 were shifted to align their first value with the corresponding AVV obtained from NSCF.

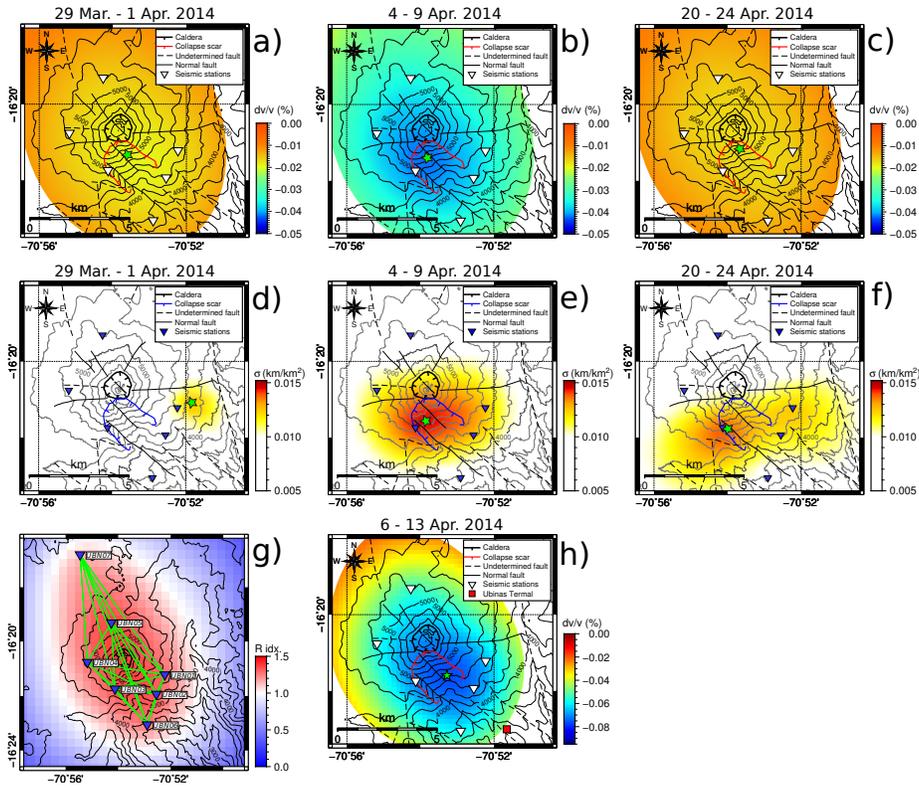


Figure 12: Maps of velocity variations (first row) and scattering cross-section density (second row) for periods P_1 (a, d), P_2 (b, e) and P_3 (c, f) for 0.1 – 1 Hz, respectively. The maps display also the topography and the main craters and faults. The position of the maximum values of velocity variation and scattering cross-section density is indicated by green stars. g) Map of corresponding restitution index, with green lines showing station pairs. The velocity changes are plotted only for pixels where restitution index are ≥ 1 . h) Maps of velocity variations after the Iquique earthquake for 3 – 5 Hz and period P_{AEQ} .

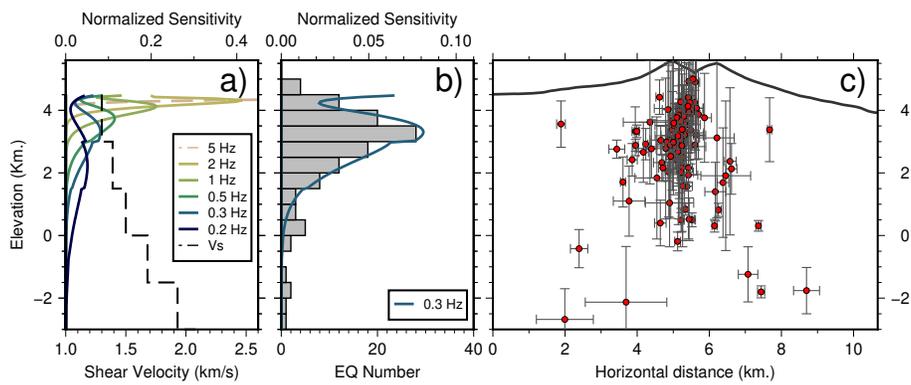


Figure 13: a) Sensitivity kernels of Rayleigh wave phase velocities to shear wave velocity for several frequencies used in this study. b) Histogram of VTs depths and sensitivity kernel of Rayleigh wave velocity for frequency of 0.3 Hz. c) Source locations of VT events for the period 26 March to 5 December 2014.

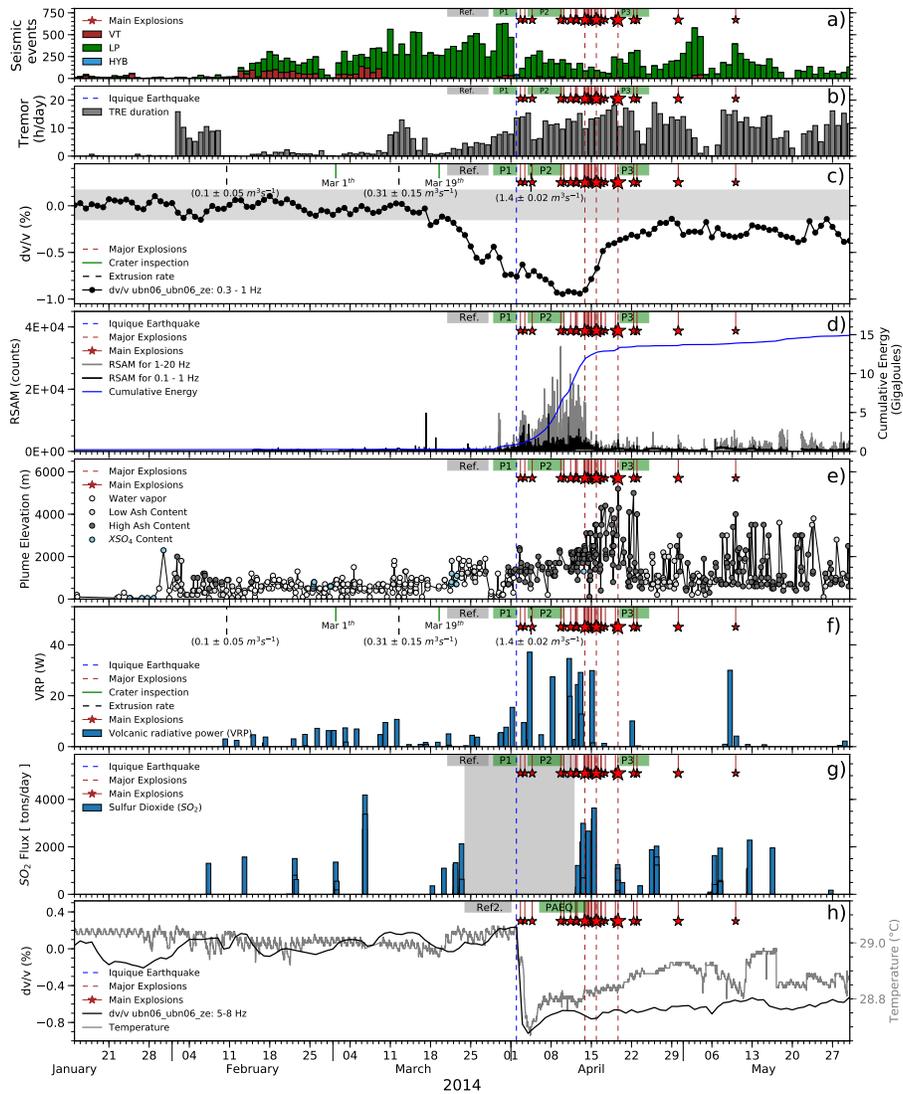


Figure 14: a) Daily count of seismic events recorded at UBN02 station. b) Duration of tremor activity. c) Velocity variations obtained by single-station cross-components between Z and E at UBN06. Estimations of extrusion rate are indicated at 3 dates (data from Copola et al., 2014[14]). d) RSAM calculated at station UBN02 in 0.1 – 1 Hz and 1 – 20 Hz frequency bands and cumulative seismic energy. e) Plume elevation with respective ash content. f) Thermal anomaly measured by Volcanic Radiative Power (VRP). g) SO₂ flux determined from mobile scanning DOAS measurements. Shaded area indicated period with no measurements. h) Temperature of spring water “Ubinas Termal” (gray line) and apparent velocity variation in the 5 – 8 Hz range (dark line).

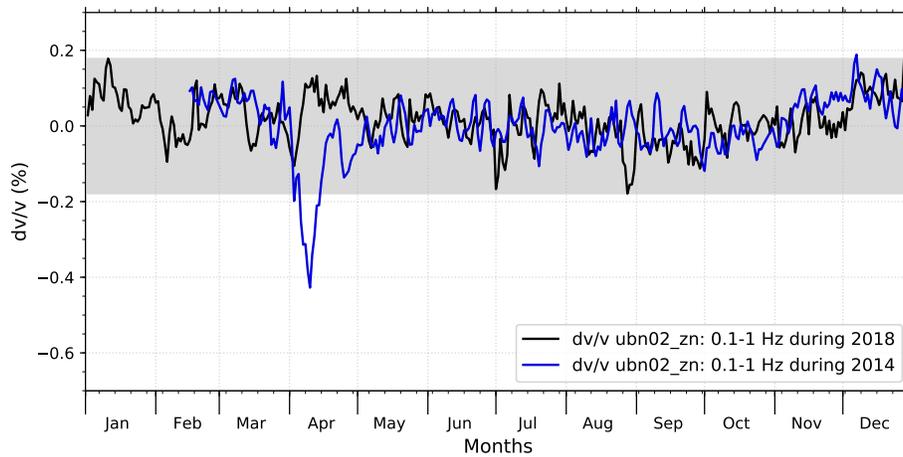


Figure 15: Comparison of dv/v obtained during a period of eruptive quiescence (year 2018, black line) with dv/v calculated for 2014 (blue line). The velocity variations associated with the eruptive crisis in April 2014 are clearly larger than the seasonal fluctuations (grey shaded area).

976 **Appendix A. Localization of velocity and structural changes**

977 *Appendix A.1. Sensitivity kernels*

978 Apparent velocity variations $\delta v/v(t)_{app}$ estimated between station pairs are
 979 related to the distribution of velocity perturbations $\delta v/v(x_0)$ in the medium S
 980 by:

$$\frac{\delta v}{v}(t)_{app} = \frac{1}{t} \int_S K(s_1, s_2, x_0, t) \frac{\delta v}{v}(x_0) dS(x_0) \quad (\text{A.1})$$

981 where t is the travel time, s_1 and s_2 are the positions of the stations, x_0 is
 982 the location of the perturbations, and K is a sensitivity kernel introduced by
 983 Pacheco and Snieder, (2005):

$$K(s_1, s_2, x_0, t) = \frac{\int_0^t p(s_1, x_0, u) p(x_0, s_2, t - u) du}{p(s_1, s_2, t)} \quad (\text{A.2})$$

984 The sensitivity kernel is a statistical measure of the time spent in each part of
 985 the region under test. $p(s_1, s_2, t)$ is the probability that the wave has traveled
 986 from s_1 to s_2 during time t , which can be approximated by the intensity of the
 987 wavefield from s_1 to s_2 at time t . Since surface waves are the dominant wave
 988 type, we use the analytic two-dimensional solution of the radiative transfer for
 989 isotropic scattering for the intensity propagator (Obermann et al., 2013):

$$p(r, t) = \frac{\exp(-ct/\ell)}{2\pi r} \delta(ct-r) + \frac{1}{2\pi \ell ct} \left(1 - \frac{r^2}{c^2 t^2}\right)^{-1/2} \exp\left(\frac{\sqrt{c^2 t^2 - r^2} - ct}{\ell}\right) \Theta(ct-r) \quad (\text{A.3})$$

990

991 where ℓ is the scattering mean free path, $\ell = \ell^*$ for isotropic scattering, ℓ^* is
 992 the transport mean free path $\ell^* = 1/(1 - \langle \cos(\theta) \rangle)$, where θ is the angle between
 993 the vector of the incident wave and the vector of the scattered wave, $\langle \cos(\theta) \rangle$
 994 is the anisotropy of diffusion and the brackets denote averaging over all solid
 995 angles. r is the distance between source and receiver, c is the wave velocity, and
 996 Θ is the Heaviside (or step) function. The first term of Equation A.3 describes
 997 the coherent part of the intensity that decreases exponentially with the distance
 998 relative to the transport mean free path. The second term describes the diffusion

999 intensity. Note that the diffusion solution is reached when $t \gg r/c$. An example
 1000 of the sensitivity kernel is displays in Figure A.1.

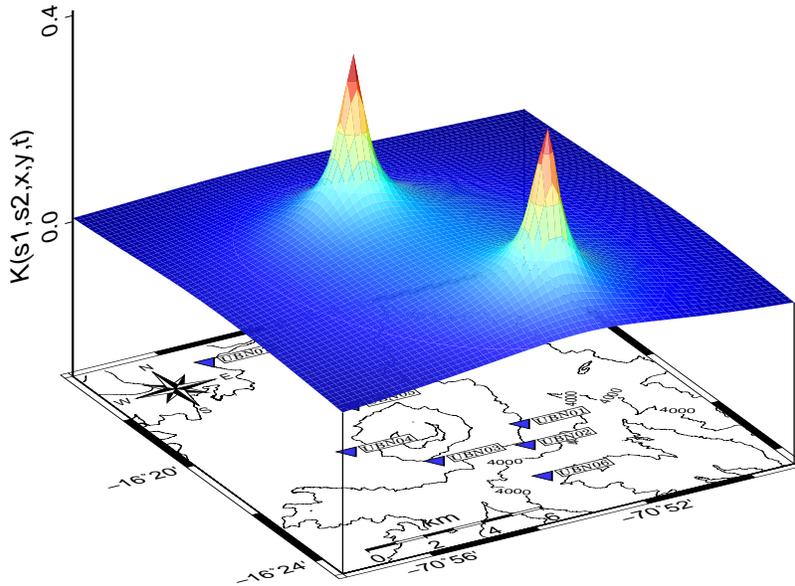


Figure A.1: Sensitivity kernel for [UBN05 - UBN06] station pair. The two peaks are located at the station positions.

1001 For the present work, we took $c = 1.38 \text{ km/s}$ and we choose a value of mean
 1002 free path $\ell = 5 \text{ km}$, as reported by Rossetto et al., (2011), who demonstrated
 1003 that this location method is not very sensitive to the value of the mean free path.

1004

1005 *Appendix A.2. Inverse problem*

1006 We follow the procedure described by Froment (2011), Obermann et al.
 1007 (2013), and Lesage et al. (2014). Equation A.1 can be rewritten as:

$$d = Gm \tag{A.4}$$

1008 where d is the data vector, the components of which are the measured apparent
 1009 velocity variations between two dates, G is the matrix of the sensitivity kernels
 1010 weighted by elementary surface over time t in the coda $G = \frac{\Delta S}{t} K$ and m is the

1011 model vector of the velocity perturbations $\delta v/v(x_0)$ for each cell:

1012 $d = \varepsilon$, $G = \frac{\Delta S}{t}K$ and $m = \delta v/v$

1013 Because the problem is linear, we can use a formulation of least square inverse
1014 method proposed by Tarantola and Valette (1982):

1015

$$m = m_0 + C_m G^t (G C_m G^t + C_d)^{-1} (d - G m_0) \quad (\text{A.5})$$

1016
1017 where m_0 is the initial model (a zero vector) and C_d is the diagonal covariance
1018 matrix of the data. C_m is the covariance matrix for the model which is intro-
1019 duced in order to produce smooth models:

1020

$$C_m(i, j) = \left(\sigma_m \frac{\lambda_0}{\lambda} \right)^2 \exp \left(- \frac{d(i, j)}{\lambda} \right) \quad (\text{A.6})$$

1021 where $d(i, j)$ is the distance between two cells i and j , λ is the correlation length,
1022 λ_0 is the cell length, and σ_m is an a priori standard deviation of the model. We
1023 used values of $\lambda = 2$ km, $\lambda_0 = 0.5$ km and $\sigma_m = 0.02$. The resolution matrix R
1024 is given by:

$$R = C_m G^t (G C_m G^t + C_d)^{-1} G \quad (\text{A.7})$$

1025 The sum of the elements of row j of the matrix is the restitution index of the
1026 j^{th} cell of the model (Vergely et al., 2010). Values close to one indicate good
1027 recovery of the model in the corresponding cells. In our case we computed the
1028 sensitivity kernels in an area of 16 km x 18 km around the crater. We divided
1029 this area into 1152 cells with $\Delta S = 0.25 \text{ km}^2$. We used several time lag in the
1030 coda waves with 20 s-long moving windows and 5 s of steep. Data are discarded
1031 when corresponding CC is < 0.75 . For periods P_1 , P_2 and P_3 we used 312, 284,
1032 and 274 data respectively for the inversion. We did not consider topography in
1033 this procedure.

1034 *Appendix A.3. Localization of structural changes*

1035 The structural changes in the medium can be localized by a similar ap-
1036 proach as that used for velocity variations (Planès et al., 2014; Obermann et

1037 al., 2013). The data are the decorrelation (DC) values that are obtained directly
 1038 from the stretching method as:

$$DC = 1 - CC \quad (\text{A.8})$$

1039 where CC are the correlation coefficients calculated between the stacked corre-
 1040 lation functions of the current study period and those of the reference period
 1041 (21 to 27 March 2014). To estimate the horizontal distribution of the decorre-
 1042 lation $DC(x_0)$, we adapt the least-squares inversion described in the previous
 1043 subsection (Equation A.5) to the decorrelation measurements:

1044

$$1045 \quad d_i = DC_i ; G_{i,j} = \frac{c\Delta S}{2} K_{i,j} \text{ and } m_j = \sigma_j$$

1046

1047 where DC_i is the vector of the decorrelation values for each station pairs. $G_{i,j}$
 1048 is the matrix of the sensitivity kernels $K_{i,j}$ for station pair i and cell j , weighted
 1049 by the area of the cells ΔS and the Rayleigh wave group velocity ($c/2$). m is the
 1050 vector of the scattering cross-section density changes that we want to estimate
 1051 at each cell j . The initial model m_0 is again null everywhere and we used an
 1052 iterative inverse procedure to constrain the values of m to be positive (Oberman
 1053 et al., 2013). As for velocity changes, we used a 20 s-long moving windows with
 1054 steps of 5 s, in the band 0.1 – 1 Hz. For P_1 , P_2 and P_3 , we used 175, 174 and
 1055 201 observations respectively as data for the inversions.