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Tracking changes in the co-eruptive seismic tremor associated with magma degassing at Piton de la Fournaise volcano

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1 Highlights

2 **Tracking changes in the co-eruptive seismic tremor associated with**
3 **magma degassing at Piton de la Fournaise volcano**

4 Cyril Journeau, Nikolai M. Shapiro, Aline Peltier, Valérie Ferrazzini, Jean
5 Soubestre, Zacharie Duputel, Andrea Di Muro, Claude Jaupart, Diego Cop-
6 pola

- 7 • Seismic monitoring network data analysis enables to measure spatial,
8 temporal and spectral properties of co-eruptive tremor
- 9 • Shallow magma degassing beneath the eruptive sites controls the Piton
10 de la Fournaise seismic tremor generating mechanism and characteris-
11 tics
- 12 • Tremor data along with others observatory data reveal different erup-
13 tive phases linked to the evolution of degassing regimes in the eruptive
14 conduit

15 Tracking changes in the co-eruptive seismic tremor
16 associated with magma degassing at Piton de la
17 Fournaise volcano

18 Cyril Journeau^{a,b}, Nikolai M. Shapiro^a, Aline Peltier^{c,d}, Valérie
19 Ferrazzini^{c,d}, Jean Soubestre^{a,e}, Zacharie Duputel^{c,d}, Andrea Di Muro^f,
20 Claude Jaupart^c, Diego Coppola^g

^a*Univ. Grenoble Alpes, Univ. Savoie Mont Blanc, Institut des Sciences de la Terre (ISTerre), CNRS (UMR 5275), Grenoble, France*

^b*Earth Sciences Department, University of Oregon, Eugene, OR, USA*

^c*Université de Paris, Institut de Physique du Globe de Paris (IPGP), CNRS (UMR 7154), Paris, France*

^d*Observatoire Volcanologique du Piton de la Fournaise, Institut de Physique du Globe de Paris (IPGP), La Plaine des Cafres, France*

^e*Icelandic Meteorological Office, Reykjavík, Iceland*

^f*Laboratoire de Géologie de Lyon : Terre, Planètes, Environnement, CNRS (UMR 5276), Ecole Normale Supérieure de Lyon, Université Claude Bernard Lyon 1, Villeurbanne, France*

^g*Department of Earth Sciences, University of Torino, Torino, Italy*

21 **Abstract**

22 Volcanic eruptions in basaltic systems represent the most frequent expression
23 of volcanic activity in the World, whose surface manifestations range from
24 effusive lava flows to more explosive events. Their eruptive dynamics and
25 style are largely controlled by the degassing of magma. In this study, we
26 analyze seismic tremors recorded during 23 eruptions that occurred at Piton
27 de la Fournaise volcano (La Réunion, France) during 2014-2021 and show
28 that their properties are strongly linked to the magma degassing. We apply
29 the network covariance matrix method to build a catalog of tremor sources
30 associated with all 23 eruptions, to measure their time-frequency properties,
31 and to locate their sources. We then conduct a multi-disciplinary analysis by
32 comparing the seismic tremor observations with images of the eruptive sites,
33 lava flow rate measurements, impulsive earthquakes, as well as deformation
34 and magma composition data. The tremor depth distribution is found to be
35 well correlated with magma gas content, indicating that the tremor generat-

36 ing mechanism is largely controlled by the shallow magma degassing under
37 the eruptive site. The resulting seismic tremor signals analysis enable to
38 track changes in degassing regimes and associated eruptive styles at the sur-
39 face, thus improving monitoring and knowledge of the eruptive dynamics of
40 the studied volcano.

41 *Keywords:* Seismo-volcanic tremor, Magma degassing, Piton de la
42 Fournaise, Eruptive dynamics, Volcano monitoring

43 1. Introduction

44 Basaltic systems exist in all types of geodynamic settings (hot spot, rift
45 zone, subduction zones) (Walker, 1993) and their eruptions are the most
46 common manifestation of surface volcanic activity on Earth (Houghton and
47 Gonnermann, 2008). The main driver of these eruptions is the formation
48 of gaseous phase during the exsolution of the volatile elements dissolved in
49 liquid magma (dominantly H_2O and CO_2) when it ascents toward the surface
50 and decompresses through nucleation of gas bubbles (Sparks, 2003; Vergnolle
51 and Gaudemer, 2015).

52 This degassing in low-viscosity basaltic systems can give rise to several
53 flow regimes in volcanic conduits (Jaupart and Vergnolle, 1988; Gonner-
54 mann and Manga, 2013), generating a wide range of eruptive styles going
55 from effusive lava flows to mild explosive Strombolian explosions and Hawai-
56 ian fountaining and more rarely intensively explosive Vulcanian eruptions.
57 The corresponding gas-liquid flow dynamic and possible transitions between
58 regimes depend on several parameters, including physico-chemical magma
59 properties (temperature, pressure, composition, viscosity, initial content of
60 dissolved volatiles elements), conduit geometry (James et al., 2004), conduit
61 roughness, and gas flow rate (Spina et al., 2019).

62 Therefore, in order to better understand the mechanisms involved and to
63 better apprehend the volcano behavior and its potential evolution towards
64 phases of unrest, we need to monitor active basaltic systems by acquiring
65 multi-disciplinary data and cross-referencing them.

66 Among several tools, analysis of seismovolcanic signals has proven very ef-
67 ficient for improving our understanding of volcanic systems and to infer their
68 complex dynamics (Chouet and Matoza, 2013). In particular, long-period
69 (LP) seismicity is thought to be associated with the gas-liquid dynamics and
70 their interaction with solid rock. This category of seismicity encompasses

71 both continuous tremor and impulsive LP events and is characterized by fre-
72 quency content between 0.5 and 5 Hz (Chouet, 1996). Physical mechanisms
73 generating volcanic tremors are not yet completely understood and they can
74 vary depending on the volcanic system and depth of source activation (Kon-
75 stantinou and Schlindwein, 2003; Chouet and Matoza, 2013). However, most
76 of existing source models involve pressure fluctuations at the origin of the
77 sustained long-period oscillations characterizing tremor signals. In partic-
78 ular, degassing is often invoked as a possible source for the generation of
79 the LP seismicity in different volcanic systems, from their deep roots, for
80 example at Klyuchevskoy (Melnik et al., 2020), to their shallow parts, for
81 example at Kilauea (Chouet and Dawson, 2011) or at Etna (Patanè et al.,
82 2008; De Barros et al., 2011).

83 In this study, we focus on recent eruptions of the Piton de la Fournaise
84 (PdF) volcano and investigate the co-eruptive seismic tremors and the asso-
85 ciated degassing dynamics. To better understand the origin of the volcanic
86 tremor at PdF and to extract information on the dynamics of its eruptions,
87 we analyze in detail the continuous data recorded by its dense seismic net-
88 work (Figure 1). We then compare the outputs with others available multi-
89 disciplinary eruptive parameters. In particular, we want to analyze spatial,
90 spectral and temporal patterns in tremor data that may indicate change in
91 the gas-liquid dynamics regime and to investigate their relationship with the
92 lava discharge rate at the vent, petrological information on lava products, pic-
93 tures of the eruptive cone, deformation of the volcanic edifice and associated
94 co-eruptive seismicity.

95 PdF is an ideal natural laboratory for testing these ideas, being one of
96 the most active, with more than two eruptions per year in average (Roult
97 et al., 2012; Staudacher et al., 2016; Chevrel et al., 2021), and most densely
98 instrumented volcano in the World, with more than 100 geophysical and geo-
99 chemical sensors distributed on the island. Each PdF eruption is associated
100 with a co-eruptive seismic tremor and thanks to good instrument mainte-
101 nance, all these eruptive seismic signatures have been recorded with a dense
102 seismic network in the vicinity of the volcano (from 16 to 20 seismic sta-
103 tions depending on data availability, within a 15 km diameter circle) for the
104 last 23 eruptions that occurred in the 2014-2021 time period. The resulting
105 large tremor database contains thus a wealth of important information and
106 requires advanced network-based methods for its exploration.

107 In section 2, we introduce the PdF eruptive dynamics with associated
108 degassing and seismic tremor patterns. Then, we present in section 3 our

109 methodological approach allowing to infer the spatial, temporal and spectral
110 properties of the seismic tremor and to compare our outputs in a multi-
111 disciplinary dataset. The obtained results are presented in section 4 and
112 finally discussed in section 5.

113 **2. Regional settings: the Piton de la Fournaise eruptive dynamics** 114 **and associated eruptive tremors**

115 *2.1. Piton de la Fournaise volcanism origin and current dynamics*

116 La Réunion is a volcanic island located on the African plate, southwest of
117 the Indian ocean in the Marscarene Basin (Figure 1). Based on dating and
118 geodynamic reconstruction, its volcanism is thought to have a hot spot origin,
119 being generated by the same mantle plume that produced the Deccan Trapp
120 about 65 Ma ago (Courtilot et al., 1986). The head of this mantle plume
121 would have passed beneath La Réunion about 5 Ma (Bonneville, 1990), and
122 the plume conduit would be now located at 100-200 km north of the island
123 (Barruol and Fontaine, 2013). Tsekhmistrenko et al. (2021) present a high
124 resolution body-wave tomography of the whole mantle column beneath the
125 western Indian Ocean providing a more complex view of the plume beneath
126 La Réunion. Their model reveals that the mantle upwellings present a tree-
127 like structure and show considerable tilt in the lower mantle instead of being
128 near-vertical, splitting into branches near the surface. Wamba et al. (2021,
129 2023) derived a multi-mode waveform shear wave tomography revealing that
130 the hotspot feeding source is anchored at the core-mantle boundary, with
131 a deep mantle plumbing system including alternating vertical conduits and
132 horizontal ponding zones.

133 The Piton de la Fournaise volcano is an intraplate shield basaltic volcano
134 of Hawaiian type located in the southeastern part of the island. It is 2631 m
135 high and active since about 530 ka. Throughout its existence, this volcano
136 has been marked by the formation of successive concentric calderas, the latest
137 being the Enclos Fouqué, formed about 4500 years ago (Bachèlery, 1981), in
138 which there is a summit cone of 400 m high and 3 km wide at its base.

139 The eruptive activity of PdF can be described as cycles with periods of
140 high activity with an average of two eruptions per year, separated by periods
141 of three to six years without activity. After a break of more than five years
142 between 1992 and 1998, the volcano has become very active again. There was
143 then a 3.5 year break between 2010 and 2014, with a resumption of activity in
144 June 2014. The volcano is still in an active period at the time of writing this

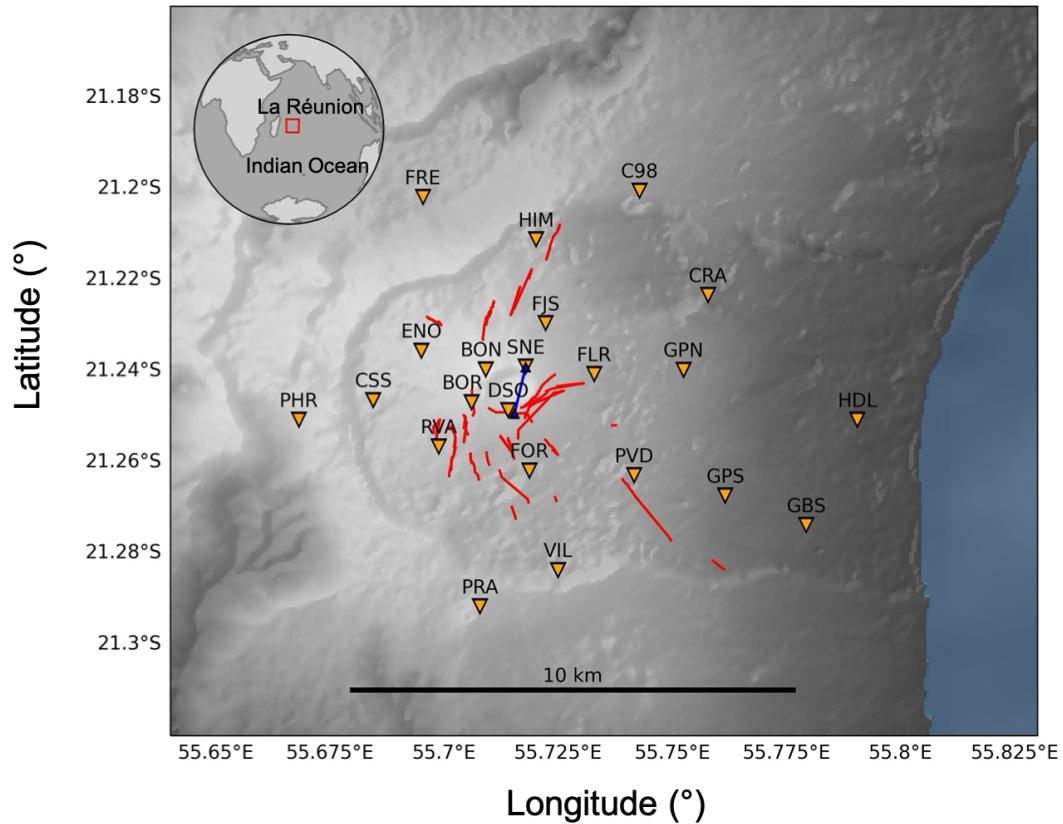


Figure 1: Grey shaded relief map of the Piton de la Fournaise volcano, located on the southeastern part of La Réunion island. The top left inset shows the location of La Réunion island in the Indian Ocean. Seismic stations used in this study are represented with orange inverted triangles. The blue line connecting the blue triangles correspond to the baseline computation between SNEG-DSRG GNSS stations. Red lines stand for the position of the eruptive fissures corresponding to all eruptions analyzed in this study.

145 paper and the last eruption considered in this analysis is that of December
 146 2021 - January 2022. For the 23 eruptions that occurred during our study
 147 period (2014-2021), the average rate of magma released is $0.63 \text{ m}^3/\text{s}$.

148 During inter-eruptive periods, one of the long-term precursor signals of an
 149 eruption is the inflation of the volcanic edifice seen by the GNSS deformation

150 sensors. This inflation is interpreted as the expansion of the shallow mag-
151 matic reservoir located at 1500-2500 m below PdF summit (Peltier et al.,
152 2005, 2008) due to a replenishment from the deep system. Such a deep
153 recharge can be resolved over ten days thanks to the inversion of GNSS data
154 (Beauducel et al., 2020). Some precursors are also visible in seismic veloc-
155 ity changes measured by cross-correlation of ambient noise. Indeed, several
156 studies have been conducted at PdF, showing decreases in seismic veloc-
157 ity correlated with edifice inflation, likely linked to some magma pressure
158 buildup before an eruption. (Brenquier et al., 2008; Duputel et al., 2009;
159 Rivet et al., 2014; Sens-Schönfelder et al., 2014; Rivet et al., 2015).

160 The short-term precursor signals are the triggering of a seismic crisis
161 composed of hundreds of volcano-tectonic earthquakes (VTs) and a strong
162 increase in the rate of deformation of the volcano (Roult et al., 2012). The
163 resulting seismic swarm is interpreted as the activation of a ring-fault portion
164 located above the shallow magma reservoir while magma starts its upward
165 propagation (Duputel et al., 2019).

166 Magma is then transported via the propagation of dykes or sills through
167 the shallow system until the fissure opening at the surface marks the be-
168 ginning of an eruption, either on the summit or on one of the flanks of the
169 volcano (Peltier et al., 2005; Smittarello et al., 2019; Dumont et al., 2022).
170 This propagation phase is marked by a decrease in the number of observed
171 VTs and it is generally accompanied by a higher frequency micro-seismicity.
172 Locating this micro-seismicity tracks the progress of the magma towards the
173 eruptive site (Taisne et al., 2011; De Barros et al., 2013; Journeau et al.,
174 2020).

175 Melt and fluid inclusion studies show that PdF magmas undergo extensive
176 CO₂ degassing at mantle level and that water and SO₂ exsolution dominate
177 degassing at shallow crustal level, halogens being released only at very shal-
178 low level (Di Muro et al., 2016). On average, PdF magmas contain low to
179 moderate volatile contents, due to their multistep ascent and degassing at
180 shallow depth (Verdurme et al., 2022). Magmas emitted after long phases
181 of quiescence like in June 2014 have experienced extensive degassing and
182 crystallisation at shallow level (Gurioli et al., 2018).

183 As generally observed at basaltic volcanoes, PdF eruptions start with
184 opening of hundred-meters-long eruptive fissures associated with lava foun-
185 taining. After a few hours, eruptions focalize to a single vent, generally at the
186 lowest point of the eruptive fissures (Fukushima et al., 2010). At this stage,
187 magma degassing typically changes from a Hawaiian to a Strombolian regime

188 (Thivet et al., 2020a). Such activity can either persist a few hours/days or a
189 few weeks/months when associated with magma re-feeding phases.

190 Eruptions typically end when the overpressure driving the magma/gas is
191 exhausted, with no further lava emission at the surface, a process commonly
192 accompanied with a transition toward an intermittent degassing regime. This
193 last phase is often reported in the IPGP-OVPF volcanological bulletins
194 (<https://www.ipgp.fr/en/observatory-releases-and-bulletins/>) as seismic se-
195 quences of “gas-pistons” marking the end of the eruptions, the last gas puffs
196 being seen in the seismic records as discrete tremor bursts.

197 Gas pistoning is a style of degassing that was first described by Swanson
198 et al. (1979) at Kilauea volcano during the 1969-1971 Mauna Ulu eruption.
199 Gas acts as a piston promoting the cyclic rise of the lava level in a conduit
200 until bubbling and gas-jetting are observed, followed by an abrupt fall of
201 the lava column. Two different source mechanisms can explained such gas-
202 piston activity: a cyclic effusion–gas-jetting–drain-back model and a foam
203 layer collapse mechanism (Marchetti and Harris, 2008). In the first model,
204 the lava column is pushed upwards due to the build-up of gas beneath a
205 relatively impermeable magma layer, until lava spreads on the surface, de-
206 creasing pressure on top of the gas layer and thus inducing gas release by
207 gas-jetting (Swanson et al., 1979; Johnson et al., 2005). The second model
208 is based on the accumulation of gas bubbles at magmatic chamber roof or
209 conduit constriction, forming a foam that collapses when reaching a critical
210 thickness. This foam collapse generates a rising gas slug in the conduit that
211 bursts at the free surface and release the over-pressurized gas (Jaupart and
212 Vergnolle, 1988, 1989; Vergnolle and Jaupart, 1990).

213 Short-period seismic records of gas-pistoning at Kilauea volcano have
214 been analyzed by Chouet and Shaw (1991); Ferrazzini et al. (1991); Goldstein
215 and Chouet (1994). A clear link has been established between those seismic
216 signals and vigorous degassing episodes, with the paroxysmal gas release
217 corresponding to the maximum observed seismic amplitudes. More recently,
218 Chouet and Dawson (2015) analyzed broadband seismic records of Kilauea
219 gas-pistoning activity which is simulated with a foam growth and collapse
220 model.

221 More rarely, the ending phases of Piton de la Fournaise eruptions are
222 characterized by “effusive paroxysms”, with an intensification of the eruptive
223 activity at the surface and a return of the lava fountains followed by a very
224 abrupt end within a few minutes (Coppola et al., 2017; Verdurme et al.,
225 2022).

226 *2.2. Co-eruptive seismic tremors*

227 The PdF eruptions thus exhibit a wide range of eruptive styles associated
228 with different degassing regimes with a more or less complex time evolution.
229 All these eruptions are recorded by the PdF seismic network and correspond-
230 ing seismic tremor signal can be seen as the eruptive seismic signature. Given
231 the spatially dense seismic network and the high temporal resolution sam-
232 pling compared to other observatory data, the seismic tremor database is
233 ideal for monitoring eruptions, catching fine patterns of their dynamics, and
234 inferring the underlying magmatic processes at play.

235 It is one of the first times that we can study the seismic records of so
236 many eruptions at a single volcano with a stable and dense network. It
237 is important to scan long time series of data including several eruptions to
238 investigate the different observed patterns and to have a broader vision re-
239 garding the different PdF eruptive manifestations and mechanisms governing
240 their dynamics.

241 Using a method based on measuring the tremor amplitude at different
242 stations of a network, Battaglia and Aki (2003); Battaglia et al. (2005b) lo-
243 cated sources of several co-eruptive tremors at the PdF for eruptions between
244 1998 and 2000. They found a shallow source correlating with the eruptive
245 vents position for frequencies above 1.5 Hz and suggested that lower tremor
246 frequencies could be related to deeper processes. Using a different approach
247 based on seismic network cross-correlations, Journeau et al. (2020) located
248 tremor sources corresponding to the three eruptions of 2010, the first ones
249 recorded by a dense seismic network of 21 stations at PdF. The resulting
250 locations are thus more precise and show a very shallow position correlat-
251 ing perfectly with that of the corresponding eruptive site. In this study, we
252 systematically locate all co-eruptive tremor sources associated with the fol-
253 lowing 23 eruptions between 2014 and 2021 and represent the results in the
254 section 4.

255 As it is difficult to deduce a unique physical tremor mechanism and to
256 interpret its signal fluctuations, it is essential to ground our observations in
257 a larger database including different disciplines, allowing them to be put into
258 perspective and to have several points of view on the eruptive processes. This
259 knowledge acquired by cross-referencing different information can ultimately
260 be re-used at volcano observatories in real-time, to read more easily the
261 seismic tremor data and translate them into physical eruptive parameters
262 (physico-chemical magma properties, size of the dyke/conduits, output lava

263 flow, etc), allowing a quantitative monitoring of the PdF eruptive dynamics
264 evolution.

265 Investigating relations between seismic tremor amplitude and lava dis-
266 charge rate can be useful to better apprehend the mechanism linking them.
267 Power law scaling relations with different power indices have been proposed
268 based on various data sets and are reviewed by Ichihara (2016). These rela-
269 tions have been extensively studied at PdF (Battaglia et al., 2005; Coppola
270 et al., 2009; Hibert et al., 2015). It is suggested that these relations depend
271 on the style of activity and that tremor may be associated with both fric-
272 tional dissipation of the flow inside the conduit and rate of burst and size of
273 bubbles. In the continuity of these works, we also report some relationships
274 between lava discharge rate, tremor amplitude and frequency content for a
275 certain number of eruptions occurring in the 2014-2021 time period.

276 **3. Data and Methods**

277 *3.1. Seismic data analysis*

278 We based our analysis on the seismic network covariance matrix approach
279 allowing an ensemble representation of inter-stations cross-correlations as
280 suggested by Seydoux et al. (2016a,b, 2017). We use the vertical component
281 records of the seismic network stations in the vicinity of PdF (PF network,
282 <https://doi.org/10.18715/reunion.PF>). The data are stored in 24-hr-
283 long time series in mseed format and downsampled from 100 to 20 samples
284 per second. We apply demeaning, linear detrending, and bandpass filtering
285 between 1 and 10 Hz. Then, some spectral whitening is applied. This data
286 analysis was performed using the CovSeisNet software (available at: <https://github.com/covseisnet/covseisnet>).

288 We first describe the covariance matrix approach that is well suited for
289 the analysis of continuous tremor-like signals. We then present a template-
290 matching approach to detect change from continuous to discrete tremor.
291 Those two methods are applied to the 23 eruptions of our tremor catalog.
292 We finally applied a template-matching and relocation method to the long-
293 duration July-August 2017 eruption in order to better characterize its corre-
294 sponding co-eruptive seismicity composed of different impulsive earthquakes
295 signals.

296 *3.1.1. Detection, location, classification and characterization of tremors based*
 297 *on the covariance matrix analysis*

298 *Covariance matrix and its eigenvalues: distinguishing seismic tremor from*
 299 *seismic ambient noise.*

300

301 According to Seydoux et al. (2016a), the covariance matrix $\mathbf{C}(f)$ is es-
 302 timated in an averaging time window of length Δt , from the time average
 303 of the Fourier cross-spectra matrices computed over a set of M overlapping
 304 subwindows of length δt , such as:

$$\mathbf{C}(f) = \langle \mathbf{u}(f)\mathbf{u}^\dagger(f) \rangle_{\Delta t} = \frac{1}{M} \sum_{m=1}^M \mathbf{u}_m(f)\mathbf{u}_m^\dagger(f), \quad (1)$$

305 with $\mathbf{u}_m(f)$ the data Fourier spectra vector in the subwindow m for the entire
 306 network.

307 We applied this calculation on consecutive averaging windows over the
 308 whole 2014-2021 time period, setting $M = 20$, $\delta t = 60s$ and the overlapping
 309 ratio to 50%, to obtain 10-min-long time windows.

310 Once estimated, this matrix can be decomposed on the basis of its com-
 311 plex eigenvectors \mathbf{v}_n associated with real positive eigenvalues λ_n :

$$\mathbf{C}(f) = \sum_{n=1}^N \lambda_n(f)\mathbf{v}_n(f)\mathbf{v}_n^\dagger(f). \quad (2)$$

312 Then eigenvalues are arranged in decreasing order and the covariance
 313 matrix spectral width is computed as a function of frequency as per Seydoux
 314 et al. (2016a):

$$\sigma(f) = \frac{\sum_{i=1}^N (i-1)\lambda_i(f)}{\sum_{i=1}^N \lambda_i(f)}. \quad (3)$$

315 This spectral width metric corresponds to the width of the covariance
 316 matrix eigenvalues distribution and allows a compact time-frequency repre-
 317 sentation of the wavefield as it is recorded by the ensemble of stations com-
 318 posing the network. A low spectral width value indicates a tight eigenvalues
 319 distribution, meaning that the wavefield is dominated by one localized source
 320 generating coherent waves. On the contrary, a large spectral width value re-
 321 flects a wavefield composed by multiple well-distributed sources generating
 322 incoherent waves. Therefore, the resulting spectral width value can be seen

323 as a proxy of the number of independent sources composing the wavefield. It
324 is thus effective in identifying temporal windows containing seismic energy
325 emitted by a spatially well-located source (e.g. tremor), contrasting from
326 windows dominated by distributed ambient seismic noise.

327 The parameter settings for calculating the covariance matrix depend on
328 the signal of interest. However, as a general rule, the number of overlapping
329 subwindows M should be at least equal to the number of stations of the net-
330 work, to ensure statistically robust estimation of the covariance matrix. In
331 addition, the duration of this sub-window δt , must be long enough to allow
332 the wave to propagate through the entire network. So, with its 20 stations
333 and aperture of around 15 km, the configuration of the PdF seismic network
334 requires the large averaging time windows Δt to be at least a few minutes.
335 This enables us to achieve satisfactory temporal resolution for the study of
336 seismic tremor. However, for the study of co-eruptive impulsive earthquakes,
337 we need a higher temporal resolution, and we therefore decide to apply a
338 template matching method. It is indeed important to apply complementary
339 detection methods together for a broader illumination of the different mech-
340 anisms at play during eruptions producing the large range of seismic signals
341 recorded by the seismic network.

342 *Locating seismic tremor using inter-station cross-correlations differential travel*
343 *times.*

344

345 When a window containing seismic energy associated with a well-defined
346 source is detected, the corresponding source can be located using the dif-
347 ferential travel times contained in inter-station cross-correlations in order to
348 “back-project” the signals energy toward its source via a grid-search approach
349 (Ballmer et al., 2013; Droznin et al., 2015). The source location quality can
350 be altered from scattering due to local heterogeneities and from imprecision
351 of the velocity model described hereafter. In order to mitigate these effects,
352 we apply a smoothing operation to derive the smooth envelope $S^{i,j}$ of the
353 cross-correlations waveforms. To do so, we compute the absolute value of the
354 analytic signal derived from the Hilbert transform, that we convolve with a
355 Gaussian window using a 1.5 s width following Journeau et al. (2020). For
356 every grid point $\mathbf{r} = x\mathbf{e}_x + y\mathbf{e}_y + z\mathbf{e}_z$, each inter-station cross-correlation
357 envelop $S^{i,j}$ is shifted by the time difference needed for the wave to travel
358 from the tested source to the two stations i and j . We finally compute the
359 function $R(\mathbf{r})$ by stacking at zero lag time the value of the shifted envelopes

360 for all stations pairs as:

$$R(\mathbf{r}) = \sum_{i=1}^N \sum_{j=i+1}^N S^{i,j} [t^i(\mathbf{r}) - t^j(\mathbf{r})] \quad (4)$$

361 with N the number of stations and $t^i(\mathbf{r})$ the travel time between the
362 tested source and the station i . We compute those travel times using the
363 Taup package of the ObsPy python toolbox (Beyreuther et al., 2010). The
364 method for ray tracing and travel time computation is explained in Buland
365 and Chapman (1983) and was originally implemented as the Taup toolkit
366 by Crotwell et al. (1999). Topographic effects are considered by taking into
367 account the seismic station elevations when computing the travel times. As
368 we explore the full depth range extending from the volcano summit to 5.6
369 km below, we implement our 3D back-projection based on a body-wave as-
370 sumption, considering that the tremor wavefield is dominated by shear waves.
371 Therefore, we use a 1D S-wave velocity model of PdF derived by Mordret
372 et al. (2015), represented in Figure S1. This function $R(\mathbf{r})$ can be interpreted
373 as a spatial likelihood function that defines the most probable location of a
374 seismic source in a particular position of our 3D grid having a 100 m x 100
375 m x 100 m spatial resolution.

376 This tremor location workflow was systematically applied to all co-eruptive
377 periods as illustrated in Figure S2. Setting the covariance matrix computa-
378 tion parameters to $M = 20$ and $\delta t = 20s$, we obtain its estimation over
379 200s-long overlapping windows. We then filter the covariance matrix by
380 its 1st eigenvector as suggested by Soubestre et al. (2019) before applying
381 the back-projection grid-search approach. This filtering removes the noise
382 spanned by higher-order eigenvectors and thus to focus on the dominant
383 tremor source. We then retrieve the filtered inter-station cross-correlations
384 in the time domain and stack them for five consecutive windows. We back-
385 project the envelopes of those filtered cross-correlations in our 3D grid-search
386 using the 1D S-wave velocity model described above. Finally, we extract the
387 3D tremor coordinates maximizing the resulting spatial likelihood function
388 computed in 10 min-long time windows.

389 Assessing the accuracy of our inter-station cross-correlation location method
390 can be done in two ways. It is first possible to measure the distance between
391 the hypocenter of an earthquake estimated using the P and S wave arrival
392 times, and the coordinates of the maximum of the spatial likelihood function
393 computed for a time window containing this earthquake. While the resulting

394 average misfit is about 7 km using DLP earthquakes located around 30 km
 395 below the Klyuchevskoy volcano (Journeau et al., 2022), the misfit is about
 396 2 km or less using shallower VT hypocenters below the Izu-Oshima volcano
 397 (Permana et al., 2019). Then, we can also measure the spatial distribution
 398 of the highest values of the tremor spatial likelihood function above a certain
 399 threshold, that we fix here to 98%. Journeau et al. (2022) shows an increase
 400 of the average vertical 98% error from 2-3 km for shallow tremor sources to
 401 about 10 km towards 30 km depth. We measured this vertical error for the
 402 first 10-min window of each Pdf eruptions analyzed in this study, and we
 403 obtained an averaged error of 0.6 km. We illustrate this in Figure S3 show-
 404 ing examples of the 98% contour values of the spatial likelihood functions for
 405 four different eruptions.

406 We note that the accuracy of our location method based on the back-
 407 projection of inter-station cross-correlations is sufficient for the study of seis-
 408 mic tremor. However, we need better spatial resolution for the study of
 409 co-eruptive impulsive earthquakes. We therefore decided to apply a double-
 410 difference algorithm to precisely re-locate earthquakes detected by template
 411 matching.

412 *Classifying seismic tremor combining spectral width and network response*
 413 *function metrics.*

414
 415 The network response function (NRF) is another covariance matrix metric
 416 designed to evaluate the level of focusing of the spatial likelihood function.
 417 We first normalize $R(\mathbf{r})$ by its sum over all grid points:

$$\bar{R}(\mathbf{r}) = \frac{R(\mathbf{r})}{\sum_{\mathbf{r}} R(\mathbf{r})} \quad (5)$$

418 and then add a temporal normalization by the maximum over the total time
 419 period:

$$\tilde{R}(\mathbf{r}) = \frac{\bar{R}(\mathbf{r})}{\max_t(\bar{R}(\mathbf{r}))}. \quad (6)$$

420 We defined $NRF(t)$ as the maximum of $\tilde{R}(\mathbf{r})$, which is expected to be high
 421 for well localized sources and low for a random incoherent wavefield.

422 The comparison of the resulting NRF with $\bar{\sigma}$, the spectral width function
 423 averaged in a given frequency band ([0.5-5] Hz), defines a plane where all
 424 the detected signals can be classified between continuous tremor-like signals

425 and impulsive earthquake-like signals thanks to their contrasting wavefield
426 characteristics measured by the seismic network (Journeau et al., 2020, 2022).
427 We show the resulting wavefield classification for the 2014-2021 period at PdF
428 in Figure S4.

429 Note that the red lines in Figure S4 are manually defined to visually
430 separate the different detection groups observed in the 2D NRF-SW space.
431 We consider a temporal window as a detection if $\bar{\sigma} < 0.78$ and $NRF > 0.2$.
432 Then, we defined the oblique red line following this equation: $SW_{tremor} =$
433 $-0.6 \times NRF + 0.9$, and a detection is classified as tremor if $\bar{\sigma} < SW_{tremor}$, or
434 as earthquake if $\bar{\sigma} > SW_{tremor}$. Tremor-earthquake detections follow such a
435 distribution in this NRF-SW plane because, compared to earthquake, tremor
436 sources emit more energy in the [0.5 - 5] Hz frequency band where $\bar{\sigma}$ had been
437 averaged (leading to lower $\bar{\sigma}$ values for tremor), and tremor wavefield back-
438 projection is less well-focused (leading to lower NRF values for tremor). In
439 order to provide an automatic wavefield classification scheme effective for
440 monitoring purposes, future work would be required taking advantage of
441 machine learning methods.

442 *Seismic tremor characterization from its time-frequencies properties and cross-*
443 *spectra phases.*

444
445 As shown in Soubestre et al. (2021), the detailed analysis of the spectral
446 width function corresponding to a detected tremor source enables to follow
447 the time-frequency properties of the associated tremor signal.

448 Moreover, it has been shown that the inter-station cross-correlation wave-
449 form is very sensitive to modifications of the tremor source mechanism or
450 change of its position and can be used as a “fingerprint” of the tremor source
451 (Droznin et al., 2015). Therefore, this information encoded in the phase
452 of the cross-correlation waveforms is complementary to the spectral content
453 information obtained from the eigenvalue-based analysis described above.
454 Droznin et al. (2015) computed a Pearson correlation coefficient between dif-
455 ferent consecutive windows of single-pair cross-correlation as a measure of
456 their waveform similarity. Soubestre et al. (2018) extended this approach to
457 the whole network using the first eigenvector of the covariance matrix as a
458 fingerprint. These vectors being complex, they defined the correlation coef-
459 ficient at each frequency as the absolute value of the complex scalar product
460 normalized by the respective norms. They finally obtained a 2D time-time
461 matrix containing all the correlation coefficients values by computing an av-

462 erage correlation coefficient in a given frequency band ([0.5-5] Hz in our case).

463 We apply this method for the December 2020 PdF eruption, using same
464 parameterization as the one used for the spectral width computation ($M =$
465 20 , $\delta t = 60s$) in order to compare the results from these two methods and to
466 characterize the seismic tremor and its evolution during the eruption.

467 3.1.2. *Template-matching for detecting the discrete “gas-pistons” signals*

468 We have developed a new approach to detect the transition between con-
469 tinuous tremor and intermittent tremor composed of individual gas-pistons
470 appearing during the final phases of eruptions. Namely, we use an automatic
471 single station three-component template-matching method for gas-pistons
472 detection and counting. It is based on the fact that each of these observed
473 seismic events (lasting approximately 10-s) are very similar between each
474 other and, therefore, behave as multiplets.

475 Selecting a 15-s template and performing a classical template-matching
476 method scanning the continuous data, we can detect additional gas-piston
477 events in the sequence. However, we also observe in some cases that frequency
478 content of these gas-piston events evolves with time. Therefore we adjusted
479 the basic template-matching procedure in order to not have to manually
480 select a single template.

481 As a first step, we select the closest seismic station to the active eruptive
482 vent to enhance signal-to-noise ratio. We then cut 1-day data stream into
483 30-min-long time windows, each of them being sub-divided into 15s long seg-
484 ments overlapping at 10% of their size, obtaining $N=11901$ subwindows. We
485 compute correlation coefficient between all 15-s subwindows of the large 30
486 min window and therefore end up with a 2D matrix ($N \times N$) containing the
487 correlation coefficient values. These operations are done for the 3 compo-
488 nents of the selected seismic station and we stack the resulting correlation
489 coefficient matrices.

490 Fixing a threshold for a correlation coefficient value of 0.4, for each 15-s
491 subwindow, we can count the number of similar 15-s events occurring within
492 the long 30-min window and thus obtain a 1D function quantifying the num-
493 ber of similar events in the 30-min window. Figure S5 summarizes all of
494 these operations.

495 Finally, we smooth in time the 1D function and we only count the number
496 of maxima of this function to obtain a proxy of the number of individual
497 gas-piston events in a given time window. We systematically apply this
498 method over the 2014-2021 time period to better constrain the transition

499 from continuous tremor toward the gas-piston phenomenon.

500 In addition to the tremor properties mentioned above, we also track the
501 seismic tremor amplitude using the RSAM (raw seismic amplitude measure-
502 ment) (Endo and Murray, 1991).

503 *3.1.3. Detection and relocation of co-eruptive earthquakes*

504 We also track the co-eruptive impulsive seismic signals that can be of
505 different types and generated in different locations of the shallow system.
506 We detect these earthquakes with a template matching approach following
507 the same approach as Lengliné et al. (2016) and Duputel et al. (2019). We
508 use 25 templates signals of 5.12 s starting 1 s before the P-wave arrival with
509 a large signal/noise ratio during the considered period. Seismic waveforms
510 and spectra of these templates are shown in Figure S6. Template matching
511 is conducted using three 3-component stations in the vicinity of the eruptive
512 site (FOR, VIL and CSS, see Figure 1 for location). The detection threshold
513 is set from a fixed probability of false detection of 10^{-5} per year per template,
514 following Lengliné et al. (2021). We show four seismic waveforms examples of
515 newly detected earthquakes in Figure S7. The detected earthquakes are then
516 re-located using a double-difference relative location algorithm (Waldhauser
517 and Ellsworth, 2000).

518 We thus detect and re-locate co-eruptive earthquakes during the July-
519 August 2017 eruption as an example.

520 *3.2. Estimation of lava discharge rate from satellite data*

521 Once the effusive phase of an eruption has started, the quantitative eval-
522 uation of the lava flow at the eruptive event, which we will refer to as TADR
523 (Time averaged Discharge Rate), becomes available. Knowledge of the lava
524 flux time evolution throughout an eruption is important to better under-
525 stand its shallow dynamics, and better interpret the corresponding tremor
526 observations.

527 For our study, we considered TADR data measured from thermal data ac-
528 quired by two systems: MIROVA (Coppola et al., 2009, 2020) and HOTVOLC
529 (Gouhier et al., 2016). The MIROVA system (Middle Infrared Observation of
530 Volcanic Activity) - an automatic volcano hot spot detection system based
531 on the analysis of MODIS data (Moderate Resolution Imaging Spectroradi-
532 ometer) - provides approximately 2-4 images daily at a resolution of 1km.
533 HOTVOLC - a web-based satellite-data-driven monitoring system using data

534 from the geostationary satellite Meteosat - provides images with higher sam-
535 pling rate (around 15 min) but with less precision.

536 We selected ten eruptions (May 2015, August-October 2015, September
537 2016, January-February 2017, July-August 2017, April-June 2018, September
538 2018, February-March 2019, April-May 2021 and December 2021 - January
539 2022) for which robust TADR measurements are available. We then use
540 these 10 eruptions as a sub-database for which we also look at geodetic and
541 petrologic data in addition to the seismic data previously introduced.

542 *3.3. Volcanic edifice deformation from GNSS baselines measurement*

543 We track the time evolution of the volcanic edifice deformation by com-
544 puting the baselines, i.e., the distances between summit stations DSRG and
545 SNEG (see location on Figure 1). The increase or decrease of these baselines
546 over time can highlight the inflation or deflation of the volcanic edifice, po-
547 tentially linked to a replenishment of the shallow magmatic reservoir or its
548 emptying, respectively. Thus, the monitoring of these baselines allows us to
549 have a simple proxy of the state of the shallow system during the eruption.
550 Daily GNSS solutions have been computed by OVPF-IPGP with the GipsyX
551 precise point positioning (PPP) software (Bertiger et al., 2020).

552 *3.4. Insights into PdF dynamics from lava products analysis*

553 We then compare the information drawn from the previously introduced
554 geophysical data with that obtained from the analysis of the lava. This
555 allows us to have a more complete vision of the eruptive dynamics at the
556 PdF, such as the H_2O degassing depth and the original depth of erupted
557 magma (shallow reservoir or deeper).

558 *3.4.1. Gas volume fraction estimation from inclusions*

559 We use the Equation 7 defined in Villemant and Boudon (1998) to obtain
560 an estimate of the percentage of gas volume fraction (χ) in the melt as a
561 function of the depth during its ascent, reflecting the amount of exsolved gas
562 bubbles.

$$\chi(\%) = \frac{K_R \times \Delta X_{H_2O}}{P}, \quad (7)$$

563 where $\Delta X_{H_2O} = X_{H_2O}(P_0, T_0) - X_{H_2O}(P, T)$, with $X_{H_2O}(P, T)$ the dissolved
564 H_2O content in magma at pressure P and temperature T , and $X_{H_2O}(P_0, T_0)$
565 the initial water content at the shallow magma reservoir level, setting $T_0 =$

566 $1100^{\circ}C$ and $P_0 = 49MPa$ corresponding to a depth of 2000 m below the
 567 surface (using $P_0 = \rho_s gz$, with $\rho_s = 2500kg/m^3$ the medium density and
 568 $g = 9.81m/s^2$ the gravity acceleration). We set the constant value $K_R = 15$,
 569 corresponding to a magma temperature $T = 1100^{\circ}C$ and a magma density
 570 $\rho_s = 2500kg/m^3$.

571 Given that CO_2 has already more or less completely degassed at this
 572 depth, the dominant gaseous species in the shallow system is H_2O (Sparks
 573 (2003); Di Muro et al. (2016)). We therefore only consider the degassing of
 574 dissolved H_2O as a contribution of bubble fraction in rising magma.

575 The amount of dissolved H_2O in the PdF magma is inferred from the
 576 analysis of primary melt inclusions hosted in olivines (Di Muro et al. (2016))
 577 using the Equation 8 fitting a population of 130 MIs (Melt Inclusions) anal-
 578 yses:

$$X_{H_2O}(P, T)(wt\%) = C1 \times \log(P) + C2 \times e^{C3 \times T} + C4 \quad (8)$$

579 where $C1 = 0.1759$, $C2 = 4.019e - 31$, $C3 = 0.0556$, $C4 = 0.164$ are con-
 580 stants and $T = 1100^{\circ}C$ is the magma temperature that we consider constant,
 581 assuming that the magma does not cool down significantly during its ascent
 582 towards the eruptive site.

583 3.4.2. Magma chemical composition estimation from petrologic data

584 A final piece of information that we wish to compare with the seismic ob-
 585 servations is the chemical composition of the eruptive products from analysis
 586 of major element compositions in bulk rocks (Coppola et al., 2017; Gurioli
 587 et al., 2018; Vlastélic et al., 2018; Thivet et al., 2020a,b). We focus in partic-
 588 ular on evolution of MgO (wt%) content of these inclusions, which is propor-
 589 tional to melt temperature and crystal accumulation, and CaO/Al_2O_3 ratio,
 590 which is proportional to melt temperature and evolution (Di Muro et al.,
 591 2016). At PdF, the phases of high lava output rate are usually associated
 592 with both hotter melts and a higher content of crystals (Coppola et al., 2017).

593 4. Results

594 4.1. Co-eruptive seismic tremor catalog

595 First, we systematically applied the covariance matrix method described
 596 in the previous section to the continuous seismic data of the PdF network
 597 throughout the 2014-2021 time period. We show the resulting 8-years time

Table 1: List of the 23 eruptions that occurred in the 2014-2021 time period (and December 2009 eruption) and specification of their characteristics. *Note that the terminal phases of the eruption of August-October 2015 are not associated with gas-piston. The gas-pistons are observed only between the phases of renewed eruptive activity (see Figure S16).

Eruption timing (dd.mm.yyyy hh:MM)	Duration (days)	Distance from summit (m)	Gas-piston
14.12.2009 14:40 - 15.12.2009 06:30	0.7	525	yes
20.06.2014 21:35 - 21.06.2014 17:09	0.8	570	yes
04.02.2015 06:50 - 15.02.2015 18:30	11.5	770	yes
17.05.2015 09:45 - 30.05.2015 16:50	13.3	1100	yes
31.07.2015 05:20 - 02.08.2015 07:15	2.1	2000	yes
24.08.2015 14:50 - 31.10.2015 04:02	58.3	1135	no*
26.05.2016 04:05 - 27.05.2016 07:40	1.1	2900	yes
11.09.2016 04:41 - 18.09.2016 00:18	6.8	1425	no
31.01.2017 15:40 - 27.02.2017 15:30	27.0	2000	yes
13.07.2017 20:50 - 27.08.2017 23:00	45.1	1700	yes
03.04.2018 06:40 - 04.04.2018 00:00	0.7	3450	yes
27.04.2018 19:50 - 01.06.2018 10:30	34.6	1550	yes
12.07.2018 23:30 - 13.07.2018 18:00	0.8	1680	yes
15.09.2018 00:25 - 01.11.2018 00:00	47.0	2010	yes
18.02.2019 05:48 - 10.03.2019 18:00	19.1	2500	no
11.06.2019 02:35 - 13.06.2019 08:00	2.2	500	yes
29.07.2019 08:00 - 30.07.2019 00:30	0.7	2250	yes
11.08.2019 12:20 - 15.08.2019 00:20	4.1	3500	yes
25.10.2019 10:40 - 27.10.2019 12:30	2.1	6250	yes
10.02.2020 06:50 - 16.02.2020 10:12	6.1	600	yes
02.04.2020 08:20 - 08.04.2020 09:30	4.2	1940	no
07.12.2020 00:40 - 08.12.2020 03:15	1.1	1700	yes
09.04.2021 15:00 - 23.05.2021 22:00	44.3	2771	yes
21.12.2021 23:30 - 16.01.2022 22:10	26	1700	no

598 series of covariance matrix spectral width in Figure 2, highlighting the seismic
599 signature of the 23 eruptions (see Table 1) and to characterize their spectral
600 properties.

601 Tremor can be separated from other seismic signals computing the NRF
602 and comparing it to the 1D averaged spectral width function (as explained
603 in section 3). We show in Figure S4 the resulting distribution of all analyzed
604 temporal windows in the NRF versus spectral width plane. Windows located
605 in the “noise” area are characterized by a low spatial coherence level with
606 respect to the “tremor” and “earthquakes” areas for which corresponding
607 windows are defined such as “detection”.

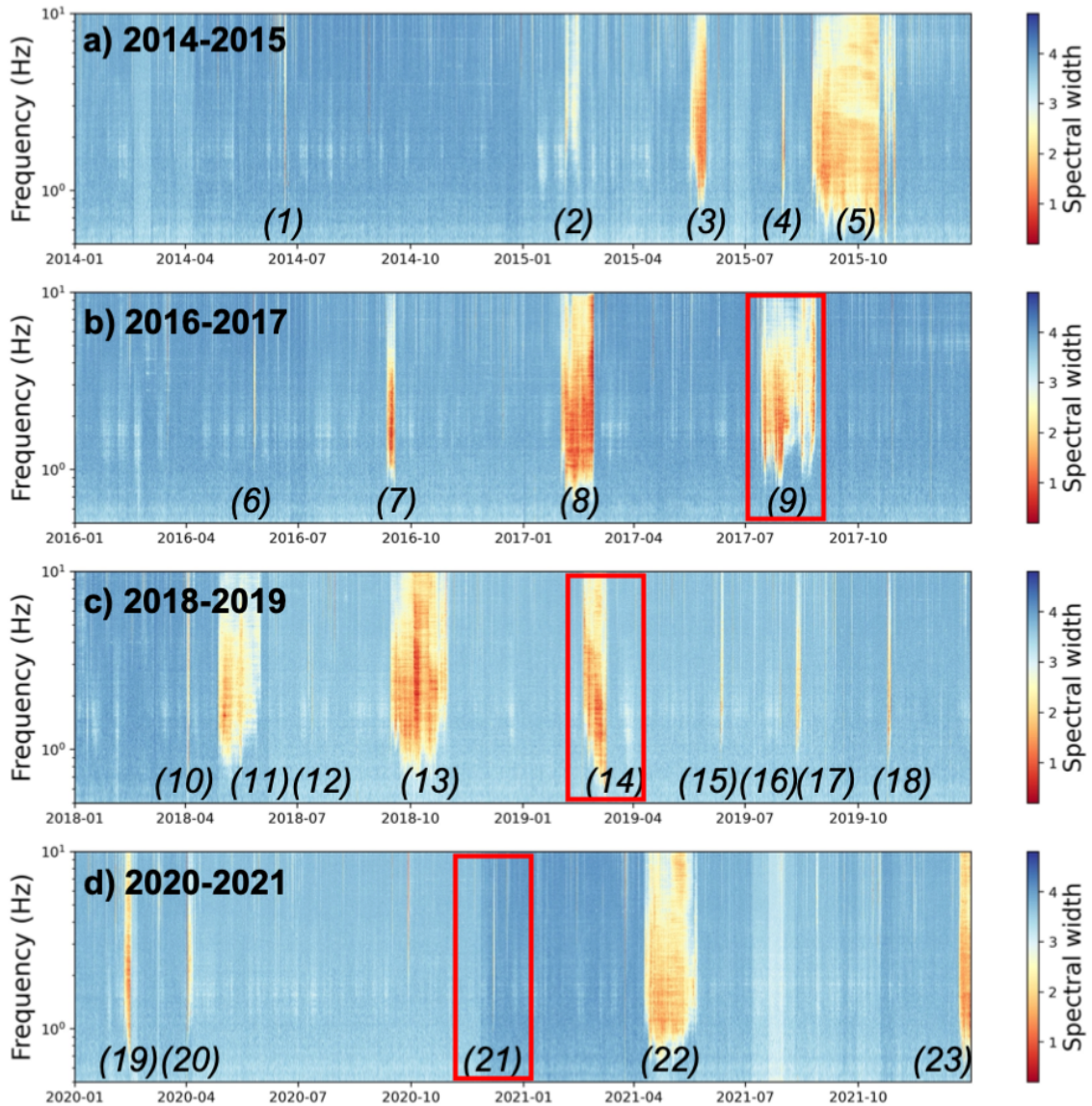


Figure 2: **Co-eruptive tremor seen as high spatial coherence seismic signals.** Seismic network covariance matrix spectral width computed for the a) 2014-2015, b) 2016-2017, c) 2018-2019, d) 2020-2021 time periods. The numbers from 1 to 23 correspond to the 23 eruptions that took place in this time period. Red rectangles highlight three selected eruptions for which we show zoom of their spectral width in Figure 4 (December 2020), Figure 6 (July-August 2017) and Figure 11 (February-March 2019).

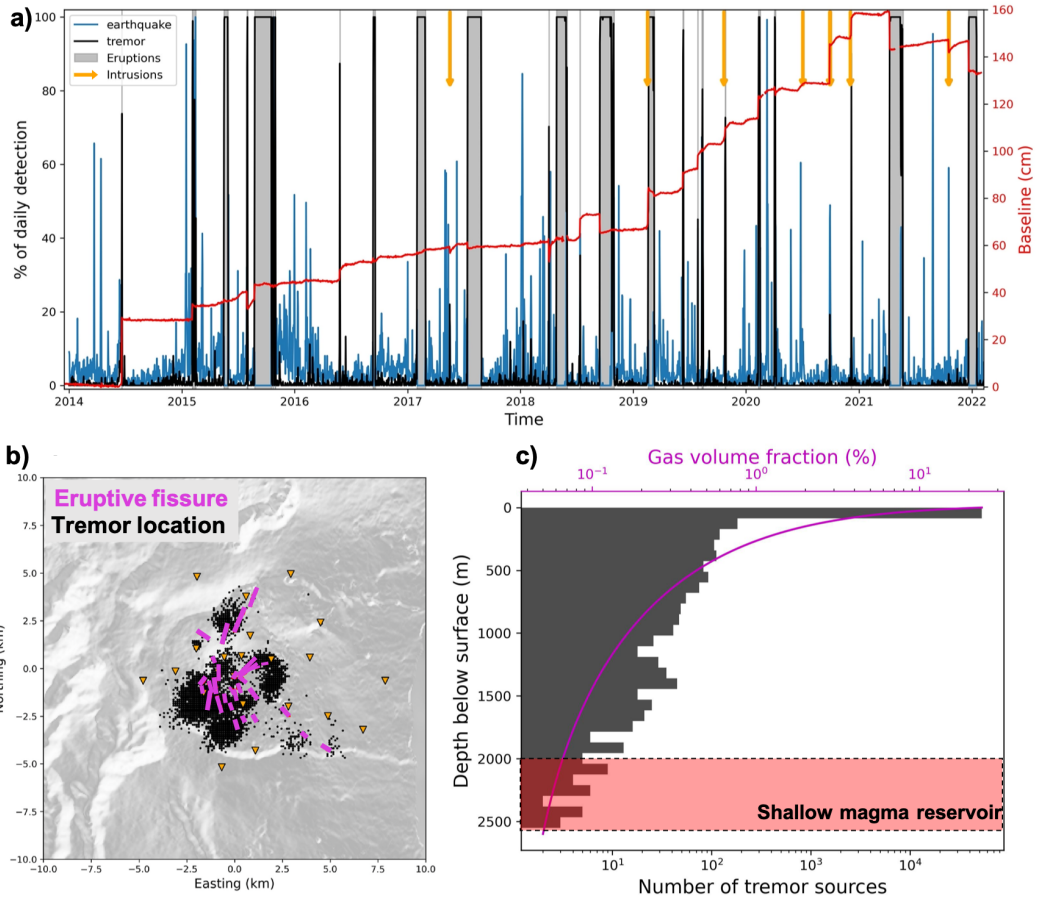


Figure 3: **Shallow co-eruptive tremor marks strong magma degassing beneath eruptive sites.** a) Temporal evolution of daily detection percentage of tremor (in black) and earthquake (in blue) signals. Grey rectangles highlight the eruptive time periods while vertical downward orange arrows stand for intrusions (i.e. magma injection that did not reach the surface). The red curve represents for the DSRG-SNEG GNSS stations baseline time series. b) Tremor sources locations corresponding to the 23 eruptions that occurred in the January 2014 - February 2022 time period. Black dots stand for the tremor source position and magenta lines for the eruptive fissures. Orange triangles indicate the seismic stations used during the cross-correlation back-projection procedure allowing tremor source location. c) Histogram of tremor source depths (logarithmic scale). The magenta curve indicates the modeled amount of exsolved gas from the magma as a function of depth during the magma ascent from the shallow reservoir, whose depth is indicated with a red rectangle.

608 We show in Figure 3a the temporal evolution of the daily percentage
609 of detection. Starting with 860,860 10-min-long overlapping windows, we
610 obtain 103,446 tremor detections and 41,415 earthquake detections. Our
611 analyzed time period span 23 eruptions corresponding to a total of 358.7
612 days of eruptions. In comparison, our automatic method detects in total
613 345.8 days of tremor during co-eruptive periods, or 96.4 % of the cumulative
614 eruptive time in our considered time period. The eruptive periods without
615 tremor detection (about 12 days in total) might be linked to phases of low
616 level of volcanic activity with weak surface manifestations due to low amount
617 of gas in the conduits. Indeed, such configuration might in turn generate a
618 tremor source too weak to exhibit strong coherence across the 15-km network,
619 which consequently falls below the detection threshold defined by our metrics
620 derived from the network covariance matrix.

621 In contrast, 4.3 % of our total number of tremor detection occur during
622 inter-eruptive periods. These windows classified as tremor outside eruptions
623 may be associated with periods of high swell, cyclone passage, or volcanic
624 tremor-like signals previously unidentified. More detailed and specific studies
625 of these windows will be required in the future.

626 Figure 3b and 3c show the epicenters and depths of the located co-eruptive
627 tremor sources, respectively. We observe that co-eruptive tremor sources are
628 systematically located close to eruptive sites and mainly concentrated a few
629 hundreds of meters beneath the surface (98 % of tremor sources are located
630 between 200 m depth and the surface). This shallow level beneath erup-
631 tive sites correspond approximately to the depth range where gas exsolution
632 accelerates strongly going from about 1% to more than 20% of gas volume
633 fraction in a few hundred of meters. Such correlation between the depth
634 profile of gas volume fraction and number of tremor source suggests that the
635 mechanism of tremor generation is strongly controlled by the shallow magma
636 degassing when it approaches the surface and the eruptive vent.

637 *4.2. Time-frequency properties of co-eruptive seismic tremor*

638 We selected as examples two eruptions whose details are presented in
639 this section. We focus on the covariance matrix spectral width time evolu-
640 tion of a short-lived eruption (December 2020) and a long-lived eruption
641 (July-August 2017). We show how tracking their time-frequency properties
642 help to infer changes in the underlying degassing regimes. We can therefore
643 compare the resulting eruptive dynamics evolution for those two different
644 scenario. While the short-lived December 2020 eruption involves simpler

645 eruptive phases transitions, the July-August 2017 one is more complex. For
646 this latter, we compare its tremor properties variations with other obser-
647 vations presented previously. We finally describe the February-March 2019
648 eruption to illustrate an example of an end-of-eruption characterized by an
649 effusive paroxysm contrasting with the more usual gas pistons events.

650 *4.2.1. Short duration December 2020 eruption*

651 The December 2020 eruption lasted approximately one day between 07.12.2020
652 (00:40) and 08.12.2020 (03:15). We show in Figure 4a-c the time evolution
653 of the co-eruptive tremor spectral content, amplitude, state (when detected
654 seismic gas-pistons swarm marks transition from continuous to intermittent
655 tremor) and temporal stability (via first eigenvector similarity). We can thus
656 observe the non-stationarity of volcanic tremor during the eruption evolving
657 through different phases.

658 Two photos of the eruptive site are shown in Figures 4d,f, taken in two
659 distinct eruption phases (initial phase (1) and intermediate phase (2)) with
660 contrasting seismic tremor properties (Figure 4a-c). The image shown in
661 Figure 4d illustrates the Hawaiian lava fountain activity all along the opened
662 eruptive fissure, as it is usually observed during initial phases at PdF. The
663 onset of this phase was characterized by a large activated frequency band, in
664 particular with low frequencies excited below 1 Hz, associated with a high
665 tremor amplitude. Those lowest frequencies gradually ceased to be excited,
666 simultaneously with a tremor amplitude decrease.

667 A stabilization of tremor properties marked a transition in eruption phases.
668 Figure 4f at the end of this transition phase, reveals that the eruptive activity
669 has focused on a single vent. At that time, we observe a Strombolian eruptive
670 activity when the activated tremor frequency band starts to widen (without
671 the initial low frequencies), with small tremor amplitude fluctuations and
672 new phase of tremor stability (Figure 4a-c).

673 The tremor state can be considered as continuous during the initial phase
674 (1) and the beginning of the intermediate phase (2). However, at the end of
675 the second phase, we start to detect some discrete seismic gas-piston events
676 within the tremor. This marks a transition toward the final phase (3) in
677 which tremor spectral content evolves toward higher frequencies and tremor
678 amplitude decreases until reaching the background noise level. In this last
679 phase, no more lava effusive activity was observed at the surface, only inter-
680 mittent degassing occurred (as illustrated in Figure 4f for another eruption),
681 probably linked with each tremor burst detected as gas-piston events.

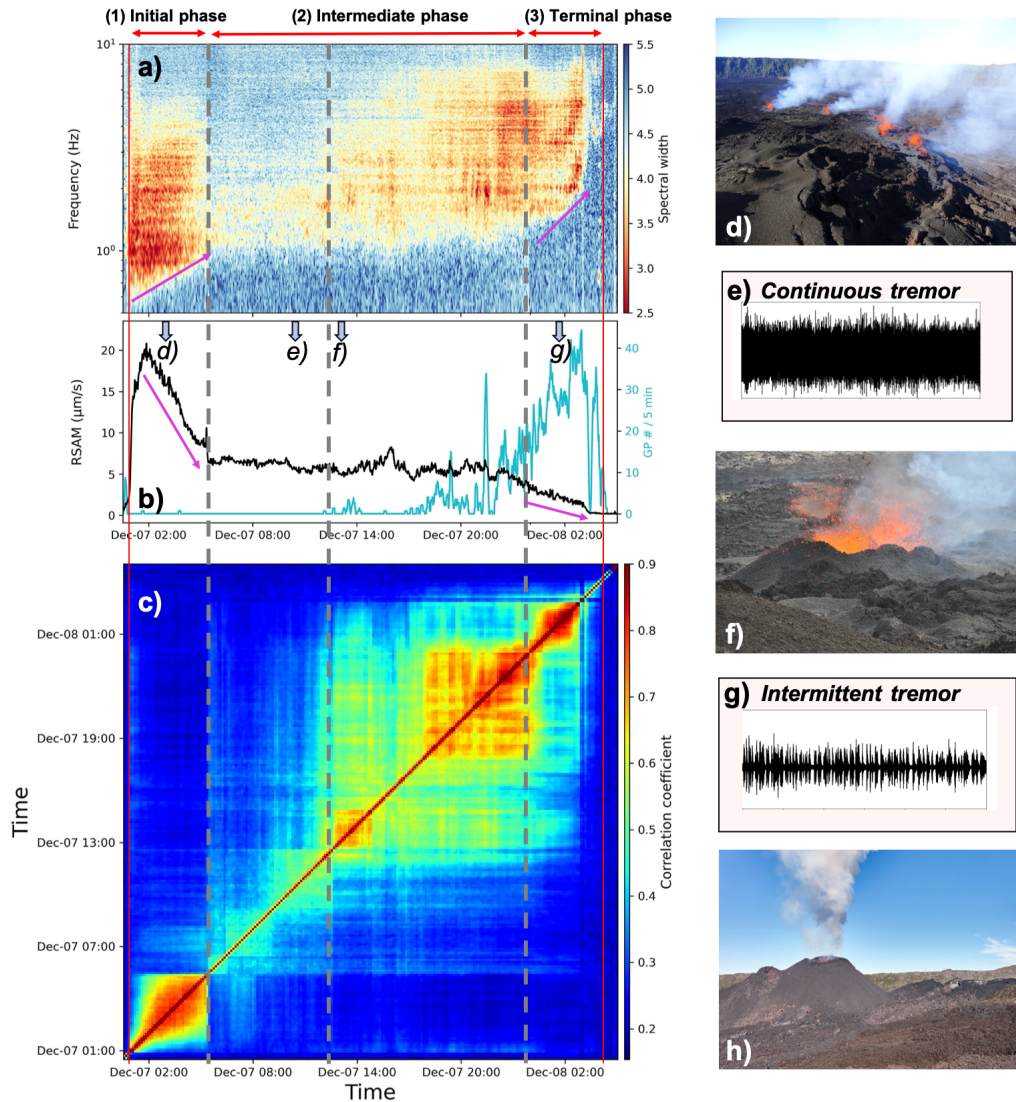


Figure 4: Shallow liquid-gas dynamic controls time-frequency co-ruptive tremor properties, example from the short-time December 2020 eruption. (Continued on the following page.)

682 The different magma degassing regimes associated with the eruption phases
 683 (1), (2), (3) and with changes in tremor characteristics are schematized in

Figure 4: (Continued caption.) a) Covariance matrix spectral width computed for 07.12.2020 and 08.12.2020. b) Black line represents tremor RSAM computed for the vertical component of the BOR station. Cyan line shows the result of the procedure of detection and counting of multiplets (seismic gas-piston events) in a final window of 5 minutes using BOR as a reference station. c) Matrix of correlation coefficients between covariance matrix's first eigenvector obtained by averaging all matrices computed in the [0.5 - 5] Hz frequency band. Vertical red lines in a), b) and c) indicate the beginning and the end of the eruption. The gray dashed lines indicate notable changes in tremor properties associated with changes of the temporal stability of the tremor interpreted as degassing regime transitions. Three identified phases ((1) initial, (2) intermediate and (3) final) are reported on top of a). Purple arrows in a) and b) highlight simultaneous increase in tremor spectral content and decrease in tremor amplitude, during the initial and final phases. d) Photo showing the opening of the eruptive fissure over approximately 500 meters with lava fountains. e) and g) are one hour seismograms recorded on the vertical component of BOR station, showing continuous and intermittent (composed of gas-pistons) seismic tremor in the initial and final phase of the eruption, respectively. f) Photo testifying the focus of the eruptive activity on a single vent with construction of a small eruptive cone and gradual transition toward a Strombolian degassing regime. Timing of photos d) and f) and seismograms zoom e) and g) are indicating in b) with light blue vertical downward arrows. h) Photo of a well-developed eruptive cone during a gas-piston degassing phase without lava effusion. It was taken during another eruption in October 2018 and we show it because such a photo was not available for the December 2020 eruption.

684 Figure 5a, using illustrations from Spina et al. (2019) and typical gas volume
685 fraction values from Vergnolle and Gaudemer (2015). Note that these gas
686 volume fraction values are only indicative and do not take into account the
687 potential accumulation of pre-eruptive bubbles in the shallow reservoir, nor
688 the significant expansion of bubbles very close to the surface.

689 It is known from several volcano geodesy studies that deformation induced
690 by magma transport through dykes can be modeled by rectangular
691 tensile cracks (Okada, 1985). We therefore decided to consider the shallow
692 dyke portion beneath eruptive sites as such and to interpret our observations
693 of tremor spectra fluctuations during the initial and final phases using a res-

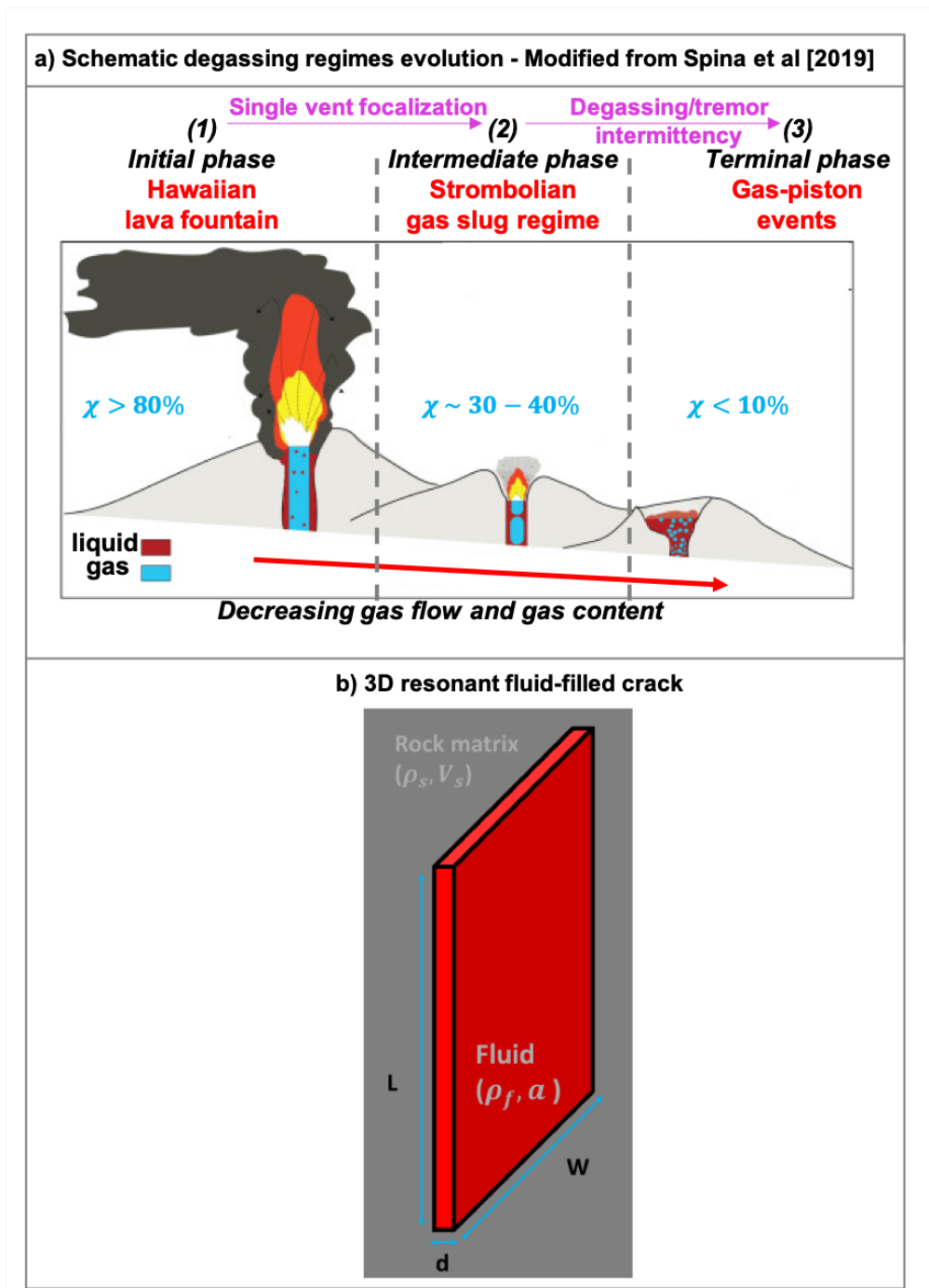


Figure 5: Schematic representation of different phases of magma degassing regimes and tremor spectra fluctuations interpretative framework. a) Figure modified from ²⁶Spina et al. (2019) showing evolution of degassing regimes while gas flow and gas content decrease over time. b) Schematic 3D fluid-filled crack with main parameters, namely crack geometry and magma properties, controlling the crack resonance frequencies.

694 onant fluid-filled crack model (Chouet, 1996; Maeda and Kumagai, 2017)
695 schematized in Figure 5b. Corresponding equations from Maeda and Kumagai
696 (2017) are detailed in the supplementary file.

697 The resonance frequency is mainly controlled by the crack geometry and
698 the properties of its filling magma, in particular its bulk modulus and sound
699 velocity associated with the gas volume fraction. The observed tremor fre-
700 quency increase in phases (1) and (3) can be linked to either a decrease in the
701 dyke dimension (associated with the focus of the eruptive activity on a single
702 vent, dyke closing, and/or decrease in resonating lava column length) or a
703 decrease in conduit gas content (explaining changes in the observed varying
704 degassing regimes).

705 *4.2.2. Long-duration and complex July-August 2017 eruption*

706 We represent in Figure 6 the temporal evolution of a set of 5 parameters
707 characterizing the dynamics of the 47-days-long July-August 2017 eruption.
708 In addition to seismic tremor properties (Figures 6a-b) that we present in the
709 previous section for the short December 2020 eruption, we also track in time
710 the lava discharge rate at the eruptive vent (Figure 6c) as well as the volcanic
711 edifice deformation and co-eruptive impulsive seismicity (Figure 6d). This
712 multi-disciplinary dataset enables to gain further insights into the functioning
713 of long-lived and more complex eruptions.

714 *Eruptive initial phase and focus of the activity at a single vent.*

715

716 Figures 7b-d represent a zoom of the initial phase of the July-August
717 2017 eruption lasting slightly more than one day, which is characterized by
718 gradual loss of low tremor frequencies excitation, and simultaneous decrease
719 of tremor amplitude and lava discharge rate.

720 As it was the case for the short December 2020 eruption, the images
721 of the eruptive site shown in Figure 7a illustrate distributed Hawaiian lava
722 fountains activity during the initial phase while the activity has focused on
723 a single vent at the end of this first phase, with the start of an eruptive cone
724 edification resulting from the accumulation of volcanic products.

725 We report in Figure 7e the initial minimum excited tremor frequency as
726 a function of the length of the eruptive fissure opened at the surface for the
727 23 eruptions of our catalog. From this distribution, a decrease of the tremor
728 frequency with an increase of the fissure length is observed, suggesting that
729 initial tremor spectra are controlled by a characteristic size over which shallow

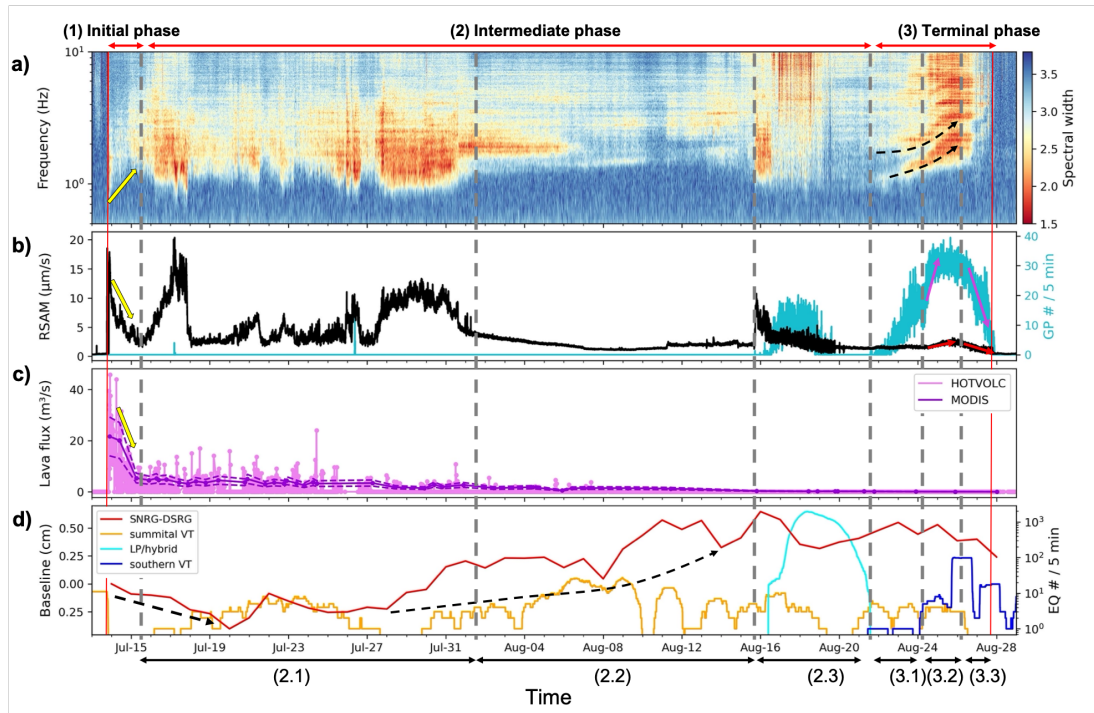


Figure 6: **Co-eruptive tremor fluctuations during the long-lived July-August 2017 eruption.** a) Covariance matrix spectral width as a function of time and frequency computed in 2-min-long time windows. The two dashed black arrows highlight some spectral gliding during the final phase.

b) Black curve represents the amplitude of the tremor estimated by calculating the RSAM on the FOR station (vertical component, filtered between 0.5 and 5 Hz). Cyan curve represents the number of detected gas-pistons in 5-min-long time windows. No gas piston events are detected during periods of continuous eruptive tremor. c) Lava discharge rate (TADR) data obtained by 2 different satellite platforms: MIROVA (purple line) and HOTOVOLC (light purple line). Yellow arrows during the initial phase in a), b) and c) highlight the cessation of lowest tremor frequencies activation, and the decrease of RSAM and TADR, respectively. d) Red line shows summit deformation illustrated by the daily baseline between 2 summit GNSS stations (DSRG and SNEG, see location on Figure 1). Orange, cyan and blue lines represent the number of detected and located summit VTs, LP/hybrids and southern VTs earthquakes, respectively. The dotted black arrows highlight a tendency of deflation of the edifice at the beginning of the eruption and inflation in the middle of the eruption. The two vertical red lines indicate the start and end of the eruption, while grey dashed vertical lines stand for notable changes in the eruptive dynamics.

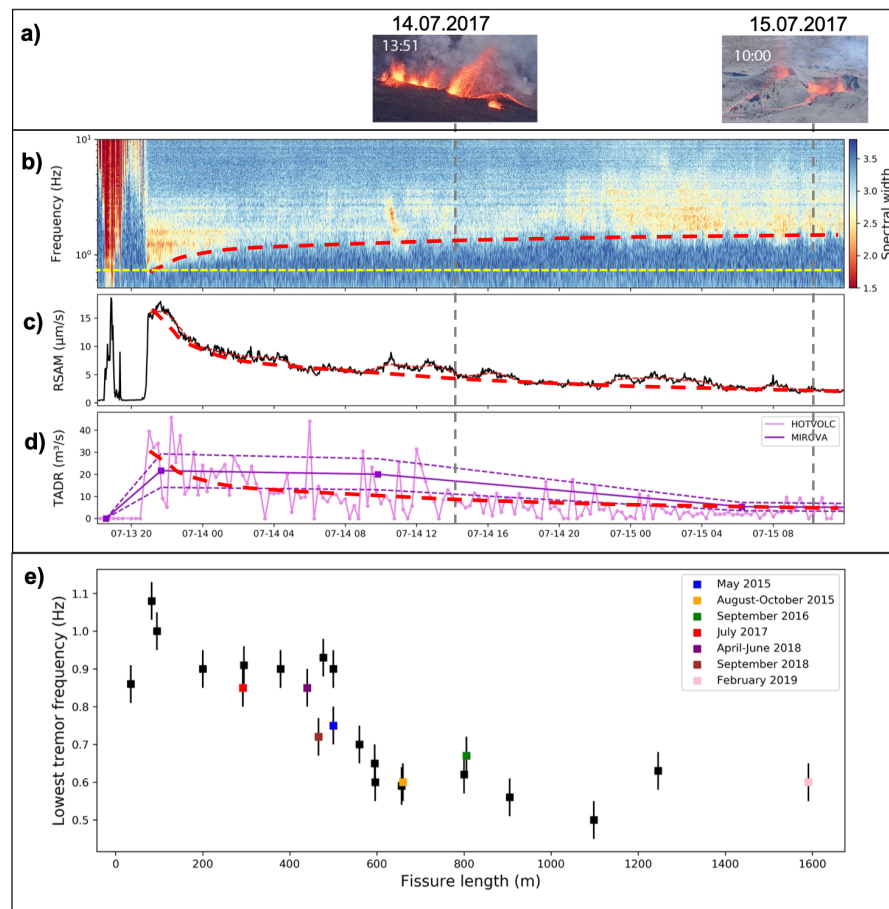


Figure 7: **Tracking seismic tremor changes while eruptive activity focus in a single vent during initial phase.** a) Two photos of the eruptive site showing lava fountains on the initial fissure (14.07.2017 - 13:51) and the eruptive cone, which started to build about 20 hours later (15.07.2017 - 10:00). These photos were taken during OVPF reconnaissance missions (© OVPF/IPGP). b) Covariance matrix spectral width as a function of time and frequency computed in 2-min-long time windows. The yellow horizontal dashed line indicates the minimum tremor frequency excited in the initial phase. c) Black curve represents the amplitude of the tremor estimated by calculating the RSAM on the FOR station (vertical component, filtered between 0.5 and 5 Hz). d) Lava discharge rate (TADR) data obtained by 2 different satellite platforms: MIROVA (purple line) and HOTVOLC (light purple line). The red curves in b), c) and d) represent interpretative trends following the gradual cessation of low frequencies excitation and the decreases in tremor amplitude and lava discharge rate. e) Minimum tremor frequency excited in the initial phase as a function of the total initial opened fissure length reported for the 23 eruptions of the 2014-2021 time period. Colored and labelled points indicate eruption for which TADR data are available and attest a lava discharge rate decrease during the initial phase.

730 degassing occurs, that might reflect the width of the dyke head arriving at the
731 shallow gas exsolution level. Then, the gradual increase of tremor frequency
732 content during initial phases can be linked to the progressive focus of the
733 activity on the eruptive fissure.

734 Selecting seven eruptions for which TADR data are available during their
735 initial phases (Figure S8a), we show in Figures S8b,c the relationships be-
736 tween tremor minimum excited frequency, tremor amplitude and the lava
737 discharge rate, revealing a systematic tremor frequency increase concomi-
738 tant with tremor RSAM and TADR decrease over time. We therefore suggest
739 that observations reported in Figures 7b-d can be generalized to others PdF
740 eruptions in their initial phases.

741 *Eruptive intermediate phase: cone instabilities and magma re-feeding.*

742
743 Intermediate phases of PdF eruptions are more complex for long-lived
744 rather than short-lived eruptions. This phase lasted 37 days for the July-
745 August 2017 eruption and was characterized by strong fluctuations in tremor
746 spectra and amplitude (Figures 6a,b) based on which we can cut this long
747 phase in three sub-phases, namely (2.1), (2.2) and (2.3) indicated at the
748 bottom of Figure 6.

749 At the beginning of phase (2.1), we note an edifice deflation (Figure 6d),
750 which started with the eruption, suggesting the shallow magma reservoir
751 was still being emptied to feed the dyke and the eruption in surface. During
752 that time, lava flux at the focused eruptive vent had still significant values
753 (Figure 6c). Therefore the eruptive cone started to grow at the end of the
754 initial phase (Figure 7a), and grew up more rapidly during the increase in
755 RSAM (Figure 6a). We note that the end of the volcanic edifice deflation
756 episode is marked by the occurrence of shallow VTs earthquakes above the
757 shallow reservoir (Figure 6d) whose location are shown in Figure 8.

758 We show in Figure S9 a zoom on phase (2.1) with photos of the eruptive
759 cone exhibiting several periods of destabilization. We highlight periods of
760 increased amplitude and low frequency activation in the tremor that appears
761 to be associated with cone building/closing. During periods when the cone
762 flank collapsed, the amplitude of the tremor returns to a lower level and the
763 low frequencies are no longer excited. In this phase, the increases in RSAM
764 are not correlated with TADR increase. However, tremor properties might
765 be modulated by non-stationary magma re-feeding. Such magma pulses in
766 the dyke can result in eruptive cone growth and in the increase of resonating

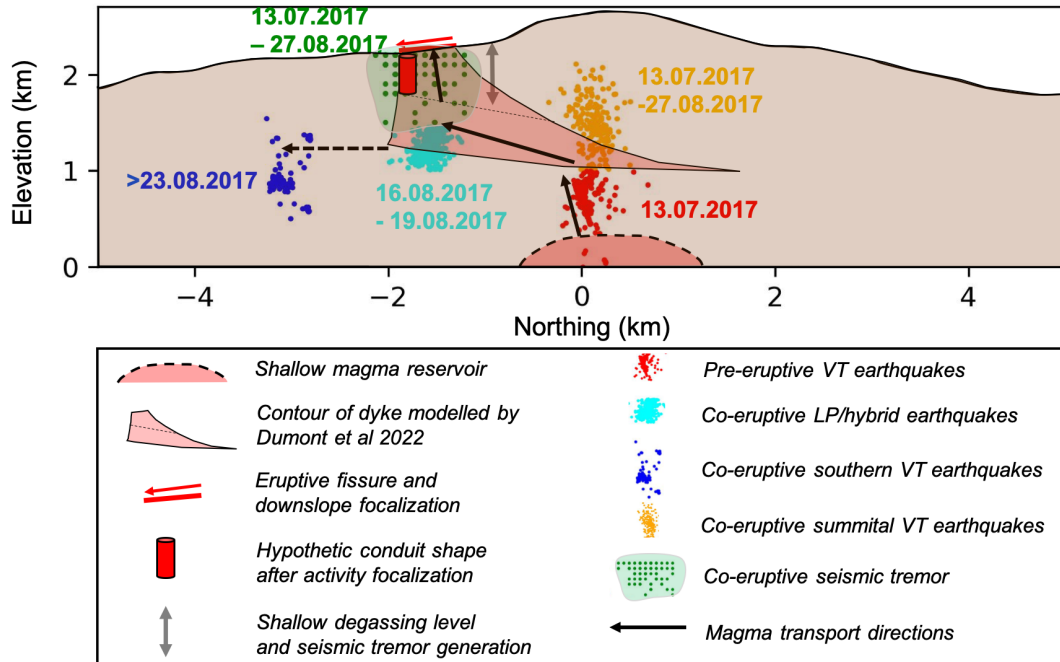


Figure 8: **Shallow magma transport and degassing with associated seismicity.** Location of seismo-volcanic sources during the July-August 2017 eruption from pre-eruptive dyke propagation to co-eruptive seismic tremor. Pre-eruptive seismic swarm location is shown with red circles. The following dyke propagation is represented as the red incurved trapezoid according to Dumont et al. (2022) who inverted InSAR data to obtain this dyke geometry. The final vertical part of the dyke coincides approximately with seismic tremor sources location indicated with green circles and associated shallow degassing level. Orange circles stand for co-eruptive shallow VTs beneath the volcano summit that accommodated phase of deformation of the underlying shallow reservoir. Cyan circles represent location of LP/hybrid earthquakes taking place after an eruptive site re-feeding phase the 15.08.2017. Blue circles stand for southern VT earthquakes occurring close to the end-of-the eruption during the final phase associated with shallow gas-piston seismic events at the eruptive site. Horizontal red line represents the eruptive fissure and the above red arrow its progressive closure when eruptive activity focused in a single vent. The underlying red conduit illustrates possible shallow geometrical changes that happened following closure and cooling of shallow dyke head and explaining changes in seismic tremor properties and degassing regimes.

767 lava column length potentially explaining the intermittent activation of low
768 tremor frequencies associated with increases in tremor amplitude.

769 In phase (2.2), we observe a weakening in co-eruptive tremor intensity
770 during approximately two weeks (Figures 6a-b), while TADR values were
771 very weak (Figure 6c) and notable changes in the summit GNSS stations
772 baseline occurred (Figure 6d). The corresponding inflation of the volcanic
773 edifice can be interpreted as magmatic recharge of the shallow reservoir while
774 the eruptive activity in surface was low. Shallow VTs earthquakes were
775 located between the volcano summit and the shallow reservoir during that
776 period (see Figure 6d and Figure 8).

777 At the end of this volcanic edifice inflation on 15.08.2017, phase (2.3)
778 started with an abrupt tremor amplitude increase and reactivation of its low
779 frequencies (at around 1 Hz). According to the OVPF's volcanological activ-
780 ity bulletin, weather conditions were too cloudy to allow reliable estimates of
781 lava flux that day, but tremor changes could be associated with a resumption
782 of ephemeral lava fountain activity at the eruptive site, observed by one of
783 the observatory's cameras ([http://volobsis.ipgp.fr/volcano-reports/
784 2017/08/OVPF_20170816_01_eruption.pdf](http://volobsis.ipgp.fr/volcano-reports/2017/08/OVPF_20170816_01_eruption.pdf)). Such increase of the eruptive
785 activity in surface was probably associated with a new arrival of magma from
786 the shallow reservoir. As seen in Figure 6b, about one day after the start of
787 phase (2.3), multiplets are detected in the continuous seismograms, reveal-
788 ing a seismic tremor composed of intermittent bursts (lasting from four to
789 one minute) and intersected by repetitive broadband impulsive signals with
790 LP/hybrid waveforms type (Figure 6d). Those latter are located approxi-
791 mately 1 km beneath the eruptive site (Figure 8).

792 *Eruptive final phase and seismic gas-piston swarm.*

793
794 A few days later, on 22.08.2017, the usual end-of-eruption gas-piston
795 seismic events has started to be detected, marking the final phase of the
796 eruption (Figure 6b). This phase lasting about five days was associated with
797 a gradual and slow gliding of the dominant frequency observed in the spectral
798 width on two distinct modes (Figure 6a).

799 Unlike the gas-piston sequence of the short eruption of December 2020,
800 this one is more complex and can be divided into three phases, namely,
801 (3.1), (3.2), (3.3), indicated at the bottom of Figure 6. We first observe in
802 phase (3.1) the transition between the weak residual continuous tremor of
803 the previous phase and the intermittent tremor composed of gas-piston. The

804 increase in the number of detected multiplets of this first phase (Figure 6b)
805 seems to be linked to the “tremor discretization”, when the seismic gas piston
806 events become more and more individualized and are thus better detected
807 by our template-matching method.

808 Then, in the phase (3.2), we observe a rapid increase in the gas-piston
809 events recurrence frequency accompanied by a slower increase of their am-
810 plitudes (Figure 6b). Note that those changes in gas-piston properties were
811 accompanied by the occurrence of VTs earthquakes a few kms to the south of
812 the eruptive site (see Figure 6d and Figure 8). Such seismicity might reflect
813 the closing of the feeding dyke pushing stagnant magma further south, or a
814 new batch of magma leaving from the shallow reservoir that did not erupt.
815 Besides magma transport, it could also potentially be explained by stress
816 transfer from the main magma system. This southern VT seismicity over-
817 laps a local deformation source revealed by InSAR data (© OVPF/IPGP -
818 August 2017 monthly bulletin, available for consultation: [http://volobsis.
819 ipgp.fr/volcano-reports/2017/08/OVPF_20170831_bullmensu.pdf](http://volobsis.ipgp.fr/volcano-reports/2017/08/OVPF_20170831_bullmensu.pdf)).

820 Finally, the eruption stopped on 27.08.2017, at the end of phase (3.3),
821 during which the number of detected gas-piston, and their amplitude, de-
822 crease until reaching the background noise level.

823 We provide a zoom of the spectral width function during the whole final
824 phase in Figure 9a, as well as examples of seismic gas-piston signals in 15-sec
825 windows in Figures 9b-c. We can clearly observe the spectral gliding in both
826 spectral width and gas-piston signals.

827 *4.2.3. Explaining spectral gliding with a fluid-filled crack model*

828 In order to better understand the origin of the observed spectral gliding,
829 we now use the analytical equations of the resonant frequency of the rect-
830 angular fluid-filled crack derived by Maeda and Kumagai (2017) enabling to
831 infer eruptive parameters changes as detailed in the methodological annex
832 in the supplementary material. We decide to invert the ratio between the
833 two observed excited modes visible in Figure 9a (see Figures S10 and S11)
834 to infer possible values of the crack stiffness and crack aspect ratio (Figure
835 S12). We then select some of possible values in a defined range to infer the
836 resonating crack length and its gas volume fraction by inverting the absolute
837 values of varying frequencies of the two modes (see Figures S13 and S14 and
838 Table S1).

839 We show the results of our inversion procedure in Figure 9c by synthe-
840 sizing the different possible paths in the parameters plane “crack length” vs

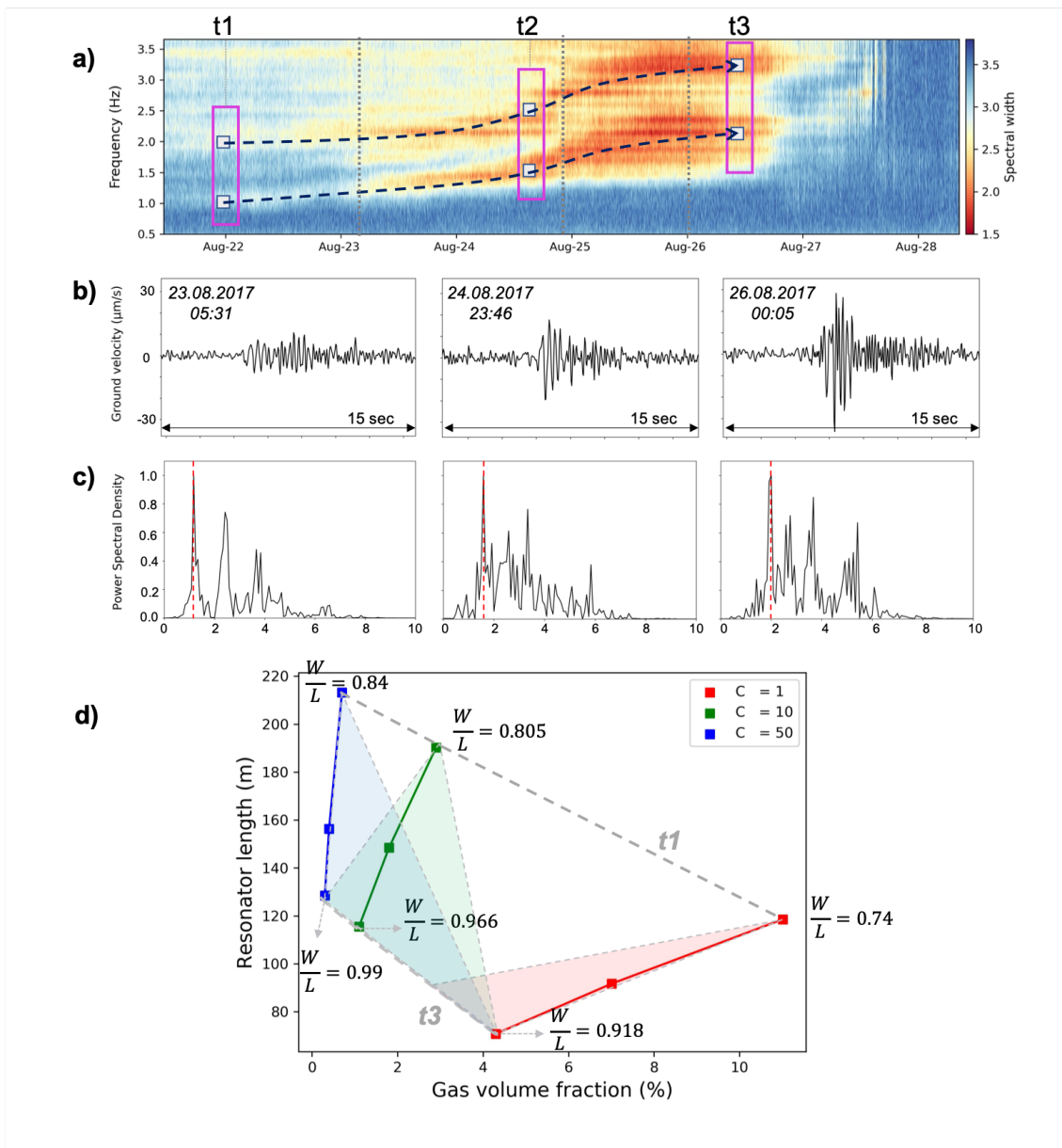


Figure 9: Tracking seismic gas-pistons spectral changes to monitor eruptive parameters during July-August 2017 final eruptive phase. (Continued on the following page.)

Figure 9: (Continued caption.) a) Zoom of the covariance matrix spectral width for the final phase of the July-August 2017 eruption. Purple rectangles indicates the time t_1 , t_2 and t_3 selected to invert observed gas-pistons frequencies (light blue rectangles) on the two activated modes highlighted with the black dashed arrows. b) and c) Examples of three seismic gas-piston event 15-sec-long seismograms and corresponding spectra, respectively. Red vertical dashed lines indicate the dominant excited frequency, showing positive gliding through time. Dates of these three signals are represented with grey vertical dashed lines in a). d) Summary of possible paths explaining the observed spectral gliding in the “resonator length” vs “gas volume fraction” plane. The values obtained by setting the crack stiffness values $C = 1$, $C = 10$ and $C = 50$ are shown in red, green and blue respectively. The 2 gray dotted lines indicate the values obtained for the date t_1 of 22.08.2017 and t_3 26.08.2017 12:00. The crack aspect ratio values associated with these crack stiffness values for dates t_1 and t_3 are indicated.

841 “gas volume fraction” that could explain the observed spectral gliding.

842 In all cases, the gliding is accompanied by an increase in the crack aspect
 843 ratio, which can be accommodated in different ways. Visual observations of
 844 the eruptive site indicated that in the period preceding and accompanying
 845 the transition (from 19.08.2017 to 21.08.2017) from the continuous residual
 846 tremor to the intermittent gas-piston regime, a weak effusive activity was
 847 observed over the entire width of the cone. Then, an image obtained by
 848 drone on 24.08.2017 reveals that there is only degassing on a reduced part of
 849 the cone in a “hornito” type crater of small width (probably a few meters).
 850 Thus we prefer a path not involving an increase in the width of the crack
 851 over time to explain the increase in W/L but rather a reduction in the length
 852 of the crack.

853 Without more external evidences, we can not determine an unique path
 854 explaining our observations, but only describe different possible scenarios.
 855 We can distinguish two extreme scenarios. The first one involves an im-
 856 portant resonator length shortening without significant gas volume fraction
 857 changes. On the contrary, in the other one, a large part of the spectral glid-
 858 ing is explained by a decreasing gas volume fraction, that can be obtained
 859 by the almost complete and final degassing of the remaining magma beneath
 860 the eruptive site. Note that all intermediate paths are valid to explain our
 861 observations.

862 The dyke shortening can be of different nature. We can imagine a “down-
863 ward dyke shortening” if the resonating bubbly lava column is drained as the
864 magma is transported to the south flank during the VT earthquakes swarm
865 observed in the final phase. In that case, this VT swarm could be explained
866 by the feeding dyke closure, pushing remaining magma to the south. An-
867 other hypothesis can be that the southern magma migration is associated
868 to a new batch of magma coming from the shallow reservoir. This might
869 cause a re-pressurization of the shallow part of the system near the eruptive
870 site, inducing a rise of the bubble nucleation level, and therefore an “upward
871 shortening” of the resonant bubbly magma portion (Maryanto et al., 2008).

872 The observed increase in number of detected gas-piston events (Figure 6b)
873 can reflect a gas flux increase at the vent. This might explain the associated
874 increase of seismic gas-piston amplitude (Spina et al., 2019). Such gas flux
875 increase can be obtained either in the “upward shortening” case if the south-
876 ern magma migration is due to a new batch of magma releasing new amount
877 of gas beneath the eruptive site. Or it can be obtained in the case of the
878 “downward shortening” inducing a depressurization of underlying magma
879 with a degassing intensification. In this latter case, we note that degassing
880 can also be the result of stagnant magma cooling via crystallization-induced
881 degassing.

882 Finally, as the observed end-of-eruption gas-piston sequence was quite
883 long (five days) and intense, a path starting with a relatively high gas content
884 (around 10%) might be more plausible, therefore requiring a significant drop
885 in this gas content value through time, as inferred in Figure 9d.

886 *4.2.4. Gas-pistons vs effusive paroxysm: example of the February-March* 887 *2019 eruptive final phase*

888 In this section, we provide an overview of the end-of-eruption final phases
889 of PdF eruptions. We systematically applied our template-matching method
890 to scan the continuous seismograms corresponding to the 23 eruptions of our
891 catalog. We can therefore attest if their final phase is characterized by a
892 transition from a continuous seismic tremor to a seismic gas-piston swarm or
893 not. We indicate the presence or absence of gas-piston events in the Table 1.

894 From 23 eruptions, only 5 final phases are not completely associated with
895 gas-piston events. In order to better understand the mechanisms governing
896 degassing regime transitions leading to the end of eruptions, we compare the
897 properties of seismic tremor in the final phases (tremor spectra, amplitude
898 and state) with data giving insights into lava discharge rates, edifice defor-

899 mation and the chemical composition of the erupted lava. We selected in
900 total 10 eruptions (see Figure S15) for which we have enough available data
901 to perform our multi-disciplinary analysis.

902 We show in Figures 10a,b,d the resulting relationships between seismic
903 tremor minimal excited frequency, its amplitude, and the lava discharge rate,
904 while Figure 10c illustrates the magma composition of the eruptions. From
905 this database, we notice that the four eruptions not exhibiting final phase gas
906 pistonning were characterized by larger values of TADR, magma MgO content,
907 tremor RSAM and with activation of lower frequencies in the tremor spectra
908 (> 1 Hz), compared to the other six eruptions with gas-pistons.

909 Eruptions without gas-piston are associated with a heterogeneous magma
910 composition as seen in Figure 10c with the increase in *MgO* content, at-
911 testing that the magma becomes more olivine-rich towards the end of the
912 eruption, which probably indicates a deep magmatic recharge of hotter, less
913 evolved, volatile-rich magma. Such deep magma pulses have been suggested
914 by Coppola et al. (2017) to explain phases of “effusive paroxysm” during the
915 August-October 2015 eruption final phase, characterized by increase in values
916 of TADR (with return of lava fountain at the eruptive site), tremor ampli-
917 tude and associated with a volcanic edifice deflation. In addition of these
918 characteristics, we also report in Figure S16a that these effusive paroxysms
919 episodes are also visible in the tremor spectral content with an activation of
920 the low frequency below 1 Hz, potentially explained by an increase of the gas
921 volume fraction in the conduit and an increase of the length of the resonating
922 lava column. We also note in Figure S16b the activation of gas-piston swarm
923 during the rest periods intersecting the three effusive paroxysm episodes.

924 Based on the increase in TADR and SO_2 emissions, Verdurme et al.
925 (2022) suggests that the final phase of the long-duration February-March
926 2019 eruption might also be fed by deep, volatile-rich magma. We show
927 in Figure 11 the temporal evolution of seismic tremor properties as well as
928 TADR and GNSS baseline for this eruption. Besides a very brief peak in the
929 multiplet detector function in Figure 11b, our comparative analysis reveals
930 that this final phase with effusive paroxysm (high TADR values in Figure 11c)
931 has similar characteristics than the one of August-October 2015, i.e. a low-
932 frequency tremor activation (Figure 11a) with high amplitude (Figure 11b),
933 and linked to an edifice deflation. This latter is probably induced by the
934 emptying of the newly filled shallow reservoir by the deep magma pulse.
935 This last magma input was then erupted during the final phase explaining
936 the high values of TADR and magma *MgO* content.

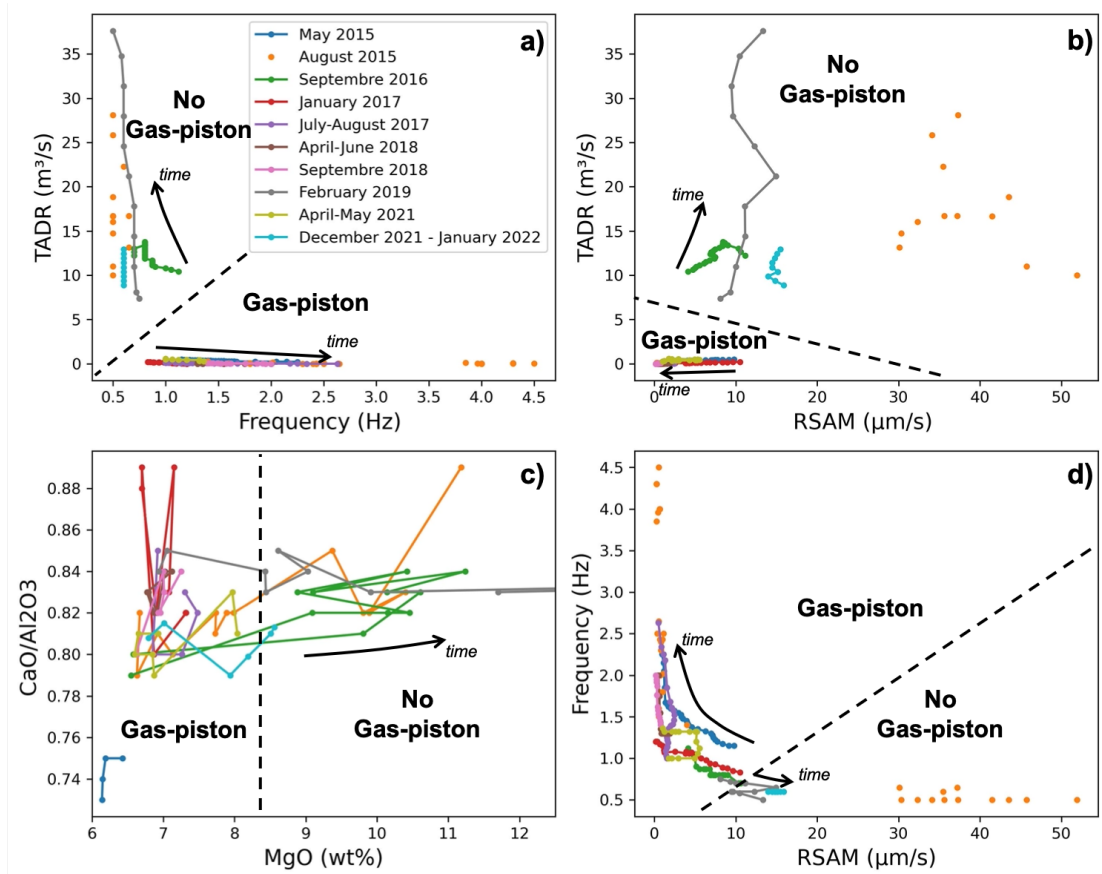


Figure 10: **Overview of Piton de la Fournaise end-of-eruption characteristics.** Geophysical and geochemical measurements of ten selected final phases displayed with different colors indicated in the caption. Dashed black lines separate the final phases characterized by gas-piston swarm from the others and black arrows indicate typical time evolution of the measured parameters. a) and b) TADR as a function of the minimal excited tremor frequency and tremor RSAM, respectively. c) Ratio of $\text{CaO}/\text{Al}_2\text{O}_3$ as a function of the amount of MgO in wt%. d) Minimum excited tremor frequency as a function of tremor RSAM. Note that the August-October 2015 eruption is a particular case with values in both “Gas-piston” and “No Gas-piston” areas because its final eruptive phase was characterized by final effusive bursts intersected by two phases of arrest associated with gas pistoning.

937 The December 2021 - January 2022 eruptive final phase was also charac-
 938 terized by the same parameters evolution. However, the two other eruptions
 939 not characterized by end-of-eruption gas-pistoning, namely the September
 940 2016 and April 2020 eruptions, are not associated with a significant edifice

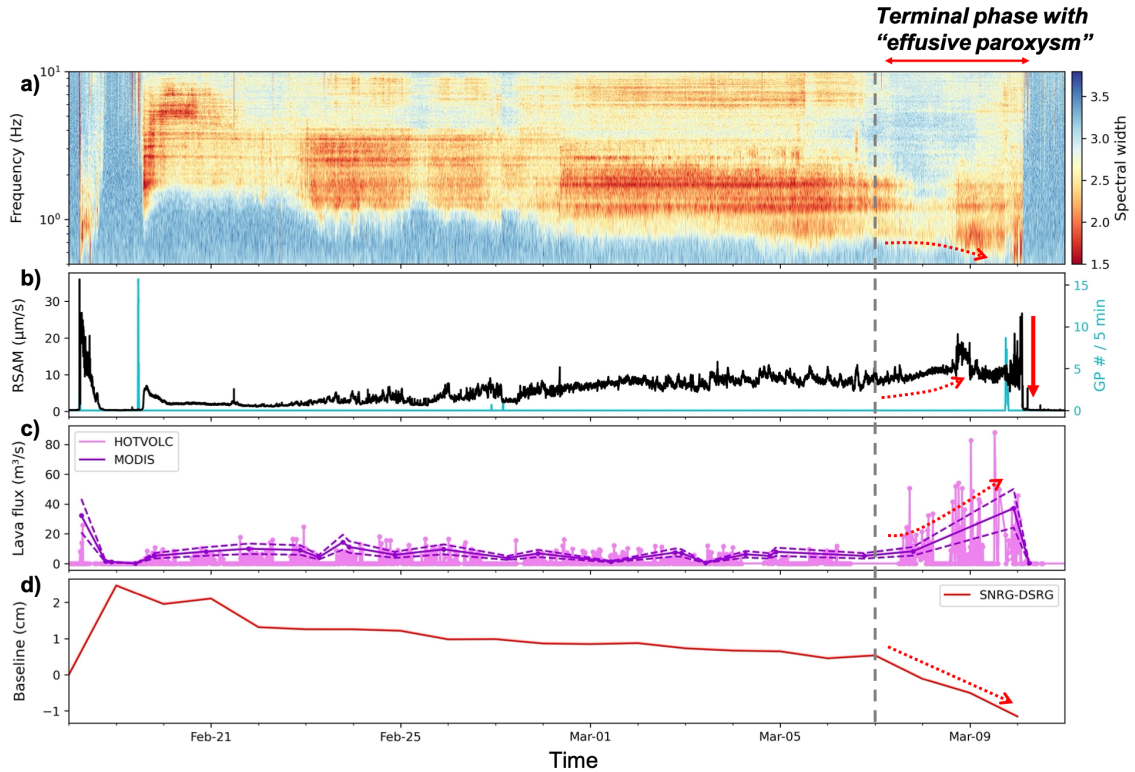


Figure 11: **Time evolution of the February-March 2019 eruption with “effusive paroxysm” during its final phase.** a) Spectral width function computed in 2-min-long time windows. b) Black curve represents the tremor RSAM computed in the [0.5 - 5] Hz frequency band. Cyan curve indicates number of detected gas-piston event in 5 min-long time windows. c) Lava discharge rate at the eruptive vent. d) Baseline between the SNRG and DSRG summit GNSS stations. Vertical dashed grey line in a), b), c) and d) represents the start of the final eruptive phase while the red dashed arrows highlight the corresponding changes in measured parameters. Downward red arrow in b) indicates the very abrupt end of the eruption.

941 deflation. These eruptions are shorter than the three previously mentioned
942 ones. They lasted 6.8 days and 4.2 days, respectively, and they are both as-
943 sociated with irregular highly explosive behavior and TADR increase during
944 their final phases, potentially explaining their absence of gas-piston swarm.
945 The last day of September 2016 eruption was characterized by simultaneous
946 Hawaiian fountaining at the main vent and transient explosive activity at a
947 second vent (Thivet et al., 2020b). During the two last days of the April
948 2020 eruption, an unusual highly explosive event occurred, with the emission
949 of large quantities of Pele’s hair (Peltier et al., 2020).

950 The five eruptions without gas-pistoning sequence are all characterized
951 by a very abrupt drop in seismic tremor RSAM and eruptive activity at their
952 end lasting a few minutes. On the contrary, the other 18 eruptions with gas-
953 piston events are characterized by a slow decaying tremor amplitude lasting
954 from hours to days associated to very low TADR values, probably reflecting
955 a slow final intermittent degassing of the remaining stagnant magma beneath
956 the eruptive site. We illustrate this tremor amplitude behavior for the end of
957 four eruptions in Figure S17 (eruptions June 2014 and December 2020 with
958 the slow decaying tremor amplitude associated with gas-pistoning activity
959 in Figure S17a and S17c, respectively; and eruptions February-March 2019
960 and December 2021 - January 2022 with abrupt tremor amplitude drop and
961 without gas-pistoning activity in Figure S17b and S17d, respectively).

962 5. Discussion

963 5.1. Shallow dyke-head magma degassing and co-eruptive tremor generation

964 Our co-eruptive seismic tremor signals study from the 2014-2021 period
965 at the PdF reveals that its generation mechanism is strongly controlled by
966 the shallow magma degassing at the head of the dyke under the eruptive site.

967 We shown that the tremor sources location are correlated with the strong
968 increase of degassing at shallow depth (Figure 3c). Those results are interest-
969 ing to compare with results of shallow intrusions imaging via the inversion
970 of InSAR data by Dumont et al. (2022). They show that eruptions away
971 from the summit area are fed by dykes forming a spoon-shaped structure
972 with a horizontal and vertical curvature (example in Figure 8 for the July
973 2017 magma injection). This result is consistent with ours revealing that the
974 tremor source originates at the transition where the magma transport be-
975 comes vertical a few hundred meters below the eruptive site, probably driven
976 by the increase in buoyancy of the magma caused by its degassing.

977 Moreover, we also show that the initial tremor spectra seems to be con-
978 trolled by the initial dyke head width arriving at the surface, of which we
979 estimate a value by measuring the size of the eruptive fissure opened at the
980 eruption start (Figure 7e). The greater the fissure length, the more the ini-
981 tial tremor spectra is shifted towards low frequencies. Similarly, a previous
982 study also showed that the fissure length increased with the amplitude of the
983 tremor at PdF (Derrien, 2019).

984 *5.2. Conduit geometry change from dyke to pipe*

985 A few hours after the eruption start, as often observed in basaltic systems,
986 the volcanic activity that was occurring all along the open fissure, focused in
987 a single vent. At the PdF, this phenomenon is accompanied by a decrease
988 in the lava flow and tremor amplitude, as well as the progressive loss of the
989 low frequencies in the tremor spectra (Figure 7a-d). This drop in TADR can
990 be explained by the drop of the overpressure in the magma reservoir, when
991 the magma is transported to the eruptive site within the dyke. This pressure
992 drop may induce a partial closure of the dyke and focusing of volcanic activity
993 by cooling and solidification. This surface focusing then causes a decrease
994 in the tremor amplitude and the loss of excitation of low frequencies in its
995 spectrum.

996 The spatio-temporal evolution of the fissure is governed by the thermo-
997 rheological behavior of the magma (Bruce and Huppert, 1989) as well as the
998 organization of the flow in the dyke (Pioli et al., 2017). This focusing process
999 of volcanic activity represents an important phenomenon in volcanology and
1000 detailed monitoring of the co-eruptive seismic tremor properties enable to
1001 add some temporal constraints to improve its understanding.

1002 However, the modeling of the corresponding tremor spectrum changes
1003 based on the fluid-filled crack model is complex because the focusing is also
1004 accompanied by a change in the eruptive conduit geometry from a rectangular
1005 dyke-like shape to a cylindrical pipe-like shape, the latter being mechanically
1006 and thermally more efficient for the magma transit (Fukushima et al., 2010).
1007 Also, the progressive decrease of magma gas volume fraction over time may
1008 have a role in the change in the tremor spectrum observed during the PdF
1009 initial eruptive phase.

1010 *5.3. Tracking changes in gas-liquid dynamics from measuring seismic tremor*
1011 *properties*

1012 The time-frequency properties of the tremor and images of the eruptive
1013 site indicate that fluctuations in the degassing regime cause changes in the
1014 tremor behavior. Tracking these changes enables monitoring the dynamics
1015 of the ongoing eruption.

1016 *Degassing regimes transition during short-lived eruptions.*

1017
1018 During short-lived eruptions, we show that the spectral properties and
1019 temporal stability of the tremor clearly outline in time three distinct eruptive
1020 phases (initial, intermediate and final, see Figure 4).

1021 The initial phase is associated with the emission of lava fountains first
1022 along the entire fissure, and then where the activity has focused to a single
1023 vent (Figure 4 and Figure 5).

1024 We show in Figure S8 that during this initial phase, although the range
1025 of values varies between eruptions, tremor RSAM and TADR seem to be cou-
1026 pled, both decreasing over time from an initial high value. Such RSAM/TADR
1027 coupling can be interpreted as the manifestation of a homogeneous degassing
1028 regime (Coppola et al., 2009), where gas and liquid are coupled and rise at
1029 the same speed, forming the Hawaiian degassing regimes in the conduit with
1030 corresponding lava fountains at the surface.

1031 When the overpressure in the dyke decreases with the lava discharge rate
1032 at the vent and the dyke is not significantly replenished, the magma begins
1033 to cool, crystallize, its ascent rate slow down with increasing viscosity. Thus
1034 a decoupling between gas and liquid occurs. The bubbles have more time
1035 to coalesce and gas accumulation processes take place, marking a transition
1036 from the Hawaiian regime to a so-called Strombolian regime characterized
1037 by gas slug bursts at the surface. Thivet et al. (2020a) highlight this regime
1038 transition from the textural analysis of the eruptive products of two short
1039 PdF eruptions in June 2014 and July-August 2015. For the short December
1040 2020 eruption, we highlight this transition from seismic tremor analysis at
1041 the beginning of the intermediate phase (Figure 4).

1042 The final phase is characterized by an intermittent degassing regime as-
1043 sociated with gas-piston events. All short eruptions we analyzed between
1044 2014-2021 end with a brief gas-piston swarm lasting a few hours with a de-
1045 crease in the associated tremor amplitude.

1046 The increase in viscosity and crystallinity of the nearly stagnant and
1047 highly degassed magma that resides in the dyke at the end of this type of
1048 eruption probably plays a key role in this final transition from continuous
1049 to discrete degassing/tremor. This effect was highlighted recently by Barth
1050 et al. (2019), suggesting that highly crystalline shallow portions of basaltic
1051 conduits may act as a flow valve transforming a steady gas flux into a series
1052 of discrete gas slugs.

1053 From an analog approach through laboratories experiments in cylindrical
1054 conduits, Spina et al. (2019) reported a link between gas flux and length of
1055 slug bubbles, and that an increase in viscosity induces a longer recurrence
1056 time between slug bursts. Therefore, the reduction of gas flux and increase
1057 in magma viscosity can be associated with a decrease of bubble length and
1058 an increase of their bursting recurrence time. At some point bubbles stop
1059 bursting while the amplitude of the associated seismic events reach the back-
1060 ground noise level. Spina et al. (2019) notes that an increase in absolute
1061 viscosity of up to two orders of magnitude is expected upon cooling and
1062 crystallization of primitive basaltic melt from 1250 down to 1150 °C with a
1063 3K/min cooling rate expected in shallow dykes (Kolzenburg et al. (2016)).
1064 This could explain rapid transitions to the final gas-piston regime.

1065 *Complex and long-lived eruptions.*

1066
1067 Between the short and long-lived eruptions, the initial phase shows similar
1068 variations in tremor properties, except that it lasts longer in the case of the
1069 long eruptions. However, when the activity focuses on a single vent, the
1070 intermediate phase of long eruptions is more complex. This is explained
1071 as the short eruptions are fed by a single batch of magma whose evolution
1072 of temperature, viscosity, and gas content modifies the shallow gas-liquid
1073 organization, accompanied by transitions of degassing regime and associated
1074 eruptive style. On the contrary, long-duration eruptions are characterized
1075 by episodes of re-feeding during the intermediate phase, explaining their
1076 longevity.

1077 Nevertheless, most long eruptions, after their re-feeding phases, end with
1078 a gas-piston phase. On the other hand, this last phase can be more complex
1079 than for those of short eruptions (Figure 6 and Figure 9).

1080 Finally, a last possible and rarer degassing regime transition is a return
1081 to more explosive activity at the end of the eruption, reported by Coppola
1082 et al. (2017) as “effusive paroxysm” for the August-October 2015 eruption.

1083 We report in this study similar characteristics for the end-of-eruptions of
1084 August-October 2015, February-March 2019 and December 2021 - January
1085 2022, with an increase in TADR and tremor RSAM, associated with lava
1086 fountaining, low frequency activation of tremor, eruption of slightly more
1087 mafic and phenocryst-rich magma and deflation of the volcanic edifice.

1088 As suggested by Coppola et al. (2017) for the August-October 2015 erup-
1089 tion, these observations can be explained by deep magma pulses feeding these
1090 episodes of paroxysms. Such pulses can be caused by the progressive empty-
1091 ing and depressurization of the shallow magma reservoir during the previous
1092 eruptions. Therefore, detection and characterization of such effusive parox-
1093 ysm are useful to integrate into our global knowledge of the PdF magmatic
1094 system dynamics. If final phases of the February-March 2019 and December
1095 2021 - January 2022 eruption have the same origin, this can reveal some
1096 cyclic shallow reservoir emptying.

1097 Regarding the observation of the renewal of lava fountains at the surface
1098 during these final phases, this can be explained by an increased gas flux in the
1099 eruptive conduit (from the degassing of new deep magma), inducing longer
1100 slug bubbles, then merging and becoming an annular lava fountain regime
1101 (Spina et al., 2019).

1102 Another characteristic of these final phases with the effusive paroxysm
1103 is that they end abruptly (tremor RSAM drop in a few minutes) compared
1104 to that with the gas-piston (tremor RSAM drop in a few hours to days).
1105 A hypothesis is that the transit of the last pulses of magma generates a
1106 significant pressure drop in the reservoir and feeding dyke, inducing a closure
1107 of the latter and a sudden end of the eruption.

1108 **6. Conclusion**

1109 We analyzed in this study the entire co-eruptive seismic tremor database
1110 available at Piton de la Fournaise between 2014 and 2021, using the covari-
1111 ance matrix method. The measurement of the spatial, temporal and spectral
1112 properties of the tremor informs us about its origin and its close link with the
1113 shallow magma degassing at the head of the dyke under the eruptive site. The
1114 time-frequency analysis of the tremor and its state (continuous-intermittent)
1115 combined with a multi-disciplinary study of other parameters of the system
1116 (eruptive site images, lava flow rate, deformation, magma composition) al-
1117 lowed us to translate the varying tremor properties into different eruptive

1118 parameters. It appears that the tremor data encodes several eruptive pro-
1119 cesses, such as volcanic activity focusing, and notable changes in degassing
1120 regimes, at the origin of the observed fluctuations in eruptive style at the
1121 surface. Therefore, the resulting seismic database provides extremely valu-
1122 able and high temporal resolution information regarding the PdF eruptive
1123 dynamics. Finally, the seismic methods used in this paper can be transposed
1124 to other basaltic systems with an adapted seismic monitoring network, thus
1125 enabling to advance in the future on a better, more global understanding
1126 of the physical mechanisms governing these eruptions and associated seismic
1127 tremor signals.

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1137 toire Volcanologique du Piton de la Fournaise (OVPF, doi:10.18715/REUNION.OVPF).
1138 The permanent network data (<https://www.fdsn.org/networks/detail/PF/>)
1139 can be downloaded from the IPGP Data Center (<http://ws.ipgp.fr/fdsnws/dataselect/1/>),
1140 the RESIF data center (<http://ws.resif.fr/fdsnws/dataselect/1/>), or the IRIS
1141 Data Management Center (IRISDMC, <http://service.iris.edu/fdsnws/dataselect/1/>).
1142 This is IPGP contribution number xxxx.

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