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## Magma propagation at Piton de la Fournaise from joint inversion of InSAR and GNSS

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1                   **Magma Propagation at Piton de la Fournaise from joint**  
2                   **inversion of InSAR and GNSS.**

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9                   **Key Points:**

- 10                   • Magma feeding the May 2016 eruption propagated laterally as a sill before turning  
11                   into a dike.
- 12                   • The sill propagation is step-wise with an initial acceleration followed by a 5 hours  
13                   pause.
- 14                   • The eruption was fed by a single batch of magma quickly disconnected from its  
15                   source.

## Abstract

Magma propagation is an unsteady process controlled by magma-crust interaction. To provide information on its dynamics, we invert complementary ground deformation data spanning the 8 hours preceding the May 26, 2016 eruption at Piton de la Fournaise volcano (La Réunion, France). Data are inverted using 3-D boundary element models combined with a Monte Carlo inversion method. The final geometry of the displacement source is determined based on four interferograms spanning the whole propagation phase while the dynamics of the propagation is inferred from temporal inversion of continuous GNSS data, using the final geometry as an *a priori* to constrain the source. The best modeled magma path consists in a 2700-meter-long sill located 800 meters above sea level and connected to the eruptive fissure by a sub-vertical dike. The quick opening of the horizontal part of the intrusion could have been favored by limited flank sliding during the early stage of propagation. The intrusion then stalled for  $\sim 5$  hours, while pressure increased slightly, until final upward propagation and eruption. Volume budget suggests that the eruption was fed by a single batch of magma quickly disconnected from its source. The delay prior to the eruption may reflect a limited magma supply. Finally, two mechanisms, potentially acting together, might have favored the eruption : a driving role of magmatic gas and/or, as often observed at Piton de la Fournaise, an eastward flank slip.

## 1 Introduction

Dike and sill intrusions are the dominant mechanisms for basaltic magma transport within the shallowest region of the Earth's crust. They sometimes propagate tens of kilometers away from the main magma reservoir. Reaching the Earth's surface, they often lead to fissural eruptions. Near to towns or man-made structures, fissural activity and lava flows present a risk, as experienced, for instance during the May 2018 eruption of Kilauea volcano in the Lower East Rift Zone [*Global Volcanism Program*, 2018] or at Piton de la Fournaise (PdF) in 1977 and 1986 [*Villeneuve and Bachèlery*, 2006]. At a given volcano, intrusions often vary in style, length, orientation and direction [*Chadwick Jr and Dieterich*, 1995; *Peltier et al.*, 2009a]. Consequently, distances between the summit and the vents, and vents' elevations, vary from one eruption to another, as observed at PdF [*Roult et al.*, 2012; *Peltier et al.*, 2009a]. Moreover, magma ascent can stop before reaching the ground and erupting (see *Moran et al.* [2011] for a review). Such non-eruptive intrusions were described, for instance at Kilauea by *Cervelli et al.* [2002], or at Natron Rift by *Calais et al.* [2008]. At PdF, between 1985 and 2016, at least 15% of the detected intrusions did not lead to eruptions [*Roult et al.*, 2012]. To mitigate the hazards induced by eruptions, it is important to understand the parameters responsible for these variations. The purpose of our work is to use *in situ* measurement to understand the physical and mechanical parameters controlling magma propagation.

The magma propagation toward the surface is complex and might be controlled by several physical and mechanical parameters. These parameters were studied using field mapping, analog laboratory experiments and numerical studies (see *Rivalta et al.* [2015] for a review). It was proven that the shape, orientation and dynamics of intrusions are strongly influenced by the host medium (rheological properties as well as local stress field), by the magma's physical properties (density and viscosity) and by the pressurization of the feeder storage zone when the intrusion remains connected with the reservoir. Geological [*Gudmundsson*, 2006], analog [*Rivalta et al.*, 2005; *Kavanagh et al.*, 2006] and numerical [*Dahm*, 2000; *Maccaferri et al.*, 2010] studies evidenced the role of heterogeneities in mechanical properties on dike paths, shape, opening and kinematics. Besides, pre-existing fractures have the potential to be re-used [*Delaney et al.*, 1986; *Ruch et al.*, 2016]. More generally, it has been demonstrated that the local stress field strongly influences magma path and velocity [*Watanabe et al.*, 2002; *Sigmundsson et al.*, 2015; *Pinel et al.*, 2017]. For instance, the stress field is locally influenced by the topography. When a caldera collapse occurs, this stress field becomes such that dike to sill transition

is favored at shallow depth. [Corbi *et al.*, 2015]. Melt properties are also known to influence magma propagation. In particular, magma density controls the buoyancy force available to drive vertical propagation. Magma being a complex mix of fluid, solid particles and dissolved gas, strong density variations are potentially induced by gas exsolution or crystals sedimentation during ascent. Gas accumulation at the tip may form a pocket that favors vertical migration [Lister, 1990; Menand and Tait, 2001; Taisne and Jaupart, 2011]. The magma's driving force is also determined by its initial overpressure, provided by the feeder storage zone. Studying the pressure variations of the intrusion through time provides information on the efficiency of the connection with the magma feeding reservoir [Segall *et al.*, 2001]. Due to the lack of information on the intrusion source, models usually consider simple end members. These include either a crack disconnected from its source [Maccaferri *et al.*, 2010], a crack fed by a constant inflow of magma [Pinel and Jaupart, 2004; Kavanagh *et al.*, 2006] or by a source at constant pressure [Pinel and Jaupart, 2000; Menand and Tait, 2001]. Such approximations may not be fully realistic. More generally, due to the multiple factors that may impact the shape and dynamics of magma during propagation, more observations, providing both high spatial and temporal resolution, are required to improve models and better determine *in situ* parameters and boundary conditions.

The analysis of seismic signals recorded during magma propagation might provide key additional information. The maximum shear stress is expected to be close to the tip of the propagating dike, potentially inducing micro-earthquakes [Rubin and Gillard, 1998]. Thus micro-earthquakes migration can be used to track magma propagation, sometimes revealing variations in the propagation velocity [Sigmundsson *et al.*, 2015]. However, locating the very shallow and low magnitude seismicity induced by magma intrusions using P and S-waves is challenging. In addition, even though, alternative methods [Taisne *et al.*, 2011] exist, the crack tip propagation does not always produce seismicity. The preexisting deviatoric stress may not be large enough to generate shear failure or the magnitude of induced earthquakes may be too low to be detected by certain seismic network configurations. Besides, even when present, seismic activity does not always reveal the specific location of the dike tip, but rather a volume of maximum stress change [Dieterich *et al.*, 2000]. At PdF, micro-seismicity sometimes shows that there has been deep vertical propagation [Battaglia *et al.*, 2005] and activation of a volume beneath the summit crater [Lengliné *et al.*, 2016; Duputel *et al.*, 2018], but no lateral migration.

In addition to seismic data, high rate geodetic observations have the potential to track the magma propagation, providing additional key information on the volume involved [Einarsson and Brandsdóttir, 1978; Aoki *et al.*, 1999; Segall *et al.*, 2001; Sigmundsson *et al.*, 2015]. Toutain *et al.* [1992] used tilt records to track the maximum uplifted area during the April 18, 1990 PdF eruption. Similarly, combining tilt data and seismic background noise, Peltier *et al.* [2005] looked into the migration of the uplifted area during 9 eruptions at PdF from 2000 to 2003, revealing a recurrent pattern: fast vertical propagation ( $2 \text{ m.s}^{-1}$ ) followed by slower lateral migration ( $0.2\text{-}0.8 \text{ m.s}^{-1}$ ). Nevertheless, the link between the maximum uplift and the dike tip location is not straightforward. Using high-Rate GNSS measurements at Mount Etna, Cannavò *et al.* [2015] determined a distribution of point pressure sources, but such sources are not able to account for the complex deformation pattern produced by dike intrusions and might lead to erroneous results [Menssian, 2013]. An alternative method to determine the evolution of volume change consists of a two-step procedure. The intrusion geometry is first determined using all available geodetic data, then linear inversions are used to retrieve the opening distribution [Aoki *et al.*, 1999; Segall *et al.*, 2001].

However, despite their high temporal resolution, continuous GNSS and tilt data are spatially sparse. When inverted, they lead to oversimplified source geometries. On the other hand, Interferometric Synthetic Aperture Radar (InSAR) data provides information on the displacement field, in the Line of Sight (LOS) direction, which has very good spa-

121 tial resolution (some meters), but poor temporal resolution (usually not better than 6 days).  
 122 Thus, InSAR data can be used to image complex displacement patterns. Addressing this  
 123 complexity with simple analytical models is not satisfactory and requires the combina-  
 124 tion of numerical models with inversions to reveal the complexities of geometric sources  
 125 [Trasatti *et al.*, 2008; Fukushima *et al.*, 2010]. Because temporal resolution of InSAR is  
 126 low (of the order of days) while the intrusion process is generally short (some tens of  
 127 minutes to hours at PdF), the opportunity to capture an ongoing eruption is rare. One in-  
 128 teresting example concerns an intrusion at Fernandina (Galapagos), where Bagnardi *et al.*  
 129 [2013] used an InSAR image acquired during the propagation phase to reveal the intrusion  
 130 geometry at an early stage. They showed that the intrusion orientation had switched from  
 131 a sill to a radial dike. In another rare case in May 2016, such an acquisition by Sentinel-1  
 132 at PdF also imaged the displacement field produced by propagating magma.

133 In order to overcome limitations specific to point measurements and InSAR, a combi-  
 134 nation of these two types of measurements have been proposed to track the opening of  
 135 dikes [Hamling *et al.*, 2009; Fukuda and Johnson, 2010; Auriac *et al.*, 2013; Sigmunds-  
 136 son *et al.*, 2015]. The proposed methods are derived from the Okada formulation [Okada,  
 137 1985] and ignore topography, and medium heterogeneities, sometimes providing erroneous  
 138 source determinations [Masterlark and Lu, 2004; Fukushima *et al.*, 2005]. This led Cur-  
 139 renti *et al.* [2011] to propose fully 3D heterogeneous elastic models to determine the dis-  
 140 tribution of openings on a planar vertical dike during the 2008 intrusion at Mount Etna.  
 141 However, inverting surface displacement allows either overpressure or opening distribu-  
 142 tions on fractures to be determined. It is more meaningful physically to determine the  
 143 overpressure for the following reasons. Pressure boundary conditions are inherently in  
 144 agreement with the physics of intrusions, where hydraulic connectivity results in homoge-  
 145 neous overpressure. Homogeneous overpressure leads to smoothly varying openings with  
 146 no need for smoothness-constraints to regularize the inversion [Zeller and Pollard, 1992].  
 147 Lastly, fewer parameters are required for overpressure determination than opening deter-  
 148 minations, making these models more likely from a statistical point of view [Tridon *et al.*,  
 149 2016].

150 In this study, we propose an integrated approach which complements the high spatial  
 151 resolution of InSAR with the high temporal resolution of GNSS data in order to (i) de-  
 152 termine the complex geometry of an intrusion located beneath a realistic topography, (ii)  
 153 resolve pressure changes as the intrusion propagates. We apply this strategy to the May  
 154 26, 2016 eruption at PdF, a highly active volcano characterized by a dense monitoring net-  
 155 work (Fig. 1) and routine InSAR monitoring, providing one of the best spatio-temporal  
 156 ground deformation data coverage available for a volcano. Taking advantage of the SAR  
 157 image that was acquired during the magma propagation, we discuss the importance of an  
 158 integrated approach to solve trade-offs between inverted parameters. Focusing on the lat-  
 159 eral propagation phase, we determine the time evolution of the pressure source. We then  
 160 discuss the characteristics of the intrusion propagation path, geometry and dynamics and  
 161 their implication in terms of physical processes.

## 162 **2 Geological and historical background**

### 163 **2.1 Piton de la Fournaise volcano**

164 PdF (La Réunion Island, France) is a basaltic intra-plate volcano reaching 2632 m  
 165 above sea level (a.s.l.) where fissure eruptions resulting from dike intrusions frequently  
 166 occur. The Central Cone of PdF (400 m high, 1500 m radius) is located within a major U-  
 167 shaped structure called the Enclos Fouqué caldera (Fig. 1). Based on the distance of the  
 168 eruptive fissures from the summit, eruptions are classified as summit, proximal (when they  
 169 occur on the flanks close to the summit), or distal (for distances greater than 4 km from  
 170 the summit). Since 1972, the percentage of eruptions classified as each type are 21%,  
 171 68%, and 11%, respectively [Peltier *et al.*, 2009a].

172 Since 1979, the Observatoire Volcanologique du Piton de la Fournaise managed by  
 173 the Institut de Physique du Globe de Paris (OVPF–IPGP) has monitored the volcano to  
 174 follow seismicity (40 seismometers), ground deformation (via 24 GNSS stations, 9 tilt-  
 175 meters, 3 extensometers) and gas emission. The network of permanent stations is now one  
 176 of the densest in the world, with distances between stations of less than 1 km in the sum-  
 177 mit area, and less than 3 km in the proximal area, (Fig. 1C) making the volcano one of  
 178 the best monitored and allowing to anticipate [Peltier *et al.*, 2018] and follow eruptions  
 179 with precision. In addition to *in situ* networks, the Indian Ocean InSAR Observatory Ser-  
 180 vice (OI<sup>2</sup>), carries out systematic acquisitions of InSAR data. Intermittent InSAR monitor-  
 181 ing was performed within this framework at the volcano from 1998, and this monitoring  
 182 became systematic in 2003.

183 In April 2007, a distal eruption at low elevation drained the shallow magma reser-  
 184 voir, leading to the most voluminous lava flow in historical records, and collapse of the  
 185 summit caldera associated with a 1 km × 0.8 km wide and 330 m deep crater [Michon  
 186 *et al.*, 2007; Peltier *et al.*, 2009b]. A large seaward displacement (up to 1.4 m) of the east  
 187 flank, evidenced by InSAR, has been associated with this major eruption [Froger *et al.*,  
 188 2015; Tridon *et al.*, 2016]. After a series of small summit eruptions and non-eruptive in-  
 189 trusions in 2008-2009, the volcano remained quiet for 3.5 years. But from 2014 to Septem-  
 190 ber 2018, activity picked up, with 14 further eruptions which alternated between north and  
 191 south flanks location (Fig. 1C) [Peltier *et al.*, 2016, 2018; Coppola *et al.*, 2017; Gurioli  
 192 *et al.*, 2018].

## 193 2.2 The May 2016 Eruption

194 On May 25, 2016 at 19:40 (all times are in Coordinated Universal Time (UTC)),  
 195 after ten days of slight inflation and low-level seismic activity (~5-10 Volcano-Tectonic  
 196 earthquakes (VT) per days), a seismic crisis began, characterized by ~2000 VT, with small  
 197 magnitude  $M < 1$ . Eight hours and 25 minutes later, on May 26 at 04:05, a 100 m long  
 198 eruptive fissure opened. This proximal eruption occurred along the Southeast Rift Zone,  
 199 2800 m southeast of the summit cone, at an elevation of 1890 m. A small volume of lava,  
 200 estimated around 0.5 Mm<sup>3</sup> from satellites and photogrammetry measurements was em-  
 201 placed (Fig. 1C). This eruption lasted for 27 hours, and only a few direct observations  
 202 were available due to its location and unfavorable weather conditions. Witnesses reported  
 203 moderate lava fountain activity a few hours after the eruption onset.

204 We focus on the May 2016 eruption for two main reasons. First, a Sentinel–1 SAR  
 205 image was acquired on May 26 at 01:45 UTC, i.e. during the seismic crisis that pre-  
 206 ceded the eruption onset. By combining this image with the previous and the following  
 207 Sentinel–1 acquisitions for PdF, we were able to compute two interferograms. The first  
 208 interferogram (hereafter referred to as S1 D1 for Sentinel–1 Descending 1) imaged the dis-  
 209 placement related to the earliest part of the magma propagation. The second interferogram  
 210 (hereafter referred to as S1 D2 for Sentinel-1 Descending 2) imaged the displacement re-  
 211 lated to the shallowest part of the magma propagation to the surface and to the opening  
 212 of the eruptive fissure. Such data provide unique constraints that may provide insight into  
 213 the magma propagation. Second, considering the location of the vent, the seismic crisis  
 214 preceding the eruption was notably longer than usual. Based on a compilation of distal  
 215 and proximal eruptions that occurred at PdF from 1985 to 2016, seismic crisis generally  
 216 last from 20 minutes to 18 hours (Fig. 2A). The longer the seismic crises, the further it is  
 217 from the summit and the lower in elevation the fissure location [Aki and Ferrazzini, 2000;  
 218 Peltier *et al.*, 2005]. A linear trend can be plotted which corresponds to a mean veloc-  
 219 ity for lateral propagation of 0.19 m.s<sup>-1</sup>, (Fig. 2A). For the May 2016 eruption, consider-  
 220 ing the summit-vent distance (2800 m), the seismic crisis was expected to last ~3h30min,  
 221 which is significantly shorter than the observed 8h25min duration. Another central ques-  
 222 tion concerns the reason why the magma stalled for 5 hours before erupting. If this delay  
 223 was due to a slower than usual propagating intrusion then, according to Corbi *et al.* [2016]

224 it should have been more sensitive to the stress field. We postulated that this would make  
 225 the study of this intrusion particularly informative.

### 226 3 Available observations for the May 2016 Eruption

#### 227 3.1 Final ground surface displacement distribution

228 Thanks to the routine InSAR monitoring at PdF, the May 2016 eruption was im-  
 229 aged by acquisitions from the Cosmo-Skymed (CSK, X- band) and the Sentinel-1 (S1,  
 230 C-band) satellites in Stripmap mode along ascending and descending tracks (Figs. 3 and 4,  
 231 Table S1). The computed interferograms are high spatial resolution (5 m by 5 m pixels).  
 232 The main processing steps are described in the Supporting Informations Text S1 [CNES,  
 233 1996; *Massonnet and Feigl*, 1998; *Chen and Zebker*, 2000].

234 Descending interferograms (CSK D and S1D) show a single lobe centered to the  
 235 southeast of the Central Cone, with maximum Line Of Sight (LOS) displacements of  
 236 about 40 cm and 35 cm toward the Sentinel-1 and CSK satellites, respectively (Fig. 4).  
 237 Ascending interferograms (CSK A and S1A) show patterns with two lobes. The western  
 238 lobe has maximum LOS displacements of 17 cm and 12 cm towards the Sentinel-1 and  
 239 CSK satellites, respectively, while the eastern lobe has displacements of up to 10 cm and  
 240 13 cm away from the Sentinel-1 and CSK satellites, respectively. The small differences  
 241 between Sentinel-1 and CSK interferograms are well explained by the differences in LOS  
 242 (see Table S1).

243 Based on deformation patterns (<https://www.obs.univ-bpclermont.fr/casoar>), the May  
 244 2016 eruption shares strong similarities with 5 eruptions which have taken place since  
 245 1998 (Fig. S1), that in June 2000, October 2010, May 2015, January and July 2017. For  
 246 the June 2000 eruption, vertical ascent beneath the Dolomieu crater followed by lateral  
 247 migration of the intrusion to the surface was determined from tilt and seismic data by  
 248 *Peltier et al.* [2005]. Ascending and descending InSAR data which captured this eruption  
 249 were inverted by *Fukushima et al.* [2010], showing that the intrusion responsible for the  
 250 displacement was a curved dike inclined to the northeast. All the others occurred after the  
 251 major April 2007 collapse and no study of their deformation sources has been published  
 252 yet.

253 Ten continuous GNSS stations (cGNSS) of the OVPF network (Fig. 1) recorded dis-  
 254 placements due to the May 2016 magma propagation. The daily position of each station is  
 255 calculated using the GAMIT/GLOBK software package [*Herring et al.*, 2010], providing  
 256 solutions with standard deviations of 0.5 cm and 1 cm for horizontal and vertical com-  
 257 ponents, respectively. GNSS data processing is described in the Supporting Information  
 258 Text S2 [*Herring et al.*, 2010; *Nikolaidis et al.*, 2001]. Displacements between August 28,  
 259 2015 and May 30, 2016 have also been measured in static-rapid mode at 80 locations of  
 260 the campaign GNSS network, on the central cone. Comparing cGNSS data with data from  
 261 this campaign GNSS network (Fig. 5), we find that it provides less precise solutions. For  
 262 example, horizontal and vertical uncertainties are around 1-2 cm and 4-5cm respectively.  
 263 However, it improves the spatial resolution of GNSS measurements. Records from daily  
 264 GNSS measurements show that the displacements recorded during the inter-eruptive pe-  
 265 riod were low with respect to the co-eruptive displacements (3-5 cm versus tens of cm)  
 266 and can be neglected. Consequently, we assume that the deformation measured with the  
 267 campaign GNSS and detected by InSAR is mainly due to the May 25 magma propagation.  
 268 The cGNSS stations which recorded the largest displacement are located at the south-  
 269 east border of the Dolomieu crater (DERG  $21.2 \pm 2.2$  cm, DSRG  $18.5 \pm 2.3$  cm, SNEG  
 270  $13.5 \pm 2.2$  cm) or at the eastern base of the cone (FERG,  $21.2 \pm 3.3$  cm) (see Table S2). Be-  
 271 cause the campaign GNSS network has more stations on the volcano than the continuous  
 272 network, and due to its different spatial distribution, it recorded larger displacements, of  
 273 up to  $24.5 \pm 5.5$  cm to the east of Dolomieu summit crater.

274 Comparing InSAR data with the campaign GNSS and the cGNSS data, both pro-  
 275 jected along the LOS, we assess the consistency between these data. On Figure. 4, this  
 276 consistency is illustrated by the similarity between the background color, representing In-  
 277 SAR displacement, and colors filling the squares and circles representing the GNSS con-  
 278 tinuous and campaign measurements, respectively, with the difference being less than the  
 279 standard deviation on measurements. Such consistency was also established by *Peltier*  
 280 *et al.* [2017] for the 4 eruptions of 2015 at PdF. However, even though both GNSS net-  
 281 works recorded significant deformation, the stations are far from the area of maximum  
 282 deformation measured by InSAR. The campaign GNSS network is centered on the Central  
 283 Cone while the deformation is the greatest in the southeast area. Therefore, we postulate  
 284 that since GNSS and InSAR data are consistent and campaign GNSS have large uncertain-  
 285 ties, GNSS data do not provide any useful constraints on the final location of the intrusion  
 286 (see section 5.1). In the following, to simplify, we do not include GNSS data when mod-  
 287 eling the final geometry of the intrusion.

### 288 3.2 Temporal observations

289 From the temporal evolution of the tilt, seismicity and Real-time Seismic Amplitude  
 290 Measurement (RSAM) ratio between DSM (summit) and FOR (southeast flank) stations,  
 291 the unusually long May 2016 pre-eruptive seismic crisis can be divided to four phases  
 292 (Fig 2B). It started at 19:40 UTC (phase Ia) with an intense seismic swarm, mostly af-  
 293 fecting the summit as shown by RSAM ratio increase, during which about 430 micro-  
 294 earthquakes occurred. We managed to locate twenty three of these events which were all  
 295 beneath the summit. Thirty minutes later (i.e. at 20:10, phase Ib), tilt inside the Enc-  
 296 los Fouqué accelerated (Fig. 2B), showing an inflation of the summit crater. From 20:40  
 297 (phase II), tilt direction at a station on the southeastern border of the Dolomieu crater  
 298 (DSOi) abruptly changed while seismicity rates decreased up to 22:00 UTC with, 2 out  
 299 of 370 micro-earthquakes located. Seismicity then stabilized (phase III) at a lower, but  
 300 still above baseline, level of approximately 100 earthquakes per hour for about 4 hours,  
 301 until the onset of a second intense seismicity swarm from 02:15 to 03:15 UTC on May 26  
 302 (phase IVa). The decrease of the seismicity rate (phase IVb) coincided with a strong de-  
 303 crease in RSAM ratio between DSM and FOR stations, in agreement with the location in  
 304 seismicity on the southeast flank (9 out of 600 events recorded were located). The erup-  
 305 tion started at 04:05. During phases III and IV, from 22:00 to the eruption onset, tilt sig-  
 306 nal variations decreased rapidly and no significant signal was recorded.

307 A seismic swarm such as observed in phase I, followed by a decrease of seismic  
 308 activity and tilt rotation as seen in phase II is often observed at PdF [*Roult et al.*, 2012].  
 309 The classic interpretation is that magma firstly propagates vertically beneath the summit  
 310 and then laterally up to the eruption onset or the end of the intrusion [*Peltier et al.*, 2005;  
 311 *Fukushima et al.*, 2010]. However, phase III and especially phase IV are unusual for this  
 312 volcano. Another uncommon point is the occurrence of many Long Period events (LP)  
 313 recorded during phases II (31 events) and III (18 events). Such events at PdF are generally  
 314 associated with lateral propagation [*Aki and Ferrazzini*, 2000].

315 Tilt data provide valuable insights into the very beginning of the propagation (phases  
 316 I and II), but because the tiltmeters are centered on the summit, this dataset is then in-  
 317 sensitive to changes that may have occurred during lateral propagation. By chance, an  
 318 additional Sentinel-1 image along a descending track was acquired during phase III at  
 319 01:45 on 26 May, allowing two interferograms to be produced covering the beginning (S1  
 320 D1) and the end (S1 D2) of the propagation (Figs. 3, 4E-F, S2 and Table S1). Compar-  
 321 ing S1 D1, which recorded the first part of the intrusion propagation, with the Sentinel-1  
 322 descending (S1 D) interferogram covering the whole propagation (Fig. 4D and E and  
 323 S2), shows that most displacement had already occurred before 01:45 : the maximum  
 324 LOS displacement of S1 D1 (38 cm) achieved almost the maximum displacement of S1  
 325 D (40 cm). S1 D2, which recorded the final part of the propagation (Fig. 4F and S2) only



326 shows a small asymmetric displacement reaching  $\sim 20$  cm toward the satellite, close to the  
 327 eruptive fissure, and a large wavelength phase shift which corresponds to a maximum of  
 328  $\sim 5$  cm toward the satellite east of the summit cone between May 26 at 01:45 and June 7.  
 329 However, such a displacement is not shown by the cGNSS measurements, for which only  
 330 1 cm of eastward displacement was recorded, along with negligible northward displace-  
 331 ment and subsidence. This suggests that at least part of the phase shift revealed by InSAR  
 332 could be of atmospheric origin.

333 cGNSS data are also processed with the TRACK software providing epoch by epoch  
 334 solutions (i.e. one position every 30s). The GITg station, located 4 km from the summit,  
 335 is used as a reference. In order to increase the signal to noise ratio, we corrected the time-  
 336 series from multi-paths [Nikolaidis *et al.*, 2001] and performed a Principle Component  
 337 Analysis (PCA). The standard deviation on the post-processing cGNSS data was computed  
 338 for horizontal (1-2 cm) and vertical (3-4 cm) components (see details in the Supporting  
 339 Information Text S2). Both co-eruptive displacements measured with daily solutions and  
 340 post-processed cGNSS, taking the reference on May 25, at 17:00 UTC, are in very good  
 341 agreement. Time evolution of the cGNSS displacements (Figs. 5 and S3) confirms that  
 342 most of the deformation occurred at the beginning of the crisis. This timing is consis-  
 343 tent with the four phases previously described based on the time evolution of seismicity,  
 344 RSAM and tilt. From 20:10 (phase Ib), summit stations (SNEg, DERg, DSRg) moved in  
 345 a centrifugal pattern. From 20:25, stations further to the east (FERg) and south (FOAg)  
 346 started moving northeastward and southwestward, respectively. At that time, displacements  
 347 were mainly horizontal, with amplitudes of the order of their standard deviation. This re-  
 348 sult is in agreement with the classic pattern for the volcano where tilt and seismic data are  
 349 the first to record unrest. From 20:40 (phase II), all summit displacement vectors turned to  
 350 the northwest then stopped between 21:00 and 21:15, while the stations at the base of the  
 351 cone, FERg and FOAg, were still moving slowly to the east and south, respectively. Coin-  
 352 cident with this change in direction, the stations started moving vertically. DSRg reached  
 353 an amplitude of 12 cm, greater than the horizontal displacements. No significant deforma-  
 354 tion was recorded after 22:00 (phases III and IV) by the cGNSS network. This is consis-  
 355 tent with InSAR data, which showed that the final part of the eruption (S1 D2) took place  
 356 in an area lacking cGNSS monitoring (Fig. 4F and S2). Despite its density, the permanent  
 357 network is too sparse to comprehensively sample the displacement field produced during  
 358 the May 2016 eruption. This observation supports the idea that an integrated approach is  
 359 required to fully interpret the ground deformation time series in term of magma propaga-  
 360 tion dynamics.

#### 361 **4 Inverse Modeling Methods and Strategies**

362 In order to determine the source of ground deformation measured by InSAR and  
 363 GNSS, we use a 3D-Mixed Boundary Element forward model [Cayol and Cornet, 1997,  
 364 1998a] combined with a neighborhood inversion algorithm [Sambridge, 1999a; Fukushima  
 365 *et al.*, 2005]. We first invert the 2 descending (CSK D and S1 D) and 2 ascending (CSK A  
 366 and S1 A) interferograms covering the whole eruptive crisis (see Figure 3 and Table S1)  
 367 to determine the final shape of the magma intrusion (Fig. 6A). In most cases at PdF, once  
 368 the eruptive dike is emplaced, even if an eruption is ongoing, no significant deformation  
 369 is observed, probably indicating that the conduit used to supply lava remains the same  
 370 [Fukushima *et al.*, 2010]. Because the cGNSS time series evidence no significant deforma-  
 371 tion except during the 8 hours preceding the onset of the eruption, we consider that  
 372 deformation for the whole eruption is the same as deformation resulting from the propaga-  
 373 tion of the intrusion to the surface. We then track lateral magma propagation by inverting  
 374 displacements recorded by the cGNSS network for progressively longer durations, such  
 375 that the determined area corresponds to the cumulative pressurized part of the intrusion.  
 376 Our approach, hereafter called the Projected Disk method, consists of using the mesh de-  
 377 termined by the inversion of the four interferograms spanning the whole eruptive crisis as

378 an *a priori* for the source location and shape, inverting for a circular area projected on this  
379 mesh (Fig. 6B).

#### 380 4.1 Forward modeling

381 Ground displacement for linear elastic homogeneous media is computed using a 3-  
382 D Mixed Boundary Element Method [Cayol and Cornet, 1997]. This method combines a  
383 direct method with a displacement discontinuity method to allow for fast and precise com-  
384 putations. The method considers tensile cracks and shear fractures as well as realistic sur-  
385 face topographies, as neglecting topography was shown to induce errors in source depth  
386 and volume characterization [Cayol and Cornet, 1998b; Fukushima et al., 2005; Master-  
387 lark, 2007]. The edifice is assumed to have a Young’s Modulus of 5 GPa and a Poisson’s  
388 ratio of 0.25, as proposed by Fukushima et al. [2005] for PdF. Boundaries are discretized  
389 with triangular elements. The topographic mesh is constructed from a Digital Elevation  
390 Model that includes the bathymetry. The mesh extension is about five times greater than  
391 the deformed area in order to limit edge effects to a few percent [Cayol, 1996]. To pro-  
392 vide the best compromise between computation time and accuracy, the mesh is dense  
393 close to the eruptive fissure (50 m between nodes) and progressively coarser further away  
394 (Fig. S4).

395 To determine the geometry of the displacement source, only 9 geometrical param-  
396 eters (Table 1 and Fig. 6A) are required. Because of the reduced number of parameters  
397 required compared to kinematic inversions for instance, this method is statistically more  
398 significant [Tridon et al., 2016]. The location of the eruptive fissure is known from field  
399 observations and used as an input to the model. Because the May 2016 eruption has one  
400 eruptive fissure, we search for a quadrangle-shaped intrusion linked to this fissure by a  
401 single *echelon*. This quadrangle could be curved along the strike and the dip direction.  
402 Preliminary models show that, as for the June 2000 eruption [Fukushima et al., 2010],  
403 such a curvature significantly increases the data model agreement. To follow the propa-  
404 gation of the deformation source before it reaches the surface, we use the Projected Disk  
405 method which only requires 3 geometrical parameters as an *a priori* on the geometry is  
406 provided of the curved quadrangular fracture, previously determined from InSAR data in-  
407 version (Table 1 and Fig. 6B). We define the mean plane for the *a priori* curved quadran-  
408 gular fracture. On this plane a circular source is defined which is projected to the curved  
409 quadrangular fracture. Parameters for the inversion of this source are the radius and lo-  
410 cation of the source before projection. This method allows us to determine a source con-  
411 sistent with both InSAR and cGNSS data, reducing the number of inverted parameters  
412 and fixing the depth by the use of an *a priori* mesh. Because the disk can be partially pro-  
413 jected beyond the mesh, we interpret the area change and not the radius change.

414 Prescribed boundary conditions are pressures changes. The ground surface is as-  
415 sumed to be stress-free and intrusions are assumed to open as a response to the difference  
416 between the magma pressure and the normal stress exerted by the host rock to their sur-  
417 face. We assume these stress changes are homogeneous. Linearly varying overpressure  
418 could also be assumed, but a previous study [Fukushima et al., 2005] showed that over-  
419 pressure gradients could not be constrained at PdF.

#### 420 4.2 Non linear Inversions

421 Following, Fukushima et al. [2005, 2010] and Wauthier et al. [2012, 2013], we use  
422 a Monte Carlo Neighborhood Algorithm to invert the deformation data. The inversion  
423 process involves determining the forward models which best fit the data. The fit between  
424 modeled and observed data is expressed through the misfit function,  $\chi^2$ , calculated in the  
425 least square sense for each forward model as:

$$\chi^2 = (\mathbf{d}_o - \mathbf{G}(\mathbf{m}))^T \mathbf{C}_d^{-1} (\mathbf{d}_o - \mathbf{G}(\mathbf{m})) \quad (1)$$

where  $\mathbf{d}_o$  represents the vector of observed displacements,  $\mathbf{G}(\mathbf{m})$  is the vector of modeled displacements such that  $\mathbf{m}$  is the set of model parameters and  $\mathbf{G}$  represents the model prediction.  $\mathbf{C}_d$  is the covariance matrix on the data. It accounts for correlated noise related to the atmospheric contribution to InSAR data, and for the variance of cGNSS data, as well as for the modeling errors. For explanations about construction of the cGNSS and InSAR data vector, as well as the covariance matrix, refer to Supporting Information Text S3 [Tarantola, 1987; Fukushima et al., 2005, 2010; Welstead, 1999; Jónsson et al., 2002; Sudhaus and Jónsson, 2009; Tridon et al., 2016] and Figure S5.

We also calculate the % explained data for the  $i^{th}$  interferogram ( $\%Ed^i$ ) as:

$$\%Ed^i = \left( 1 - \sqrt{\frac{(\mathbf{u}_{obs}^i - \mathbf{u}_{mod}^i)^T (\mathbf{u}_{obs}^i - \mathbf{u}_{mod}^i)}{\mathbf{u}_{obs}^{iT} \mathbf{u}_{obs}^i}} \right) * 100 \quad (2)$$

where  $\mathbf{u}_{obs}^i$  and  $\mathbf{u}_{mod}^i$  are the data and model vectors, respectively, corresponding to all pixels in the masked interferograms. This value is independent of the covariance matrix, the subsampling and the dataset used to run the inversion.

The inversion consists of two stages: a search and an appraisal stage. The search stage [Sambridge, 1999a] aims to find an ensemble of models that preferentially sample the good data-fitting regions of the parameter space. The appraisal stage [Sambridge, 1999b] concerns the estimation of model uncertainties. Following the Bayesian inference, the population of models calculated during the first stage is resampled with a Monte Carlo integration procedure, enabling the calculation of the posterior probability density function (PPD) without any new calculation of the forward problem (see details in the Supporting Information Text S4 [Sambridge, 1999a, 1998; Fukushima et al., 2005; Tridon et al., 2016; Fukushima et al., 2010; Wauthier et al., 2012]).

### 4.3 Strategy for the inversion of the GNSS time series

The intermediate S1 D1 interferogram provides good spatial-resolution information on the displacement with a good spatial resolution, while the intrusion was underway. We used this data independent of the GNSS time series to validate our Projected Disk method. We invert the shape and location of the intrusion using either GNSS, InSAR (S1 D1) or both datasets together. We find that introducing *a priori* on the source location is needed to be able to invert cGNSS data and that the Projected Disk method leads to consistent results between InSAR and cGNSS data even when one datatype is omitted (Table 2). We next perform independent inversions considering progressively longer durations of the cGNSS recording starting at  $t_0 = 17:00$  UTC. The first time step in the inversion corresponds to the moment when the cGNSS signal surpasses the standard deviation, i.e. at 20:40. Length of the time steps is adjusted to the displacement change, using smaller time steps when larger changes are observed. We invert sources for 29 successive time intervals, with 5 min time steps between 20:40 and 21:30, then 10 min time steps up to 22:30 and finally 30 min time steps until the eruption onset at 04:05. As the intrusion proceeds, changes in the GNSS signal become smaller (Fig. 5). One more calculation at 01:45 was made for comparison with InSAR data (S1 D1) which lead to 30 time steps computed.

## 5 Results

### 5.1 Final geometry of the intrusion derived from four interferograms

Inverting the 4 InSAR data spanning the whole eruption, we find that the best-fit final geometry for the intrusion is a low dipping intrusion at a mean depth of  $800 \pm 60$  m a.s.l., slightly dipping to the northeast ( $\sim 8^\circ$ ). We refer to this intrusion as a sill for two reasons: i) the dip is of the order of the topographic slope above the intrusion; ii) in a

471 similar basaltic shield volcano context at Fernandina, Galapagos, such a low dipping in-  
 472 trusion was also referred to as a sill. This sill is rooted to the southeast of the Dolomieu  
 473 crater, and covers a quadrangular area of about 2700 m  $\times$  600 m. It reaches the surface  
 474 turning into a sub-vertical 880 $\pm$ 16 m high dike (Tables 2 and S3, Fig. 7) feeding the erup-  
 475 tive fissure. A plot of the opening shows that the northern corner of the sill did not really  
 476 open (Fig. 7A) and that the intrusion shape is probably less angular at depth than pre-  
 477 dicted by the model. Tests of joint inversions conducted using GNSS data in addition to  
 478 InSAR, even if their weight is increased to be the same as InSAR data, do not modify the  
 479 obtained model or the data fit (see Supporting information S5 [Tarantola, 2005; Dupu-  
 480 tel et al., 2014; Auriac et al., 2013; Yabuki and Matsu'Ura, 1992; Akaike, 1980; Fukuda  
 481 and Johnson, 2010; Sigmundsson et al., 2015]) confirming that the addition of GNSS data  
 482 is not necessary. Our best-fit model well explains the 4 interferograms (Table 2), with a  
 483 better fit (95%) for the descending tracks than for the ascending ones (82%) and an un-  
 484 derestimation of the eastward displacement east of the summit cone (Fig. 8). The ap-  
 485 praisal (Fig. 7B) does not evidence any trade-off between parameters. We estimate the  
 486 volume of this intrusion to be 2.5 Mm<sup>3</sup> with an average opening of 0.5 m (Table 2). Con-  
 487 sequently, 2.5 Mm<sup>3</sup> of the magma was stalled at depth, while 0.5 Mm<sup>3</sup> of lava was emit-  
 488 ted. Neglecting volume changes due to compressibility and degassing, an approximation  
 489 of the total volume supplied to the volcano leads to 3 Mm<sup>3</sup>. We estimate that 80% of  
 490 the magma remains trapped at depth, which is unusual at this volcano, where most of the  
 491 magma is usually emitted. Inversions of InSAR displacements for 5 eruptions from 1998  
 492 to 2000, and inversions of campaign GNSS data for 6 eruptions between 2004 and 2006,  
 493 [Fukushima et al., 2010; Peltier et al., 2008] led to estimate that 17% and 15 %, respec-  
 494 tively, of the magma supplied remained to contribute to the intrusive growth of the edifice.

## 495 **5.2 Temporal evolution**

### 496 **5.2.1 Initial upward propagation from tilt and seismicity**

497 From 20:10 to 20:40, vertical cGNSS displacements were four times smaller than  
 498 their standard deviation, and only recorded at the 4 summit stations. As a consequence,  
 499 the source can not be constrained from these data. However, tilt and seismicity [Peltier  
 500 et al., 2005; Battaglia et al., 2005; Fukushima et al., 2010], which usually record unrest  
 501 earlier than the cGNSS data, suggest that PdF intrusions first propagate upward below the  
 502 summit crater then laterally. In coherence with these authors, we will assume that the first  
 503 minutes of the crisis correspond to an upward dike propagation (phase Ib). Assuming the  
 504 reservoir is around sea level [Peltier et al., 2007, 2009a; Lengliné et al., 2016, e.g.] and  
 505 the intrusion stops propagating vertically when it reaches the lowest point in the identi-  
 506 fied sill (600 m a.s.l.), we can compute a vertical propagation velocity of 0.3 m.s<sup>-1</sup>. As  
 507 noted before for five eruptions that took place between 1998 and 2000 [Fukushima et al.,  
 508 2010], InSAR data show no evidence of the vertical conduit connecting the reservoir and  
 509 the modeled intrusion. This can be explained either by the conduit being too narrow to  
 510 be distinguished from the signal of the intruded sill, or by the closure of this conduit as  
 511 magma propagated upward, consistent with the propagation of a buoyant magma.

### 512 **5.2.2 Beginning of the propagation from the first intermediate InSAR data : a vali- 513 dation of the Projected Disk method**

514 We take advantage of the S1 D1 interferogram computed using a Sentinel-1 image  
 515 acquired during the propagation, 2h20 before the eruption onset, to validate our Projected  
 516 Disk method and to gain information on the position of the intruded magma at this stage.  
 517 We perform 3 inversions, first using only cGNSS data, second using only the S1 D1 In-  
 518 SAR data and last inverting both datasets jointly. We postulate that the robustness of the  
 519 method is reflected by its ability to retrieve data omitted in the inversions. Thus we also  
 520 compute the % of explained data on the data (InSAR or cGNSS) omitted in the inversion.  
 521 We find that the Projected Disk method works well for explaining data (either InSAR or

cGNSS) omitted in the inversion (Table 2). For all dataset used, the horizontal location of the fracture is well retrieved and parameters are close (Fig. 9). When only cGNSS data are used, the source overpressure and volume are overestimated as 60% and 35%, respectively, and the source area is underestimated by 10% compared to the joint inversion of cGNSS and InSAR data. This is probably related to the spatial distribution of cGNSS data with respect to the intrusion. Inverting only InSAR data leads to equally low misfits for cGNSS or InSAR data as for jointly inverted cGNSS and InSAR data (Table 2), confirming that when InSAR data are available, adding the cGNSS data does not improve the fit (Supporting information Text S5). Finally, joint inversion of these two datasets leads to the shortest confidence intervals (Table. S4). However, using only the cGNSS data, the fit of the omitted InSAR data remains acceptable (77% Table 2). This, together with the ability of the method to determine the model parameters (Fig. S6), indicates that this method is a satisfactory way of tracking magma from cGNSS data.

The inversion shows that, at 01:45, 77% of the area had already opened and that this opening mainly concerned the sill part of the intrusion. This sill had an average opening of 1 m, while the dike part of the intrusion remained closed. We estimate the volume of magma in the intrusion to be 2.6 Mm<sup>3</sup> (Table 2) which is consistent but slightly larger than the volume of magma (2.5 Mm<sup>3</sup>) for the whole intrusion (Table 2).

### 5.2.3 Insights into the magma propagation dynamics from cGNSS data

At 20:40, the intrusion started as a sill with a 0.5 Mm<sup>2</sup> area, located southeast of the Dolomieu Crater, at 800 m a.s.l. (Figs 10, S7 and movie S1 in the Supporting Information). The source area slowly increased until 21:10. At 21:15, a sudden southward propagation, with an area increase of up to 3.8 Mm<sup>2</sup> was associated with a sudden decrease in pressure. At 21:30, the intrusion had reached its maximal extension and the area became stable. We estimate the mean velocity of the propagation front as being 0.6 m.s<sup>-1</sup> between 20:40 and 22:00, with a peak of 2 m.s<sup>-1</sup> around 21:15. From 21:30, while the area remained unchanged, pressure and volume slowly increased up to midnight. No significant evolution was evidenced after that time. In particular, neither the opening of the vertical fracture nor the opening of the eruptive fissure were recorded on the cGNSS network as stations are too far from the most deformed area. However, the interferogram S1 D2 covering the end of the eruption reveals that deformation was focused close to the location where the fissure later reached the surface.

Differentiating volume with respect to time provides insight into the magma inflow evolution (Fig. 10B). After a brief increase from 400 ± 100 m<sup>3</sup>.s<sup>-1</sup> to 900 ± 100 m<sup>3</sup>.s<sup>-1</sup> from 20:40 to 21:15, the flow rate strongly decreased up to 22:00 and stabilized at 30 ± 10 m<sup>3</sup>.s<sup>-1</sup>, without any significant variations until the eruption onset at 04:05.

### 5.2.4 End of the propagation from the second intermediate InSAR data

Inverting the S1 D2 interferogram spanning the end of the propagation with the Projected Disk method, the best model is a 880 m high subvertical dike with an average opening of 0.4 m (Fig. 11A, Table S5), which feeds the eruptive fissure. We determined a dike volume of 0.3 Mm<sup>3</sup> (Table 2). The source covering the beginning (blue sill) and the end (red dike) of the eruption cover complementary areas (Fig. 11) and are consistent with the seismicity that could be located (Fig. 7A). Consistency of both sources is also shown by the fact that the interferograms covering the whole eruption are well explained by the sum of the models (Fig. 11, last row) corresponding to the first part (Fig. 11, first column) and the second part (Fig. 11, last column) of the eruption derived from the S1D1 and S1D2 InSAR data, respectively. Our best model (Fig. 11B and Table 2) explains 83% of the InSAR data and is also consistent with a model derived using a quadrangular source (Supporting information Text S6).

We assume that the decrease in seismicity after the second major peak at 03:15 (Fig. 2B, phase IVb) coincided with the onset of vertical propagation, for the following reasons: (1) the second peak occurred at a time intermediate between the sill propagation and the final dike propagation as indicated by the intermediate SAR acquisition, (2) at PdF earthquake swarms are often associated with the onset of propagation, (3) laboratory hydraulic fracturing experiments show that the onset of acoustic emission (the equivalent of our *in situ* micro-earthquakes) is an indication of the initiation of fracturing and that acoustic emissions drop as soon as the fracture starts propagating [Zoback *et al.*, 1977]. Based on this assumption, we determine a vertical propagation velocity of  $0.3 \text{ m.s}^{-1}$  corresponding to 880 m of propagation in 50 min.

## 6 Discussion

### 6.1 Modeling limitations

Forward modeling used in this study relies on linear elasticity, and a homogeneous and isotropic medium. Solution of the inverse problem solved here is non unique, such that one displacement field could be explained by combinations of different source geometries or by numerous mechanical properties, for instance, inclusion of anisotropy or plasticity. Because no inclined dikes are visible on the walls of the PdF summit crater, Got *et al.* [2013] assumed that the observed asymmetric displacements could be attributed to vertical dikes emplaced in a plastic medium. However their models are only 2D. Also, most intrusions found in the plumbing system of the dormant neighbor Piton des Neiges volcano are inclined outward, which is consistent with a brittle edifice [Chaput *et al.*, 2017] and dikes determined from the inversion of Piton de la Fournaise intrusions [Fukushima *et al.*, 2010; Peltier *et al.*, 2008]. Thus, an elastic and homogeneous medium explains, at least, the first order displacement field. However, we acknowledge that, at the dike tip, plastic behavior probably occurs due to stress singularities [Jaeger *et al.*, 2009]. Moreover, for the May 2016 eruption, residuals east of summit cone have a different sign for the descending and ascending interferograms (Figs. 7B and 11B). This is consistent with an eastward residual displacement of the flank that our model is not able to explain. Considering the short time-scale studied and the fact that an elastic framework seems sufficient to explain most of the observations with a minimum number of parameters, this hypothesis is the most likely from a statistical point of view. However, we also note the need for models able to properly take into account more complex mechanical properties.

### 6.2 Integration of InSAR and cGNSS data

Despite the fact that PdF has one of the densest cGNSS networks in the world, this study shows some limitations of using only this cGNSS network when attempting to track proximal or distal intrusions. It must be noted that large areas are unmonitored to the north and southeast of the Central Cone where, in the past decades, numerous fissures opened and lava flows were emplaced [Roult *et al.*, 2012]. Because of this configuration, cGNSS measurements during the May 2016 proximal eruption missed the maximum deformation and provided little information on the deformation gradient. Another limitation of the summit coverage of the cGNSS network is that it is often blind to shallow distant processes, such as the final opening of the eruptive fissures. While, deformation associated with this opening is large (tens of centimeters) in amplitude but it only affects a small area (Fig. 4F). In order to combine the advantages of the high temporal resolution of cGNSS data and the good spatial resolution of InSAR data, we developed an approach in which the temporal evolution of the source is tracked by inverting for a circular pressure source projected on the *a-priori* source determined from InSAR data covering the whole eruption (Projected Disk method). We also tried another method which was free of such *a priori*. This method consisted of inverting the migrating source as a planar ellipse (see Supporting Information Text S7, Fig. S8 and Table S6). Both methods were tested

621 for their ability to determine the source characteristics from cGNSS data alone. Taking  
 622 advantage of the Sentinel 1 image acquired when the intrusion was underway, we com-  
 623 pared models from inversion of cGNSS data only with models from joint inversion of  
 624 InSAR and cGNSS for the same time step. This validation test revealed that the Ellipse  
 625 method was unable to determine sources consistent with InSAR data from cGNSS data  
 626 only (Figs. S9 and S10, Table S7 and S8). However, despite the cGNSS network configu-  
 627 ration not being optimal for this proximal eruption, using the Projected Disk method, the  
 628 inversion of cGNSS data led to source parameters consistent with inversions on InSAR  
 629 data and joint datasets and the fit of InSAR data omitted in the inversion is acceptable  
 630 (77%) (Figs. 9 and S9). The main limitation concerns the source pressure (60%) and vol-  
 631 ume (35%) with respect to those estimated using both cGNSS and InSAR data.

632 Monitoring with high spatial and temporal resolution analogue experiments of dikes  
 633 and cone sheet propagation, *Guldstrand et al.* [2018] proposed a method which used the  
 634 detection of the maximum uplift to predict the vent location. However, even at some of  
 635 the best monitored volcanoes, this approach is far from being applicable : (i) Propaga-  
 636 tion duration is usually too short to be detected by InSAR monitoring. Moreover, the pro-  
 637 cessing of InSAR data is not instantaneous. (ii) cGNSS networks would need to be very  
 638 dense. Here, with one of the densest networks worldwide, part of the propagation of prox-  
 639 imal or distal eruptions is still undetected by the network. (iii) the maximum uplift does  
 640 not always provide information on the vent location. The May 2016 eruption demonstrates  
 641 that fissure eruptions sometimes start as subhorizontal sills. In this case, the maximum up-  
 642 lift is roughly central to the source rather than being located at the intersection between  
 643 the fracture and the ground.

### 644 **6.3 Origin of the step-wise propagation**

645 The May 2016 intrusion is characterized by two changes in direction (vertical to  
 646 horizontal, then horizontal to vertical), as well as a step-wise horizontal propagation, with  
 647 one acceleration (21:15) and a pause before the final vertical propagation. Changes in the  
 648 propagation velocity can be attributed to different propagation directions (vertical / lateral  
 649 / vertical). Other changes, occurring while magma is propagating laterally along a planar  
 650 intrusion, probably have different origins. We investigate here the origin of the dike to  
 651 sill rotation, whether observed changes in lateral velocity are real and, if so, their origin,  
 652 the possibility that the intrusion was fed by a single batch of magma quickly disconnected  
 653 from the reservoir, and finally the potential role of gas exsolution and east flank slip.

#### 654 **6.3.1 Required conditions for dike to sill rotation**

655 A sudden change from vertical to lateral propagation is commonly observed at PdF  
 656 [*Peltier et al.*, 2005; *Fukushima et al.*, 2010]. In the May 2016 case, changes in intrusion  
 657 directions are an interesting feature. Here we investigate whether this dike to sill rotation  
 658 was a response to a rotation of the minimum principal stress [*Menand et al.*, 2010; *Macca-*  
 659 *ferri et al.*, 2011] or whether its geometry is influenced by a preexisting discontinuity.

660 Sills can alternate with dikes, due to stress rotation induced by successive intrusions.  
 661 If stresses in the edifice are close to isotropic, as assumed for the dormant neighbor vol-  
 662 cano Piton des Neiges [*Chaput et al.*, 2014a] or for Fernandina in the Galapagos [*Chad-*  
 663 *wick Jr and Dieterich*, 1995], dike intrusions would provide enough stress change to induce  
 664 a rotation of the minimum principal stress, further inducing sill intrusions. A relaxation  
 665 mechanism such as slip on the east flank would then be required to start initiate dike in-  
 666 trusions again. While relaxation of the east flank was demonstrated in 2007 [*Brenguier*  
 667 *et al.*, 2012; *Froger et al.*, 2015] and confirmed to be ongoing [*Peltier et al.*, 2015; *Chen*  
 668 *et al.*, 2017], systematic studies of eruption sequences [*Fukushima et al.*, 2010; *Peltier*  
 669 *et al.*, 2008] show no clear evidence of an alternation of dikes and sills at a given location,  
 670 though these studies covered only limited periods of time.

671 The rotation could also be favored by the unloading induced by successive caldera  
 672 collapses [Corbi *et al.*, 2015]. Among the 6 eruptions we identified as displaying sim-  
 673 ilar displacement patterns (see Fig. S1), both modeled intrusions (June 2000 and May  
 674 2016) [Fukushima *et al.*, 2010] present similar shapes and it is likely that all these erup-  
 675 tions share the same source shape, of a sill turning into a dike on its western edge. Only  
 676 one occurred before the last major collapse which affected the summit area in 2007 [Stau-  
 677 dacher *et al.*, 2009] by deepening the crater by 330 m. The other five took place after this  
 678 event. Thus, it is likely that the 2007 caldera collapse created a stress field more favorable  
 679 to sill intrusions.

680 Another possibility is that dikes turn into sills, when they encounter a more rigid  
 681 material, as shown by analog experiments [Kavanagh *et al.*, 2006] and numerical experi-  
 682 ments [Maccaferri *et al.*, 2010], or because they encounter a preexisting discontinuity [De-  
 683 laney *et al.*, 1986; Maccaferri *et al.*, 2011]. Because the modeled structure has a slight  
 684 northeast dip, its geometry could correspond to the upper limit of a flank sliding structure  
 685 repeatedly activated during magma injection. Because the intrusion geometry is aligned  
 686 with large scarps in the caldera rim, this structure perhaps corresponds to the limit of the  
 687 Grand Brûlé depression (Fig. 1) [Bachelery and Mairine, 1990; Merle and Lénat, 2003],  
 688 and may connects further east to the flank detachment associated with the 2007 eruption  
 689 [Tridon *et al.*, 2016].

### 690 **6.3.2 Step-wise lateral propagation**

691 The mean velocity estimated for the lateral propagation is  $0.6 \text{ m.s}^{-1}$ , which lies in  
 692 the range of previous estimates of  $0.2\text{--}0.8 \text{ m.s}^{-1}$  made for PdF, obtained from tilt and  
 693 seismicity data [Peltier *et al.*, 2005]. However, both the location of the propagating frac-  
 694 ture front (Fig. 10A) and the temporal evolution of the area and pressure, (Fig. 10B) re-  
 695 veal sudden variations in lateral propagation velocity, corresponding to an instantaneous  
 696 velocity of  $2 \text{ m.s}^{-1}$  between 21:10 and 21:15, then a pause from 22:00 up to the final ver-  
 697 tical propagation that started at 3:15.

698 One of the questions is whether the 21:15 velocity jump is real whether it is an ar-  
 699 tifact induced by the limited number of stations to the southeast of the volcano. In order  
 700 to test this hypothesis, we performed synthetic simulations where we simulated the cGNSS  
 701 displacements induced by a source spreading and/or migrating at a constant rate (Details  
 702 are provided in the Supporting Information Text S8, Figs S11 and S12). We determined  
 703 that displacement changes similar to those observed could be obtained provided the source  
 704 was both extending and migrating at the same time. Thus, considering the uncertainties  
 705 on cGNSS measurements, we cannot completely exclude that a continuous evolution of  
 706 the area and pressure induced the observed cGNSS data, and that the interpretation of a  
 707 velocity jump in the magma propagation is an artifact.

708 However, the number of earthquakes provides a strong constraint, and can be used as  
 709 a proxy to the propagation history. Laboratory experiments of hydraulic fracturing show  
 710 that propagation of a hydraulic fracture proceeds in several stages [Zoback *et al.*, 1977].  
 711 The fracture first initiates when the tensile strength is overcome (the initiation stage), and  
 712 it takes some time and a higher pressure for a macroscopic fracture to form and for fluids  
 713 to start propagating in the fracture (the breakdown stage). At this stage, pressure drops.  
 714 These different stages are associated with different rates of micro-fracturing as recorded  
 715 by acoustic emissions. The initiation stage corresponds to an increase in micro-fracturing,  
 716 and the breakdown stage corresponds to a decrease in seismicity rate. Using rate and state  
 717 friction [Dieterich, 1994], it is possible to relate the micro-fracturing to the stress history.  
 718 Thus, during injection, prior to the onset of fracture propagation, stress and seismicity rate  
 719 increase. When the fracture starts propagating, pressure and host rock stress decrease, and  
 720 seismicity returns to a background rate. Note that, after a stress decrease, seismicity does  
 721 not immediately stop, as the time to return to a background rate is an inverse function of



the stress rate [Toda *et al.*, 2002]. During the May 2016 PdF magma propagation, seismicity showed two peaks during the lateral magma propagation phase (Fig. 2B) at 21:15 and at 21:45. The first peak correlates with a rapid front advance (Fig. 10A), corresponding to an area increase and a pressure decrease (Fig. 10B). The second peak does not correspond to any area or pressure change, but this lack of evidence might be related to the cGNSS network configuration, and the lack of stations southeast of the volcano. Because of the simultaneous timing of seismicity rate and propagation steps inferred from cGNSS, we can conclude that the 21:15 propagation step is not an artifact induced by the cGNSS network configuration. It is also likely that there was another propagation step at 21:45.

The residuals observed in the InSAR measurement (Figs. 7B and 11B) suggest an eastward displacement of the east flank that the model is not able to explain. Strain acceleration evidenced by surface displacements and seismicity rates during the intrusion initiation indicates that damage and strain softening might occur in the edifice [Carrier *et al.*, 2015]. Similarly, as proposed by [Got *et al.*, 2013] for 2007, and frequently suspected during eruptive crises at PdF, stress accumulation may cause the eastern flank to exceed a plastic threshold, leading to eastward slip which may further facilitate lateral migration of an intruded sill. It is likely that the stress decrease due to eastward flank slip led to a burst in magma inflow (Fig. 10B) which could explain the jump in lateral propagation observed at 21:15.

### 6.3.3 Limited magma supply and rapid disconnection from the reservoir

An unusual feature of the magma propagation during the May 2016 eruption is the 5 hours pause (phase III) observed from 22:00 (Fig. 10), leading to an exceptionally long duration of the intrusion phase of 8h25min. Generally, to reach that distance from the summit, it should have only lasted 3h30 (Fig. 2). The small volume lava flow ( $0.5 \text{ Mm}^3$ ) could have been fed with no or little input of new magma from the reservoir after the sill injection. A rough estimation of the volume of magma involved in the whole eruption ( $3 \text{ Mm}^3$ ) corresponds to the sum of the magma volume estimated from the inversion of interferograms covering the whole eruption ( $2.5 \text{ Mm}^3$ , Table 2) and the erupted volume ( $0.5 \text{ Mm}^3$ ). Since we ignore the effect of magma compressibility, while summing estimated volume of a source at depth and volume of lava flows, it is likely that this total volume is slightly overestimated. However, this volume is only slightly ( $0.4 \text{ Mm}^3$ ) larger than the volume emplaced in the sill ( $2.6 \text{ Mm}^3$ , Table 2) prior to the final dike propagation and eruption. Moreover, the overpressure for the whole eruption (1.5 MPa, Table 2) is 0.7 MPa smaller than the overpressure (2.2 MPa, Table 2) estimated for the sill part of the eruption, indicating that the eruption opened the system and released some pressure. Lastly, the lava flow volume emitted only represents 20% of the magma supplied to the volcano during this eruption, whereas it usually represents 85% [Peltier *et al.*, 2008; Fukushima *et al.*, 2010], suggesting that the volume supplied for this eruption was small and mostly trapped at depth.

Because the magma inflow rate was low, and the pressure increase from 22:00 was small (Fig. 10), the intrusion might have lacked the overpressure required to overcome the barriers to propagation and thus been more prone to stop than an intrusion with a larger overpressure [Maccaferri *et al.*, 2011]. The low magma supply rate favors a scenario with a single small batch of magma disconnected from its feeding reservoir. This bottom closure could be due to a small reservoir rapidly depressurized by the intrusion propagation. This intrusion could have stopped before reaching the surface either due to the surrounding stress field [Maccaferri *et al.*, 2010] or to cooling [Taisne and Tait, 2011], leading to a failed eruption. However, propagation restarted vertically, without any evidence of a new burst in influx of magma.

### 6.3.4 Final dike propagation : gas exsolution or East flank slip ?

The final dike is radial and rooted at the western edge of the deformation pattern where the thickness of rocks above the sill is at its minimum (880 m below the surface). As a consequence, the dike initiated where the lithostatic stress was minimum, and where it was easiest to intrude. Two factors might have triggered the final intrusion either independently or in conjunction. One internal factor is related to the role of gas, while the other external factor is related to the role of the east flank slip (Fig. 12).

As intrusions propagate and reach higher elevations, gas exsolves, migrates and accumulates at the intrusion tip. In basaltic magmas, accumulated buoyant bubbles can even separate from the surrounding melt [Menand and Tait, 2001; Maimon *et al.*, 2012]. Menand and Phillips [2007] demonstrated that, because of cooling and solidification, gas segregates more efficiently in sills than dikes, and that gas segregation is more efficient for low supply rates. Here, the conditions are met for efficient gas segregation : the intrusion is a sill and the supply rate is low. Moreover, the dike part of the intrusion is located at the highest elevation of the sill, where gas would tend to accumulate. Although, there was little visual observation for this eruption, lava fountaining activity was reported, characteristic of a high gas content [Coppola *et al.*, 2009]. Laboratory observations show that the part of fractures filled with gas have less lateral extension than the liquid filled part of fractures [Menand and Tait, 2001]. Here, the intrusion geometry corresponds to a large sill, connected to a narrow dike. This narrowing pathway is consistent with upward fracture propagation triggered by gas segregation. Another argument in favor of the role of gas, is that 49 Long Period (LP) events were recorded during phases II and III of the seismic crisis. According to Zecevic *et al.* [2013], LP events are not common at PdF. When they do occur, they are generally associated with lateral propagation which leads to proximal or distal eruptions [Aki and Ferrazzini, 2000]. The source of LP events at PdF is not well understood, but the resonance of gas filled fractures is an efficient way of generating them [Chouet, 1996].

Sill intrusion is expected to favor flank slip [Famin and Michon, 2010; Got *et al.*, 2013; Chaput *et al.*, 2014b]. It relaxes stress perpendicular to the rift and may allow the dike to intrude along the rift. Since 2007 the east flank is known to slip at a steady rate of  $0.02 \text{ m.yr}^{-1}$  [Chen *et al.*, 2017], and transient slip sometimes occurs as a response to intrusions [Brenquier *et al.*, 2012; Peltier *et al.*, 2015; Froger *et al.*, 2015]. Three independent data suggest that a limited transient east flank slip carried on after the initial slip at 21:15: records on FERG station (up to 1 cm of eastward displacement during the week after the eruption), second intermediate InSAR S1 D2, and residuals from the inversion of the four interferograms spanning the whole eruption. We assume that such a transient slip would favor gas exsolution leading in turn to reactivation of the propagation vertically.

## 7 Conclusion

In this study, we combined InSAR and continuous GNSS (cGNSS) data to gain insights into the dynamics of a magmatic intrusion at Piton de la Fournaise. An InSAR image was acquired as the May 2016 eruptive crisis was underway, allowing us to test a specific approach dedicated to integrating InSAR and cGNSS data. We use the deformation source determined from the inversion of InSAR data covering the whole eruption as an *a priori* for tracking a circular pressurized area from cGNSS data. This method estimates the source depth intrinsically, while this depth is wrongly estimated when only the cGNSS data is used without any *a priori* derived from InSAR.

The best modeled intrusion corresponds to a 2700 m long sill, dipping to the northeast and turning into a 880 m high dike located on the western limit of the sill, along the south east rift zone. This intrusion has a similar sill to dike shape as the one determined for the June 2000 event. Because InSAR patterns are similar for the June 2000, October

821 2010, May 2015, May 2016, January and July 2017 eruptions, these eruptions probably  
822 correspond to the same intrusion geometry. Because most of these intrusions occurred  
823 after the 2007 caldera collapse, it is possible that they were guided by the stress field re-  
824 sulting from the collapse. Alternatively, their geometry, as well as their location, suggest  
825 that they might follow a pre-existing discontinuity, which might correspond to the upper  
826 limit of the Grand-Brûlé depression.

827 The May 2016 intrusion is characterized by two changes in direction from vertical  
828 to lateral, then from lateral to vertical. From the seismicity, it is likely that the reservoir  
829 started fracturing at 19:40 (Fig. 12, phase Ia) and from the tilt, that the dike started prop-  
830 agating upward at 21:10 (phase Ib). Then, at 20:40, the intrusion turned laterally to the  
831 southeast as a sill (phase II). Inverting cGNSS data at 30 successive time steps, we found  
832 that lateral magma propagation was unsteady. Magma slowly propagated until 21:15, then  
833 the sill area suddenly increased to reach its maximum dimension at 22:00. The velocity  
834 jump is confirmed by a peak in seismicity and could be associated with east flank sliding.  
835 Pressure and volume kept on increasing until midnight, while the intrusion was paused.  
836 After 5 hours of stand-by, the increase in rate of micro-earthquakes marked fracture initia-  
837 tion (phase IVa). This was followed by a decrease in the rate of micro-earthquakes which  
838 marked the onset of upward propagation at 03:15 (phase IVb), leading to a fissural erup-  
839 tion at 04:05. From volume budgets and the estimated magma flux, it is likely that the  
840 eruption was fed by little or no input of new magma after the initial supply of a single  
841 batch of magma. This could explain why the intrusion lasted longer than expected con-  
842 sidering the distance between the eruptive fissure and the summit. This intrusion could  
843 have been non-eruptive if it had not been for the additional factors favoring the ultimate  
844 eruption. Several arguments indicate that the final part of propagation might have been  
845 triggered by gas segregation in the sill. Consistent with observations of intrusions trigger-  
846 ing flank slip, the dike opening and upward propagation might also have been favored by  
847 an east flank slip as indicated by residual displacement on the InSAR data.

851 The proposed method relies on the integration of InSAR and cGNSS data. Since it  
852 requires InSAR data covering an intrusion, so far, it can not be used in a predictive way.  
853 However, it is a powerful way to track magma propagation fronts, and pressure changes  
854 after an eruption. It provides a valuable picture of an unsteady process characterized by  
855 a complex 3D geometry and potentially controlled by multiple factors. Assessing the rel-  
856 ative contributions of the proposed processes i.e. stress inherited from the 2007 caldera  
857 collapse, the east flank instability and the gas exsolution, on the control of the magma  
858 propagation dynamics will be the subject of further work.

Source	Parameter Name	Description
<b>Quadrangle linked to the surface</b>	Dip (°)	Dip of the dike (vertical=90°)
	Shear (°)	Shear angle in the direction of strike
	Botelev (m)	Above sea level elevation of the bottom side midpoint
	Botlen	Length of bottom side relative to the top
	Twist (°)	Twist angle (horizontal angle between top and bottom sides)
	Botang (°)	Vertical angle of the bottom side
	Dtop (m)	Vertical length of the segmented part
	Botcurv (°)	Bottom side curvature
	Vertcurv (°)	Dike vertical curvature
<b>Projected Disk</b>	$X_c, Y_c$ (km)	NS and EW coordinates of the center of the projected disk in a mean plane computed from <i>a priori</i> mesh
	Radius (m)	Radius of the projected disk

848

**Table 1.** Definition of the geometrical parameters used to define the fracture meshes used as deformation

849

sources. These parameters were first established by *Fukushima et al.* [2005] to restrict the search to mechani-

850

cally feasible models.

	Whole intrusion		Before 01:45		After 01:45	
<b>Quadrangle</b>	Overpressure (MPa)	1.5			1.1	
	Average opening (m)	0.5			0.2	
	Area (10 <sup>6</sup> m <sup>2</sup> )	4.8			1.1	
	Volume (10 <sup>6</sup> m <sup>3</sup> )	2.5	Not	Not	Not	0.2
	<b>% Explained data</b>		considered	considered	considered	
	InSAR CSK A	<b>82%</b>				
	InSAR CSK D	<b>95%</b>				
	InSAR S1 A	<b>82%</b>				
	InSAR S1 D	<b>95%</b>				
	InSAR S1 D2					<b>81%</b>
		<b>cGNSS only</b>	<b>InSAR only</b>	<b>cGNSS+InSAR</b>	<b>InSAR only</b>	
<b>Projected Disk</b>	Overpressure (MPa)		3.5	1.9	2.2	3.0
	Average opening (m)		1.0	0.6	0.7	0.4
	Area (10 <sup>6</sup> m <sup>2</sup> )		3.4	3.9	3.7	0.6
	Volume (10 <sup>6</sup> m <sup>3</sup> )	Not	3.5	2.4	2.6	0.3
	<b>% Explained data</b>	considered				
	cGNSS		<b>84%</b>	83%	<b>83%</b>	
	InSAR S1 D1		77%	<b>96%</b>	<b>96%</b>	
InSAR S1 D2					<b>83%</b>	

859

**Table 2.** Modeling results comparing different datasets and modeling strategies. Results obtained modeling

860

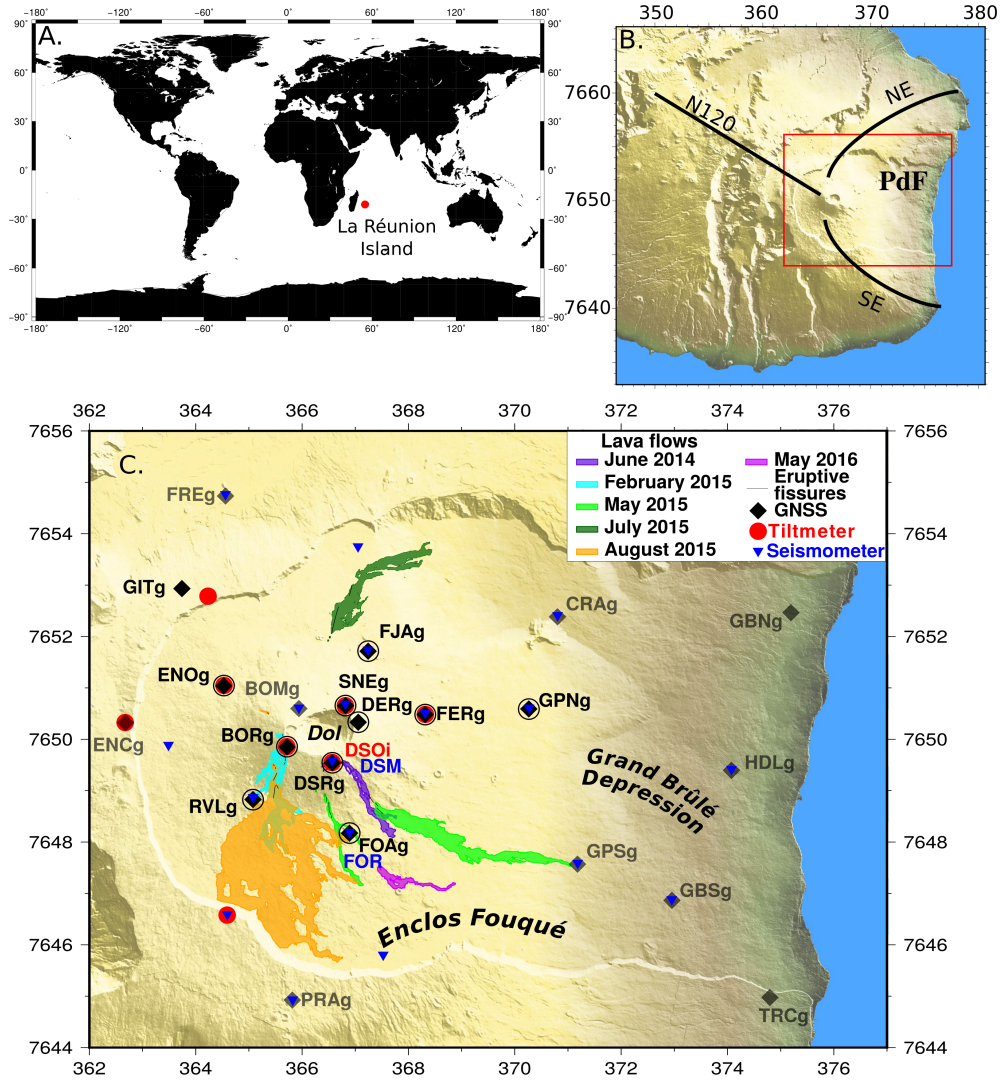
(top) the May 2016 intrusion with a quadrangle linked to the surface, (bottom) both parts of the eruption

861

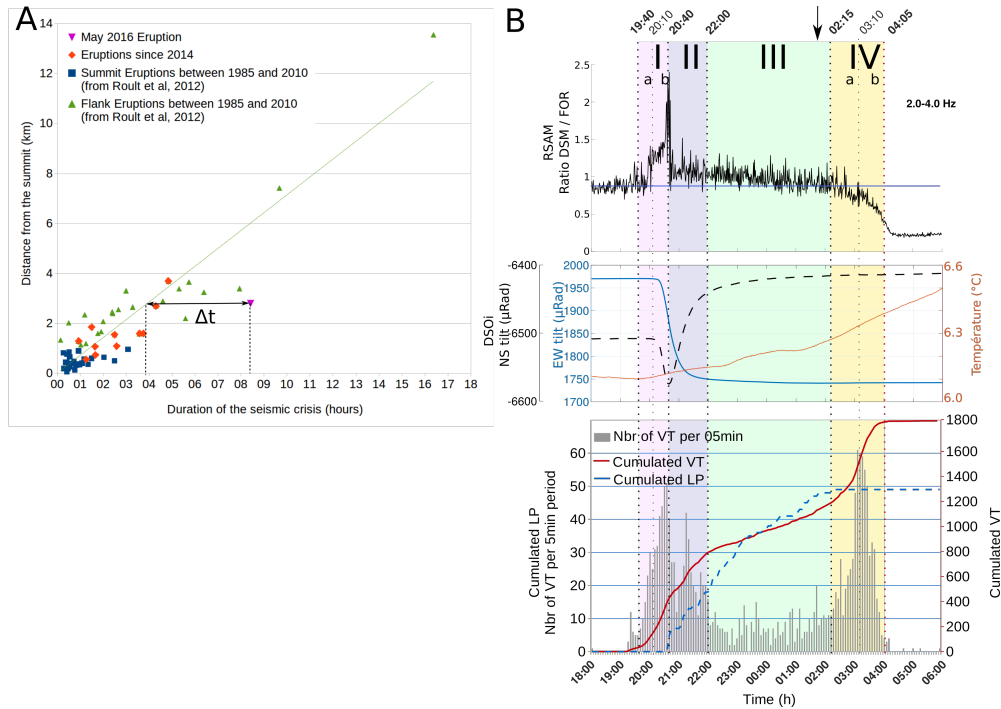
(before and after 1:45) with the Projected Disk method. % explained data computed with data used in the

862

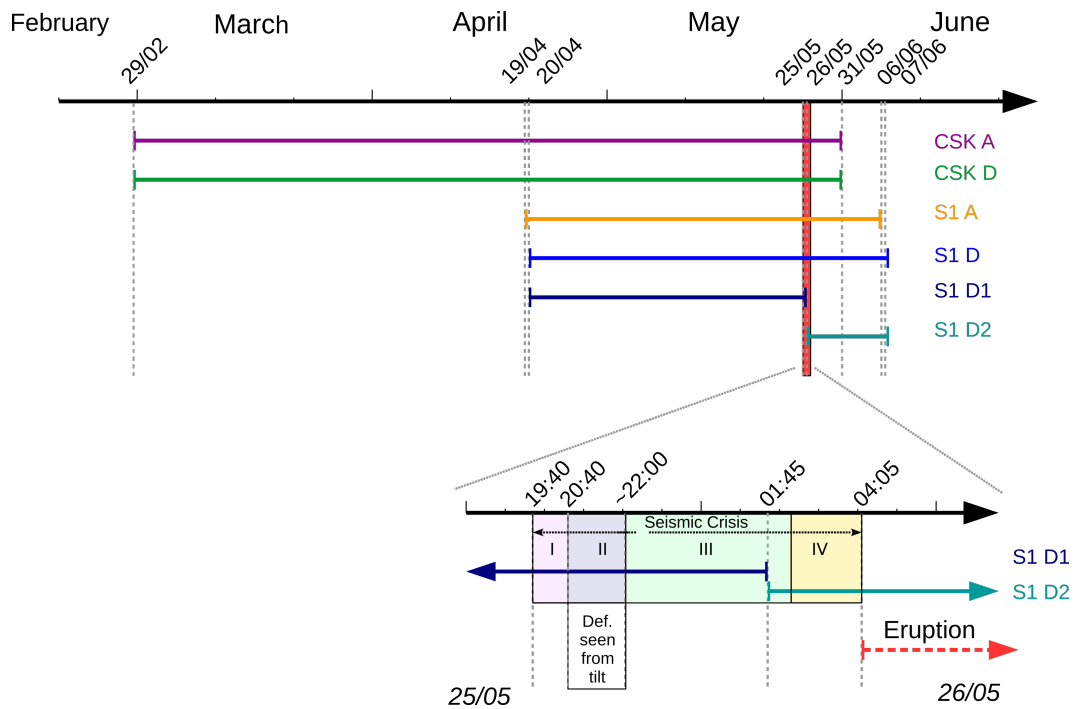
inversions are in bold, while those computed using data omitted in the inversion are in italic.



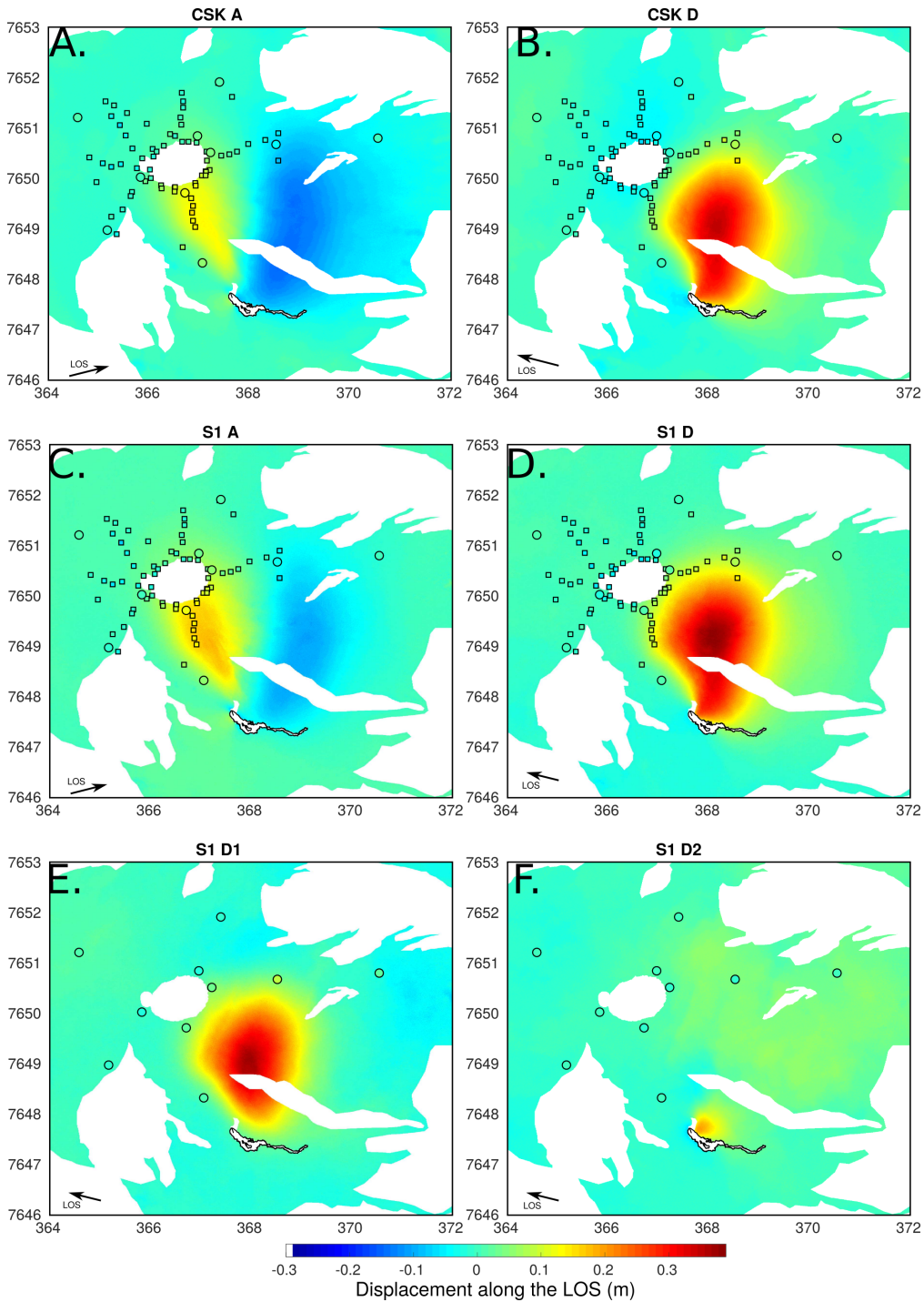
863 **Figure 1.** Piton de la Fournaise (PdF) setting. (A) Location of La Réunion Island. (B) Relief of the PdF  
 864 volcano. Black lines refer to the axe of the 3 rift zones. (C) Locations of the lava flows emplaced from June  
 865 2014 to May 2016 and a portion of the OVPF monitoring network. Black and grey diamonds represent the  
 866 locations of the permanent GNSS stations. The black circled diamonds refer to the GNSS stations used in  
 867 this study, the grey diamonds are stations not used here. Red dots and blue triangles represent the location  
 868 of the OVPF tiltmeters and seismometers, respectively. 'Dol' refers to the Dolomieu crater. Coordinates in  
 869 kilometers (WGS84, UTM 40S).



870 **Figure 2.** (A) Distance of the highest point of eruptive fissures from the center of the Dolomieu crater as  
 871 a function of the seismic crisis durations. Blue squares and green triangles represent summit eruptions and  
 872 flank eruptions between 1985 and 2010 (from *Roult et al.* [2012]), respectively. Orange diamonds represent  
 873 eruptions from 2014 and 2015. The purple triangle represents the May 2016 eruption. The green line rep-  
 874 represents the linear trend. (B) *Top* Ratio of RSAM amplitude between DSM and FOR stations (see Fig. 1 for  
 875 location). *Middle* NS (black dashes) and EW (blue line) tilt and temperature (orange line) variations at DSOI  
 876 station (see Fig. 1 for location). *Bottom* Histogram of the number of volcano-tectonic (VT) events per 5 min  
 877 periods during the May 2016 seismic crisis (left axis). The red curve represents the cumulated number of  
 878 VT events(right axis). The blue curve represents the cumulated number of long period events (LP, left axis).  
 879 Colored areas with numbers I, II, III and IV refer to the different stages of the intrusion propagation described  
 880 in 3.2. The black arrow represents the acquisition time of the Sentinel image used to compute interferograms  
 881 S1 D1 and S1 D2 (Fig. 3). Times are UTC.

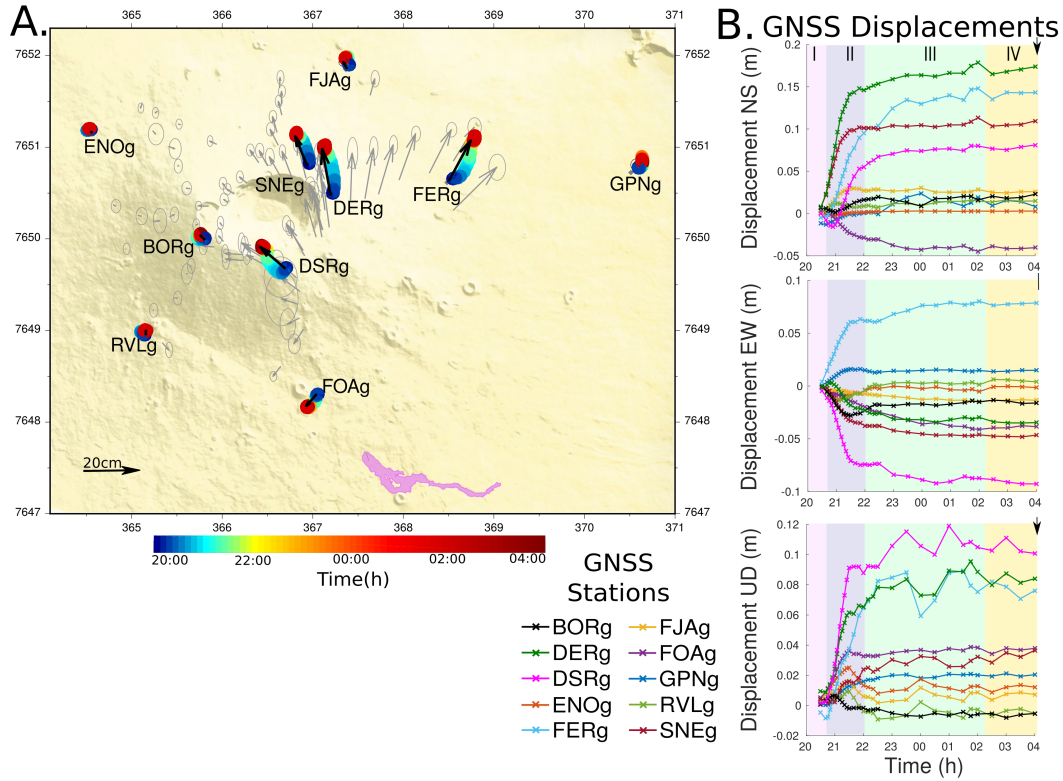


882 **Figure 3.** (Top) Time spanned by each interferogram used in this study. The vertical red area represents  
 883 the eruption. (Bottom) Zoom, colored areas corresponding to numbers I, II, III and IV represent the different  
 884 phases of the intrusion propagation as revealed by the seismic and tilt data (see Fig. 2 and paragraph 3.2 for  
 885 a description). Phase II also corresponds to the period when significant deformation was recorded on the  
 886 cGNSS stations. The vertical dashed line at 01:45 marks the time of the intermediate SAR acquisition by  
 887 Sentinel-1

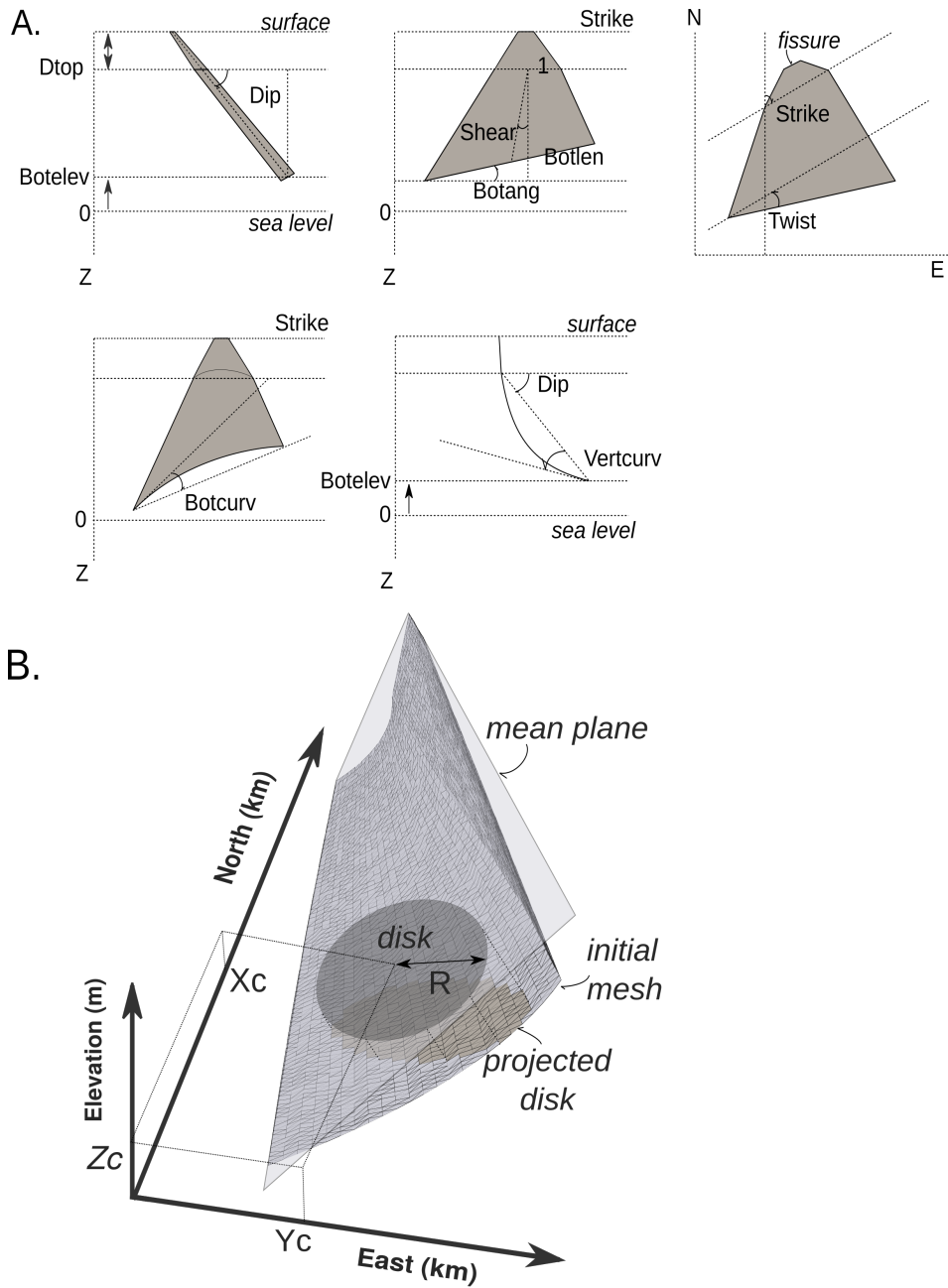


888 **Figure 4.** Comparison between InSAR and GNSS data. LOS displacement (in meters) imaging the May  
 889 2016 eruption. (A and B) Cosmo-SkyMed ascending and descending data covering the whole eruption. (C  
 890 and D) Sentinel ascending and descending data covering the whole eruption. (E and F) Sentinel descen-  
 891 ding data covering the first and last parts of the intrusion propagation. GNSS displacements are represented  
 892 by squares and circles with colors representing the amount of displacement projected along the LOS of the  
 893 InSAR data. Squares in A, B, C, and D represent displacement from the campaign GNSS measurements  
 894 between August 31, 2015 and May 27, 2016. Circles represent displacements of the 10 permanent cGNSS  
 895 stations in the area. For A, B, C, and D, cGNSS displacements between May 25, 2016 17:00 and May 26,  
 896 2016 04:05 are shown. For E and F cGNSS displacements are measured between May 25, 2016 17:00 – May  
 897 26, 2016 01:45 May 26, 2016 01:45 – May 26, 2016 04:05, respectively. Times are UTC.

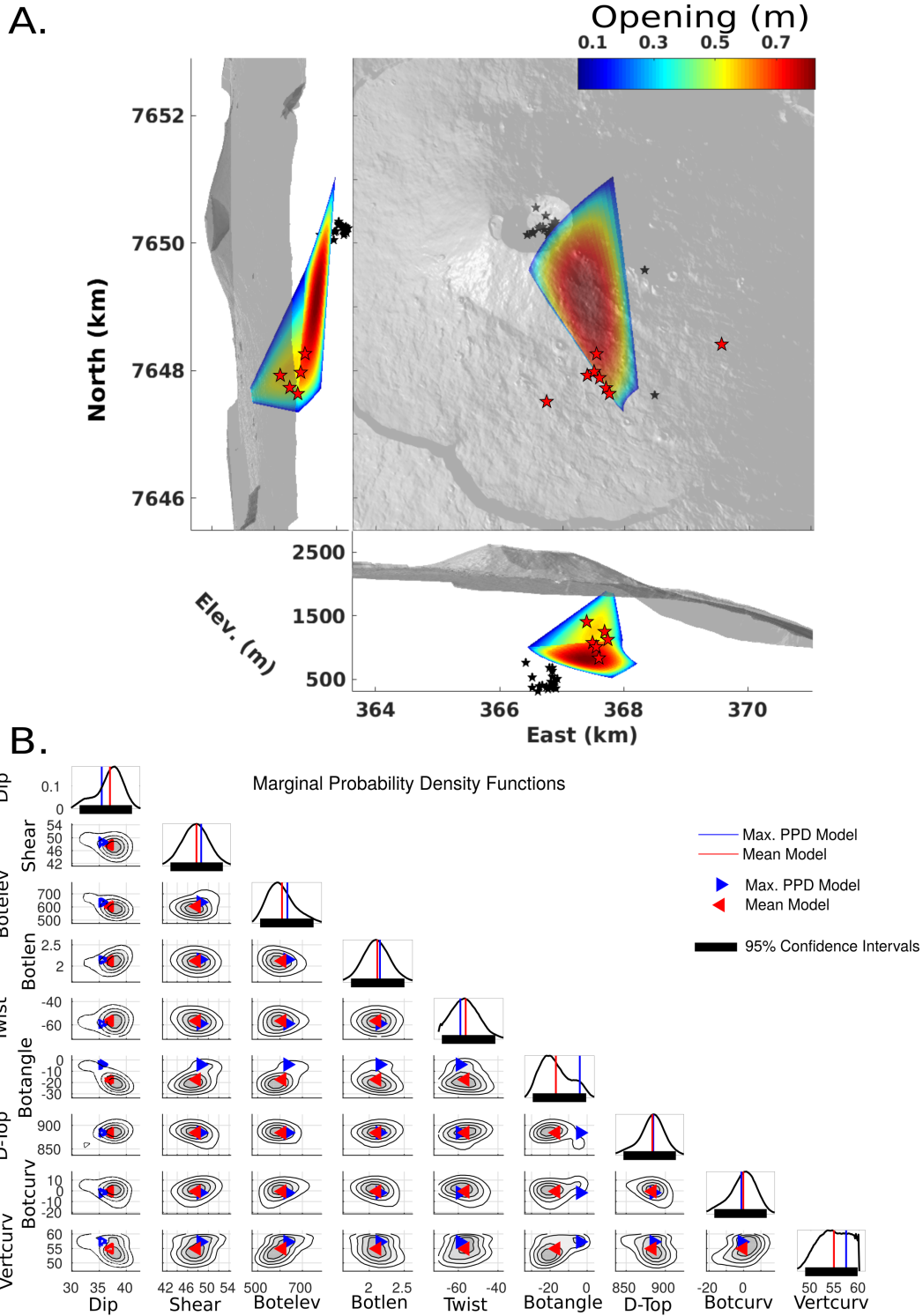




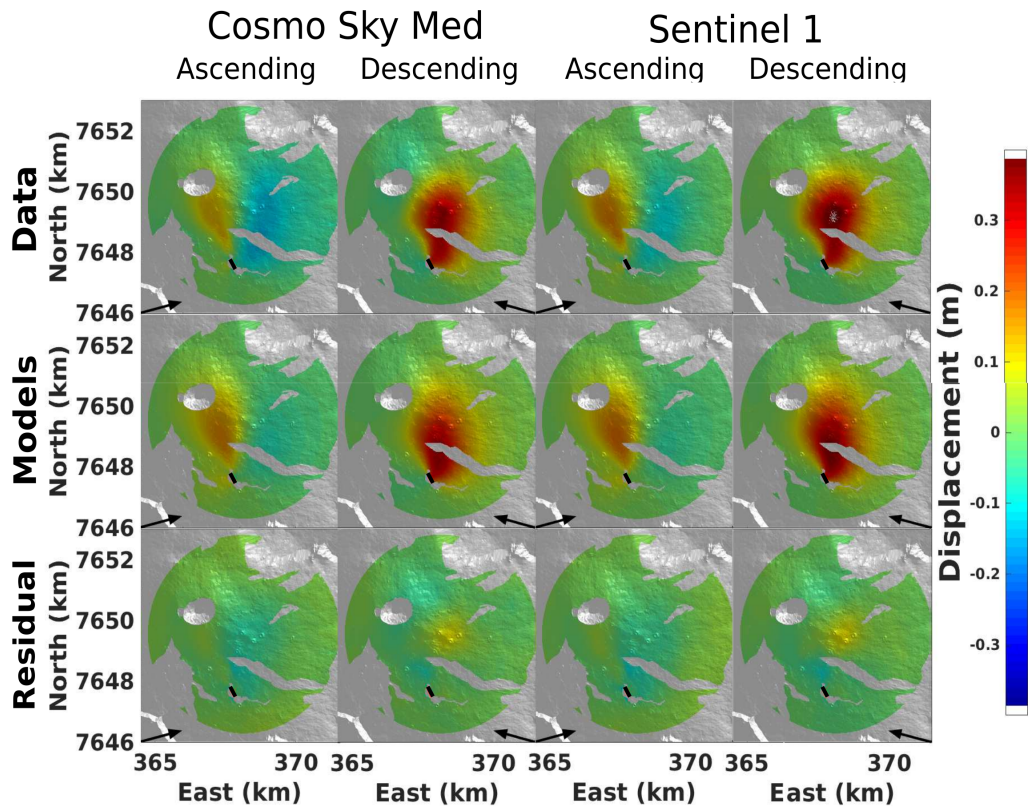
898 **Figure 5.** GNSS data for the May 2016 eruption. (A) Map view of the horizontal displacements from  
 899 cGNSS data and campaign GNSS data. Colored circles represent the evolution of the horizontal displacement  
 900 with time for the 10 cGNSS stations. Black arrows represent the cumulated displacement of the 10 cGNSS  
 901 stations between May 25 at 17:00 and May 26 at 04:05. Standard deviations for horizontal and vertical compo-  
 902 nents are 1-2 cm and 3-4 cm, respectively. Grey arrows represent displacements from the campaign GNSS  
 903 network between August 31, 2015 and May 27, 2016. Ellipses represent their 95% confidence intervals.  
 904 The lava flow is shown in purple. Coordinates are in kilometers (WGS84, UTM 40S). (B) Cumulated dis-  
 905 placement (in meters) is plotted as a function of time for the 10 cGNSS stations which recorded the intrusion.  
 906 Crosses represent the sampling dates used in the time step inversion. Black arrows mark the eruption onset.  
 907 Colored areas with numbers I, II, III and IV refer to the different stages of the intrusion propagation described  
 908 in section 3.2. Times are UTC.



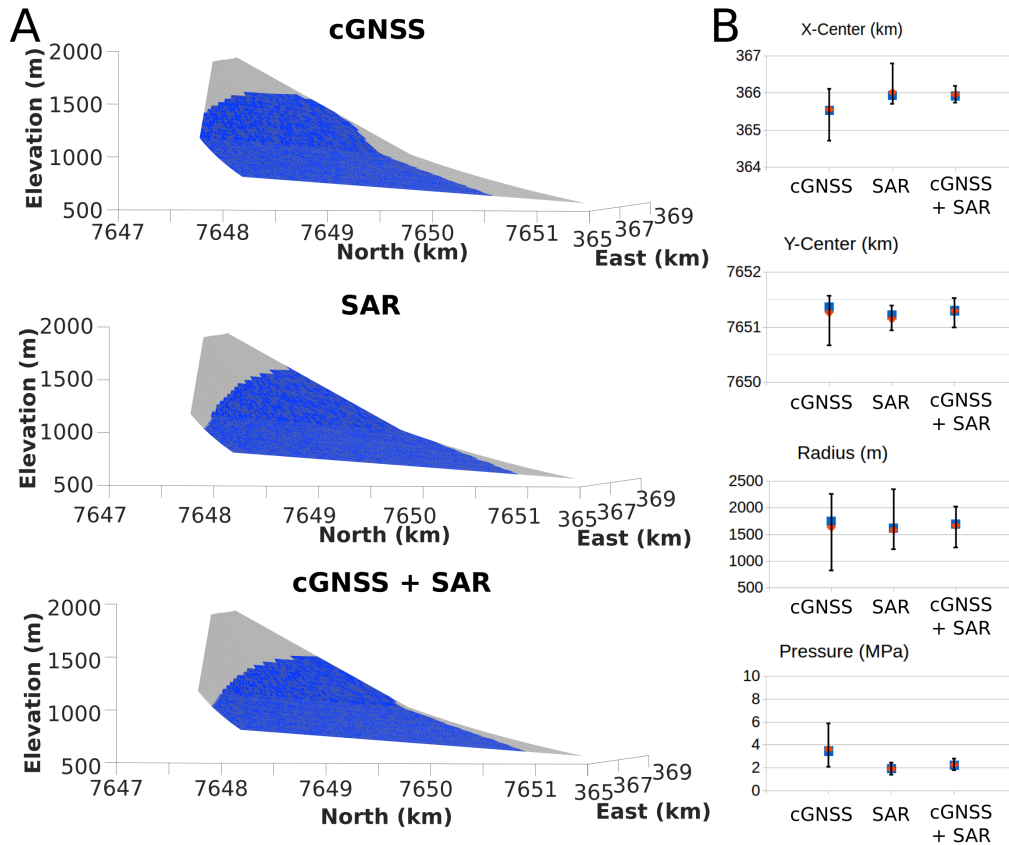
909 **Figure 6.** Representation of the parameters defining the fractures used as deformation sources. (A) 9 pa-  
 910 rameters for a curved quadrangular-shaped intrusion linked to the eruptive fissure by a single echelon. (B) 3  
 911 parameters for a disk projected on an initial mesh (Projected Disk method). (See Table 1 for the description of  
 912 all parameters.)



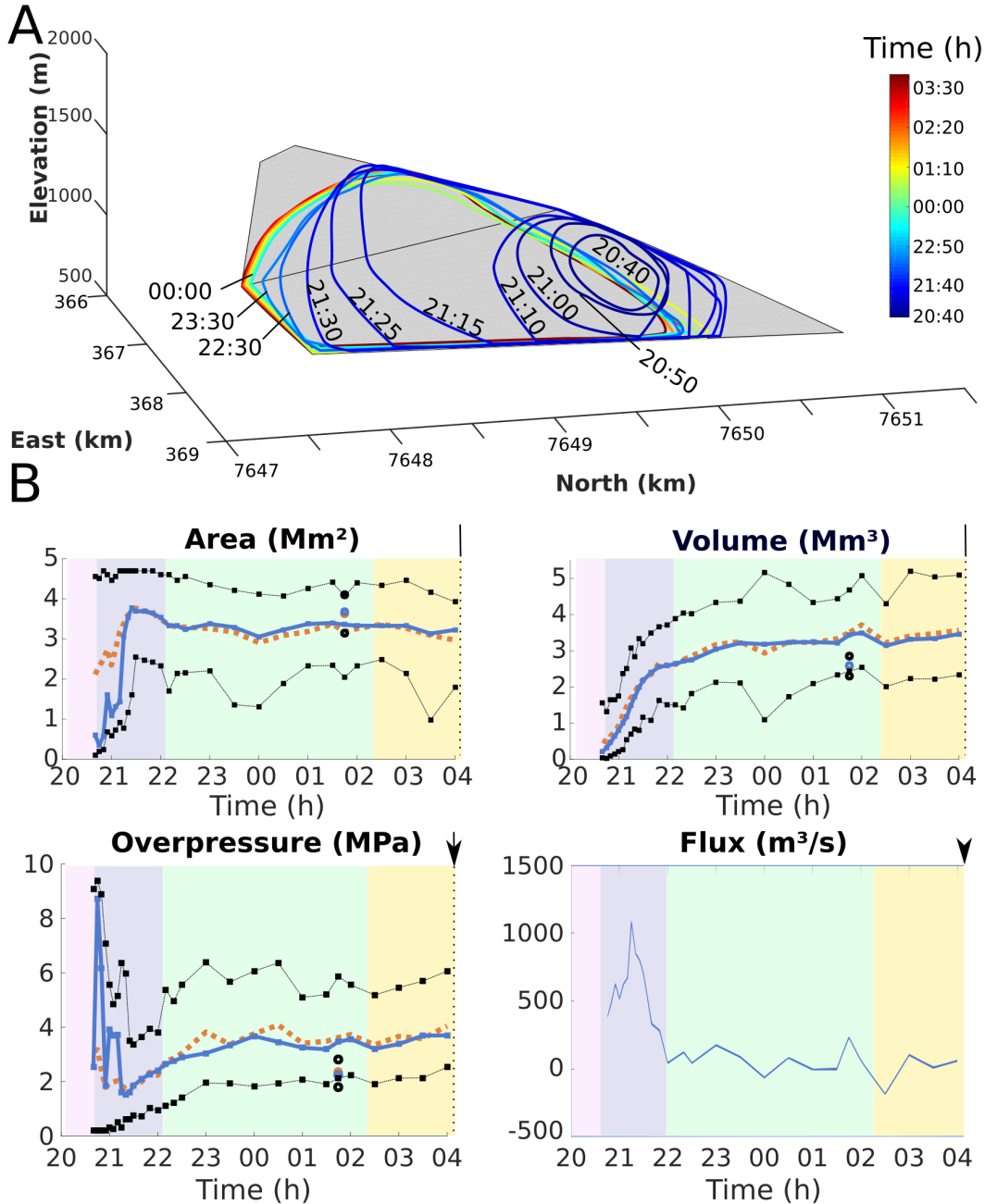
913 **Figure 7.** A. Geometry and opening of the source (best-fit model) determined by inversion of the 2 de-  
 914 scending and 2 ascending interferograms spanning the whole eruption. Black and red stars represent hypocen-  
 915 ters of located micro-earthquakes occurring on May 25 and 26, respectively. B. One-dimensional (diagonals)  
 916 and two-dimensional (off-diagonals) marginal posterior probability density (PPD) functions plotted with the  
 917 maximum PPD (blue) and mean (red) models. Note that the PPD are zoomed in on the best fitting parameter  
 918 range which is much smaller than the explored intervals (see Table S3).



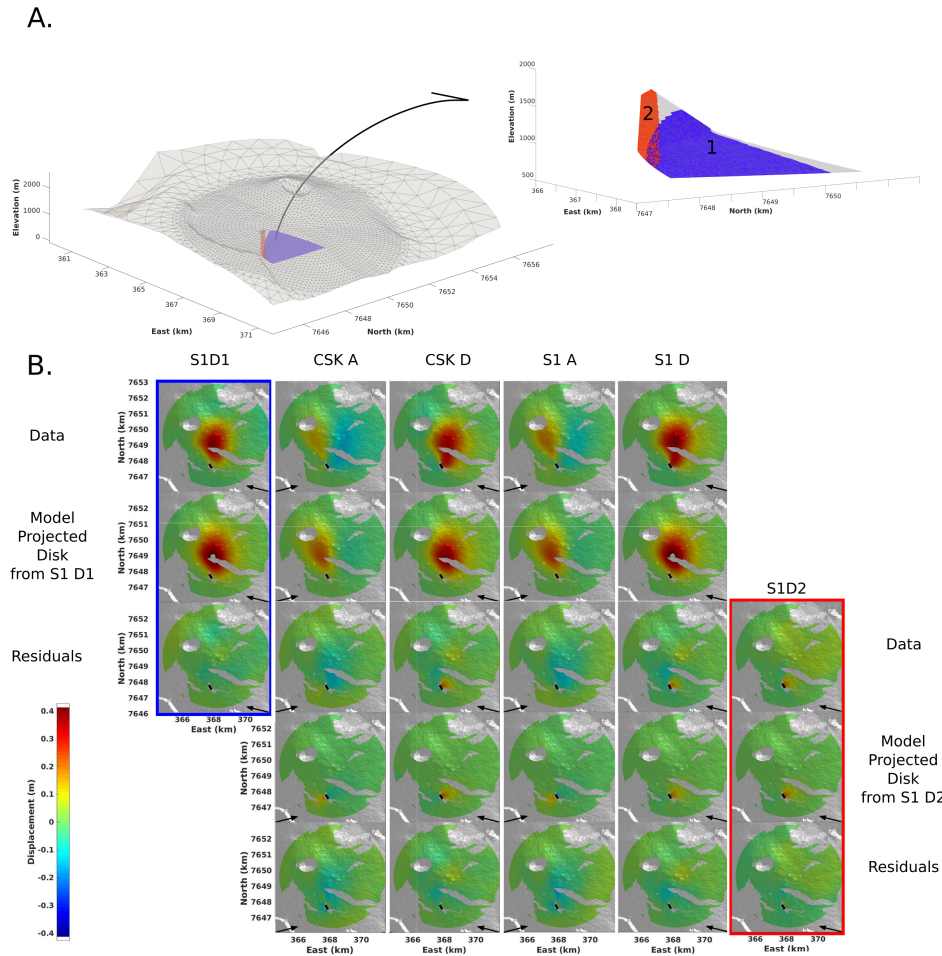
919 **Figure 8.** Unwrapped displacements along the LOS in meters, from the interferograms (*top*), model (*mid-*  
 920 *dle*) and residuals (*bottom*). The black segment indicates the location of the eruptive fissure. Black arrows  
 921 represents the LOS direction. Coordinates are UTM.



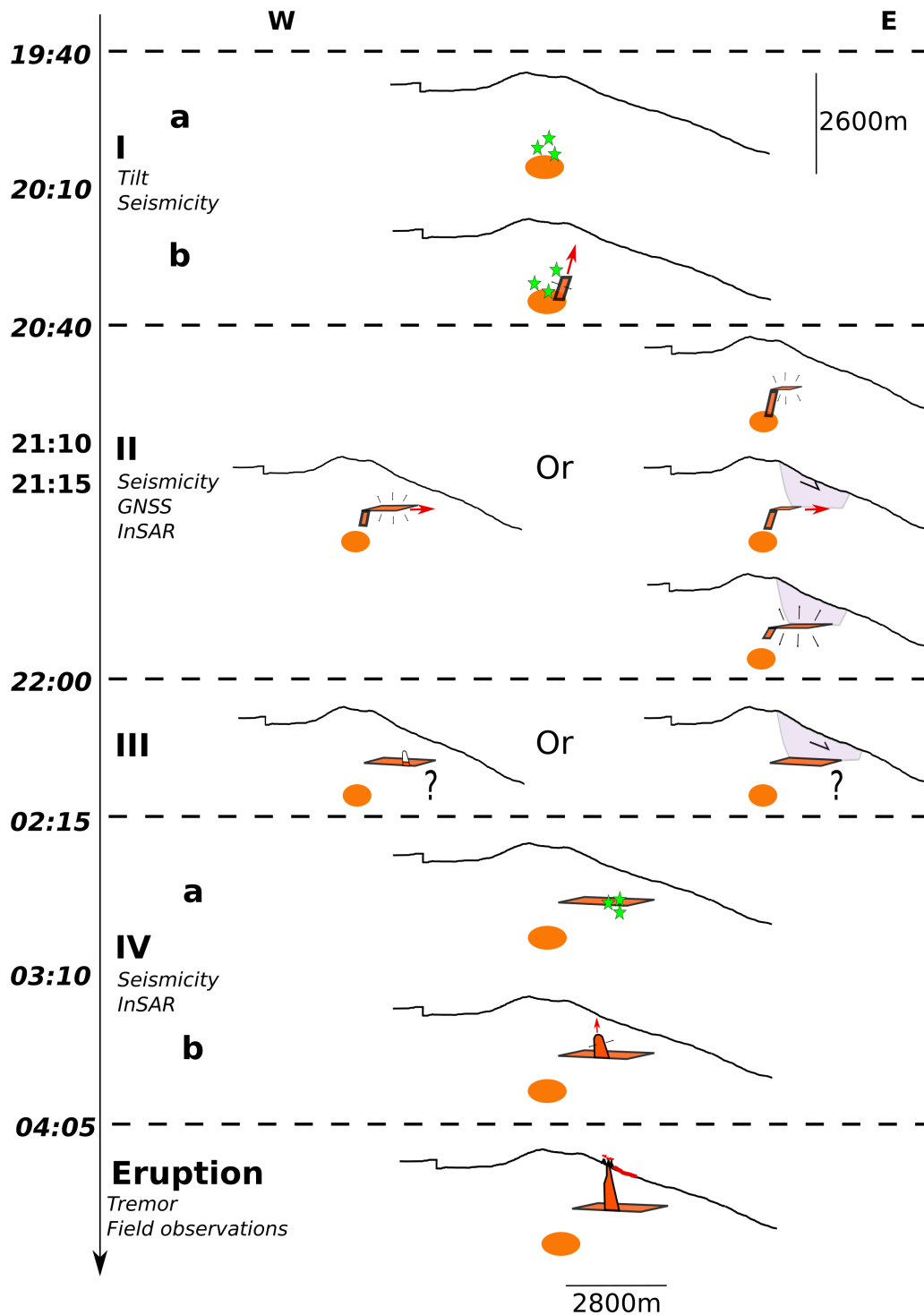
922 **Figure 9.** Projected Disk method results. A. Comparison between the intrusion determined from InSAR  
 923 data covering the whole eruption (grey) and the intrusion determined for the first part of the eruption using  
 924 the Projected Disk method with cGNSS data, InSAR S1 D1 data or both datasets (blue) B. Best (blue squares)  
 925 and mean (orange diamonds) model parameters for the Projected Disk method using either cGNSS, InSAR S1  
 926 D1 data and both datasets. Error bars represent the 95% confidence intervals.



927 **Figure 10.** A. Location of the overpressure source determined from the temporal inversion (colors) for 14  
 928 time steps between 20:40 and 4:00 out of the 30 time step inverted. The pressurized area is assumed to be a  
 929 disk projected on the *a priori* mesh (in grey) determined from the inversion of the four InSAR data covering  
 930 the whole eruption. B. Results of the temporal inversion. Time evolution of the overpressure, the area and  
 931 the volume of the source inverted for each time step. The best-fit model (blue line), the mean model (orange  
 932 dashes), and the lower and upper limits of the 95% confidence interval (black dots), respectively, are shown.  
 933 Circles shown at 01:45 represent results from joint inversion of InSAR and GNSS data. The flux is obtained  
 934 by deriving the best-fit volume curve. Colored areas with numbers I, II, III and IV refer to the different stages  
 935 of the intrusion propagation described in 3.2. Times are UTC time.



936 **Figure 11.** A. Deformation sources determined from the inversion of Sentinel-1 interferograms covering  
 937 the beginning (S1 D1, blue) and the end (S1 D2, red) of the propagation using the Projected Disk method.  
 938 Inset represents a zoom on of the intrusion. The grey mesh is the *a priori* curved quadrangular source de-  
 939 termined from the inversion of the four InSAR data covering the whole eruption. B. Comparison of data,  
 940 modeled displacements and residuals for the best-fitting inverted sources of both parts of the eruption (before  
 941 and after 01:45). The black segment indicates the location of the eruptive fissure. The first and last columns  
 942 concern the inversions at the beginning (blue) and end (red) of the propagation. The four columns in the mid-  
 943 dle concern the whole intrusion. The first row is data, the second row is the model displacements determined  
 944 for the beginning of the eruption (blue source in A). The third row compares the residuals between the inter-  
 945 ferograms covering the whole eruption (second to fourth columns) to the interferogram corresponding to the  
 946 final part of the eruption (last column). The fourth row presents the model displacements obtained for the end  
 947 of the propagation (red source in A). The bottom row is the remaining residuals. Arrows represent the LOS.



948 **Figure 12.** Summary diagram illustrating the scenario proposed for the May 2016 intrusion propagation.  
 949 Phase I : Vertical ascent first evidenced by seismicity increase (a) then in addition by tilt (b). Phase II : Lateral  
 950 propagation (left) potentially favored by flank slip (right). Phase III : Pause with eventual gas accumulation  
 951 (left)and/or flank creeping (right). Phase IV : Final dike propagation marked by seismicity increase (a) then  
 952 decrease (b) before the eruption onset.



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## References

- Akaike, H. (1980), Seasonal adjustment by a Bayesian modeling, *Journal of time series analysis*, *1*(1), 1–13.
- Aki, K., and V. Ferrazzini (2000), Seismic monitoring and modeling of an active volcano for prediction, *Journal of Geophysical Research: Solid Earth*, *105*(B7), 16,617–16,640.
- Aoki, Y., P. Segall, T. Kato, P. Cervelli, and S. Shimada (1999), Imaging magma transport during the 1997 seismic swarm off the Izu Peninsula, Japan, *Science*, *286*(5441), 927–930.
- Auriac, A., K. Spaans, F. Sigmundsson, A. Hooper, P. Schmidt, and B. Lund (2013), Iceland rising: Solid Earth response to ice retreat inferred from satellite radar interferometry and viscoelastic modeling, *Journal of Geophysical Research: Solid Earth*, *118*(4), 1331–1344.
- Bachèlery, P., and P. Mairine (1990), Evolution volcano-structurale du Piton de la Fournaise depuis 0.53 Ma, *Le volcanisme de la Réunion, Monographie. Cent. Rech. Volcanol., Clermont-Ferrand, France*, pp. 213–242.
- Bagnardi, M., F. Amelung, and M. P. Poland (2013), A new model for the growth of basaltic shields based on deformation of Fernandina volcano, Galápagos Islands, *Earth and Planetary Science Letters*, *377*, 358–366.
- Battaglia, J., V. Ferrazzini, T. Staudacher, K. Aki, and J.-L. Cheminée (2005), Pre-eruptive migration of earthquakes at the Piton de la Fournaise volcano (Réunion Island), *Geophysical Journal International*, *161*(2), 549–558.
- Brenguier, F., P. Kowalski, T. Staudacher, V. Ferrazzini, F. Lauret, P. Boissier, P. Catherine, A. Lemarchand, C. Pequegnat, O. Meric, et al. (2012), First results from the Un-

- 1005 derVolc high resolution seismic and GPS network deployed on Piton de la Fournaise  
1006 volcano, *Seismological Research Letters*, 83(1), 97–102.
- 1007 Calais, E., N. d’Oreye, J. Albaric, A. Deschamps, D. Delvaux, J. Déverchère, C. Ebinger,  
1008 R. W. Ferdinand, F. Kervyn, A. S. Macheyeki, et al. (2008), Strain accommodation by  
1009 slow slip and dyking in a youthful continental rift, East Africa, *Nature*, 456(7223), 783.
- 1010 Cannavò, F., A. G. Camacho, P. J. González, M. Mattia, G. Puglisi, and J. Fernández  
1011 (2015), Real time tracking of magmatic intrusions by means of ground deformation  
1012 modeling during volcanic crises, *Scientific reports*, 5, 10,970.
- 1013 Carrier, A., J.-L. Got, A. Peltier, V. Ferrazzini, T. Staudacher, P. Kowalski, and P. Boissier  
1014 (2015), A damage model for volcanic edifices: Implications for edifice strength, magma  
1015 pressure, and eruptive processes, *Journal of Geophysical Research: Solid Earth*, 120(1),  
1016 567–583.
- 1017 Cayol, V. (1996), Analyse élastostatique tridimensionnelle du champ de déformations des  
1018 édifices volcaniques par éléments frontières mixtes, Ph.D. thesis, Paris 7.
- 1019 Cayol, V., and F. Cornet (1997), 3D mixed boundary elements for elastostatic deformation  
1020 field analysis, *International journal of rock mechanics and mining sciences*, 34(2), 275–  
1021 287.
- 1022 Cayol, V., and F. H. Cornet (1998a), Three-dimensional modeling of the 1983–1984 erup-  
1023 tion at Piton de la Fournaise Volcano, Réunion Island, *Journal of Geophysical Research:  
1024 Solid Earth*, 103(B8), 18,025–18,037.
- 1025 Cayol, V., and F. H. Cornet (1998b), Effects of topography on the interpretation of the  
1026 deformation field of prominent volcanoes—Application to Etna, *Geophysical Research  
1027 Letters*, 25(11), 1979–1982.
- 1028 Cervelli, P., P. Segall, F. Amelung, H. Garbeil, C. Meertens, S. Owen, A. Miklius, and  
1029 M. Lisowski (2002), The 12 September 1999 upper east rift zone dike intrusion at Ki-  
1030 lauea volcano, Hawaii, *Journal of Geophysical Research: Solid Earth*, 107(B7).
- 1031 Chadwick Jr, W., and J. Dieterich (1995), Mechanical modeling of circumferential and  
1032 radial dike intrusion on Galapagos volcanoes, *Journal of Volcanology and Geothermal  
1033 Research*, 66(1-4), 37–52.
- 1034 Chaput, M., V. Famin, and L. Michon (2014a), Deformation of basaltic shield volcanoes  
1035 under cointrusive stress permutations, *Journal of Geophysical Research: Solid Earth*,  
1036 119(1), 274–301.
- 1037 Chaput, M., V. Pinel, V. Famin, L. Michon, and J.-L. Froger (2014b), Cointrusive shear  
1038 displacement by sill intrusion in a detachment: A numerical approach, *Geophysical Re-  
1039 search Letters*, 41(6), 1937–1943.
- 1040 Chaput, M., V. Famin, and L. Michon (2017), Sheet intrusions and deformation of Piton  
1041 des Neiges, and their implication for the volcano-tectonics of La Réunion, *Tectono-  
1042 physics*, 717, 531–546.
- 1043 Chen, C. W., and H. A. Zebker (2000), Network approaches to two-dimensional phase  
1044 unwrapping: intractability and two new algorithms, *JOSA A*, 17(3), 401–414.
- 1045 Chen, Y., D. Remy, J.-L. Froger, A. Peltier, N. Villeneuve, J. Darrozes, H. Perfettini, and  
1046 S. Bonvalot (2017), Long-term ground displacement observations using InSAR and  
1047 GNSS at Piton de la Fournaise volcano between 2009 and 2014, *Remote Sensing of En-  
1048 vironment*, 194, 230–247.
- 1049 Chouet, B. A. (1996), Long-period volcano seismicity: its source and use in eruption fore-  
1050 casting, *Nature*, 380(6572), 309.
- 1051 CNES (1996), Philosophy and Instructions for the Use of the DIAPASON Interferometry  
1052 Software System Developed at CNES.
- 1053 Coppola, D., D. Piscopo, T. Staudacher, and C. Cigolini (2009), Lava discharge rate and  
1054 effusive pattern at Piton de la Fournaise from MODIS data, *Journal of Volcanology and  
1055 Geothermal Research*, 184(1-2), 174–192.
- 1056 Coppola, D., A. Di Muro, A. Peltier, N. Villeneuve, V. Ferrazzini, M. Favalli,  
1057 P. Bachèlery, L. Gurioli, A. Harris, S. Moune, et al. (2017), Shallow system rejuvena-  
1058 tion and magma discharge trends at Piton de la Fournaise volcano (La Réunion Island),

- 1059 *Earth and Planetary Science Letters*, 463, 13–24.
- 1060 Corbi, F., E. Rivalta, V. Pinel, F. Maccaferri, M. Bagnardi, and V. Acocella (2015), How  
1061 caldera collapse shapes the shallow emplacement and transfer of magma in active volca-  
1062 noes, *Earth and Planetary Science Letters*, 431, 287–293.
- 1063 Corbi, F., E. Rivalta, V. Pinel, F. Maccaferri, and V. Acocella (2016), Understanding the  
1064 link between circumferential dikes and eruptive fissures around calderas based on nu-  
1065 merical and analog models, *Geophysical Research Letters*, 43(12), 6212–6219.
- 1066 Currenti, G., R. Napoli, and C. Del Negro (2011), Toward a realistic deformation model  
1067 of the 2008 magmatic intrusion at Etna from combined DInSAR and GPS observations,  
1068 *Earth and Planetary Science Letters*, 312(1-2), 22–27.
- 1069 Dahm, T. (2000), Numerical simulations of the propagation path and the arrest of fluid-  
1070 filled fractures in the Earth, *Geophysical Journal International*, 141(3), 623–638.
- 1071 Delaney, P. T., D. D. Pollard, J. I. Ziony, and E. H. McKee (1986), Field relations between  
1072 dikes and joints: emplacement processes and paleostress analysis, *Journal of Geophys-  
1073 ical Research: Solid Earth*, 91(B5), 4920–4938.
- 1074 Dieterich, J. (1994), A constitutive law for rate of earthquake production and its appli-  
1075 cation to earthquake clustering, *Journal of Geophysical Research: Solid Earth*, 99(B2),  
1076 2601–2618.
- 1077 Dieterich, J., V. Cayol, and P. Okubo (2000), The use of earthquake rate changes as a  
1078 stress meter at Kilauea volcano, *Nature*, 408(6811), 457.
- 1079 Duputel, Z., P. S. Agram, M. Simons, S. E. Minson, and J. L. Beck (2014), Accounting  
1080 for prediction uncertainty when inferring subsurface fault slip, *Geophysical Journal In-  
1081 ternational*, 197(1), 464–482.
- 1082 Duputel, Z., O. Lengliné, and V. Ferrazzini (2018), Constraining spatio-temporal charac-  
1083 teristics of magma migration at Piton de la Fournaise volcano from pre-eruptive seismicity,  
1084 *Geophysical Research Letters*.
- 1085 Einarsson, P., and B. Brandsdóttir (1978), Seismological evidence for lateral magma in-  
1086 trusion during the July 1978 deflation of the Krafla volcano in NE-Iceland, *Tech. rep.*,  
1087 University of Iceland, Reykjavik, IS.
- 1088 Famin, V., and L. Michon (2010), Volcano destabilization by magma injections in a de-  
1089 tachment, *Geology*, 38(3), 219–222.
- 1090 Froger, J.-L., V. Famin, V. Cayol, A. Augier, L. Michon, and J.-F. Lénat (2015), Time-  
1091 dependent displacements during and after the April 2007 eruption of Piton de la Four-  
1092 naise, revealed by interferometric data, *Journal of Volcanology and Geothermal Re-  
1093 search*, 296, 55–68.
- 1094 Fukuda, J., and K. M. Johnson (2010), Mixed linear–non-linear inversion of crustal de-  
1095 formation data: Bayesian inference of model, weighting and regularization parameters,  
1096 *Geophysical Journal International*, 181(3), 1441–1458.
- 1097 Fukushima, Y., V. Cayol, and P. Durand (2005), Finding realistic dike models from in-  
1098 terferometric synthetic aperture radar data: The February 2000 eruption at Piton de la  
1099 Fournaise, *Journal of Geophysical Research: Solid Earth*, 110(B3).
- 1100 Fukushima, Y., V. Cayol, P. Durand, and D. Massonnet (2010), Evolution of magma con-  
1101 duits during the 1998–2000 eruptions of Piton de la Fournaise volcano, Réunion Island,  
1102 *Journal of Geophysical Research: Solid Earth*, 115(B10).
- 1103 Global Volcanism Program (2018), Report on Kilauea (United States), in *Weekly Volcanic  
1104 Activity Report, 2 May–8 May 2018*, edited by S. K. Sennert, Smithsonian Institution and  
1105 US Geological Survey.
- 1106 Got, J.-L., A. Peltier, T. Staudacher, P. Kowalski, and P. Boissier (2013), Edifice strength  
1107 and magma transfer modulation at Piton de la Fournaise volcano, *Journal of Geophys-  
1108 ical Research: Solid Earth*, 118(9), 5040–5057.
- 1109 Gudmundsson, A. (2006), How local stresses control magma-chamber ruptures, dyke in-  
1110 jections, and eruptions in composite volcanoes, *Earth-Science Reviews*, 79(1-2), 1–31.
- 1111 Guldstrand, F., O. Galland, E. Hallot, and S. Burchardt (2018), Experimental Constraints  
1112 on Forecasting the Location of Volcanic Eruptions from Pre-eruptive Surface Deforma-

- tion, *Frontiers in Earth Science*, 6, 7.
- Gurioli, L., A. Di Muro, I. Vlastélic, S. Moune, S. Thivet, M. Valer, N. Villeneuve, G. Boudoire, A. Peltier, P. Bachèlery, et al. (2018), Integrating field, textural, and geochemical monitoring to track eruption triggers and dynamics: a case study from Piton de la Fournaise, *Solid Earth*, 9(2), 431.
- Hamling, I. J., A. Ayele, L. Bennati, E. Calais, C. J. Ebinger, D. Keir, E. Lewi, T. J. Wright, and G. Yirgu (2009), Geodetic observations of the ongoing Dabbahu rifting episode: new dyke intrusions in 2006 and 2007, *Geophysical Journal International*, 178(2), 989–1003.
- Herring, T., R. King, and S. McClusky (2010), Introduction to Gamit/Globk, *Massachusetts Institute of Technology, Cambridge, Massachusetts*.
- Jaeger, J. C., N. G. Cook, and R. Zimmerman (2009), *Fundamentals of rock mechanics*, John Wiley & Sons.
- Jónsson, S., H. Zebker, P. Segall, and F. Amelung (2002), Fault slip distribution of the 1999 Mw 7.1 Hector Mine, California, earthquake, estimated from satellite radar and GPS measurements, *Bulletin of the Seismological Society of America*, 92(4), 1377–1389.
- Kavanagh, J. L., T. Menand, and R. S. J. Sparks (2006), An experimental investigation of sill formation and propagation in layered elastic media, *Earth and Planetary Science Letters*, 245(3-4), 799–813.
- Lengliné, O., Z. Duputel, and V. Ferrazzini (2016), Uncovering the hidden signature of a magmatic recharge at Piton de la Fournaise volcano using small earthquakes, *Geophysical Research Letters*, 43(9), 4255–4262.
- Lister, J. R. (1990), Buoyancy-driven fluid fracture: similarity solutions for the horizontal and vertical propagation of fluid-filled cracks, *Journal of Fluid Mechanics*, 217, 213–239.
- Maccaferri, F., M. Bonafede, and E. Rivalta (2010), A numerical model of dyke propagation in layered elastic media, *Geophysical Journal International*, 180(3), 1107–1123.
- Maccaferri, F., M. Bonafede, and E. Rivalta (2011), A quantitative study of the mechanisms governing dike propagation, dike arrest and sill formation, *Journal of Volcanology and Geothermal Research*, 208(1-2), 39–50.
- Maimon, O., V. Lyakhovsky, O. Melnik, and O. Navon (2012), The propagation of a dyke driven by gas-saturated magma, *Geophysical Journal International*, 189(2), 956–966.
- Massonnet, D., and K. L. Feigl (1998), Radar interferometry and its application to changes in the Earth’s surface, *Reviews of geophysics*, 36(4), 441–500.
- Masterlark, T. (2007), Magma intrusion and deformation predictions: Sensitivities to the Mogi assumptions, *Journal of Geophysical Research: Solid Earth*, 112(B6).
- Masterlark, T., and Z. Lu (2004), Transient volcano deformation sources imaged with interferometric synthetic aperture radar: application to Seguam Island, Alaska, *Journal of Geophysical Research: Solid Earth*, 109(B1).
- Menand, T., and J. C. Phillips (2007), Gas segregation in dykes and sills, *Journal of Volcanology and Geothermal Research*, 159(4), 393–408.
- Menand, T., and S. R. Tait (2001), A phenomenological model for precursor volcanic eruptions, *Nature*, 411(6838), 678.
- Menand, T., K. Daniels, and P. Bengeriat (2010), Dyke propagation and sill formation in a compressive tectonic environment, *Journal of Geophysical Research: Solid Earth*, 115(B8).
- Menassian, S. J. (2013), Validation of a ‘displacement tomography’ inversion method for modeling sheet intrusions, Master. thesis, Michigan Technological University.
- Merle, O., and J.-F. Lénat (2003), Hybrid collapse mechanism at Piton de la Fournaise volcano, Reunion Island, Indian Ocean, *Journal of Geophysical Research: Solid Earth*, 108(B3).
- Michon, L., T. Staudacher, V. Ferrazzini, P. Bachèlery, and J. Marti (2007), April 2007 collapse of Piton de la Fournaise: a new example of caldera formation, *Geophysical Research Letters*, 34(21).

- 1167 Moran, S. C., C. Newhall, and D. C. Roman (2011), Failed magmatic eruptions: late-stage  
1168 cessation of magma ascent, *Bulletin of Volcanology*, *73*(2), 115–122.
- 1169 Nikolaidis, R. M., Y. Bock, P. J. Jonge, P. Shearer, D. C. Agnew, and M. Van Domselaar  
1170 (2001), Seismic wave observations with the Global Positioning System, *Journal of Geo-*  
1171 *physical Research: Solid Earth*, *106*(B10), 21,897–21,916.
- 1172 Okada, Y. (1985), Surface deformation due to shear and tensile faults in a half-space, *Bul-*  
1173 *letin of the seismological society of America*, *75*(4), 1135–1154.
- 1174 Peltier, A., V. Ferrazzini, T. Staudacher, and P. Bachèlery (2005), Imaging the dynamics  
1175 of dyke propagation prior to the 2000–2003 flank eruptions at Piton de La Fournaise,  
1176 Reunion Island, *Geophysical Research Letters*, *32*(22).
- 1177 Peltier, A., T. Staudacher, and P. Bachèlery (2007), Constraints on magma transfers and  
1178 structures involved in the 2003 activity at Piton de La Fournaise from displacement  
1179 data, *Journal of Geophysical Research: Solid Earth*, *112*(B3).
- 1180 Peltier, A., V. Famin, P. Bachèlery, V. Cayol, Y. Fukushima, and T. Staudacher (2008),  
1181 Cyclic magma storages and transfers at Piton de La Fournaise volcano (La Réunion  
1182 hotspot) inferred from deformation and geochemical data, *Earth and Planetary Science*  
1183 *Letters*, *270*(3-4), 180–188.
- 1184 Peltier, A., P. Bachèlery, and T. Staudacher (2009a), Magma transport and storage at Piton  
1185 de La Fournaise (La Réunion) between 1972 and 2007: A review of geophysical and  
1186 geochemical data, *Journal of Volcanology and Geothermal Research*, *184*(1-2), 93–108.
- 1187 Peltier, A., T. Staudacher, P. Bachèlery, and V. Cayol (2009b), Formation of the April  
1188 2007 caldera collapse at Piton de La Fournaise volcano: Insights from GPS data, *Jour-*  
1189 *nal of Volcanology and Geothermal Research*, *184*(1-2), 152–163.
- 1190 Peltier, A., J.-L. Got, N. Villeneuve, P. Boissier, T. Staudacher, V. Ferrazzini, and  
1191 A. Walpersdorf (2015), Long-term mass transfer at Piton de la Fournaise volcano evi-  
1192 denced by strain distribution derived from GNSS network, *Journal of Geophysical Re-*  
1193 *search: Solid Earth*, *120*(3), 1874–1889.
- 1194 Peltier, A., F. Beauducel, N. Villeneuve, V. Ferrazzini, A. Di Muro, A. Aiuppa, A. Der-  
1195 rien, K. Jourde, and B. Taisne (2016), Deep fluid transfer evidenced by surface defor-  
1196 mation during the 2014–2015 unrest at Piton de la Fournaise volcano, *Journal of Vol-*  
1197 *canology and Geothermal Research*, *321*, 140–148.
- 1198 Peltier, A., J.-L. Froger, N. Villeneuve, and T. Catry (2017), Assessing the reliability and  
1199 consistency of InSAR and GNSS data for retrieving 3D-displacement rapid changes,  
1200 the example of the 2015 Piton de la Fournaise eruptions, *Journal of Volcanology and*  
1201 *Geothermal Research*, *344*, 106–120.
- 1202 Peltier, A., N. Villeneuve, V. Ferrazzini, S. Testud, T. Hassen Ali, P. Boissier, and  
1203 P. Catherine (2018), Changes in the long-term geophysical eruptive precursors at Piton  
1204 de la Fournaise: Implications for the response management, *Frontiers in Earth Science*,  
1205 *6*, 104.
- 1206 Pinel, V., and C. Jaupart (2000), The effect of edifice load on magma ascent beneath a  
1207 volcano, *Philosophical Transactions of the Royal Society of London A: Mathematical,*  
1208 *Physical and Engineering Sciences*, *358*(1770), 1515–1532.
- 1209 Pinel, V., and C. Jaupart (2004), Magma storage and horizontal dyke injection beneath a  
1210 volcanic edifice, *Earth and Planetary Science Letters*, *221*(1-4), 245–262.
- 1211 Pinel, V., A. Carrara, F. Maccaferri, E. Rivalta, and F. Corbi (2017), A two-step model  
1212 for dynamical dike propagation in two dimensions: Application to the July 2001 Etna  
1213 eruption, *Journal of Geophysical Research: Solid Earth*, *122*(2), 1107–1125.
- 1214 Rivalta, E., M. Böttlinger, and T. Dahm (2005), Buoyancy-driven fracture ascent: Experi-  
1215 ments in layered gelatine, *Journal of Volcanology and Geothermal Research*, *144*(1-4),  
1216 273–285.
- 1217 Rivalta, E., B. Taisne, A. Bungler, and R. Katz (2015), A review of mechanical models of  
1218 dike propagation: Schools of thought, results and future directions, *Tectonophysics*, *638*,  
1219 1–42.

- 1220 Roullet, G., A. Peltier, B. Taisne, T. Staudacher, V. Ferrazzini, A. Di Muro, et al. (2012),  
 1221 A new comprehensive classification of the Piton de la Fournaise activity spanning the  
 1222 1985–2010 period. Search and analysis of short-term precursors from a broad-band seis-  
 1223 mological station, *Journal of Volcanology and Geothermal Research*, 241, 78–104.
- 1224 Rubin, A. M., and D. Gillard (1998), Dike-induced earthquakes: Theoretical considera-  
 1225 tions, *Journal of Geophysical Research: Solid Earth*, 103(B5), 10,017–10,030.
- 1226 Ruch, J., T. Wang, W. Xu, M. Hensch, and S. Jónsson (2016), Oblique rift opening re-  
 1227 vealed by reoccurring magma injection in central Iceland, *Nature communications*, 7,  
 1228 12,352.
- 1229 Sambridge, M. (1998), Exploring multidimensional landscapes without a map, *Inverse*  
 1230 *problems*, 14(3), 427.
- 1231 Sambridge, M. (1999a), Geophysical inversion with a neighbourhood algorithm -I. Search-  
 1232 ing a parameter space, *Geophysical Journal International*, 138(3), 479–494.
- 1233 Sambridge, M. (1999b), Geophysical inversion with a neighbourhood algorithm -II. Ap-  
 1234 praising the ensemble, *Geophysical Journal International*, 138(3), 727–746.
- 1235 Segall, P., P. Cervelli, S. Owen, M. Lisowski, and A. Miklius (2001), Constraints on dike  
 1236 propagation from continuous GPS measurements, *Journal of Geophysical Research:*  
 1237 *Solid Earth*, 106(B9), 19,301–19,317.
- 1238 Sigmundsson, F., A. Hooper, S. Hreinsdóttir, K. S. Vogfjörð, B. G. Ófeigsson, E. R.  
 1239 Heimisson, S. Dumont, M. Parks, K. Spaans, G. B. Gudmundsson, et al. (2015), Seg-  
 1240 mented lateral dyke growth in a rifting event at Bárðarbunga volcanic system, Iceland,  
 1241 *Nature*, 517(7533), 191.
- 1242 Staudacher, T., V. Ferrazzini, A. Peltier, P. Kowalski, P. Boissier, P. Catherine, F. Lauret,  
 1243 and F. Massin (2009), The April 2007 eruption and the Dolomieu crater collapse, two  
 1244 major events at Piton de la Fournaise (La Réunion Island, Indian Ocean), *Journal of*  
 1245 *Volcanology and Geothermal Research*, 184(1-2), 126–137.
- 1246 Sudhaus, H., and S. Jónsson (2009), Improved source modelling through combined use of  
 1247 InSAR and GPS under consideration of correlated data errors: application to the June  
 1248 2000 Kleifarvatn earthquake, Iceland, *Geophysical Journal International*, 176(2), 389–  
 1249 404.
- 1250 Taisne, B., and C. Jaupart (2011), Magma expansion and fragmentation in a propagating  
 1251 dyke, *Earth and Planetary Science Letters*, 301(1-2), 146–152.
- 1252 Taisne, B., and S. Tait (2011), Effect of solidification on a propagating dike, *Journal of*  
 1253 *Geophysical Research: Solid Earth*, 116(B1).
- 1254 Taisne, B., F. Brenguier, N. Shapiro, and V. Ferrazzini (2011), Imaging the dynamics  
 1255 of magma propagation using radiated seismic intensity, *Geophysical Research Letters*,  
 1256 38(4).
- 1257 Tarantola, A. (1987), Inverse problem theory: Method for data fitting and model parameter  
 1258 estimation, *Elsevier*, 613.
- 1259 Tarantola, A. (2005), *Inverse problem theory and methods for model parameter estimation*,  
 1260 vol. 89, siam.
- 1261 Toda, S., R. S. Stein, and T. Sagiya (2002), Evidence from the AD 2000 Izu islands earth-  
 1262 quake swarm that stressing rate governs seismicity, *Nature*, 419(6902), 58.
- 1263 Toutain, J.-P., P. Bachelery, P.-A. Blum, J.-L. Cheminee, H. Delorme, L. Fontaine,  
 1264 P. Kowalski, and P. Taochy (1992), Real time monitoring of vertical ground deforma-  
 1265 tions during eruptions at Piton de la Fournaise, *Geophysical research letters*, 19(6),  
 1266 553–556.
- 1267 Trasatti, E., C. Giunchi, and N. P. Agostinetti (2008), Numerical inversion of deformation  
 1268 caused by pressure sources: application to Mount Etna (Italy), *Geophysical Journal In-*  
 1269 *ternational*, 172(2), 873–884.
- 1270 Tridon, M., V. Cayol, J.-L. Froger, A. Augier, and P. Bachelery (2016), Inversion of coeval  
 1271 shear and normal stress of Piton de la Fournaise flank displacement, *Journal of Geo-*  
 1272 *physical Research: Solid Earth*, 121(11), 7846–7866.

- 1273 Villeneuve, N., and P. Bachèlery (2006), Revue de la typologie des éruptions au Piton de  
1274 La Fournaise, processus et risques volcaniques associés, *Cybergeo: European Journal of*  
1275 *Geography*.
- 1276 Watanabe, T., T. Masuyama, K. Nagaoka, and T. Tahara (2002), Analog experiments on  
1277 magma-filled cracks, *Earth, planets and space*, *54*(12), 1247–1261.
- 1278 Wauthier, C., V. Cayol, F. Kervyn, and N. d’Oreye (2012), Magma sources involved in the  
1279 2002 Nyiragongo eruption, as inferred from an InSAR analysis, *Journal of Geophysical*  
1280 *Research: Solid Earth*, *117*(B5).
- 1281 Wauthier, C., V. Cayol, M. Poland, F. Kervyn, N. d’Oreye, A. Hooper, S. Samsonov,  
1282 K. Tiampo, and B. Smets (2013), Nyamulagira’s magma plumbing system inferred from  
1283 15 years of InSAR, *Geological Society, London, Special Publications*, *380*(1), 39–65.
- 1284 Welstead, S. T. (1999), *Fractal and wavelet image compression techniques*, SPIE Optical  
1285 Engineering Press Bellingham, Washington.
- 1286 Yabuki, T., and M. Matsu’Ura (1992), Geodetic data inversion using a Bayesian informa-  
1287 tion criterion for spatial distribution of fault slip, *Geophysical Journal International*,  
1288 *109*(2), 363–375.
- 1289 Zecevic, M., L. De Barros, C. J. Bean, G. S. O’Brien, and F. Brenguier (2013), Investigat-  
1290 ing the source characteristics of long-period (LP) seismic events recorded on Piton de la  
1291 Fournaise volcano, La Réunion, *Journal of Volcanology and Geothermal Research*, *258*,  
1292 1–11.
- 1293 Zeller, S. S., and D. D. Pollard (1992), Boundary conditions for rock fracture analysis  
1294 using the boundary element method, *Journal of Geophysical Research: Solid Earth*,  
1295 *97*(B2), 1991–1997.
- 1296 Zoback, M., F. Rummel, R. Jung, and C. Raleigh (1977), Laboratory hydraulic fracturing  
1297 experiments in intact and pre-fractured rock, in *International Journal of Rock Mechanics*  
1298 *and Mining Sciences & Geomechanics Abstracts*, vol. 14, pp. 49–58, Elsevier.