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► To cite this version:

Agus Budi-Santoso, Philippe Lesage, S. Dwiyono, Sri Sumarti, S. Subandriyo, et al.. Analysis of the seismic activity associated with the 2010 eruption of Merapi volcano, Java. *Journal of Volcanology and Geothermal Research*, 2013, 261, pp.153-170. 10.1016/j.jvolgeores.2013.03.024 . hal-01021907

HAL Id: hal-01021907

<https://hal.univ-grenoble-alpes.fr/hal-01021907v1>

Submitted on 15 Jul 2014

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1 Analysis of the Seismic Activity Associated with the 2010 2 Eruption of Merapi Volcano, Java

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5 Agus Budi-Santoso^{1,2}, Philippe Lesage², Sapari Dwiyono¹, Sri Sumarti¹, Subandriyo¹,
6 Surono¹, Philippe Jousset^{4,5}, Jean-Philippe Metaxian².

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8
9 ¹Badan Geologi, Jalan Diponegoro No. 57, 40122 Bandung, Indonesia (agusbudisantoso@yahoo.com),
10 (s4par1@yahoo.co.id), (merapi_bpptk@yahoo.com), (jsubandriyo@gmail.com), (surono@vsi.esdm.go.id). +62 22 727 2606

11 ²ISTerre, CNRS, Université de Savoie, IRD 219 73376 Le Bourget du Lac cedex, France. (lesage@univ-savoie.fr),
12 (jean-philippe.metaxian@ird.fr).

13 ⁴BRGM, RIS, 3 Avenue Claude Guillemin, BP36009, 45060 Orléans Cedex 2, France

14 ⁵Now at Helmholtz Center GFZ, Telegrafenberg, 14473 Potsdam, Germany (pjousset@gfz-potsdam.de) +49 30 288 1299

15 16 **Keywords**

17 **Merapi Volcano, Volcano Seismology, Eruption Forecasting, Pre-eruptive Seismicity,**
18 **RSAM, Material Failure Forecast Method, Source Location.**

19 20 21 **Abstract**

22 The 2010 large explosive eruption of Merapi is the first episode of this type that has been
23 instrumentally observed on this volcano. The main features of the seismic activity during the
24 pre-eruptive period and the crisis are presented in this paper. The first seismic precursors were
25 a series of four shallow swarms 12 to 4 months before the eruption. They are interpreted as
26 resulting from perturbations of the hydrothermal system by increasing heat flow. The
27 precursory seismic activity strictly speaking started about 6 weeks before the explosion of
28 October 26th. During this period, the rate of seismicity increased almost constantly yielding a
29 cumulative seismic energy release for volcano-tectonic (VT) and multiphase events (MP) of
30 $7.5 \cdot 10^{10}$ J. This value is 3 times the maximum energy release before former effusive eruptions
31 of Merapi. The high level reached and the accelerated behaviour of both the deformations of
32 the summit and the seismic activity are distinct features of the 2010 eruption with respect to
33 previous events.

34 The hypocenters of VT events are split into two clusters with depths (below the summit) of
35 [2.5-5] km and less than 1.5 km, respectively. The aseismic zone at [1.5-2.5] km depth is a
36 robust feature that was already detected in previous studies. This could correspond to a poorly
37 consolidated layer which is part of the ‘Ancient Merapi’ structure. Most of deep VT events
38 occurred before October 17th. After that, shallow activity strongly increased. This migration
39 of the seismic sources is consistent with the final stage of a rapid magma ascent before the
40 eruption. The deep seismic activity is interpreted as associated with the failure and
41 enlargement of a narrow conduit by a large amount of rapidly ascending magma, while the
42 shallow seismicity could be related to the rupture of the summit plug.

43 Hindsight forecastings of the occurrence time of the eruption are performed by applying the
44 Materials Failure Forecast Method (FFM). They use cumulative RSAM calculated either on
45 the raw records or on signals classified according to their dominant frequency. Stable
46 estimations are obtained during the last 6 days with fluctuations as small as ± 4 hours around
47 the time of the first explosion. This approach could thus be useful to support decision making
48 in the case of future explosive episodes at Merapi assuming that similar precursory processes
49 will occur.

50

51 **1. Introduction**

52 Merapi is located on Java Island, at about 30 km north of the city of Yogyakarta. It is
53 considered to be one of the most dangerous volcanoes of Indonesia because of its densely
54 populated surroundings and its high level of eruptive activity. The recent history of Merapi
55 (Voight *et al.*, 2000) is characterized by two eruptive styles: 1) effusive growth of viscous
56 lava domes, with typical recurrence of 4 to 6 years, that gravitationally collapse producing
57 pyroclastic flows known as « Merapi-type nuées ardentes »; 2) more exceptional explosive
58 eruptions of relatively large size, associated with column collapse and pyroclastic flows
59 reaching large distances. The October – November 2010 eruption is the first explosive type
60 event of Merapi (VEI ~ 4) that has been recorded by a multiparametric monitoring network
61 and that was not preceded by emergence of lava dome. Previous instrumentally observed
62 eruptions, in 1984, 1986, 1992, 1994, 1997, 1998, 2001, and 2006 (VEI = 1 - 3) were
63 common lava extrusions followed by dome collapses. This situation offers a unique
64 opportunity to compare the seismic activity associated with the two types of eruption and to
65 look for precursory evidences of a transition between effusive and explosive styles.

66 As for most volcanoes in the world, the seismic activity of Merapi is characterized by a large
67 variety of events that correspond to different locations and physical processes of the sources.
68 Since 1984, the classification of events at Merapi includes the following types: deep (VTA)
69 and shallow (VTB) volcano-tectonic, multiphase (MP), low frequency (LF), very long period
70 (VLP) events, tremor and rock fall (Ratdomopurbo and Poupinet, 2000). Hypocenter
71 distributions of VT events display an aseismic zone at 1.5-2.5 km depth (Ratdomopurbo and
72 Poupinet, 2000; Wassermann and Ohrnberger, 2001; Hidayati *et al.*, 2008) that has been
73 interpreted as a ductile high-temperature zone.

74 Eruptions at Merapi are generally preceded by VT and MP seismicity on varying time scales
75 from weeks to months (Ratdomopurbo and Poupinet, 2000; Voight *et al.*, 2000; Suharna *et al.*,
76 2007). However, some eruptions were not preceded by seismicity increase such as in 1986
77 and 1994. These later events are interpreted to be gravitational collapses of the dome. In 1991,
78 about 25% of the shallow VT events belonged to seismic multiplets. These families of events
79 with similar waveforms correspond either to sources very close to each other with identical
80 focal mechanisms or to non-destructive and repetitive sources. Ratdomopurbo and Poupinet
81 (1995) analyzed multiplets during 1992 by using a cross-spectral method on the coda waves.
82 They detected an increase of the seismic velocity of 1.2 % inside the volcano that may be
83 related to the pressurization of the magma feeding system. Using records of a repeatable
84 controlled source, Wegler *et al.* (2006) also observed an increase of the shear velocity before
85 the 1998 eruption.

86 The velocity structure of Merapi is still poorly known. Active experiments using air gun shots
87 in water basins have been carried out to investigate this structure (Lühr *et al.*, 1998; Wegler *et*

88 *al.*, 1999). Because the direct P- and S-waves generated by superficial sources were rapidly
89 attenuated due to strong scattering by the heterogeneous medium, no velocity model could be
90 obtained by this approach. However, the spindle-like shape of the seismogram envelopes could
91 be explained by a diffusion model (Wegler and Lühr, 2001). Thus hypocenter determinations
92 generally use a homogeneous model with P-wave velocity of 3 km s^{-1} (Ratdomopurbo, 1995;
93 Hidayati *et al.*, 2008) or 2.8 km s^{-1} (Wassermann and Ohrnberger, 2001). At a larger scale, a
94 tomographic study (Koulakov *et al.* 2007, 2009; Wagner *et al.* 2007) revealed an
95 exceptionally strong velocity anomaly in the crust between Merapi and Lawu (eastern of
96 Merapi) volcanic groups, interpreted as a zone with high content of fluids and melts feeding
97 the active volcanoes in the area.

98 Focal mechanisms of VT events recorded in 2000-2001 have been estimated by Hidayati *et al.*
99 (2008) using both polarity and amplitude of P-wave first motions. For VTA and most deep
100 VTB events, they are of normal-fault types while VTB located close to the surface are of both
101 reverse and normal fault types. Hidayat *et al.* (2000; 2002) studied very-long period (VLP)
102 events that occurred in 1998. These pulses with periods of 6-7 s and displaying similar
103 waveforms from event to event are coeval with MP or LF earthquakes. Hidayat *et al.* (2002)
104 carried out moment tensor inversion of the waveforms and proposed a source model
105 consistent with a dipping crack located at about 100 m under the dome. They suggested a
106 source process involving the sudden release of pressurized gas through the crack over a time
107 span of about 6 s. No VLP events were observed during the active periods of 2001 and 2006,
108 while a significant number of VLP events were observed in 2010 prior to and during the
109 eruption (Jousset *et al.*, this issue).

110 In this paper, we present some aspects of the seismic activity of Merapi in the year preceding
111 and during the 2010 eruption. We give a description of the types of seismic events observed
112 and a detailed chronology of the seismicity during this period. The spatial and temporal
113 distribution of the VT earthquake hypocenters provides important information on the pre-
114 eruptive processes in the structure. We apply the Material Failure Forecast method (Voight,
115 1988) to the RSAM values and test the potential of this approach to forecast the time of the
116 eruption onset. The comparisons between the features of the 2010 seismicity and those of
117 preceding eruptions give some clues to distinguish explosive from effusive impending
118 eruption.

119

120 **2. Seismic network**

121 The monitoring system of Merapi, operated by BPPTK (Balai Penyelidikan dan
122 Pengembangan Teknologi Kegunungapian) – Volcano Observatory of Yogyakarta which
123 belongs to CVGHM (Centre of Volcanology and Geological Hazard Mitigation), is mainly
124 based on seismic, deformation and geochemical measurements. The permanent seismic
125 network consists of four short-period (SP) stations equipped with L4C and L22 seismometers.
126 Signals are transmitted to Yogyakarta by radio with VHF modulation and are digitized by a
127 Güralp DM16S acquisition system at a rate of 100 samples per second with 16 bits accuracy.
128 SP stations have been used as reference stations in routine analysis, such as event
129 classification and counting, source location, and seismic energy calculations. In addition up to
130 six broadband (BB) stations using Güralp CMG-40TD seismometers with period 60 s and
131 TCP/IP protocol for data transmission have been installed on July 2009 and February 2010.
132 Both types of stations use GPS clocks for synchronization and Güralp Compressed Format

133 (GCF) for data file storage. Fig. 1 shows the configuration of the monitoring network; seismic
134 stations are located on and around the volcano at distances to the crater ranging from 0 to 6
135 km.

136 Some breakdowns in stations reduced the amount of available records during the pre-eruptive
137 period. Fig. 2 summarizes the operation time intervals. Furthermore, the GPS clock of some
138 broadband stations failed during several time intervals. In order to use arrival times from these
139 stations for source location, a procedure of clock re-synchronization, based on seismic noise
140 correlation (Stehly *et al.*, 2007; Sens-Schönfelder, 2008), was applied. The cross-correlation
141 function (CCF) of the noise recorded in two stations is directly related to the Green function
142 between the two sites (e.g. Campillo, 2006). When the clock of one of the stations is drifted,
143 the CCF is delayed by the same lag with respect to that obtained when both clocks are
144 synchronized. Thus, by locking for the maximum of the correlation function between the
145 shifted and the reference CCF, it is possible to estimate the delay and to synchronize the
146 stations. An estimated precision of ~ 0.05 s is obtained with this approach which uses low-pass
147 (< 4 Hz) filtered signals (Fig. 3).

148

149 **3. Main features of the seismic events**

150 Since the installation of a telemetered network in 1982, the same classification of seismic
151 signals has been used at Merapi for sake of consistency (Ratdomopurbo, 1995; Ratdomopurbo
152 and Poupinet, 2000). The main types of signal are volcanotectonic (VT), multiphase (MP),
153 low-frequency (LF), rockfall (RF), and tremor. VT events are characterized by clear onsets
154 and high frequency content (up to 25 Hz). They are associated with brittle failure in the rock
155 and have generally simple double-couple mechanism (McNutt, 1996). VT events are similar
156 to common tectonic earthquakes. The main differences are that the former are related to
157 volcanic activity and they frequently occur in swarms (McNutt, 2000).

158 VT's at Merapi are sub-divided into deep (VTA) and shallow (VTB) events. VTA (Fig. 4a)
159 are characterized by hypocenters at depth larger than 2 km below the summit, and clear P- and
160 S-wave arrivals. VTB events have depths smaller than 2 km and more emergent onsets at
161 distant stations (Fig. 4b). For some events, S-waves are undistinguishable. VTA and VTB
162 events can be recognized by differences amplitude ratios for the first arrivals between summit
163 (PUS) and flank (DEL) stations. Their differences in waveform and amplitude are probably
164 related to a larger level of scattering and attenuation for paths in the shallow parts of the
165 structure for VTB than for deeper paths for VTA (Wegler and Lühr, 2001).

166 Multiphase earthquakes are characterized by emergent onsets, maximum frequency of 4 to 8
167 Hz, and shallow depth (Fig. 4c). This type of signal is similar to hybrid events in other
168 classification schemes (McNutt, 1996). They are related to magma flow in the upper conduit
169 and dome growth (Ratdomopurbo and Poupinet, 2000). Their rate of occurrence is sometimes
170 correlated with summit deformations (Beauducel *et al.*, 2000). Low-frequency earthquakes
171 (LF), also called long-period (LP) events, have generally emergent onset, lack of S-wave
172 arrival, and dominant peak frequency in the range [1-3] Hz (Fig. 4d). They are thought to be
173 generated by resonance of fluid-filled cavities in the structure produced by pressure
174 perturbations (Chouet, 1996). However, due to the strong attenuation of the high-frequency
175 waves, some events identified as LF at distant stations may be actually MP events (Hidayat *et al.*
176 *et al.*, 2000). Very-Long-Period (VLP) events occurred at Merapi in 1998 (Hidayat *et al.*, 2002)
177 and 2010 (Jousset *et al.*, this issue) but were not observed in 2001 and 2006. This type of

178 signal corresponds generally to the low frequency component of a MP or VT event and it is
179 interpreted as mass transfer of fluid inside the structure (Ohminato *et al.*, 1998; Legrand *et al.*
180 2000; Chouet *et al.* 2005; Waite *et al.*, 2008, Jolly *et al.*, 2012).

181 Tremor consists of long-lasting vibrations that are also associated with resonance effects in
182 cavities (Chouet B., 1988; Konstantinou and Schindwein, 2002), fluid flow (Rust *et al.*
183 2008), or degassing (Lesage *et al.*, 2006). At Merapi they are relatively sparse, of low
184 amplitude, and their spectra contain a few regularly spaced peaks, with fundamental
185 frequency of [2-5] Hz (Fig. 5). They occurred more frequently in 2010 than before previous
186 eruptions. Rockfalls (RF) are characterized by progressively increasing amplitude at the onset,
187 long duration and high frequency content (5 to 20 Hz). Pyroclastic flows (PF; Fig. 6), usually
188 generated by dome collapse, produce RF-type signals with fairly long duration (up to tens of
189 minutes) and large enough amplitudes to be recorded at the farthest stations.

190

191 **4. Chronological description of the pre- and co-eruptive seismicity**

192 This section summarizes the history of the seismic activity during the year preceding the
193 eruption, with focus on the last few weeks and during the crisis. It mainly relies on routine
194 manual counting and classification of events based on waveform shape. Daily statistics are
195 thus made in local time (GMT+7). Seismic energy given below is calculated using the
196 Gutenberg-Richter equation:

$$197 \qquad \qquad \qquad \log E = 11.8 + 1.5M \qquad \qquad \qquad (1)$$

198 where M is the magnitude (Gutenberg and Richter, 1956) and E is in ergs. Magnitude of VT
199 is calculated using local magnitude definition of Richter (1935, 1958). To minimize the
200 influence of distance on the determination of magnitude of VTA and VTB, amplitudes
201 measured at station DEL (2.6 km from summit) are used instead of that at the closest station
202 PUS (0.5 km) since DEL is at about the same distance to the clusters of VTA and VTB. On
203 the other hand, since the MP events always occurs at shallow depth and have low amplitude at
204 station DEL, PUS is used to calculate the magnitude (Ratdomopurbo, 1995). Adequate
205 amplitude corrections are applied to each station in order to get consistent magnitude
206 determinations. During inter-eruptive periods, the level of seismic activity is usually very low.
207 For example, following the 2006 eruption, an average of 5 MP and less than one VT per day
208 were registered. The total seismic energy (VT and MP) released per day was less than $0.4 \cdot 10^8$
209 J on average. The first evidence of unrest was four short duration (3 to 4 hours) VT swarms
210 that occurred on October 31st (Fig. 7), December 6th, 2009, February 1st, and June 10th 2010.
211 These swarms include a small number of detected events (14, 13, 6, and 30, respectively) with
212 maximum local magnitude of 2.5 and shallow depth (< 1 km). This kind of activity is
213 considered as an early precursor, as all the previous eruptions since at least 1992 were
214 preceded by a series of seismic swarms.

215 In early September 2010, the level of seismicity began to increase, with about 10 MP and 3
216 VT events per day and seismic energy released of $0.6 \cdot 10^8$ J per day. On the 12th at 8:23 local
217 time, a VT earthquake with local magnitude $M = 2.5$ and depth of 3 km was felt in the three
218 northernmost observation posts (Fig. 1). The earthquake was followed by a large rockfall at
219 10:21. A similar VT event occurred on September 13th with magnitude of 2.4 and the same
220 depth. As of September 19th, the rate of occurrence reached 38 MP, 5 VTA, and 6 VTB per

221 day, with a total energy of $6 \cdot 10^8$ J and a maximum magnitude of 2.6 (Fig. 8). This increasing
222 seismicity coincided with accelerating inflation of the summit, as revealed by repeated
223 distance measurements (Suroño *et al.*, 2012). On the basis of these observations, the alert
224 level was raised to II on September 20th, 2010 (Suroño *et al.*, 2012).

225 Harmonic tremors with weak amplitudes and durations of up to 70 minutes were detected
226 from September 30th to October 4th at stations closest to the crater (Fig. 5). Spectrograms
227 contain up to three regularly spaced peaks and display a phenomenon of frequency gliding
228 which corresponds to progressive decrease of the fundamental frequency, from about 5 to 3
229 Hz, and of the overtones frequency accordingly. This phenomenon forms cycles of 17 minutes
230 duration, approximately. During the intrusive phase, 1-26 October, more than 200 VLP events
231 were recorded, mostly at the summit stations and up to ~3 km from the crater. They are
232 characterized by frequency content in the range [0.01 – 0.2] Hz and they are coeval to VT,
233 MP, or LF events (Jousset *et al.* this issue).

234 The seismic activity continued to increase in October together with deformation rate, gas
235 emission, and changes in gas composition (Aisyah N. *et al.*, 2010). The daily number of
236 seismic events reached 56 VT on the 17th, 579 MP and a total energy of $51 \cdot 10^8$ J on the 20th.
237 An increasing number of rockfall also occurred with up to 85 events on the 20th (Fig. 8). The
238 alert level was raised to III on October 21st. On October 23rd – 24th, a total of 27 LF events
239 occurred with dominant frequency in the range [1.5 – 2.5] Hz. Some of them produced
240 amplitude saturation at short period stations. The largest ones were recorded at all the stations
241 and could be located at a few hundreds of meters beneath the summit. The level of seismicity
242 dramatically raised on October 24th to 26th. On the 24th, the number of VT, MP, RF and the
243 seismic energy were 80, 588, 194, and $59 \cdot 10^8$ J, respectively. On the 25th, the corresponding
244 values were 222, 624, 454, and $132 \cdot 10^8$ J. The alert was raised to level IV (evacuation) on
245 October 25th at 18:00 local time, 23 hours before the onset of the eruption. By the occurrence
246 of the first eruption on October 26th, 2010 at 17:02 local time (10:02 UTC), 232 VT, 397 MP,
247 269 RF and 4 LF were counted with an energy of $197 \cdot 10^8$ J.

248 The first phase of the eruption was phreato-magmatic explosive and produced a pyroclastic
249 flow that reached up to 5 km to the South (Suroño *et al.*, 2012). The duration of the
250 corresponding seismic signal was 330 s. On October 27th, the seismicity decreased to 7 VT,
251 34 MP, 1 LF, and 109 RF. During the two following days, the daily number of events rose
252 again to 34 VT, 129 MP, 222 RF, 7 PF, and to 67 VT, 223 MP, 354 RF, 32 small PF,
253 respectively. The eruptive activity decreased afterward and only 4 PF were observed on
254 October 31st. Meanwhile a burst of 22 LF events and a weak 13 minutes long episode of
255 tremor occurred this day.

256 High frequency tremor appeared on November 3rd in relation with more and more frequent
257 pyroclastic flows. At 11:00 local time, this tremor became continuous. At 16:05, authorities
258 decided to enlarge the restricted zone to a radius of 15 km from the summit. At 18:46 a
259 pyroclastic flow reached a distance of 9 km destroying seismic station KLA. On November
260 4th and 5th, the SP seismograms were saturated and individual events were undistinguishable.
261 However, by using low-pass filter ($f < 0.1$ Hz), it was possible to detect that the largest
262 eruption took place on November 5th at 00:01 local time (Nov. 4th at 17:01 UTC – Fig. 6).
263 This eruption lasted about 27 minutes, produced 15 km long pyroclastic flows, and destroyed
264 stations DEL and PUS and broadband stations at the summit of the volcano. The same day at
265 01:00 local time, the radius of the restricted zone was set at 20 km.

266 In complement to seismic features such as daily number of earthquakes and source location,
267 the cumulative energy of VT and MP events calculated over the preceding year has been used
268 at BPPTK for estimating the current state of activity (Fig. 9). For eruptions before 2010, this
269 energy ranged from 10^{10} J (10 GJ) in 1992 to $2.4 \cdot 10^{10}$ J in 1997 and 2006. Thus, in practice,
270 special attention is paid to the monitoring observations when this energy reaches 10^{10} J. On
271 October 16th, the cumulative energy was $2.4 \cdot 10^{10}$ J and an eruption or a dome extrusion was
272 expected. However, the energy rate increased more rapidly instead with a maximum value of
273 $0.62 \cdot 10^{10}$ J per day on October 25th. Together with the accelerating huge local deformations
274 (displacement of up to 3 meters), the very high value reached by the energy was one of the
275 key elements that pointed a much larger eruption than usual, yielding timely decision of
276 evacuation. It is interesting to note that the cumulative energy for the 2010 eruption is lower
277 than all the others between days 270 and 325 (95 to 40 days before the eruption onset). This
278 indicates that, while seismic energy is progressively released during a long period before
279 effusive eruptions, in the case of an explosive crisis, most of the energy is produced in the last
280 few days or weeks.

281

282 5. Source location

283 From the data base of seismic events of Merapi, beginning in October 2009, 679 events,
284 recorded on 4 to 9 seismic stations could be located. In the present work, the Hypoellipse
285 program (Lahr, 1999) was used with a homogeneous half space velocity model assuming $V_p =$
286 3 km s^{-1} and $V_p/V_s = 1.86$ (Ratdomopurbo and Poupinet, 2000). In order to estimate realistic
287 uncertainties on hypocenter positions, an approach by Monte-Carlo simulation was applied.
288 The observed arrival times were modified by random perturbations with Gaussian distribution
289 and standard deviation of 0.1 s, and new hypocenter positions were obtained. This procedure
290 was repeated 1000 times for each event. Outliers were removed by using the Thomson Tau
291 method (Thomson, 1985). These outliers represent a small proportion of the whole set of
292 solutions. The remaining solutions were used to calculate confidence ellipses for each event
293 by carrying out principal component analysis (Jackson, 1988) on the covariance matrix of
294 positions (Got *et al.*, 2011).

295 Figure 10 displays the results on source location. The histogram of the uncertainties on depth
296 (Fig. 10e) shows that most of them are smaller than 0.5 km, with a maximum number close to
297 0.3 km. 19 events with uncertainty on depth larger than 1 km were removed before plotting
298 the location map and cross-sections. Hypocenters are distributed at depth less than 5 km
299 below the crater, in a cylinder with elliptical section of 2 km x 1 km approximately and
300 longest axis in the NE-SW direction (Fig. 10a-c). The distribution in depth is split into two
301 separated clusters. The deepest one (about 116 events) lies between 2.5 and 5 km below the
302 summit. It contains VTA type events following the classification used at Merapi (see section
303 3). The shallowest cluster is constituted by VTB events with maximum depth of 1.5 km.
304 Consequently, it appears that an aseismic zone does exist at depth of 1.5 to 2.5 km below the
305 crater. This feature is also shown both by the histogram of the hypocenter depths and by the
306 probability density function of the source depths which display clear minima at 1.5 – 2.5 km
307 depth (Fig. 10d). In order to verify whether this gap is due to an artefact of the hypocenter
308 determination, source depths are plotted as a function of differences of P-waves arrival times
309 ($t_{DEL} - t_{PUS}$) between stations DEL (located 1.5 km below the summit) and PUS which is close
310 to and 200m below the summit (Fig. 10f). Again, two clusters can be observed in this
311 representation, separated mainly along the ($t_{DEL} - t_{PUS}$) axis. Values of ($t_{DEL} - t_{PUS}$) in the
312 range $[-0 - 0.25]$ s are associated with deep VTA events while time differences of 0.35 to 1 s

313 correspond to shallow VTB earthquakes. The relative lack of values between 0.25 and 0.35 s
314 is a robust observation and is consistent with the existence of an aseismic zone at 1.5 – 2.5 km
315 depth.

316 The four seismic swarms that occurred from October 2009 to June 2010 were located at less
317 than 1 km below the summit. They include thus mostly VTB events. In Fig 11, hypocenter
318 depths are plotted as a function of time. The numbers of VTA and VTB per day are also
319 presented. It appears clearly that the VTA events occurred during the first part of the pre-
320 eruptive period until around October 17th. After that, while VTA activity was vanishing, a
321 sharp increase of the number of VTB events was observed until the eruption. Therefore, a
322 migration of the seismic activity from deep to shallow part of the edifice seems to have
323 occurred about 10 days before the eruption.

324

325 **6. RSAM and hindsight forecasting**

326 Real Time Seismic Amplitude Measurement (RSAM) is a robust tool for monitoring volcanic
327 activity because it provides a simple indicator of the level of seismic energy released (Endo
328 and Murray, 1991). At Merapi, real-time monitoring by RSAM was carried out during the
329 critical period of the eruptive crisis of 2010. For this, a module made by BPTTK calculated
330 every 5 minutes the RSAM value from the discriminator output of station KLA (Fig. 1). It
331 provided valuable information on the increasing seismic activity before the eruption and on
332 the energy of eruptive events during the crisis which was of great help in managing the
333 situation (Fig. 8f).

334 In order to carry out more detailed analysis, we recalculated RSAM from digital raw data of
335 station PUS as

$$336 \quad RSAM = \frac{\sum_{i=1}^n |A_i - \bar{A}|}{n} \quad (2)$$

337 where A_i is the signal amplitude, \bar{A} the mean amplitude in the calculation window, and n the
338 number of samples of the window. An initial window length of two minutes was used and, for
339 long-term analysis, a mean value every two hours was calculated. Because the whole
340 hardware of station PUS was replaced on April 2010 by new equipment with different
341 sensitivity, an amplitude correction was applied for sake of consistency between data recorded
342 before and after the change. Furthermore, because the tectonic earthquakes are not related to
343 volcanic activity, they were removed from RSAM values thanks to the daily seismicity
344 counting catalogue and low pass filtering to identify their coda.

345 RSAM is calculated on the continuous record which includes signals of all types. In order to
346 get some more details and as an attempt to separate the contribution of different types of
347 source, the following procedure was applied. For each 1-mn long window, a spectrum is
348 calculated and the frequency of its maximum is determined. Then the segment of record is
349 classified according to this peak frequency among the following ranges: [0.01 – 1] Hz, [1-3]
350 Hz, [3-5] Hz, [5-10] Hz and [1-15] Hz. Because a segment generally contains no more than
351 one event, this classification of signals roughly corresponds to the different types of event
352 defined in section 3. Range [1-15] Hz includes all the events but the noise is reduced. After

353 that, a cumulative value of RSAM is calculated for each frequency range. This procedure
 354 gives different results than the Seismic Spectral Amplitude Measurement (SSAM, Stephens *et*
 355 *al.*, 1994) which is the mean level of signal filtered in different frequency ranges. For
 356 example, in the range [1-3] Hz which contains mostly LF events, it helps separating the
 357 contribution of these low amplitude events that occurred a few days before the onset of the
 358 eruption (Fig. 12).

359 Starting October 2009, the average value of RSAM is almost constant in spite of some small
 360 bursts of energy related to the seismic swarms. A slight increase of RSAM is first observed on
 361 September 12th 2010, followed by an accelerating release of energy until October 6th. This
 362 day, a marked decrease of RSAM was observed. This behaviour appears clearly on the
 363 smoother curve of cumulative values which displays a discontinuity of its slope on October 6th
 364 (Fig. 13). After that, RSAM presents again an accelerating behaviour till the first eruption on
 365 October 26th. Other accelerating phases are observed before the eruptions of October 29th and
 366 November 3rd. The maximum values of RSAM provide also qualitative indications on the
 367 relative amplitude of the different stages of the eruptive sequence. The first eruption of
 368 October 26th is associated with a maximum RSAM value of $3.7 \cdot 10^5$ arbitrary unit (A.U).
 369 However, since the onset of eruption many seismic signals are saturated. Thus the RSAM
 370 associated with the eruption phases is under estimated. The following eruptions until
 371 November 2nd produce smaller maxima. RSAM peaks at $5.7 \cdot 10^5$ A.U on November 3rd and
 372 then reaches its highest values, $6.7 \cdot 10^5$ A.U., on November 4th, when the station was
 373 destroyed. The RSAM in the bands [3-5] and [5-10] Hz displays similar behaviour before the
 374 onset of the eruption. After that, RSAM in the band [3 – 5 Hz] displays a relative decrease in
 375 comparison with the other frequency bands. As the former range contains mostly VT events,
 376 this observation suggests that the fraction of energy released by brittle fracture is lower after
 377 the eruption onset. This is consistent with a condition of open conduit.

378 Accelerated rates of seismicity have been observed over different time scales before many
 379 eruptions (see e.g. Tokarev, 1971; Voight, 1988; Cornelius and Voight, 1994; Kilburn and
 380 Voight, 1998; De la Cruz-Reyna and Reyes-Davila, 2001; Kilburn, 2003; Smith *et al.*, 2007;
 381 Arambula *et al.*, 2011; Traversa *et al.*, 2011). This behaviour is at the basis of the Material
 382 Failure Forecast Method (FFM) that has been widely used for estimating the time of an
 383 eruption (Voight, 1988; Cornelius and Voight, 1994, 1995; De la Cruz-Reyna and Reyes-
 384 Davila, 2001). First introduced for the study of landslides (e.g. Fukuzono and Terashima,
 385 1985), the FFM assumes that a pre-eruptive stage is analogous to a damaging or creep process
 386 before the failure of the material. An observable Ω related to this process, such as
 387 displacement, strain, or level of seismic activity, is governed by an empirical power law
 388 between its rate of change $\dot{\Omega}$ and acceleration $\ddot{\Omega}$:

$$389 \quad \ddot{\Omega} = A \dot{\Omega}^\alpha \quad (3)$$

390 where A and α are constants that can be estimated from the observations (Cornelius and
 391 Voight, 1995). α is found to lie between 1 and 2, generally closer to 2 (Voight, 1988). For $\alpha =$
 392 2, equation 3 can be solved by integration yielding:

$$393 \quad \Omega = \frac{-1}{A} \ln \left[\frac{A(t^* - t) + (\dot{\Omega}^*)^{-1}}{A(t^* - t_0) + (\dot{\Omega}^*)^{-1}} \right] + \Omega_0 = B \ln(1 + st) + C \quad (4)$$

394 (Cornelius and Voight, 1995), where $\Omega(t = t_0) = \Omega_0$ and $\dot{\Omega}(t = t^*) = \dot{\Omega}^*$. B and C are constants
 395 and C can be chosen null. $s = -1/t_f$, with t_f the predicted time of failure or eruption which
 396 corresponds to Ω infinite. The time of failure t_f can be used as an estimation of the time of the
 397 eruption onset. However there can be a time delay between them (Voight, 1988; Bell *et al.*,
 398 2011a)

399 Exercises of hindsight prediction of the eruption time were carried out by fitting a function
 400 given by Eq. 4 to the observed cumulative values of RSAM. Note that RSAM is
 401 approximately proportional to the seismic moment-rate and energy-rate and thus can be used
 402 as $\dot{\Omega}$ (Cornelius and Voight, 1995). Thereby cumulative values of RSAM can be modelled by
 403 function Ω in eq. 4. For each trial, constants B, s, and t_f are estimated by least squares fitting.
 404 In this task, a crucial issue is the choice of the time window used to fit the model to the data.
 405 As described in section 4, a first clear increase of the seismicity was observed on September
 406 12th. Then a sudden decrease of the slope of the cumulative RSAM occurred on October 6th
 407 followed by another acceleration stage until October 26th. A first trial was made with a fitting
 408 window from September 13th to October 5th (Fig. 14). For this interval the adjustment is
 409 excellent (correlation coefficient of 99.9%) and the predicted failure time is on October 26th at
 410 07:00, 3 hours before the eruption onset. However a clear departure between the theoretical
 411 and observed curves appears after October 6th. Another trial was thus made with a fitting
 412 window starting on October 7th and ending on the 25th (Fig. 14). The predicted time is
 413 October 26th at 19:00 (time lag of 9 hours) and again, the correlation coefficient in this
 414 interval is very close to one. These first results confirm that the FFM model used is suitable to
 415 explain the observations in the two time periods. However large modifications probably
 416 occurred in the volcanic system around October 6th and make it more difficult to apply the
 417 method. More general solutions of eq. 3 were also considered, with $\alpha \neq 2$. In this case, α is an
 418 unknown parameter which is estimated together with t_f . In almost all these trials, the resulting
 419 value of α is very close to 2.

420 In order to test more in detail the robustness of the model as forecasting tool, a series of trials
 421 was carried out using different fitting time windows. In the following discussion, all dates are
 422 in October and $t = 20$ corresponds to October 20th for example. The windows have starting
 423 time $t_{start} = 7$ and their ending times t_{end} varies up to 26. The observations are either
 424 cumulative RSAM or cumulative RSAM calculated on signals classified by frequency range.
 425 The differences between the predicted time of failure t_f and the time t_{erupt} of the first eruption
 426 (October 26th 10:02 UTC) are plotted as a function of t_{end} in Fig 15.

427 For $t_{end} < 13$, the predicted time t_f is erratic and cannot be used. However, for $t_{end} > 13$, t_f
 428 varies smoothly as a function of t_{end} , displays variations between -5 and + 1.5 days around
 429 t_{erupt} and then converges toward t_{erupt} for $t_{end} > 20$ (Fig. 15a). Similar results are obtained with
 430 $t_{start} = 6$ or 8.

431 The use of RSAM calculated on signals classified following their dominant frequency yields
 432 similar results and seems to improve the precision of the prediction. For $t_{end} > 20$, $t_f - t_{erupt}$ is
 433 positive and smaller than 0.5 and 0.7 days, for ranges [0.01-1] and [1-15] Hz, respectively
 434 (Fig. 15b&c). For the band [5-10] Hz, $t_f - t_{erupt}$ is negative and tends to zero for increasing t_{end}
 435 (Fig. 15d). In the range [3-5] Hz, the estimated time of failure varies in the interval
 436 $t_f = t_{erupt} \pm 4$ hours during the last 6 days before the first eruption (Fig. 15c).

437 Because the deformation displayed acceleration behaviour before the eruption, the same FFM
438 approach can be applied. In this case, observations are the variation of the slope distance
439 between Kaliurang observatory and a reflector located on the southern part of the summit
440 (Fig. 1). Measurements were carried out by EDM (Electronics Distance Measurement) almost
441 every day. The adjustment between these observations and function Ω given by eq. 4 is not as
442 good as that obtained for RSAM (Fig. 16) Moreover, for the deformations, the estimated
443 values of t_f increase monotonically and tends toward the time of eruption onset for t_{end} close to
444 t_{erupt} (Fig. 15f).
445

446 7. Discussion

447 The seismic activity of Merapi during the pre-eruptive period and the eruption of 2010
448 presents features commonly observed during previous eruptions and some characteristics that
449 had never been recorded before, such as its high level of energy release. The types of event
450 identified in 2010 are similar to those observed since seismic stations were installed on the
451 volcano. Although empirically designed from waveform observations, the event classification
452 reflects the diversity of physical processes and locations of seismic sources. The two types of
453 volcano-tectonic event, VTA and VTB, correspond to two depth ranges for their hypocenters.
454 They are easily distinguished by different amplitude patterns in the seismic network and by
455 distinct differences of P-wave arrival times between stations, while it is difficult to recognize
456 them with only one station. The most numerous events are multi-phase, as hundreds of MP
457 signals were counted daily before and during the eruption (Fig. 6). They are interpreted as
458 fragile ruptures that trigger resonance response of an adjacent magma-filled conduit or crack.
459 They are mainly observed accompanying magma extrusions or in association with dome
460 instabilities. However MP events can also occur during periods of quiescence. Their origins
461 and source mechanisms are thus still not well-known and require further studies including
462 precise hypocenter determinations.

463 Unlike many other volcanoes, low-frequency (LF or LP) events and tremor are relatively
464 scarce on Merapi. Most of the mechanisms that have been proposed to explain these kinds of
465 event involve fluids interacting with the surrounding medium (Chouet, 1996). In the case of
466 Merapi, LF events mainly occurred at shallow depths after the first phreatomagmatic
467 explosions (Jousset *et al.*, – this issue). They probably result from the interaction of the
468 intrusive magma body with the hydrothermal system that lies beneath the summit (Müller and
469 Haak, 2004). The few harmonic tremors detected during the pre-eruptive period are probably
470 associated with the increasing gas emission. A possible mechanism for these vibrations is the
471 periodic opening and closing of a valve in a crack that produce intermittent pulses of gas with
472 frequency stabilized by the resonance of the fluid-filled cavity in the same manner as in a
473 clarinet (Lesage *et al.*, 2006). This process generates regularly spaced spectral peaks by the
474 Dirac comb effect and is an efficient mechanism to radiate seismic waves (Rust *et al.*, 2008).
475 The frequency gliding displayed in spectrograms may result from variations of the wave
476 velocity in the resonator due to varying content of two-phase fluid (such as gas bubbles in
477 water or magma), or to changes in the cavity itself modifying its length or stiffness.

478 Hypocenter determination is a difficult task on volcanoes because of lack of clear phase
479 arrivals, especially for MP, LF, and tremor events, sharp topography, limited knowledge of
480 the velocity structure, and, eventually, a too small number of stations. These drawbacks
481 produce sometimes bad precision in source location, especially in depth, yielding fuzzy
482 patterns of hypocenter distribution that are difficult to interpret. It is thus necessary to obtain
483 reliable estimations of uncertainties on source positions. While the errors calculated by the

484 programs of hypocenter determination depend mainly on the consistency between the
485 observed arrival times, the Monte-Carlo approach gives more robust estimations as it takes
486 the geometry of seismic rays into account (Got *et al.*, 2011). However, errors due to bad
487 knowledge of the structure are not included in this procedure. Following the Monte-Carlo
488 approach, the clouds of points obtained during a simulation provides an approximation of the
489 probability density function of the source position. Its maximum can be taken as the
490 hypocenter and its spread and shape reflect the precision of this determination. Nevertheless,
491 precise hypocenter determination for a larger proportion of earthquakes will require a
492 combination of a larger number of stations, with broader band and three-components
493 seismometers, seismic arrays, and better velocity models. Automatic data processing and
494 source location will be also very useful during the next crises.

495 The aseismic zone that appears between 1.5 and 2.5 km depth is a robust feature of the
496 seismicity of Merapi. The present results confirm the findings of Ratdomopurbo and Poupinet
497 (2000), Wassermann and Ohrnberger (2001), and Hidayati *et al.* (2008), that were obtained
498 for seismic events recorded in 1991, 1998, and in 2000-2001, respectively, and show that it is
499 a permanent structure over at least 20 years. Ratdomopurbo and Poupinet (2000) postulated
500 that it could correspond to the presence of a more ductile zone related to a small shallow
501 magma reservoir. However, deformation measurements (Beauducel *et al.*, 1999) and
502 electromagnetic data (Commer *et al.*, 2006) are not consistent with such a shallow storage
503 zone. Alternatively, the aseismic zone could correspond to the part of the Ancient Merapi left
504 by the Holocene sector collapses (Newhall *et al.*, 2000). This layer is mainly composed of
505 auto-brecciated lava flows, St. Vincent-type pyroclastic flows and lahar deposits
506 (Berthommier *et al.*, 1990). It is probably poorly consolidated and thus less seismogenic than
507 the surrounding layers. Indeed, it lies between the older structure of Pre-Merapi period and
508 the series of andesitic lava flows and pyroclastic flows of the Middle and Recent Periods
509 (Camus *et al.*, 2000).

510 The first seismic observation of unrest of the volcano was a series of swarms of shallow VT
511 events in October 2009, December 2009, February and June 2010. Seismic swarms are
512 generally triggered by variations of the effective stress in fractures (Saccorotti *et al.*, 2001). In
513 the case of Merapi, they could be related to the perturbation of the hydrothermal system due
514 to the intrusion of a deep hot body or to heating by increasing gas flow through the structure.
515 Some deeper earthquakes occurred also before and after the eruption in the close vicinity of
516 Merapi. Establishing a clear relationship between these events and the eruption requires more
517 detailed studies. The precursory seismic activity strictly speaking started at the beginning of
518 September 2010, about a month and a half before the eruption onset. Most VTA events, with
519 focal depths of 2.5 to 5 km, occurred before October 17th. After this date, VTA became very
520 scarce while shallow (< 1.5 km) VTB activity strongly increased. Although the focus of
521 seismic activity is not necessarily close to the head of a magmatic intrusion, the marked
522 change in hypocentral positions is quite consistent with the final stage of a rapid ascent of
523 magma shown by petrological data (Surono *et al.*, 2012).

524 The cumulative seismic energy release through VT and MP earthquakes during the year
525 preceding the eruption reached $7.5 \cdot 10^{10}$ J. For the previous eruptions of 1992 to 2006, this
526 energy never exceeded $2.5 \cdot 10^{10}$ J. This much higher level of energy is the most important
527 seismic characteristics of the 2010 eruption and is clearly consistent with its highly explosive
528 nature. Most of this energy was emitted in the last 6 weeks before the initial eruption of
529 October 26th with a striking accelerating rate. Together with deformation and gas emission
530 measurements, this observation formed the basis of the identification of the impending large
531 eruption and the timely decision of evacuation within a more extended region than usual

532 (Surono *et al.*, 2012). Seismic activity originates mainly from mass movements inside the
533 structure, such as magma intrusion and gas release, which produces stress variations and
534 ground deformations. There is thus a relationship between seismic energy release,
535 deformation, and volume change (McGarr, 1976; Yokoyama, 1988). In 2010, the bulk volume
536 of juvenile deposits was estimated at 0.03 – 0.06 km³ (Surono *et al.*, 2012), while the
537 corresponding value was 0.01 km³ in 2006 (Sri-Sayudi *et al.*, 2007). The marked difference
538 between seismic energy release in 2010 and during previous eruptions can be thus related
539 with difference of magma volume. The high level of energy in 2010 is thus consistent with the
540 rapid ascent of a large amount of volatile-rich magma (Surono *et al.*, 2012). This much larger
541 volume of magma through the relatively narrow 2006 conduit produced rock damaging,
542 creep, connexion of pre-existing network of cracks, and failure (Voight, 1988; De la Cruz-
543 Reyna and Reyes-Davila, 2001; Kilburn, 2003), resulting in conduit enlargement and higher
544 seismic activity. From this point of view, the system could be considered as almost closed
545 before the 2010 crisis. The accelerating seismic activity is also related to the accelerating
546 deformation of part of the summit and both phenomena can be considered as precursory signs
547 of a large explosive eruption. This behaviour significantly differs from those observed before
548 the previous effusive eruptions where no strongly accelerating energy release or deformation
549 occurred during the pre-eruptive stage (Ratdomopurbo and Poupinet, 2000; Voight *et al.*,
550 2000).

551 The seismic energy release before this eruption is of medium order of magnitude compared
552 with those of other eruptions of andesitic or dacitic volcanoes. For example, it is much higher
553 than the cumulative energy release before the 1990 Kelut eruption ($5.6 \cdot 10^8$ J; Lesage and
554 Surono, 1995) and that of Redoubt in 1989-1990 ($> 10^8$ J; Power *et al.*, 1994). A similar order
555 of magnitude was obtained at Bezymianny in 1955 ($4 \cdot 10^{11}$ J), Tokachi in 1962 ($5 \cdot 10^{10}$ J), or
556 El Chichon (10^{11} J) in 1982 (Tokarev, 1985; Yokoyama, 1988). On the other hand, the energy
557 release at Merapi is more than one order of magnitude smaller than at Shiveluch in 1964 (1.2
558 10^{12} J; Tokarev, 1985) and at Mt St Helens before its large eruption in 1982 ($\sim 8 \cdot 10^{12}$ J;
559 Yokoyama, 1988; Qamar *et al.*, 1983). Note that in the latter two cases, the eruptions were
560 associated to the emplacement of a cryptodome and to the gravitational collapse of the
561 volcano flank. The corresponding mechanical behaviour of Shiveluch and Mt St Helens were
562 thus quite different from that of Merapi.

563 The accelerated rate of seismic energy was clearly reflected in the RSAM values and offered
564 an interesting opportunity to test the Material Failure Forecast method (Voight, 1988). The
565 results obtained with this model on Merapi show its ability to forecast the eruption time
566 several days before with good precision. The hindsight forecasting trials carried out in this
567 study may help describing a scenario of what would have been obtained if this FFM approach
568 were applied during the pre-eruptive period. Using a starting time on September 12th for
569 calculations, a first estimation of the eruption time would have been obtained before October
570 6th with very good precision (difference between predicted and eruption times of 3 hours).
571 After this date, because of a marked change in the RSAM tendency, the forecasted time would
572 have shown strong variations for increasing ending time of the fitting window. In addition a
573 larger and larger departure between observed and calculated curves would have appeared.
574 Then it would have been necessary to modify the starting time of the fitting window to the 7th
575 of October. In the subsequent daily trials, the estimated eruption time progressively converges
576 toward the previous estimation and, for ending time later than October 20th, it becomes quite
577 stable with a departure from the eruption time smaller than 1.5 days. The use of RSAM
578 calculated with signals classified according their dominant frequency completes and improves
579 the results. For example, for dominant frequencies in the range [3-5] Hz, the forecasted time

580 is quite stable and its difference with the time of occurrence of the first explosion is smaller
581 than 4 hours during the last 6 days of the pre-eruptive period.

582 These encouraging results are obtained with the assumption that exponent α in the basic
583 equation of FFM (eq. 3) is equal to 2. In previous studies, α was found to be close to this
584 value which yields mathematical simplifications in the solution of the equation (Voight, 1988;
585 Cornelius and Voight, 1995). In the present case, the direct estimations of α which gave
586 values close to 2 and the good fitting between the observations and the theoretical curves
587 confirms the choice of the exponent.

588 The accelerated behaviour of some parameters used in volcano monitoring has been
589 interpreted as resulting from damaging processes of solid materials before their failure
590 (Voight, 1988; 1989; Cornelius and Voight, 1994). Kilburn (2003) associated the accelerating
591 rate of seismicity with the growth and the progressive connection of arrays of pre-existing
592 fractures while magma propagates to the surface. De la Cruz-Reyna and Reyes-Davila (2001)
593 applied a Kelvin-Voigt viscoelastic model to describe the tertiary creep associated with
594 degradation and weakening of the medium preceding the rupture. They fitted a logarithmic
595 curve, similar to eq. 4, to the cumulative value of the root mean square of the seismic signal
596 recorded before eruptions of Colima volcano, Mexico, and gave correct predictions of their
597 time of occurrence. In any of these interpretative models, the system must be closed before
598 the unrest. Most of the features of the seismic activity preceding the 2010 eruption of Merapi
599 volcano indicate that the magmatic conduit was indeed closed or almost closed in relation to
600 the large volume of magma that was intruding. Therefore the physical conditions required by
601 the models to produce good estimations of the time of eruption were probably fulfilled in this
602 case. On the contrary, before preceding eruptions of Merapi, such as that of 2006, no marked
603 accelerating behaviour was observed which is consistent with a much smaller volume of
604 magma extruding through an open conduit. Thus, it appears that better forecasts could be
605 obtained with FFM for large explosive crisis than for small effusive events. Note that,
606 although FFM can provide useful indications on the onset time, it cannot forecast the size on
607 the impending eruption.

608 One of the main difficulties in using the FFM approach in real-time would have come from
609 the sharp variation in the RSAM rate that occurred around October 6th. Similar variations
610 were observed before two eruptions of Colima volcano (De la Cruz-Reyna and Reyes-Davila,
611 2001). In the case of Merapi, this change may be related with the upward migration of the
612 focal depths. This stage may be interpreted as the intrusion of the magma in the aseismic
613 ductile zone at 1.5 – 2.5 km depth followed by the progressive failure of the overlying plug
614 that was the last barrier for the magma to reach the surface. The magma progression through
615 layers of different mechanical strength may have produced variable load regimes on the
616 material yielding fluctuations in the accelerated behaviour of RSAM.

617 When using the FFM or similar methods for operational forecasting, it is of paramount
618 importance to take into account the many possible sources of uncertainty on the estimation of
619 the eruption time (Bell *et al.*, 2011b). Part of the uncertainty comes from the choice of the
620 time window used to fit the theoretical curve. In this work, several starting times and many
621 ending times of the window have been systematically tested in order to study the stability of
622 the estimations and to obtain more confident results. On the other hand, the models are
623 slightly improved when classified signals are used to calculate RSAM instead of the complete
624 raw records. The best results are obtained for signals with dominant frequency in the range [3-
625 5] Hz which contains mainly VT events. This is consistent with the interpretation of the FFM
626 method as discussed above. The very short lag between the estimated times of failure and the

627 eruption onset suggests that the first explosion occurred immediately after the rupture of the
628 last plug in the conduit. The displacement of part of the summit, measured by EDM, displays
629 also acceleration behaviour. However the use of these observations for eruption forecasting
630 does not give stable and usable solutions. This sector of the crater wall was probably partly
631 uncoupled from the rest of the volcano and it collapsed during the eruption. Its movement was
632 probably not representative of the deformation of the remaining parts of the structure.
633

634 **8. Conclusion**

635 After an exceptional eruption, it is of paramount importance to carry out a thorough analysis
636 of all the observations produced by the monitoring network and which could not be processed
637 in detail during the crisis. This paper presents some results obtained in this process for the
638 seismic data. Many aspects of the Merapi seismicity in 2010 still remain to be studied and this
639 work is in progress. However at this stage, important features have been already underlined.
640 Beside some early seismic swarms observed 12 to 4 months before the 2010 crisis, the
641 seismic activity of Merapi increased almost monotonically during the last 6 weeks. The
642 number of LF events, VLP events and tremors recorded in this period were larger than for
643 common eruptions. However the most relevant characteristics of the 2010 activity are 1) the
644 high level of seismic energy release, about three times the maximum value obtained for the
645 previous eruptions, and 2) the clear accelerated behaviour observed in the number of VT and
646 MP events, in the release of energy, and in the RSAM values. This behaviour is consistent
647 with the strong acceleration of the displacement of some benchmarks of the summit measured
648 by EDM. These features can be considered as clear evidences that the impending eruption
649 would be much larger than the frequent effusive events. Indeed, they contrast with those of
650 previous eruptions which were not preceded by such marked accelerations.

651 The good fitting and hindsight forecasting obtained in applying the FFM to RSAM calculated
652 in the pre-eruptive period result from the accelerated nature of this seismicity. This is
653 consistent with evidences that the volcanic system was almost closed with respect to the rapid
654 intrusion of a large volume of magma, in agreement with the high level of energy release and
655 the explosive eruption on 26th October. The trials of a posteriori prediction of the eruption
656 time shows that good precision can be achieved if magma and hypocenter migrations and/or
657 changes of load regime, which may modify the evolution of the observables, are identified
658 and if the forecasting strategy is adapted to this situation. The abrupt modification that
659 appeared in the RSAM rate around October 6th is probably related with the upward shift of the
660 most seismically active region from below to above the aseismic zone located between 1.5
661 and 2.5 km depth beneath the crater.

662 In the future, if an episode of unrest of the seismic activity of Merapi produces a large
663 cumulative energy release, with respect to that of the frequent effusive eruptions, with a
664 clearly accelerated rate, and if other observables, such as deformation or gas emission, present
665 similar behaviour, then a large explosion of the same type as that of 2010 should be
666 considered as highly probable. In this case, on condition that their limitations are well
667 understood, the FFM or similar methods would be of great help in supporting decision
668 making for evacuation.

669

670

671

672 **Acknowledgment**

673 We wish to acknowledge the efforts of BPPTK staff especially Ilham Nurdin, Febri Sadana,
674 and Purwoto for their struggle to ensure the seismic stations work properly. We are grateful to
675 VDAP USGS for the assistance and equipment of short period stations and to MIAVITA
676 project for the equipments of broadband stations. The MIAVITA project was financed by the
677 European Commission under the 7th Framework Program for Research and Technological
678 Development, Area “Environment”, Activity 6.1 “Climate Change, Pollution and Risks”.
679 We’d like to thank the Ministry of Energy and Mineral Resources of Indonesia for the
680 doctoral scholarship granted to Agus Budi-Santoso. This work was partially supported by the
681 Coopération Franco-Indonésienne funded by the French Ministère des Affaires Etrangères, by
682 the Université de Savoie and the Institut de Recherche pour le Développement. The catalogs
683 of seismicity of previous eruptions made by Suharna are greatly appreciated. Discussions and
684 suggestions from Jean Luc Got and Jean-Claude Thouret are gratefully acknowledged. We
685 thank Ulrich Wegler, John Pallister and an anonymous reviewer for their interesting and
686 useful comments.

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893 **Figure captions**

894 **Fig. 1.** Monitoring network of Merapi and location of short-period and broadband stations,
895 EDM reflectors and observation posts. Distance from Kaliurang observation post to reflectors
896 RK (dotted line) was measured by EDM. Another seismic station (CRM) located at 40 km
897 south from Merapi is out of range of the map.

898 **Fig. 2.** Operation intervals of seismic stations in 2009 and 2010. Black vertical line indicates
899 end of year 2009. Dotted vertical lines show first eruption onset (October 26th at 10:02) and
900 the largest eruption (November 4th at 17:01 UTC or November 5th at 00:01 local time). Most
901 of the stations were destroyed by the later eruptions.

902 **Fig. 3.** Clock synchronization by seismic noise cross-correlation. Two VT events recorded by
903 stations LBH (top) and PUS (middle) when they were synchronized (a) and while GPS clock
904 of LBH was out of order (b). Cross-correlation functions of noise (CCF, bottom panels)
905 between the two stations when clocks were either synchronized (a) or not (b). Time lag
906 between the two CCF is used to correct the clock drift.

907 **Fig. 4.** Different types of event observed on Merapi. For each sample, waveforms recorded at
908 two stations and spectrogram are displayed. a) volcanotectonic A; b) volcanotectonic B; c)
909 multiphase; d) low-frequency.

910 **Fig. 5.** An episode of tremor that occurred on October 1st, 2010 05:42. Seismogram recorded
911 at station PUS, spectrogram, and spectrum calculated on short window around 5000 seconds.

912 **Fig. 6.** Seismogram of station PUS on November 4th until station destruction (~21:30). Dotted
913 vertical line indicates onset of largest eruption at 17:01 UTC. Although record was saturated,
914 the eruption could be detected in low pass filtered ($f < 0.1$ Hz) seismogram (bottom panel).

915 **Fig. 7.** Seismogram of first swarm of October 31st, 2009. It lasted about 3 hours. Another
916 larger VT event occurred about 3 hours afterward (right edge of plot).

917 **Fig. 8.** Daily numbers of events for period September-December 2010. Dash-dot vertical lines
918 indicate date of alert level rising. Bottom panel shows daily RSAM obtained at the
919 observatory. The RSAM value on November 5th is about 5 times that of October 26th.

920 **Fig. 9.** Comparison of cumulative energy release of VT and MP earthquakes during one year
921 prior to several eruptions from 1992 to 2010.

922 **Fig. 10.** Hypocenters of VT earthquakes. a) Map of epicenters, b) NS cross section, c) EW
923 cross section. Hypocenters are indicated by crosses, with their 67 % confidence interval (pink
924 ellipses). d) Histogram of the hypocenter depths (black solid bar) and probability density
925 function of source depths (black hollow bar), calculated using Monte Carlo method. e)
926 Histogram of uncertainties on depth. f) Depths as a function of differences of P-wave arrival
927 times between stations DEL and PUS.

928 **Fig. 11.** Depth of events plotted as a function of time on periods of a) October 2009-October
929 2010 and b) September-October 2010. Daily numbers of VTA and VTB events are shown by
930 red and blue lines, respectively.

931 **Fig. 12.** a) SSAM and its cumulative value for the range [1-3] Hz. b) RSAM and its
932 cumulative value calculated with signals whose dominant frequency is in the interval [1-3]

933 Hz. c) Same as b), after removing the tectonic events. A marked increase of LF activity
 934 appears more clearly in the last few days before the eruption.

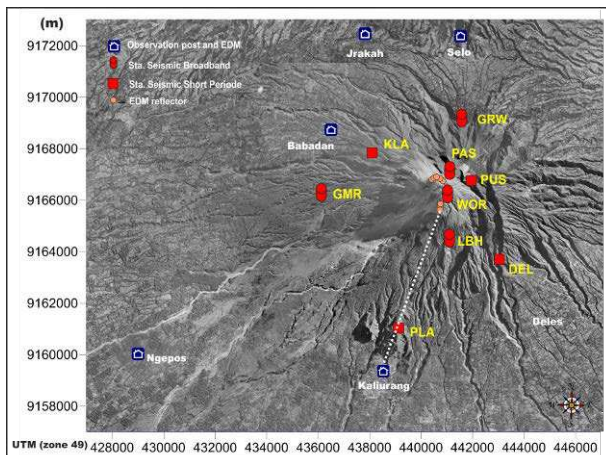
935 **Fig. 13.** RSAM calculated from station PUS (blue area) and its cumulative value (black line)
 936 during 3 months prior eruption. Cumulative RSAM for signals with dominant frequency in
 937 the ranges 0.01-0.1 Hz (yellow line), 1-3 Hz (brown line), 3-5 Hz (green line), 5-10 Hz
 938 (magenta line), and 1-15 Hz (red line). Brown dashed vertical lines and arrows indicate main
 939 explosions.

940 **Fig. 14.** Cumulative RSAM (black line) before eruptions. Theoretical curves calculated with
 941 FFM with fitting windows from September 13th to October 5th (red line) and from October 7th
 942 to 26th (blue line).

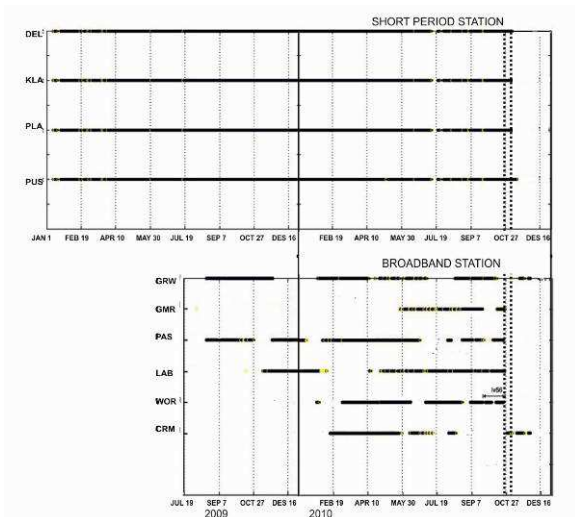
943 **Fig. 15.** Difference between predicted time t_f and time of eruption onset t_{erupt} as a function of
 944 ending time of the fitting window t_{end} , and calculated with $t_{start} = \text{October } 7^{\text{th}}$. Observations are
 945 a) unfiltered RSAM, b) RSAM calculated for signals with dominant frequency in the range
 946 0.01-1 Hz, c) same for 3-5 Hz, d) 5-10 Hz, e) 1-15 Hz, and f) variation of the slope distance
 947 measured by EDM.

948 **Fig. 16.** Variation of the slope distance between Kaliurang observatory and the southern part
 949 of the summit (circle) and theoretical curves (black line) obtained with different ending times
 950 of the fitting windows. Starting time is October 7th.

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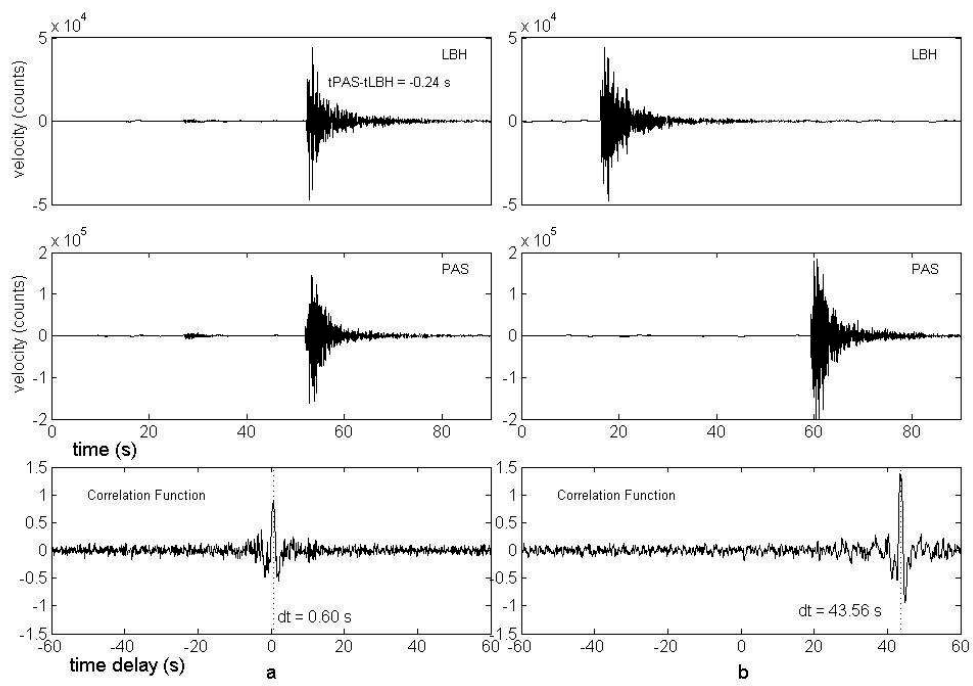
954
 955 Figure 1.



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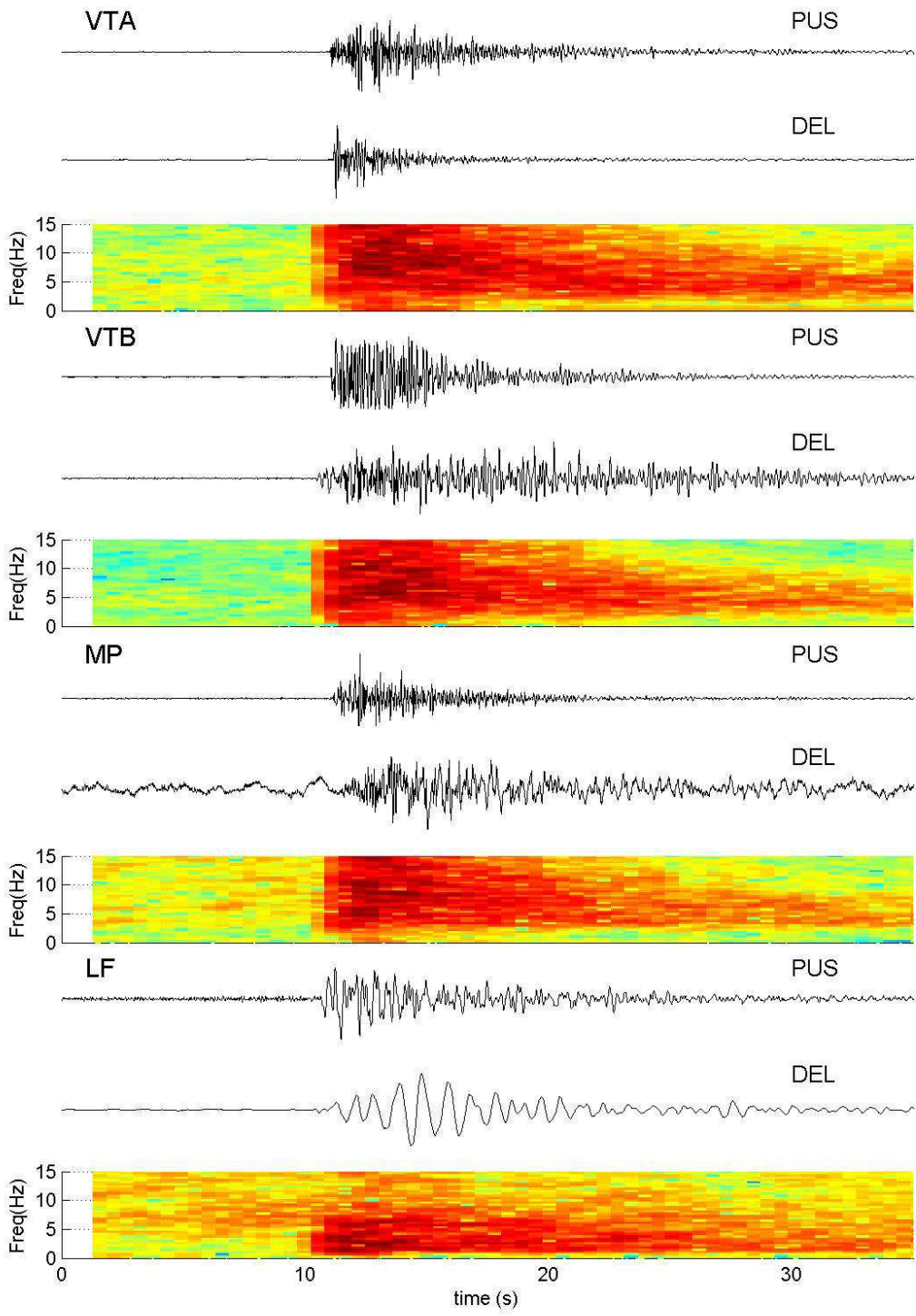
957 Figure 2.

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960 Figure 3.

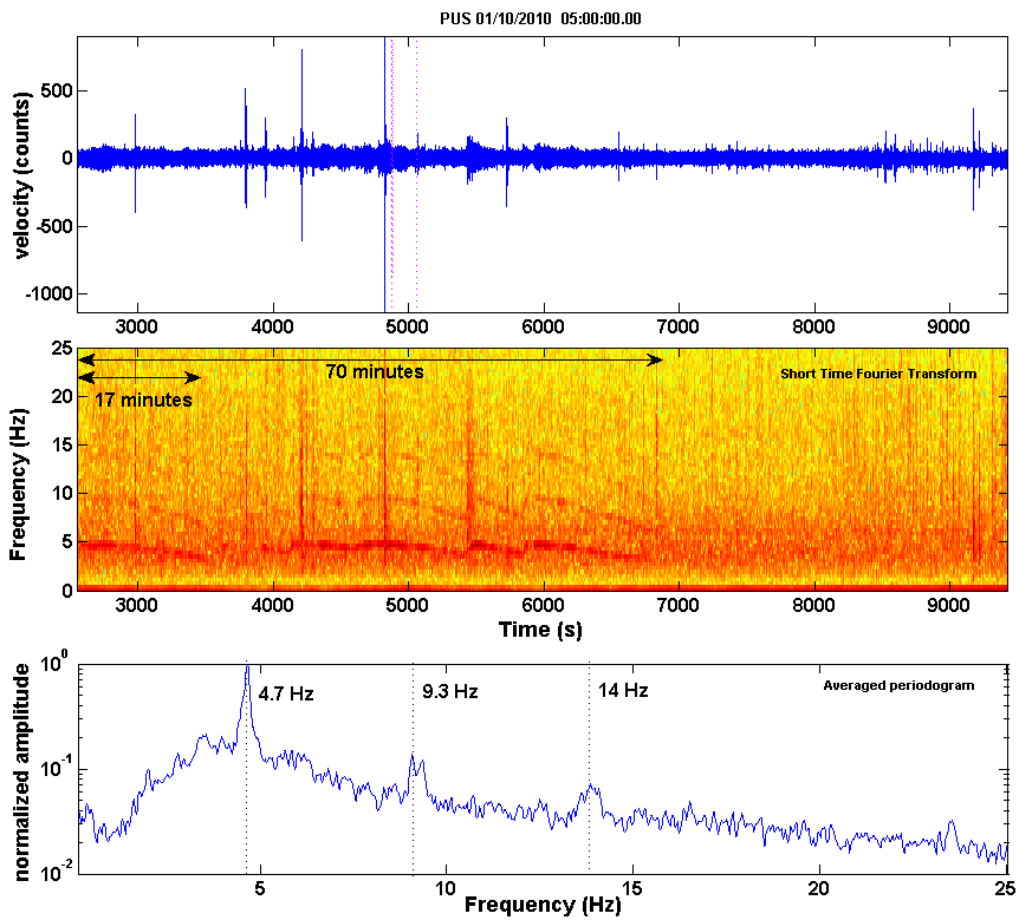


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962 Figure 4.

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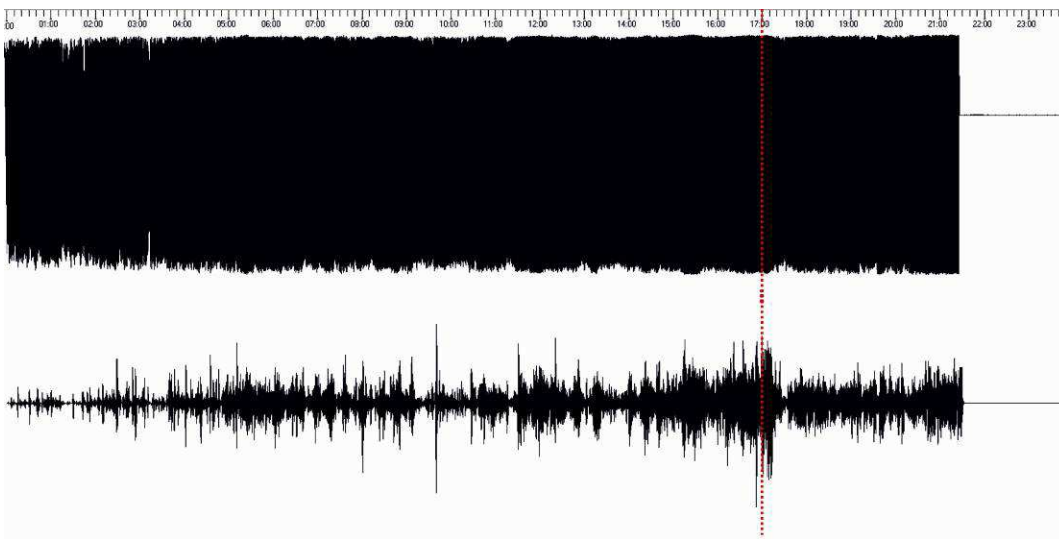
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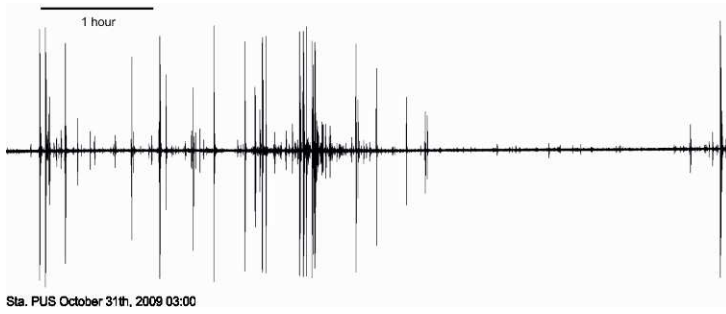
966 Figure 5.

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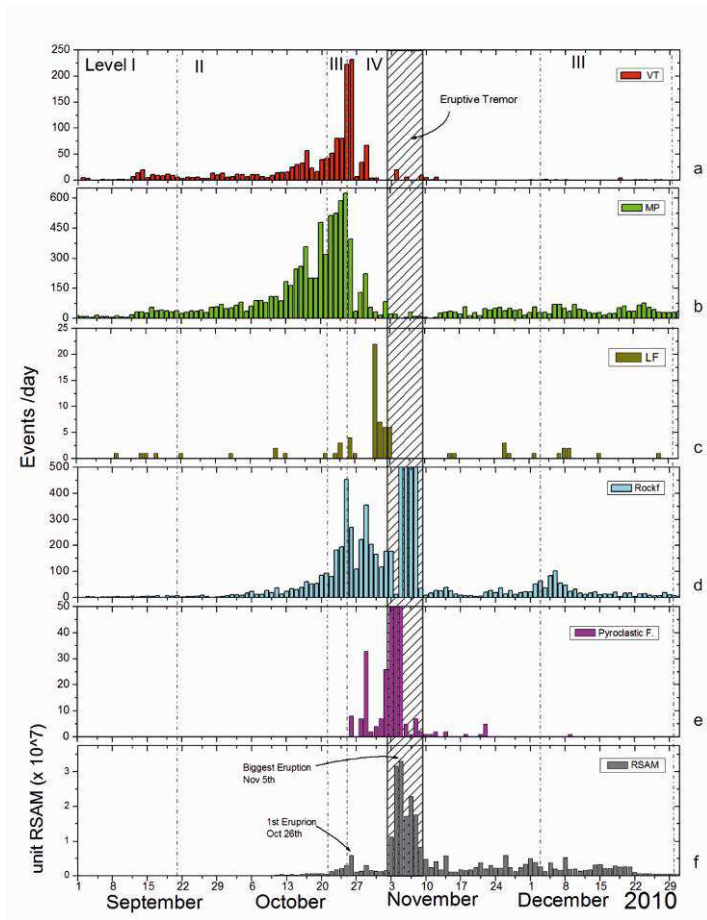
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969 Figure 6.



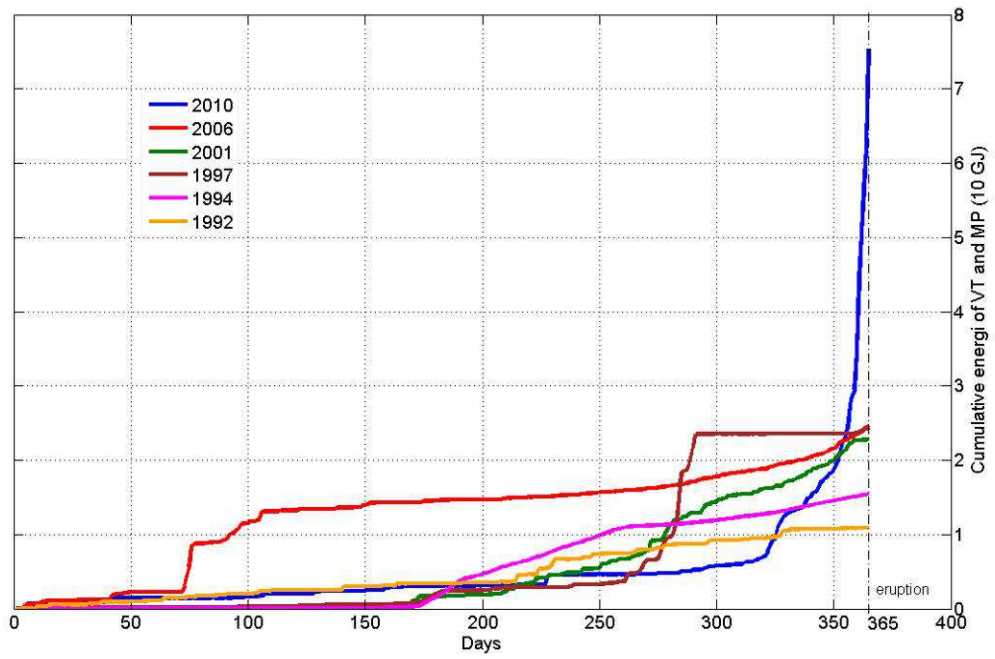
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971 Figure 7.



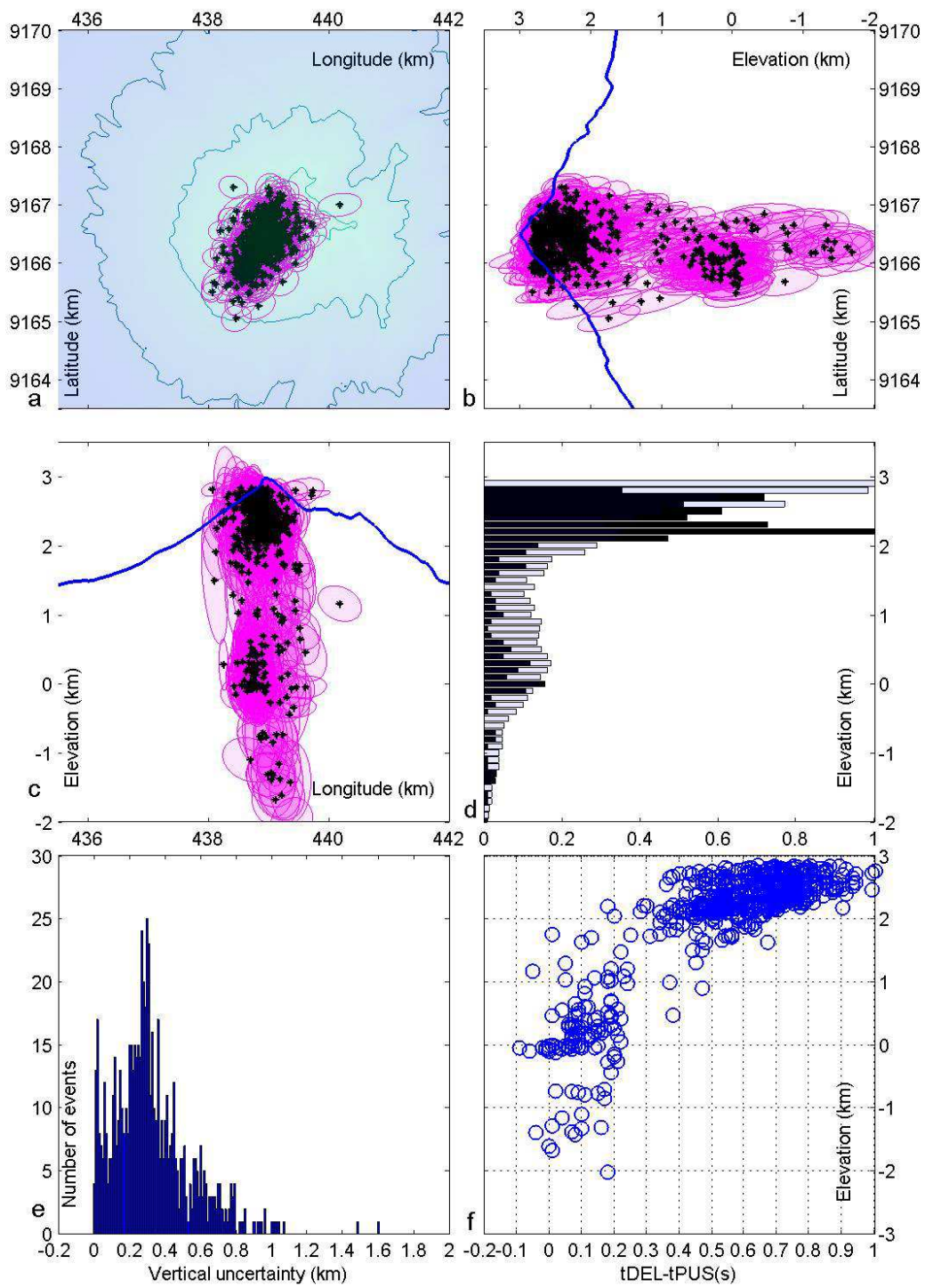
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973 Figure 8



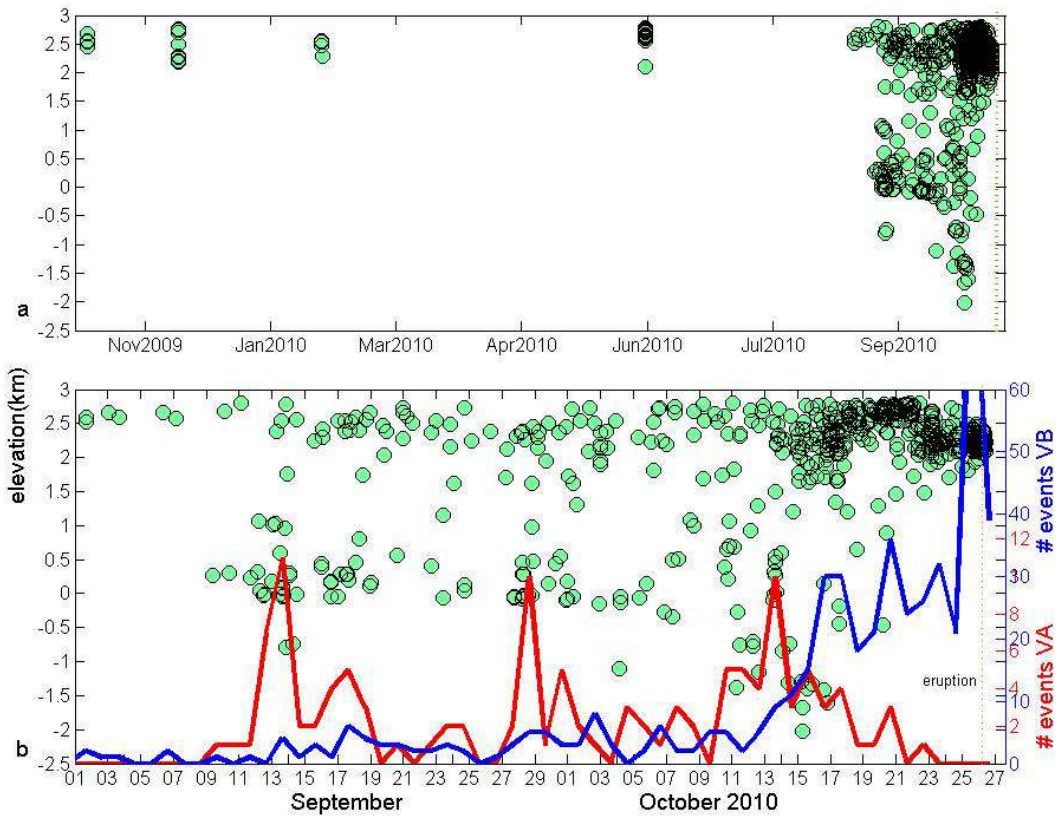
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975 Figure 9.



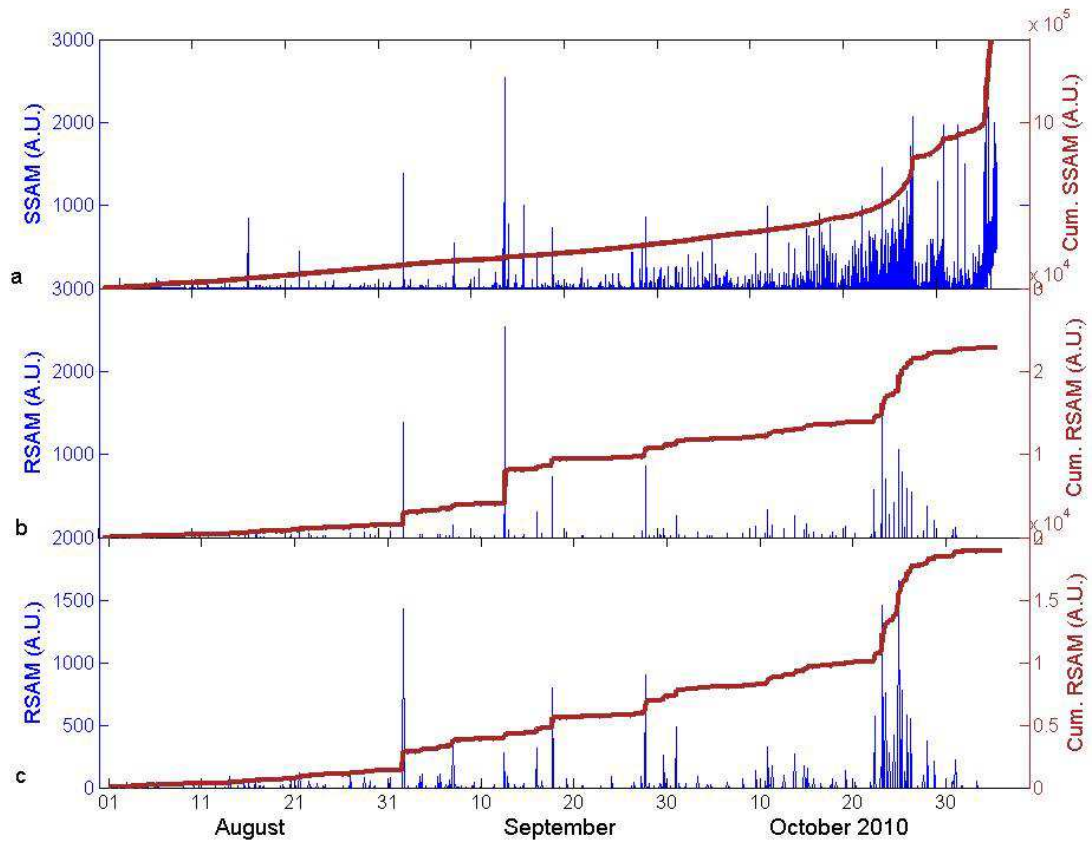
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977 Figure 10.



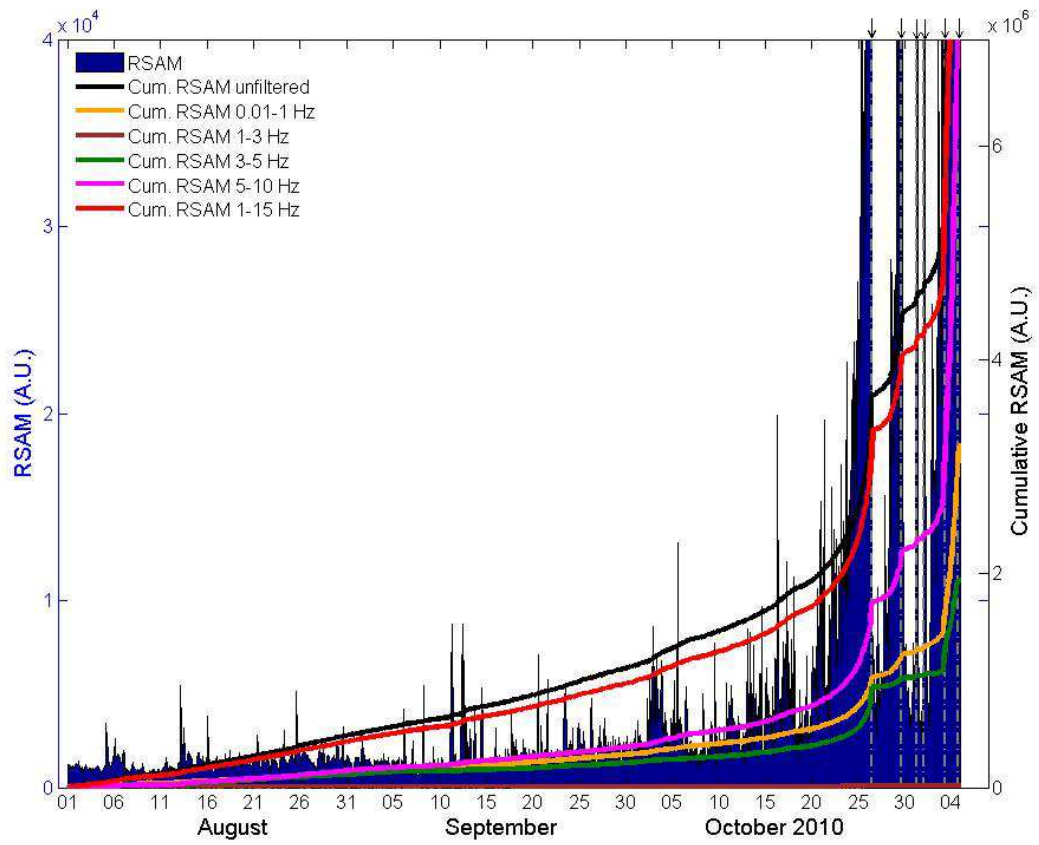
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979 Figure 11



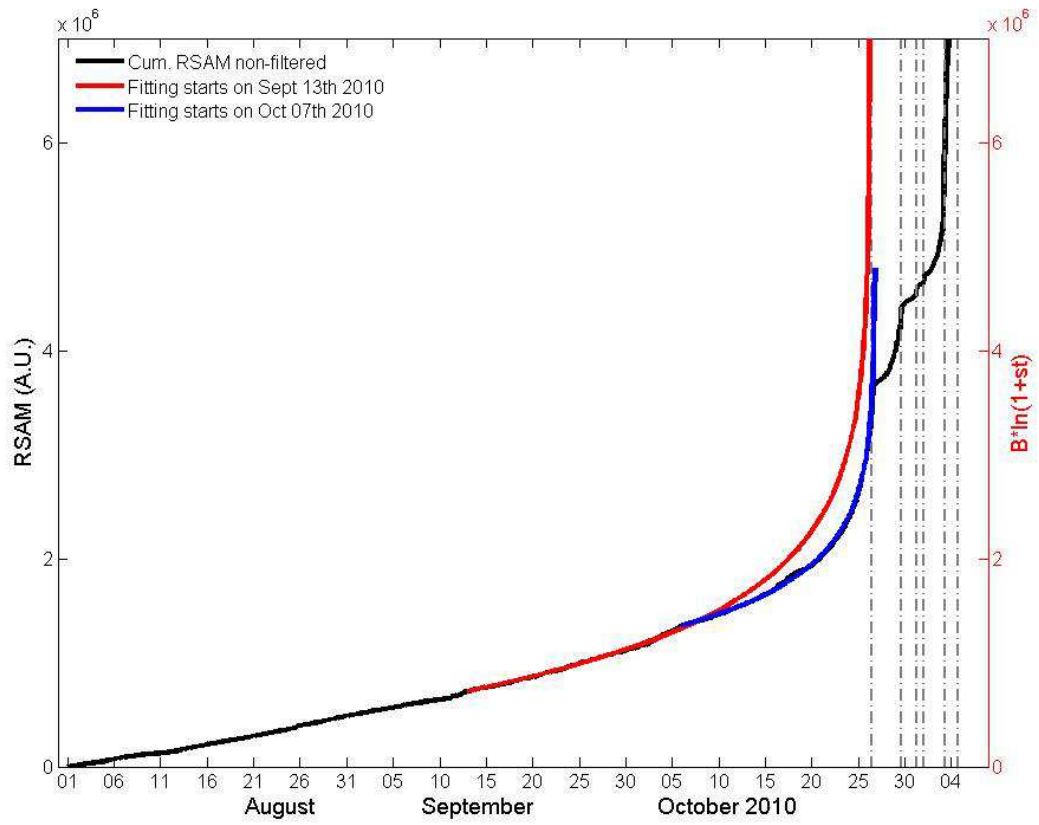
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981 Figure 12



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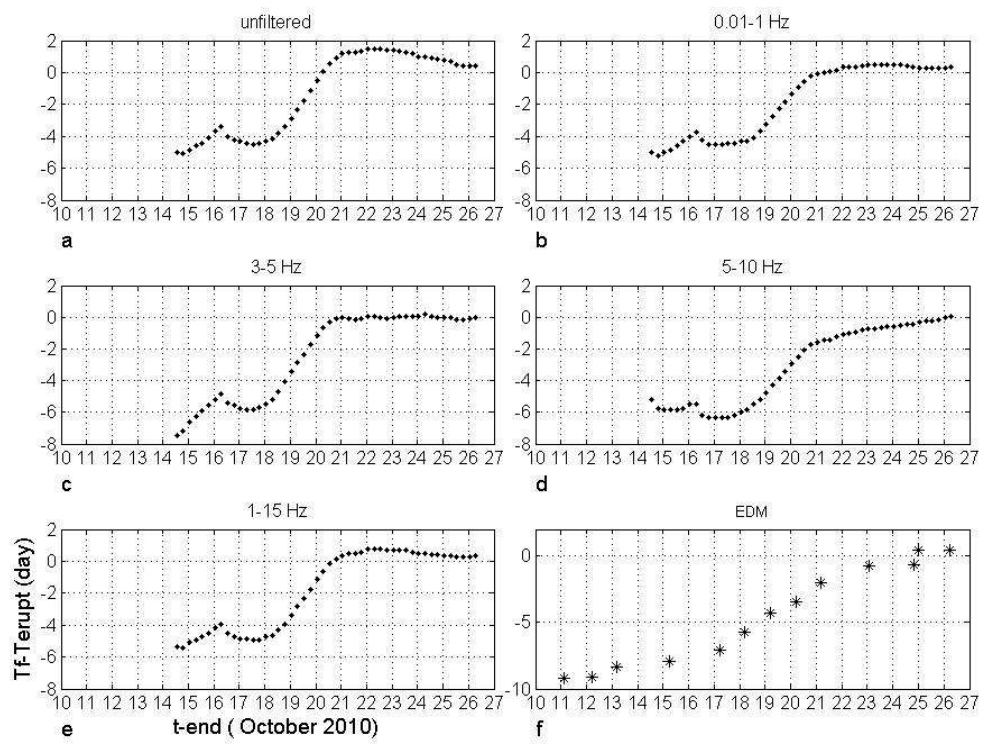
983 Figure 13



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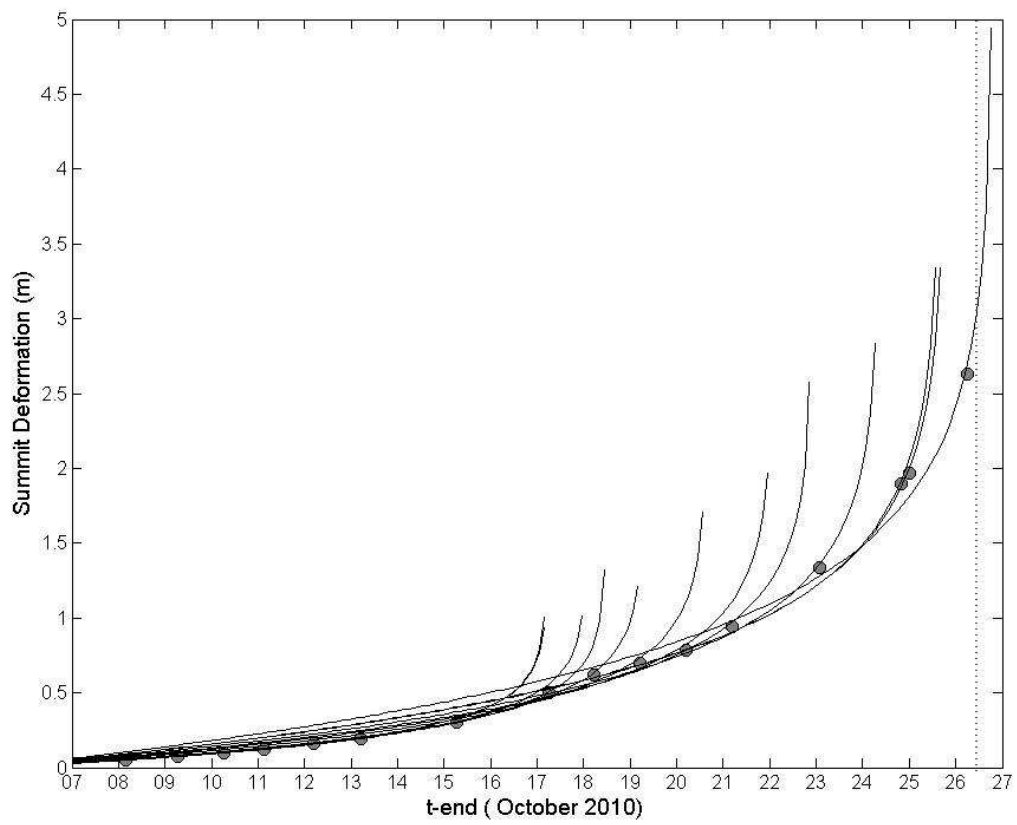
985 Figure 14

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990 Figure 16