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Temporal Variability of Ground Shaking and Stress Drop in Central Italy: A Hint for Fault Healing?

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1 Temporal variability of ground-shaking and stress
2 drop in central Italy: a hint for fault healing?

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18 Abstract

19 Ground Motion Prediction Equations (GMPEs) are calibrated to predict the
20 intensity of ground shaking at any given location, based on earthquake mag-
21 nitude, source-to-site distance, local soil amplifications and other parame-
22 ters. GMPEs are generally assumed to be independent of time; however,
23 evidence is increasing that large earthquakes modify the shallow soil condi-
24 tions and those of the fault zone for months or years. These changes may
25 affect the intensity of shaking and result in time-dependent effects that can
26 potentially be resolved analyzing between-event residuals (residuals between
27 observed and predicted ground motion for individual earthquakes averaged
28 over all stations). Here we analyze a data set of about 65000 recordings for
29 about 1400 earthquakes in the moment magnitude range 2.5-6.5 occurred in
30 central Italy from 2008 to 2017 to capture the temporal variability of the
31 ground shaking at high frequency. We first compute for each earthquake
32 between-event residuals in the Fourier domain with respect to a ground mo-
33 tion prediction equation developed ad-hoc for the analyzed data set. The
34 between-events show large changes after the occurrence of mainshocks such
35 as the Mw 6.3, 2009 L' Aquila, the 2016, Mw 6.2 Amatrice and Mw 6.5

36 Norcia earthquakes. Within the time span of a few months after the main-
 37 shocks, the between-event contribution to the ground shaking varies by a
 38 factor 7. In particular, we find a large drop in the between-events in the
 39 aftermath of the l'Aquila earthquake, followed by a slow positive trend that
 40 leads to a recovery interrupted by a new drop at the beginning of 2014.
 41 We also quantify the frequency-dependent correlation between the Brune
 42 stress-drop $\Delta\sigma$ and the between-events. We find that the temporal changes
 43 of $\Delta\sigma$ resemble those of the between-event residuals; in particular, during
 44 the period when the between-events show the positive trend, the average
 45 logarithm of $\Delta\sigma$ increases with an annual rate of 0.19 (i.e., the amplifica-
 46 tion factor for $\Delta\sigma$ is 1.56 per year). Breakpoint analysis located a change in
 47 the linear trend coefficients of $\Delta\sigma$ versus time in February 2014 although no
 48 large earthquakes occurred at that time. Finally, the temporal variability
 49 of $\Delta\sigma$ mirrors the relative seismic velocity variations observed in previous
 50 studies for the same area and period, suggesting that both crack-healing
 51 along the main fault system and healing of micro-cracks distributed at shal-
 52 low depths throughout the surrounding region might be necessary to explain
 53 the wider observations of post-earthquake recovery.

54 Introduction

55 The intensity of seismic shaking at a given site is a function of the earthquake
 56 size, style of faulting, source-to-site distance and site condition. Ground
 57 motion prediction equations (GMPEs) incorporate these functional depen-
 58 dencies and are time-independent, meaning that the intensity of shaking is

59 assumed to be independent of any process affecting the fault zone. How-
 60 ever, some observations suggest that shaking intensity changes depending
 61 on the timing of earthquakes within sequences. For example, it has been
 62 observed that aftershocks generate lower median ground motion in the high
 63 frequencies (e.g., Boore and Atkinson 1989) and, therefore, they have been
 64 flagged in the recent NGA-West2 strong motion data set (Wooddell and
 65 Abrahamson, 2014) allowing GMPE developers to treat them differently
 66 from mainshocks. Some of the variability in the observed shaking may be
 67 explained by variability in the stress drop (e.g., Wu and Chapman, 2017).
 68 Analyzing the NGA-West2 data, Baltay and Hanks (2014) found that av-
 69 erage stress drop for mainshocks is 30% larger than the average stress drop
 70 for aftershocks.

71 Identifying repeating deviations from the medial model can help distinguish
 72 processes that are not modeled but contribute significantly to ground mo-
 73 tion variability. For example, several studies correlated earthquake-specific
 74 residuals (also called between-event or inter-event) to stress drop variability
 75 (e.g., Anderson and Lei, 1994; Bindi et al., 2007; Cotton et al., 2013; Oth et
 76 al., 2017; Bindi et al., 2017; Baltay et al., 2017; Ameri et al., 2017; Trugman
 77 and Shearer, 2018). Therefore, presence of any temporal pattern in the dis-
 78 tribution of residuals can be used as diagnostic of time-dependent fault or
 79 medium properties, and ultimately help understanding how non-stationary
 80 processes, such as protracted seismic sequences or the long precursory phase
 81 of large earthquakes, affect ground motion. For example, Socquet et al
 82 (2016) and Piña-Valdés et al. (2018), in discussing the ground-shaking
 83 time-dependencies observed during the 2014 Iquique subduction sequence,

84 suggested that the temporal changes in the between-event residuals were
 85 associated with aseismic slip around the rupture area.
 86 Here we take advantage of a large data set from central Italy to investigate
 87 the temporal changes of the between-event residuals and their link with the
 88 stress drop $\Delta\sigma$ variability. The data set includes about 1400 earthquakes
 89 in the magnitude range 2.5-6.5, belonging to the main sequences of the last
 90 ten years, i.e. the 2009 L'Aquila (Chiaraluce et al., 2011) and the 2016-17
 91 Amatrice-Visso-Norcia (Chiaraluce et al., 2017) sequences. About 60 earth-
 92 quakes occurred in area of the 2013-14 Gubbio swarm (De Gori et al., 2015)
 93 are included as well. We first calibrate an ad-hoc GMPE for the Fourier
 94 amplitude spectra and we evaluate the between-event residuals at different
 95 frequencies. Then, the between-event are used as exploratory tool to detect
 96 event-dependent temporal changes in the ground shaking and we conclude
 97 presenting the temporal variability of $\Delta\sigma$.

98 Data

99 In this study, we analyze about 65000 recordings (for each component of
 100 motion) from 1400 earthquakes recorded by 340 stations installed in Central
 101 Italy (Figure 1 and Figure S1 of the electronic supplements to this article).
 102 The earthquakes cover the magnitude range from 2.5 to 6.5 and hypocentral
 103 distances from 10 to 180 km are considered. The data set includes the main
 104 sequences occurred in the area in the last 10 years, namely the 2009, Mw
 105 6.3 L'Aquila (indicated as e_1 in Figure 1); the 2016, Mw 6.1 Amatrice (e_2
 106 in Figure 1); the 2016 Mw 6.1, Visso (e_4 in Figure 1); the 2016, Mw 6.5,

107 Norcia (e_3 in Figure 1) (for a map with the time evolution of the events, see
 108 Figure S2 in the electronic supplements to this article). Following Bindi et
 109 al. (2018), in this study we consider the moment magnitude from the Geofon
 110 catalog for all events with $M_w \geq 5.7$ but the 2009, L'Aquila mainshock, for
 111 which we use the Global Centroid Moment Tensor (GCMT) value since the
 112 Geofon solution is not available (see the Data and Resources section). The
 113 data set also includes recordings from 59 earthquakes with magnitude larger
 114 than 2.5 occurred in the area of the 2013-14 Gubbio swarm; a complete
 115 description of the swarm is given by De Gori et al. (2015) and Valoroso et
 116 al. (2017). The station distribution is shown in Figure S1 in the electronic
 117 supplement to this article.

118 We analyse the Fourier amplitude spectra (FAS) of S-waves windows
 119 band-pass filtered with a variable high pass corner frequency depending on
 120 the signal-to-noise ratio. The Butterworth high-pass corner varies in the
 121 range 0.05–0.4 Hz while the low pass one was fixed to 40 Hz. The FAS are
 122 smoothed using the Konno and Ohmachi (1998) algorithm (the smoothing
 123 parameter b was set to 40). Details about the data selection and processing
 124 are provided by Pacor et al. (2016) and by Bindi et al. (2017).

125 **Source parameters**

126 For each earthquake, we consider the source parameter (i.e., stress drop and
 127 seismic moment) derived by Bindi et al. (2017) using a generalized inversion
 128 technique (GIT). In the GIT approach (e.g., Castro et al., 1990; Oth et al.,
 129 2011), the spectral values of a set of earthquakes recorded by a network of

stations are simultaneously inverted to isolate the contribution of source,
 propagation and site effects. The GIT approach exploits the redundancy
 of information (that is, the same earthquake is recorded at several stations
 located at different distances, and several earthquakes are recorded at the
 same station) to set-up an over-determined system of equations solved in a
 least-squares sense. To remove unresolved degrees of freedom which generate
 trade-offs among different components of the solution, some constraints are
 applied, such as the choice of a reference distance at which the attenuation
 is assumed to be one and a reference site condition (i.e., one or more stations
 whose site amplification is assumed to be known). In this study, we use the
 results of Bindi et al (2018) who applied a non-parametric GIT inversion
 where any a-priori seismological models for source and attenuation were
 adopted during the GIT inversion. To estimate the seismic moment and the
 corner frequency for each earthquake, the resulting non-parametric source
 spectra were fit to a Brune (1970) source model which assumes a circular
 fault with uniform stress drop. In Bindi et al. (2018), the source fit was
 performed allowing a deviation of the high frequency acceleration spectral
 level from a constant value as predicted by the Brune model. The high-
 frequency slope of the source spectrum is referred to as k_{source} . Given the
 seismic moment and the corner frequency, the stress drop was computed
 following Eshelby (1957) and Keilis-Borok (1959).

The source parameters are shown in Figure 2, in terms of scaling between
 $\Delta\sigma$, Mw and hypocentral depth. Most of the considered depths including
 those of the mainshocks are located between 5 and 10 km. The stress drop
 tends to increase with depth and has a strong magnitude dependence (Pacor

et al., 2016; Bindi et al. 2017). The mainshocks have the largest $\Delta\sigma$, around 10 MPa. The overall $\Delta\sigma$ variability covers almost three orders of magnitude. The procedure followed in this study to estimate the uncertainties on $\Delta\sigma$ is described in the electronic supplement to this article.

Ground motion model

In this study, we describe the Fourier amplitude spectra $FAS(f, R)$ at frequency f of S-waves recorded at hypocentral distance R with the following seismological model:

$$FAS(f, R) = S(f) \cdot P(f, R) \cdot Z(f) = K \frac{M_0 f^2}{1 + (\frac{f}{f_c})^2} \cdot \frac{1}{R^n} \exp(-\frac{\pi f R}{Q\beta}) \cdot Z(f) \quad (1)$$

where the acceleration source spectra $S(f)$ is parametrized considering an ω -square model (Aki, 1967) and the spectral attenuation with distance $P(f, R)$ is controlled by the geometrical spreading exponent n and the anelastic attenuation, the latter being modeled through the quality factor $Q(f)$. In equation 1, the constant K depends on the density and velocity at the source location, on radiation pattern and free surface amplification effects, whereas $Z(f)$ accounts for site amplification effects. We only consider far-field source terms and extended-source effects are not accounted for. The asymptotic form of the source spectrum is as follow:

$$S(f) \propto \begin{cases} M_0 f_c^2 & \text{if } f \gg f_c \\ M_0 f^2 & \text{if } f \ll f_c \end{cases} \quad (2)$$

172 The source spectrum depends on two parameters, the seismic moment
 173 M_0 and the corner frequency f_c connected through the stress drop $\Delta\sigma$
 174 (Brune, 1970; Eshelby, 1957) as follows:

$$\Delta\sigma \propto M_0 d^{-3} \propto M_0 f_c^3 \quad (3)$$

175 where d is the source radius. Considering equation 3, equation 2 can be
 176 re-written as:

$$S(f) \propto \begin{cases} M_0^{\frac{1}{3}} \Delta\sigma^{\frac{2}{3}} & \text{if } f \gg f_c \\ M_0 f^2 & \text{if } f \ll f_c \end{cases} \quad (4)$$

177 If the average stress drop of the analyzed earthquakes is assumed to
 178 be constant, the scaling of the source spectrum with the earthquake size is
 179 controlled only by the seismic moment (Aki, 1967). Under this assumption
 180 and considering a mixed effect regression (Bates et al., 2015), Equations 1
 181 and 4 suggest the following parametric model for $FAS(f, R)$:

$$\ln(FAS) = a_1 + a_2 Mw + a_3 \ln(R) + a_4 R + \delta B_e + \delta B_s + \epsilon \quad (5)$$

where the moment magnitude Mw is proportional to $\text{Log}(M_0)$ (Hanks and Kanamori, 1979). In equation 5, the coefficients a_i are the (frequency dependent) fixed effects which define the median prediction; δB_e , δB_s are the

random effects for the earthquake and station grouping levels, respectively; ϵ is the residual distribution. In order to allow more complex scaling with magnitude, the functional form considered in this study is the following:

$$\ln(FAS) = e_1 + b_1(Mw - M_{ref}) + b_2(M - M_{ref})^2 + [c_1 + c_2(M - M_{ref})]\ln\left(\frac{R}{R_{ref}}\right) + c_3(R - R_{ref}) + \delta B_e + \delta B_s + \epsilon \quad (6)$$

182 with $M_{ref} = 3.5$ and $R_{ref} = 1km$. In equation (6), the fixed effect
 183 coefficients describe the scaling with distance (c_1 and c_3 are connected to
 184 the geometrical spreading attenuation and the quality factor, respectively)
 185 and with magnitude (b_1 and b_2 are controlling the scaling with the seismic
 186 moment). Coefficient c_2 introduces a magnitude dependency in the attenua-
 187 tion with distance, while the off-set e_1 depends (at high frequency) on source
 188 characteristics such as the average stress drop, among other quantities. The
 189 between-event δB_e quantifies the systematic deviation of recordings for the
 190 same event with respect the median prediction. At high frequencies, the
 191 deviation of the stress drop of any earthquakes from the average of the pop-
 192 ulation is expected to contribute to the δB_e residuals while differences in
 193 the average radiation pattern among the earthquakes due to uneven station
 194 distribution can contribute to δB_e at low frequency (along with other fac-
 195 tors such as differences in the density and velocity at the source location,
 196 errors in the magnitude values, etc). The between station δB_s random ef-
 197 fects, sometimes referred to as $\delta S2S$, absorb the frequency dependent site
 198 amplification indicated with $Z(f)$ in equation 1.

199 The frequency-dependent coefficients of the model (6) and the standard

200 deviations of δB_e , δB_s , and ϵ are listed in Table S1 of the electronic sup-
 201 plement to this article. The residuals ϵ versus hypocentral distance and
 202 δB_e versus magnitude are exemplified in Figure 3 for two frequencies. Over
 203 the intervals well constrained by data, the average residuals do not show
 204 systematic trends with the predictor variables but the variability of δB_e in-
 205 creases with frequency (see also Table S1). Weak trends at short distances
 206 (at high frequency) and for large magnitude (at low frequencies), which are
 207 not impacting on the analysis performed in this study, could be removed by
 208 introducing distance and magnitude hinges in equation (5).

209 **Between-event temporal variability**

210 The temporal trend of the between-event δB_e at 10 Hz is shown in Figure
 211 4 while zooms over different time windows are presented in Figure S3 of
 212 the electronic supplement to this article. In addition to the large variability
 213 in the aftermath of the mainshock occurrence, the most striking feature in
 214 Figure 4 is the positive trend developing from the end of 2009, a few months
 215 later than the April 6, 2009 L'Aquila mainshock, to late 2013-early 2014.
 216 In the period from early 2014 to August 2016, when the Amatrice sequence
 217 started, δB_e shows a large variability with average value close to zero. When
 218 observed at low frequencies (Figure 4), δB_e shows a weak trend with time.
 219 As discussed in the model development section, δB_e is expected to absorb,
 220 at high frequencies, the effect of the stress drop variability. Figure 5 shows
 221 that the correlation between δB_e and $\Delta\sigma$ is significant at 10 Hz, whereas the
 222 correlation is low at 0.75 Hz. It is worth noting that the mainshocks and the

223 largest aftershocks deviate from the average correlation trend defined by the
 224 aftershock population. We ascribe this behavior to the fact that while $\Delta\sigma$
 225 varies over three order of magnitude for small events, it is almost constant
 226 for earthquakes above magnitude 5. Since GMPE well describe the average
 227 the ground shaking generated by the largest magnitudes, their δB_e are dis-
 228 tributed close to zero (Figure 3). The correlation of δB_e with $\Delta\sigma$ is also
 229 highlighted in Figure S4 of the electronic supplement where large positive
 230 residuals are associated to events with $\Delta\sigma$ higher than 0.6 MPa (i.e., the
 231 population average; Bindi et al., 2018) while earthquakes with lower stress
 232 drop have negative residuals. Figure 5 also shows the dependence of δB_e on
 233 hypocentral depth. The observed trend is reflecting the $\Delta\sigma$ dependences on
 234 depth as shown in Figure 2. The degree of correlation measured in terms
 235 of Pearson coefficient (Figure 6) confirms that the correlation is strongest
 236 around 10 Hz. The decrease of correlation toward low frequencies reflects
 237 the diminishing importance of $\Delta\sigma$ in determining the spectral amplitudes
 238 at frequencies lower than the corner one while the reduction above 10 Hz
 239 suggests that source-related effects other than the stress drop also affect the
 240 ground motion variability at high frequencies. The high frequencies radia-
 241 tions depends on many factors: small scale slip heterogeneity/slip roughness
 242 (Causse et al., 2010), rupture velocity and slip source function (Mai et al.,
 243 2017) and near source attenuation (Purvance and Anderson, 2003). For
 244 example, analyzing a smaller data set, Bindi et al. (2017) found a correla-
 245 tion between δB_e and the slope at high frequency of the acceleration source
 246 spectrum.

247 Stress drop temporal variability

248 The stress-drop variability with time (Figure 7a) resembles the variability
249 observed for δB_e . If the earthquakes are grouped according to the latitude
250 of their epicenters as shown in Figure 1, and focusing on the average trend,
251 we observe that:

- 252 • $\Delta\sigma$ of earthquakes located in the L'Aquila region (Figure 7b) rapidly
253 diminishes during the first month after the mainshock on April 6; the
254 recovery starts after about two months (see also Figure 8a);
- 255 • for events located in the Campotosto segment (Chiaraluce et al., 2011)
256 (Figure 7c), the recovery of the logarithm of $\Delta\sigma$ develops over a time
257 span of 4 years, from 2010 to 2013, at an annual rate of 0.17 (i.e., the
258 amplification factor for $\Delta\sigma$ is 1.5 per year). We recall that the Cam-
259 potosto segment includes the northernmost termination of the 2009
260 sequence and the southern tip of the 2016 fault system (Chiaraluce et
261 al., 2017). In particular, the four events with magnitude larger than
262 5 occurred in January 2017 are characterized by large $\Delta\sigma$ (see Figure
263 8), as for the largest aftershocks occurred over this segment during the
264 2009 sequence.
- 265 • a decrease in $\Delta\sigma$ is observed at the beginning of 2014, although no
266 large earthquakes occurred at that time. Figure 9 shows the results of
267 a breakpoint analysis (Bai, 1994; Zeileis et al., 2002; 2003) performed
268 to detect changes in the coefficients of the linear regression with time.
269 The analysis identifies a change-point within the period February 10 -

270 March 8, 2014 across which the slope of the logarithm of $\Delta\sigma$ with time
 271 reduce from $4.7e^{-04}$ to $9.4e^{-05}$ (the amplification factor per year for
 272 $\Delta\sigma$ reduces from 1.5 to 1). A detailed description of the breakpoint
 273 analysis is reported in the electronic supplement to this article. The
 274 cause driving this drop are not known. No large earthquakes occurred
 275 in the area around February-March 2014; the only notable event is the
 276 Gubbio swarm (Gualandi et al., 2017). At this stage, it is difficult to
 277 assess the plausibility of its involvement in the process we are examin-
 278 ing here. Possible connections with the seismic and aseismic moment
 279 released during the 2013-14 Gubbio swarm are worth to be the subject
 280 of future work.

- 281 • earthquakes located in the northern group (Figure 7d) mainly be-
 282 long to the 2016-17 Amatrice-Norcia-Visso sequence (Chiaraluce et
 283 al., 2017); also for these events, $\Delta\sigma$ is larger for the mainshocks and
 284 for the aftershocks above magnitude 5.5 (Figure S5 in the electronic
 285 supplement), and decreases after the mainshock occurrence (Figure
 286 8c). The events occurred in this area before the 2016-17 sequence
 287 follow the same trends observed for the Campotosto segment.

288 **Discussions and conclusions**

289 The between-event residuals δB_e computed for ten years of data in central
 290 Italy show significant temporal variability at high frequency (Figure 4). On
 291 one hand, the time dependency of δB_e imply temporal changes of the ground
 292 shaking that could have an impact over the short term hazard. In the first

couple of months, δB_e at 10 Hz varies in the range from -1 to 1, roughly
 (i.e., about a factor 0.7 for spectral amplitudes); after a couple of months
 from L' Aquila mainshock, a trend develops with δB_e increasing on average
 from about 0 to 0.8, (i.e., about factor 2 in high frequency spectral content).
 On the other hand, Figure 7 shows that the high-frequency between-events
 variability resembles the time variability of the stress drop $\Delta\sigma$. Tempo-
 ral variability of $\Delta\sigma$ has been observed in previous studies. For example,
 Abercrombie (2014) analyzed 25 earthquakes in three repeating sequences
 on the San Andreas at Parkfield, observing a long term gradual increase of
 $\Delta\sigma$ before the 2004 magnitude 6 earthquake. The values show an immedi-
 ate decrease after the mainshock occurrence before recovering to previous
 values. Using a long-term stress drop catalog, Chen and Shearer (2013)
 found relatively stable long-term average stress drop in Southern California
 but a slow increase trend after large main shocks within the Landers fault
 zone was also identified, in agreement with a possible long-term fault zone
 recovery (Li et al., 1998).
 Fault healing has been shown to promote the generation of high-frequency
 earthquakes both in laboratory experiments and on natural faults (e.g.,
 Marone, 1998; McLaskey et al., 2012, Scuderi et al., 2016). The connection
 between pore pressure and effective normal stress has been also advocated
 to explain the time variability of the stress drop. Recently, Yoshida et al.
 (2017) analyzed a swarm triggered by the 2011 Tohoku earthquake, eval-
 uating temporal changes in stress drop and b-value. They discussed the
 temporal variations of stress drop (very similar patterns to those observed
 in this study) in terms of changes in the frictional strength due to fluid

318 migration. In central Italy, pore pressure diffusion due to fluids migration
 319 played a role in the preparatory phase of the L'Aquila mainshock (e.g., Di
 320 Luccio et al 2010). However, pore pressure diffusion generally occurs over
 321 time scales of weeks to months. Thus, we reckon it is difficult to attribute
 322 solely to migration of fluids or pore pressure the variations we observe, which
 323 occur over time scales of several years.

324 Among other techniques, monitoring changes in seismic velocities has been
 325 shown to be effective in detecting fault healing and reloading processes (e.g.,
 326 Brenguier et al., 2008; Chen et al., 2010). For example, Peng and Ben-Zion
 327 (2006) investigated the temporal variations of seismic velocity along the
 328 north Anatolian fault analyzing repeating earthquake clusters in the after-
 329 shock zones of the 1999 Izmit and Düzce earthquakes. The authors observed
 330 a sharp seismic velocity reduction immediately after the Düzce main shock,
 331 followed by a gradual logarithmic-type recovery. They concluded that the
 332 temporal changes of material properties occur in the topmost portion of
 333 the crust and, although the change is more prominent at stations located
 334 close to recently ruptured fault zones, it is not limited to the immediate
 335 vicinity of the fault zone. In central Italy, Soldati et al (2015) computed
 336 the relative velocity variation from cross-correlations of noise data over the
 337 period 2008-2012, including the 2009 L'Aquila main shock. The temporal
 338 variation obtained for the relative velocity (reproduced in Figure 10) has a
 339 trend very similar to the stress drop: an abrupt co-seismic decrease at the
 340 time of the main shock occurrence, followed by an unstable behavior for a
 341 few months and, finally, a recovery of the velocity (see also in Figure 7c).
 342 Regarding the spatial distribution of the co-seismic velocity drop, Soldati

et al. (2015) compared the velocities changes averaged over a one-month time window selected before and after the main shock occurrence, and excluding the day of the main shock. They found (see their Figure 5) that the drop was maximum over the area surrounding the L'Aquila epicentre and in the northeast direction from the fault zone, including the Campotosto area. The similarities of the trends observed for the stress drop and for the relative velocity variations suggest that, in agreement with Heckels et al (2018), the recovery can be associated both to crack-healing along the main fault system and to healing of micro-cracks distributed at shallow depths throughout the surrounding region.

Data and Resources

The R software (R Development Core Team, 2008; <http://www.R-project.org>) has been used in this study to perform the regressions. In particular, the packages lme4 (Bates et al., 2015; <https://cran.r-project.org/web/packages/lme4/news.html>); ggplot (Wickham, 2009; <http://ggplot2.org>); changepoint (Killick, R., & Eckley, I.A., 2014; <https://www.jstatsoft.org/article/view/v058i03>); strucchange (Zeileis et al., 2002; <http://www.jstatsoft.org/v07/i02/>). The waveforms used in this study have been downloaded from European Integrated Data Archive-EIDA (<https://www.orfeus-eu.org/data/eida/>) and from the Italian Civil Protection (DPC) repository (<http://ran.protezionecivile.it/IT/index.php>). Regarding the permanent networks, we used data from the networks with FDSN code: MN, IV, IT (<http://www.fdsn.org/networks/>). The moment magnitude used in this study for all earthquakes larger than 5.7

366 have been taken from the Geofon moment tensor catalog ([http://geofon.gfz-](http://geofon.gfz-potsdam.de/eqinfo/list.php?mode=mt)
367 [potsdam.de/eqinfo/](http://geofon.gfz-potsdam.de/eqinfo/list.php?mode=mt) [list.php?mode=mt](http://geofon.gfz-potsdam.de/eqinfo/list.php?mode=mt)). Only for L'Aquila mainshock, we
368 used the GCMT solution ([http://www.globalcmt.org/](http://www.globalcmt.org/CMTsearch.html) [CMTsearch.html](http://www.globalcmt.org/CMTsearch.html)).
369 The earthquake locations are taken from the INGV bulletin ([http://cnt.rm.ingv.it/](http://cnt.rm.ingv.it/inside)
370 [inside](http://cnt.rm.ingv.it/inside)). All the Internet sites have last accessed on December 2017. Some of
371 the Figures were prepared with GMT (Wessel and Smith, 1991).

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Figure captions

Figure 1. Map of earthquake epicenters (circles) analyzed in this study (see Data and Resources section). Circles are filled according to the latitude of the epicenter, assuming arbitrary thresholds at latitudes 42.4 and 42.68. A few earthquakes belonging to the 2014 Gubbio swarm are also included. The focal mechanisms of earthquakes with magnitude larger than 6 are shown as beach ball taken from Geofon and from the Global Centroid Moment Tensor catalogs (see Data and Resources section). The rectangles depict the surface projection of the faults as given in Luzi et al. (2016).

Figure 2. Scaling relationships between stress drop $\Delta\sigma$, hypocentral depth and moment magnitude Mw for the earthquakes analyzed in this study (Bindi et al., 2018). The trend-lines are estimated through a local regression (Loess) performed using the ggplot2 package in R (Wickham, 2009).

Figure 3. Observation minus prediction residuals versus predictor variables for the model in equation (1). The residual ϵ versus distance and the between event δB_e versus moment magnitude Mw are shown for the regressions performed at 0.75 Hz (left) and 10 Hz (right).

Figure 4. (a) Between-event residuals (δB_e) versus time, at 10 Hz. Earthquakes belonging to the Gubbio swarm (triangles; see Figure 1) are not considered for the evaluating the local-trend analysis; zooms over different time windows are presented in Figure S3 of the electronic supplements to this article. (b) Between-event versus time, at 0.75 Hz. (c) the same as in

580 (b) but zooming over the 2016 sequence. Vertical bars represent the 95%
581 confidence interval for δB_e .

582 **Figure 5.** Between-event δB_e residuals versus stress drop $\Delta\sigma$ (top) and
583 hypocentral depth (bottom), considering the results for 0.75 Hz (left) and
584 10 Hz (right).

585 **Figure 6.** Person correlation coefficients between $\Delta\sigma$ and δB_e at different
586 frequencies.

587 **Figure 7.** Temporal variability of stress drop $\Delta\sigma$. (a) complete dis-
588 tribution of earthquakes; (b) only earthquakes located in proximity of the
589 2009 L' Aquila mainshock ; (c) only earthquakes occurred in the Campotosto
590 segment; (d) earthquakes occurred in the area corresponding to the 2016-17
591 mainshocks. For the location of the earthquakes in panels (b) through (d),
592 see Figure 1. Vertical bars represent the 95% confidence interval for $\Delta\sigma$.
593 Zooms over different windows are available in Figure 8.

594 **Figure 8.** Temporal variability of stress drop $\Delta\sigma$, different zooms of
595 Figure 7.

596 **Figure 9.** Results of the breakpoint analysis (Zeileis et al., 2002; 2003).
597 A change in the linear trend with of $\Delta\sigma$ (dashed lines) is detected between
598 February 10 and March 8, 2014 (vertical gray line). Details of the analysis
599 are reported in the electronic supplements to this article.

600 **Figure 10.** Comparison between the relative shear wave velocity vari-
601 ation computed by Soldati et al (2015) (points) and the $\Delta\sigma$ (squares) time
602 variability.

603 **Figures**

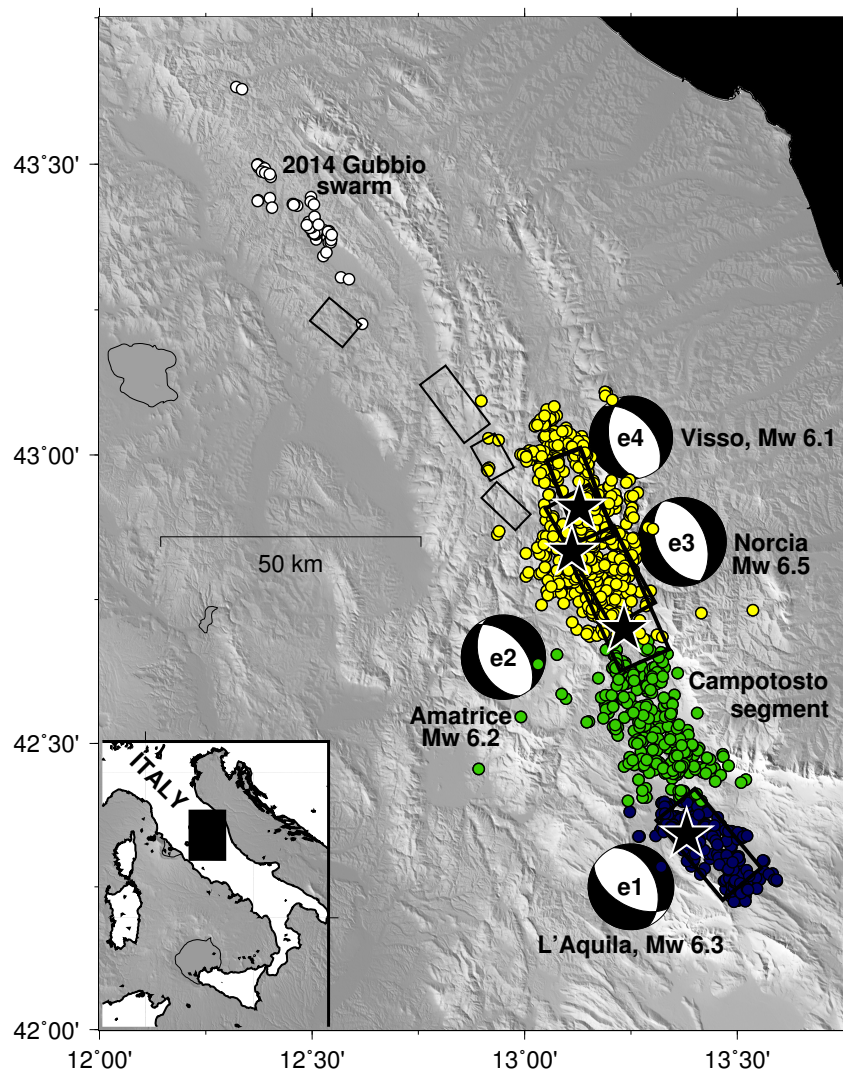


Figure 1:

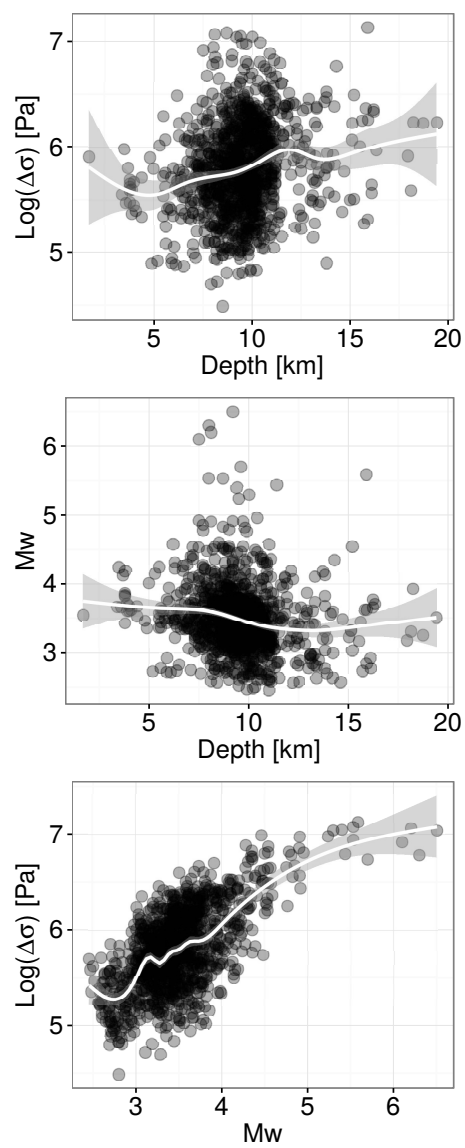


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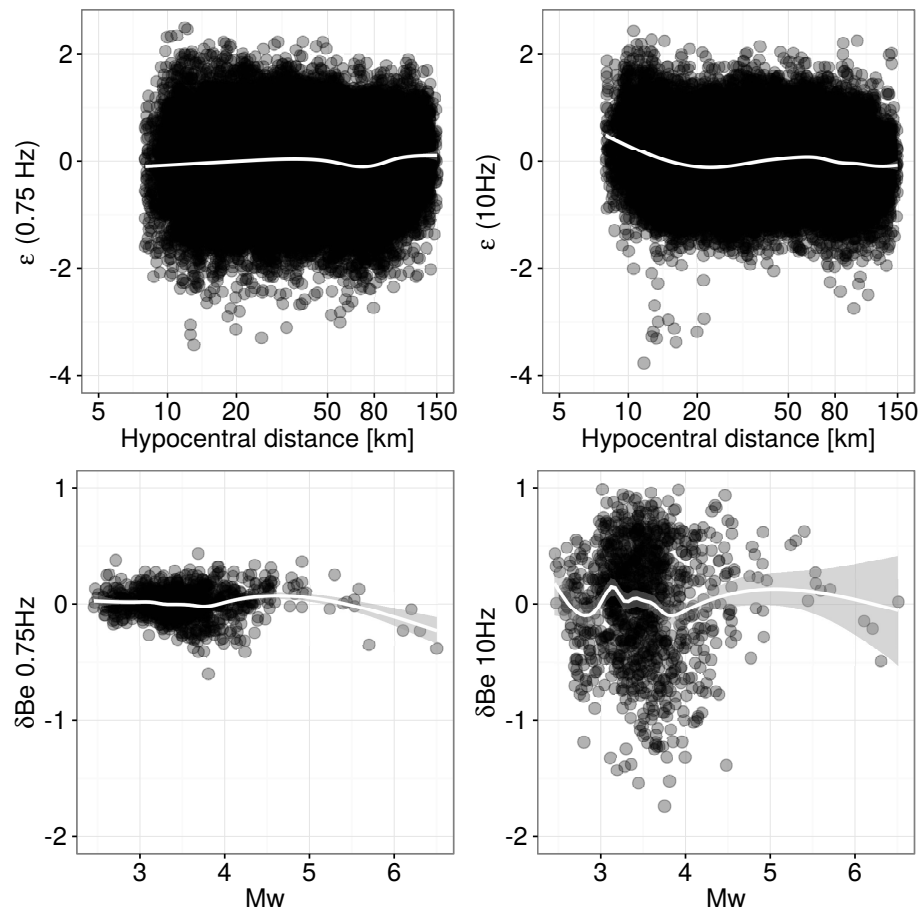


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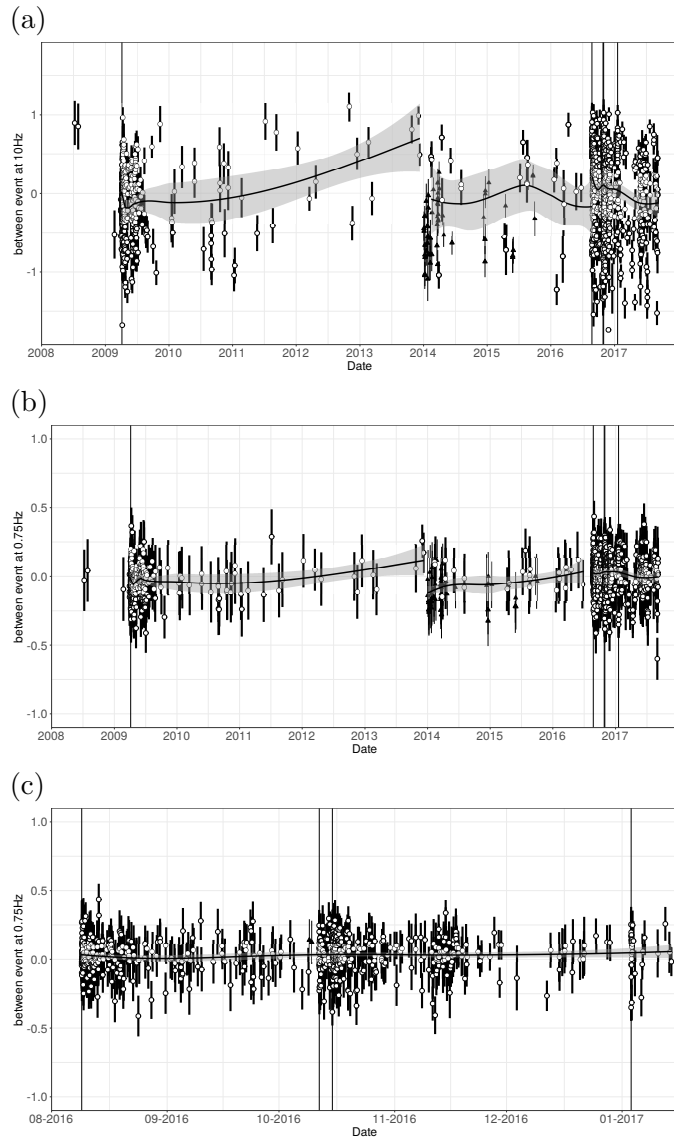


Figure 4:

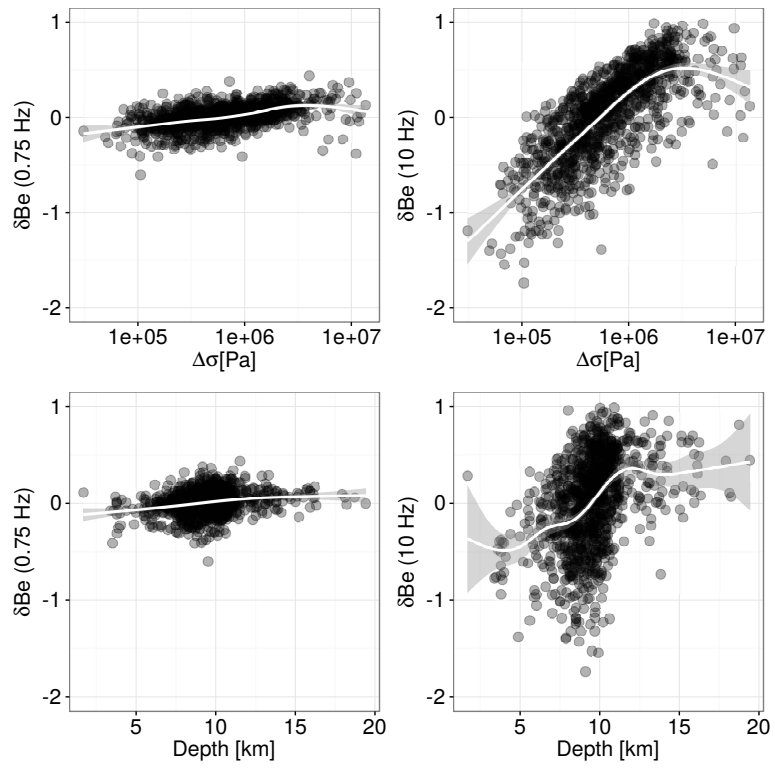


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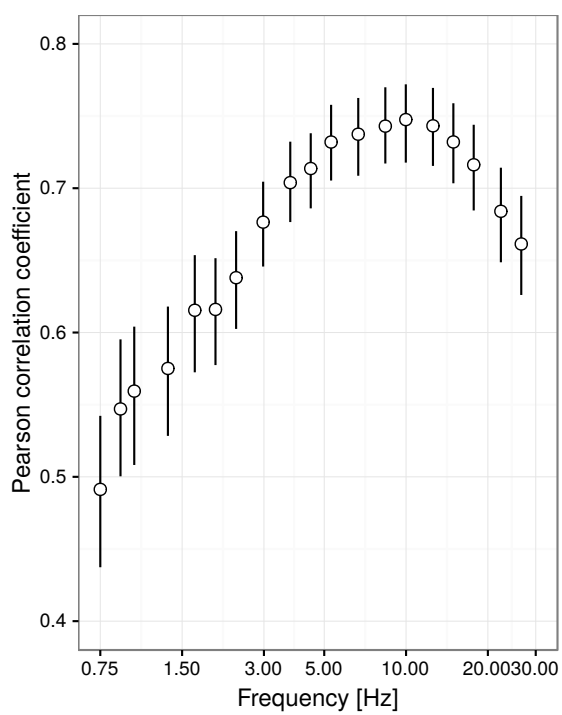


Figure 6:

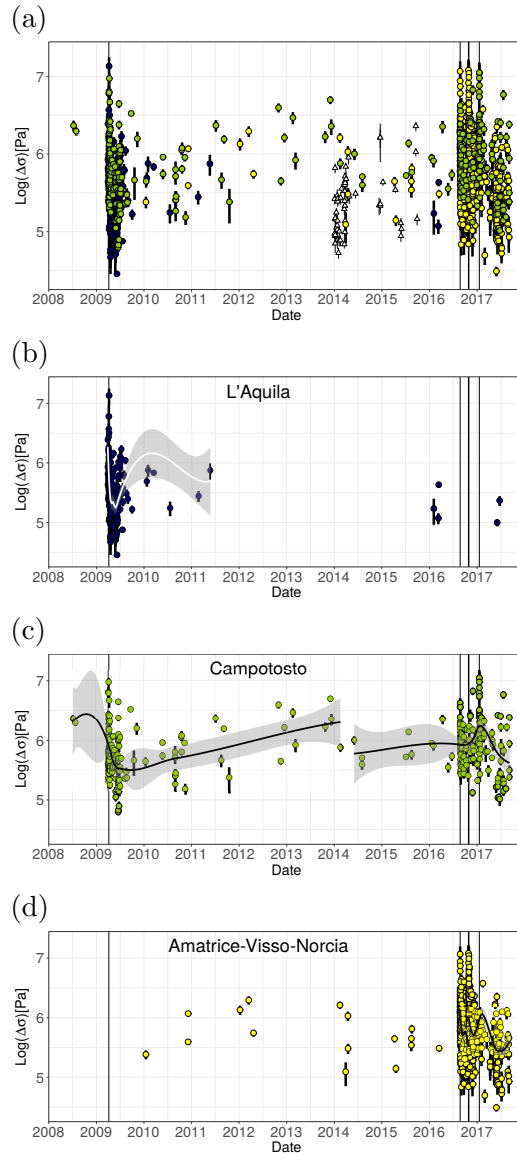


Figure 7:

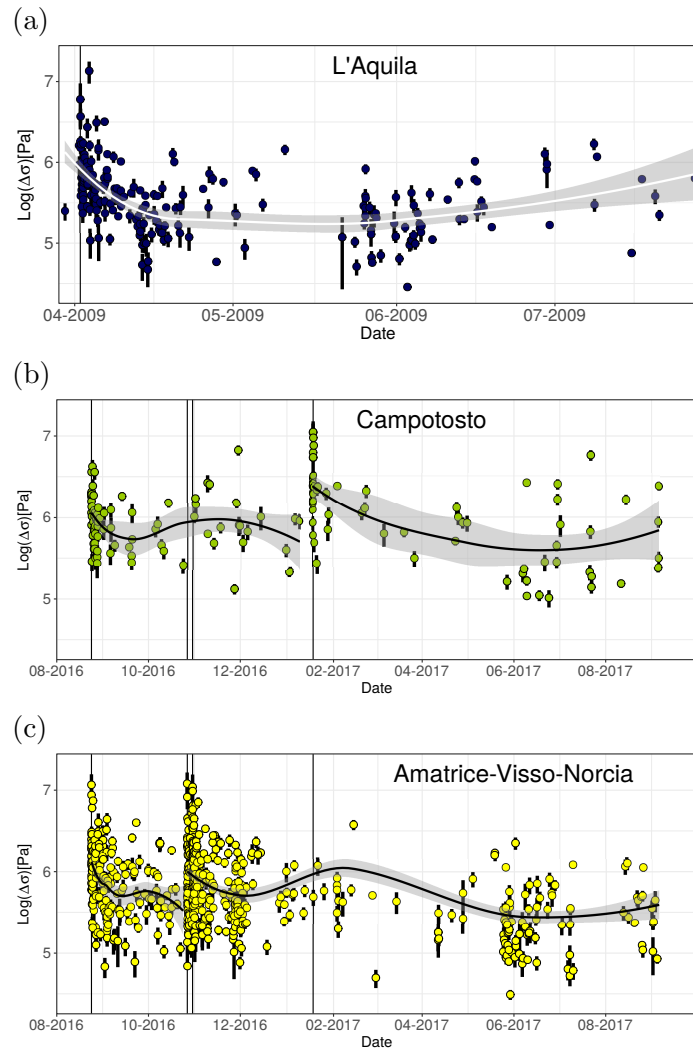


Figure 8:

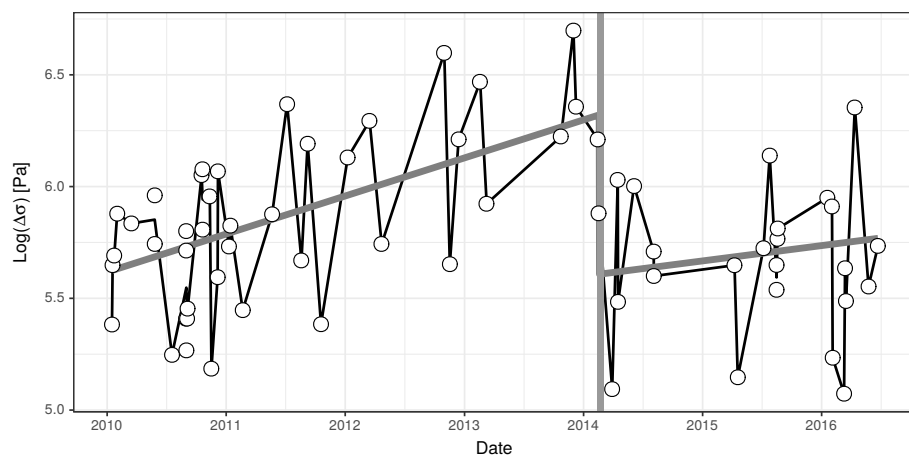


Figure 9:

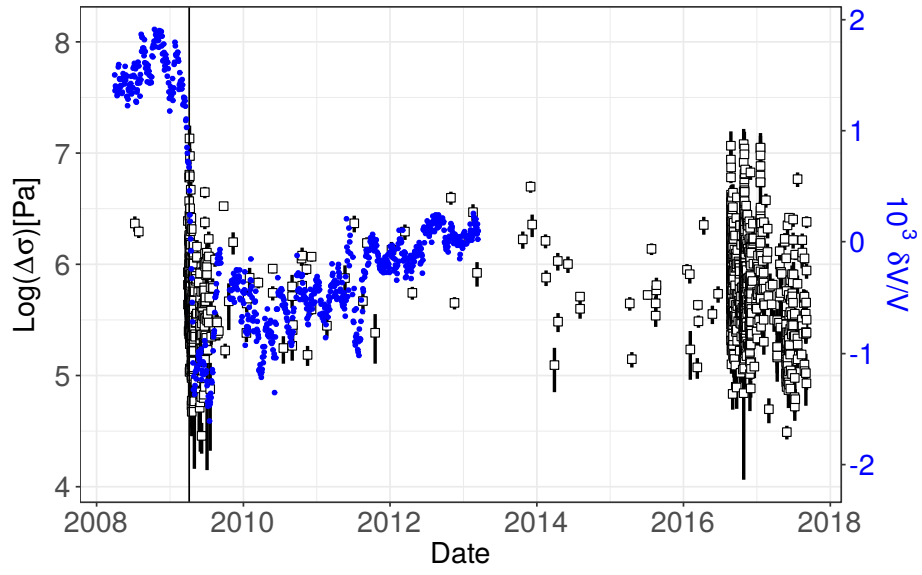


Figure 10:

Temporal variability of ground-shaking and stress drop in central Italy: a hint for fault healing?

D. Bindi, F. Cotton, D. Spallarossa, M. Picozzi, and E. Rivalta

This Supplementary Material includes additional Figures cited in the main article. In figure S1, a map with the locations of stations used in this study is provided. The map in Figure S2 shows the earthquake locations as in Figure 1 but with symbols filled according to the time elapsed since the L'Aquila earthquake. Figure S3 complements Figure 4 by showing the variability with time of the between event residuals δB_e at 10Hz considering different time windows. In Figure S4, the symbols used to depict the time dependency of δB_e at 10Hz are filled according to the logarithm of the stress drop $\Delta\sigma$, measured in [Pa]. Finally, in Figure S5, the symbols used to show the time variability of $\Delta\sigma$ are filled according to the moment magnitude M_w .

The Supplementary Material includes also Table S1, listing the frequency-dependent coefficients of the Ground Motion Prediction Equation developed in this study (equation 6) for the Fourier amplitude spectra. Finally, the Supplementary Material includes additional text describing both the error propagation applied for $\Delta\sigma$ and the break-point analysis performed for locating the $\Delta\sigma$ change-point occurred at the beginning of 2014.

Error propagation

The non-parametric source spectra $S(f)$ obtained through the non-parametric Generalized Inversion (Bindi et al., 2017) are described in terms of acceleration source model (Brune, 1970):

$$S_{Brune}(f) = const M_0 \frac{f^2}{1 + \left(\frac{f}{f_c}\right)^2} \quad (S1)$$

where M_0 is the seismic moment, f_c the corner frequency and $const$ is a constant depending on the radiation pattern, free surface amplification, density and velocity at the source location. We fit the $S_{Brune}(f)$ model to $S(f)$ by computing the logarithm (hereinafter we refer to the logarithm in base 10 as \log and to the natural logarithm as \ln) of the spectra and considering as independent variables $\log(M_0)$ and f_c . The fit is performed following a non-linear least-squares approach and the variance-covariance matrix νCov is evaluated. From the square root of the diagonal of νCov , we extracted the standard errors on f_c and on $\log(M_0)$, referred to as δ_{f_c} and $\delta_{\log M_0}$, and from the off-diagonal elements the correlation between f_c and $\log(M_0)$, indicated as $\delta_{(f_c, \log M_0)}$. The error propagation is evaluated through the variance formula for correlated variables, truncated at the first order of the Taylor's expansion:

$$\delta_{\Delta\sigma}^2 = \left(\frac{\partial \Delta\sigma}{\partial f_c}\right)^2 \delta_{f_c}^2 + \left(\frac{\partial \Delta\sigma}{\partial \log(M_0)}\right)^2 \delta_{\log M_0}^2 + 2\delta_{(f_c, \log M_0)} \frac{\partial \Delta\sigma}{\partial f_c} \frac{\partial \Delta\sigma}{\partial \log(M_0)} \quad (S2)$$

The stress drop $\Delta\sigma$ is computed from M_0 and f_c using the following relationships (Eshelby, 1957; Brune 1970):

$$\Delta\sigma = \frac{7}{16} \frac{M_0}{r^3} \quad (S3)$$

$$r = 2.34 \frac{v_s}{2\pi f_c} \quad (S4)$$

where v_s is the shear-wave velocity at the source. Considering the logarithm of $\Delta\sigma$, equations (S3) and (S4) give

$$\log(\Delta\sigma) = constant + \log(M_0) + 3\log(f_c) \quad (S5)$$

and equation (2) becomes

$$\delta_{\log(\Delta\sigma)}^2 = \left(\frac{\partial \log(\Delta\sigma)}{\partial f_c}\right)^2 \delta_{f_c}^2 + \left(\frac{\partial \log(\Delta\sigma)}{\partial \log(M_0)}\right)^2 \delta_{\log M_0}^2 + 2\delta_{f_c, \log M_0} \frac{\partial \log(\Delta\sigma)}{\partial f_c} \frac{\partial \log(\Delta\sigma)}{\partial \log(M_0)} \quad (S6)$$

From equation (5),

$$\frac{\partial \log(\Delta\sigma)}{\partial \log(M_0)} = 1 \frac{\partial \log(\Delta\sigma)}{\partial f_c} = \frac{3}{\ln(10)} \frac{1}{f_c} \quad (S7)$$

Using (7) in (6), we obtain

$$\delta_{\log(\Delta\sigma)}^2 = \frac{9}{[\ln(10)]^2} \left(\frac{\delta_{f_c}}{f_c}\right)^2 + \delta_{\log M_0}^2 + \delta_{(f_c, \log M_0)} \frac{6}{\ln(10)} \frac{1}{f_c} \quad (S8)$$

The error propagation in equation (S8) is similar to the one described by Cotton et al (2013), in their equation (9). The differences are related to a different choice of the variables, since they used $\log(f_c)$ while we used f_c , and to the fact that Cotton et al (2013) neglected the cross-term of the covariance matrix. The advantage of using $\log(f_c)$ is that equation (S5) becomes linear with respect to the variables and, therefore, equation (S2) is exact. The drawback is that in fitting the Brune model, the dependence on the variable takes the more complex form of $10^{\log(f_c)}$. Regarding the cross-term, since $\delta_{(f_c, \log M_0)}$ is negative, the choice of Cotton et al (2013) was conservative.

Finally, since $\delta_{\Delta\sigma} = \ln(10) \Delta\sigma \delta_{\log(\Delta\sigma)}$, from equation (S8) we get the standard error on $\Delta\sigma$

$$\delta_{\Delta\sigma} = \Delta\sigma \sqrt{9 \left(\frac{\delta_{f_c}}{f_c}\right)^2 + [\ln(10)]^2 \delta_{\log M_0}^2 + \frac{6\ln(10)}{f_c} \delta_{(f_c, \log M_0)}} \quad (S9)$$

Figure S6 shows the 95% confidence interval, computed as 1.96 times the standard error, for the data analyzed in this study. Zooms over different time windows are shown in Figure S7.

Breakpoint analysis

In order to assess the statistical significance of the abrupt drop in $\Delta\sigma$ occurring at the beginning of 2014, we performed a change-point analysis (Page, 1954; Jaiswal et al., 2015). The analysis was performed through the following steps, outlined in Figure S8:

- we considered $\Delta\sigma$ values between January 2010 and June 2016 (Figure S8a);
- since the time series is unevenly sampled, we applied an interpolation scheme. After some tests using different approaches, we used a simple linear interpolation between consecutive samples. The time series interpolated using a regular sampling of 1 day is shown in Figure S8b;
- A preliminary detection in terms of significant change in the average was performed following the approach of Killick and Eckley (2014). The detected change point (Figure S8c) is located on February 15, 2014;
- finally, the breakpoint analysis of Zeileis et al (2003) was applied to identify and locate a change in the coefficients of the linear regression with time. Figure S8d shows the location of the change point with its 95% confidence interval (corresponding to the period from February, 10 to March, 8) and the best linear models before and after the change point.

Table caption

Table S1. Coefficients ($e_1, b_1, b_2, c_1, c_2, c_3$) of the Fourier ground motion prediction equation described in equation (6); ϕ_{s2s} , τ , ϕ_0 are the standard deviation of the between station δBs , of the between event δBe and residual ε distributions (see equation 6).

Figure captions

Figure S1 Location of stations considered in this study.

Figure S2 Location of earthquakes considered in this study with symbols filled according to the earthquake origin time with respect to April 6, 2009 (see also Figure 1).

Figure S3 Between-event residuals δBe versus time at 10 Hz for different time intervals (see also Figure 4).

Figure S4 Between-event residuals δBe versus time with symbols filled according to stress drop $\Delta\sigma$. Triangles indicate earthquakes occurred in the area of the 2013-14 Gubbio swarm, see Figure 1.

Figure S5 Stress drop $\Delta\sigma$ versus time with symbols filled according to moment magnitude M_w . Triangles indicate earthquakes occurred in the area of the 2013-14 Gubbio swarm, see Figure 1.

Figure S6. 95% confidence intervals for $\Delta\sigma$.

Figure S7. 95% confidence intervals for $\Delta\sigma$, for different time intervals.

Figure S8. Change point analysis. a) original data. b) data linearly re-sampled at the rate of one sample per day; c) detection of the change point for the average; d) detection of the change point in the linear regression coefficients

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Table S1. Coefficients ($e_1, b_1, b_2, c_1, c_2, c_3$) of the Fourier ground motion prediction equation described in equation (6); ϕ_{s2s} , τ , ϕ_0 are the standard deviation of the between station δB_s , of the between event δB_e and residual ε distributions (see equation 6).

Freq [Hz]	e_1	b_1	b_2	c_1	c_2	c_3	ϕ_{s2s}	τ	ϕ_0
0.35	0.80146	2.21669	-0.04089	-2.00751	0.17379	0.00545	0.47522	0.21587	0.70041
0.4	1.04779	2.27639	-0.04779	-2.07823	0.17007	0.00764	0.47742	0.20514	0.6891
0.5	1.33959	2.42340	-0.06803	-2.12578	0.15654	0.00992	0.48835	0.18558	0.64771
0.59	1.56116	2.52567	-0.1043	-2.15865	0.14717	0.01189	0.51034	0.17384	0.6542
0.75	1.92521	2.68625	-0.15401	-2.18397	0.12672	0.01265	0.54612	0.14888	0.60759
0.89	2.33067	2.71259	-0.18884	-2.23543	0.1305	0.01312	0.57532	0.14129	0.58482
1	2.60114	2.72391	-0.21177	-2.27032	0.12893	0.01357	0.59266	0.14381	0.59366
1.33	3.09564	2.76270	-0.23556	-2.26438	0.10171	0.01225	0.61237	0.16665	0.56748
1.67	3.48468	2.68448	-0.24237	-2.27198	0.0962	0.01163	0.61487	0.19459	0.53169
1.99	3.67784	2.62330	-0.24584	-2.26209	0.0945	0.01103	0.6234	0.22036	0.53082
2.37	3.76324	2.55504	-0.23485	-2.20498	0.08715	0.00903	0.63017	0.25424	0.52461
2.98	3.59658	2.38582	-0.21798	-2.04151	0.0927	0.00468	0.60396	0.29829	0.52005
3.75	3.42988	2.25625	-0.19759	-1.90391	0.08802	0.00061	0.62728	0.34572	0.52529
4.46	3.28532	2.08639	-0.17907	-1.79832	0.10474	-0.00292	0.65699	0.37856	0.5284
5.3	3.07226	2.00030	-0.15746	-1.69606	0.09872	-0.00611	0.69546	0.40803	0.52817
6.67	2.87406	1.86515	-0.13773	-1.60836	0.10035	-0.00995	0.76182	0.44405	0.52234
8.39	2.57230	1.78698	-0.11523	-1.51731	0.08999	-0.01435	0.81647	0.48376	0.52027
10	2.36854	1.71378	-0.09805	-1.47415	0.08535	-0.01735	0.87734	0.51209	0.51604
12.56	2.31993	1.61380	-0.08088	-1.50875	0.08424	-0.02057	0.96463	0.56033	0.51112
14.93	2.34572	1.50170	-0.06593	-1.57316	0.09194	-0.02223	102.517	0.59434	0.51074
17.74	2.35082	1.44668	-0.03929	-1.67057	0.08297	-0.0224	108.441	0.63308	0.5131
22.33	2.52609	1.39944	-0.01618	-1.91863	0.06611	-0.01832	112.420	0.68558	0.52859
26.54	2.67305	1.39498	0.01618	-2.11285	0.04734	-0.01503	109.470	0.70831	0.52971

