Separation of source and site effects in ground motions recorded in the village of Onna during aftershocks of the 2009 April 6, $M_w$ 6.1 L’Aquila earthquake

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SUMMARY

The village of Onna, a 10 km apart from L’Aquila, central Italy, was dramatically struck by the 2009 April 6, $M_w$ 6.1 earthquake with 80 per cent of buildings collapsed or severely damaged. The high vulnerability of predominantly ancient buildings and the propensity of site geology to amplify ground motion on the Holocene sediments of the Aterno river valley were unanimously thought as responsible for the huge destruction. To quantify site effects in the damaged zone of Onna and study source scaling over a wide magnitude range (2.0 $\leq M_L \leq$ 5.4), we have used recordings of 20 stations installed in Onna and other villages around L’Aquila. We analyse more than 1000 seismograms of 202 aftershocks occurring up to source-to-receiver distances of 50 km and infer site and source parameters by means of an inversion procedure. The source spectra inferred from the data inversion conﬁrm the large variability in the high-frequency radiation already found by other authors for L’Aquila earthquakes, with Brune stress drops around 10 MPa at the highest magnitudes of the investigated range and spreading mostly between 0.1 and 1 MPa at smaller magnitudes. Moreover, the inversion of our data yields larger amplitudes of empirical transfer functions in the village of Onna conﬁrming the role of the local geology on damage. The site functions of Onna show a common resonance mode around 2.7 Hz, with amplitudes attaining a factor of 4–5. Moreover, we find that the transfer function amplitude does not decrease below 2 in a large high-frequency band above the site resonant frequency, up to more than 10 Hz. This indicates a further broad-band contribution to the ground motion amplification in Onna. In a simulation of the main shock scenario applying the estimated source scaling to aftershock records, Onna results in the highest accelerations among the villages around L’Aquila. In distinct contrast, transfer functions close to unity in the entire frequency band are found for stations installed on harder rock formations (Mesozoic limestone and Pleistocene siltstone). Their scenario accelerations result in the smallest values, consistently with the lowest macroseismic intensities.

Key words: Earthquake ground motions; Site effects.

1 INTRODUCTION

On 2009 April 6, an $M_w$ 6.1 earthquake severely damaged the town of L’Aquila, central Italy. In the town, the macroseismic intensity attained VIII–IX of the Mercalli–Cancani–Sieberg (MCS, see Sieberg 1930) scale and even larger damage (MCS intensity of IX–X) was suffered by the village of Onna, in the Aterno river valley, 10 km from L’Aquila (Tertulliani et al. 2009). Interestingly the village of Monticchio, located less than a couple of kilometres from Onna, experienced MCS intensity of VI–VII. The high damage level of Onna was investigated by several authors (Bergamaschi et al. 2011; Di Giulio et al. 2011). Although the role of the building vulnerability might be not marginal for the highly destructive effects in Onna, the difference in terms of near-surface geology between Onna and Monticchio has to be considered primarily responsible for the different damage of the two villages. The former is settled on fluvial-lacustrine sediments (Holocene–Upper Pleistocene) that are superimposed on a lower Pleistocene conglomerates substrate of the Aterno river valley (Compagnoni et al. 2011), whereas the latter is settled on a gentle mountain slope of Mesozoic limestone and Pleistocene silts.

Unfortunately, the main shock ground motion was not recorded in Onna and Monticchio, therefore only weak motions can be used to retrieve information about the site ampliﬁcation propensity of Onna, limitedly to a linear response approximation. In order to assess the difference in site response and characterize ground motions in Onna...
and nearby villages we have analysed as many as possible seismograms of aftershocks. Aftershocks come from the complex system of normal and normal-oblique faults, mainly NW–SE and NNW–SSE striking that became active after the main shock (Galadini & Galli 2000; Chiaraluce et al. 2011; Valoroso et al. 2013). The 2009 L’Aquila seismic sequence counted thousands of events and was recorded by pre-existing networks and by a large number of temporary stations deployed by different institutions in the epicentral area after the main shock (Cultrera et al. 2009; Bergamaschi et al. 2011). The availability of a large amount of weak- and strong-motion data recorded for the main shock and the entire sequence of foreshocks and aftershocks, with independent well-determined source mechanisms (Scognamiglio et al. 2010; Herrmann et al. 2011; D’Amico et al. 2013), provided a unique opportunity to investigate source scaling, crustal attenuation characteristics and site effects. Many of the previous literature results will be mentioned in the presentation of the new findings of this paper, where we focus primarily on (i) ground motion variations caused by site conditions over very small distances in and around Onna, and (ii) ground motion variations between events caused by different stress release down to magnitudes as small as $M_L \approx 2$. Acceleration spectra from 202 aftershocks are inverted to obtain individual-event source spectra and individual-station empirical transfer functions simultaneously. Data are composed of seismograms of small to moderate magnitude earthquakes recorded at temporary networks deployed by researchers of Istituto Nazionale di Geofisica e Vulcanologia in Onna and other nearby villages in the middle Aterno river valley. This data set includes seismograms of stations installed in the most intensely damaged zone of Onna between July and October 2009 that were not used in previous studies. Since the inter-station distance is of the order of few kilometres, our approach mimics the Generalized Inversion Technique (GIT) but the crustal attenuation term is taken from a recent similar study in the same region that honoured the better condition of tens of kilometres distant stations and events up to source distances of 120 km (Pacor et al. 2016). The empirical site transfer functions resulting from our inversion are characterized by large amplitude variations reflecting the large differences that ground motions show within distances as small as few kilometres during the same earthquakes, with the largest and the smallest amplitudes persisting in Onna and Monticchio, respectively. In an acceleration scenario of the main shock constructed through seismograms of aftershocks, the difference in ground motion between the two villages, characterized by a similar vulnerability, is consistent with their remarkably different macroseismic intensities. Moreover, assessed source spectra result in a magnitude-dependent stress drop with a larger increase of variability at low magnitude, confirming a trend already proposed for seismic sequences in the Apennines (Rovelli & Calderoni 2014) possibly due to the higher sensitivity of smaller size ruptures to fault strength heterogeneities.

2 DATA

Collected data consist of seismograms that were recorded in the middle Aterno river valley in two different time intervals and by different stations (Fig. 1). We have organized the data into two data sets. Data set-1 consists of 1866 waveforms from 86 earthquakes recorded at 12 stations all around the town of L’Aquila in the time interval from 2009 April 7 until May 11 (Table 1). For safety precautions access to the most damaged sector of Onna was not permitted at that time, and one station (MI03) was installed just outside the forbidden zone. Local magnitudes $M_L$ of Data set-1 span from 2.8 to 5.4 with the largest portion of data between 3.0 and 3.8 (Fig. 2, upper panels). The hypocentral distances range from 9 to 38 km with a best ray coverage in the uppermost crust (<20 km). Data set-2 consists of 2544 waveforms from 225 earthquakes recorded at eight stations in the time interval from 2009 July 23 until October 25.

Table 1. Stations of Data set-1 used in the inversion and number of records for each station.

<table>
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<th>STATION</th>
<th>LOCALITY</th>
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</table>
Separation of source and site effects in Onna

Figure 2. Number of events with given local magnitude $M_L$ (left-hand side) and $M_L$ versus source distance $R$ (right-hand side) for Data set-1 and Data set-2 (in the upper and lower panels, respectively).

Table 2. Stations of Data set-2 used in the inversion and number of records for each station.

<table>
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<td>Monticchio</td>
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(Table 2). Five stations were deployed in Onna, two in Monticchio and one few kilometres from Monticchio. One of the stations in Onna (ONN1) reoccupied the site of MI03, the other four stations were installed in the most intensely damaged zone where it was authorized us access from July 2009. Aftershocks of Data set-2 are in a narrower magnitude range ($2.0 \leq M_L \leq 3.9$) compared to Data set-1, and their source-to-receiver distances range from 7 to 46 km (Fig. 2, bottom panels). Stations of Data set-1 were equipped with either Lennartz LE-3D/5 s or LE-3D/1 s sensors, whereas all stations of Data set-2 were equipped with Lennartz LE-3D/5 s.

Recordings were baseline-corrected and the instrumental response was removed using the factory supplied zeros and poles. The resulting ground motion time series were highpass filtered applying an acausal fourth-order Butterworth filter, with cut-off frequency of 0.2 Hz. The used $S$-wave windows start 1 s before the $S$-wave onset (picked on the horizontal components) and end when 80 per cent of the record’s total energy is reached, a 5 per cent cosine taper being applied to each time-window. Fourier spectra were computed in terms of acceleration, combining the horizontal components into their root-mean-square average. The result was smoothed using the Konno & Ohmachi (1998) algorithm, fixing the smoothing parameter $b$ to 40. Spectral amplitudes of Data set-1 were discretized into 77 frequency bins equally spaced in logarithmic scale between 0.3 and 25 Hz, while spectral amplitudes of Data set-2 were discretized into 69 frequency bins between 0.5 and 25 Hz. Some specimens of seismograms and processing results are shown in Fig. 3.

Pre-event noise windows with the same length as the $S$-wave windows were used to compute the signal-to-noise ratio (SNR), and at each frequency point, only records with SNR higher than 3 were retained for the inversion. Below 0.5 Hz, too few records had an acceptable SNR (for the smaller events, the SNR was mostly smaller than 1.5–2 below 0.5 Hz). At each of the frequency bins selected, only events recorded by at least three stations and only stations that recorded at least three events with acceptable SNR were kept in the data set. This selection procedure yields 622 waveforms from 80 earthquakes of Data set-1, and 426 waveforms from 122 earthquakes of Data set-2. As a result of the adopted frequency-dependent SNR threshold, only a few spectra of the smallest events ($M_L \sim 2$) result in missing values at low and/or high frequencies in the inversion.

3 INVERSION METHOD

In the GIT approach, ground motion spectral amplitudes are written as

$$U_{ij}(f, M_i, R_{ij}) = S_i(f, M_i) \cdot A(f, R_{ij}) \cdot G_j(f)$$

(1)

where $U_{ij}(f, M_i, R_{ij})$ represents the observed spectral amplitude (of ground acceleration, in our case) at the $i$th station resulting from the $i$th earthquake with magnitude $M_i$ at hypocentral distance $R_{ij}$, $S_i(f, M_i)$, $A(f, R_{ij})$ and $G_j(f)$ are the unknown terms of the problem:
Figure 3. Example seismograms and results of data processing. Panels (a) and (d) show the horizontal components recorded by stations AQ02 and MI03, respectively, during two events of Data set-1. Red vertical lines indicate the time window selected for the spectral analysis. Fourier amplitude spectra of the horizontal (geometric average) ground acceleration are shown in panels (b) and (e), respectively. The red curve is the smoothed spectrum (Konno & Ohmachi 1998). Panels (c) and (f) are relative to the final result of the source spectra, scaled to the reference distance $R_0$ (7 km). The dotted curves indicate the omega square best-fit model.

$S(f, M_i)$ represents the source spectrum of the $i$th earthquake, $A(f, R_{ij})$ accounts for the entire-path attenuation, and $G_j(f)$ is the site response of the $j$th station. Eq. (1) is linearized by taking the logarithm of each term

$$\log_{10} U_{ij}(f, M_i, R_{ij}) = \log_{10} S_i(f, M_i) + \log_{10} A(f, R_{ij}) + \log_{10} G_j(f)$$

Eq. (2) describes a linear system of the form $Ax = b$, where $b$ is the data vector containing the logarithmic spectral amplitudes, $x$ is the vector containing the model parameters, and $A$ is the system matrix relating them. The inverse problem of eq. (2) is solved by many authors (e.g. Castro et al. 1990; Parolai et al. 2000, 2004; Oth et al. 2008, 2009) in a nonparametric approach that is especially efficient when a good distance coverage is available.

In this approach, the functional form of $A(f, R_{ij})$ has no formal dependence on specific parameters and implicitly contains all the effects that determine the attenuation along the travel path (geometrical spreading, anelastic and scattering attenuation, refracted arrivals, etc.). Since these effects vary slowly with distance, $A(f, R_{ij})$ is only constrained to be a smooth function of distance $R$ with a value of 1 at a reference distance $R_0$. Since eq. (2) still contains one undetermined degree of freedom due to the linear dependence between site and source terms, this degree of freedom is fixed by setting a reference condition either for the source or the site part (Andrews 1986; Oth et al. 2011). Usually, a common reference condition is used to fix the site function of a single rock site (e.g. Ameri et al. 2011) or the average of a set of rock sites (e.g. Poggi et al. 2011) to be equal to unity, independently of frequency. Once $A(f, R_{ij})$ has been determined, spectral amplitudes at each frequency are corrected for the effect of seismic attenuation, scaled to the reference distance, and then split into source and site terms.

In this study the interstation distance varies in a very small range, therefore ray paths between source and stations are not adequately
varied. In these conditions, the GIT method fails in separating correctly the source, site, and entire-path propagation terms (Riepl et al. 1998; Parolai et al. 2000). We have bypassed this difficulty by adopting the attenuation functions derived for the same region by Pacor et al. (2016, indicated as PAC16 hereinafter) that applied the GIT approach to a large data set composed of 9000 waveforms from 261 events of the L’Aquila sequence (including the main shock) recorded at epicentral distances smaller than 120 km, in the local magnitude range 3.0 ≤ M<sub>l</sub> ≤ 5.8. The PAC16 propagation model was carefully checked by the authors that computed the residual (log10 observation/prediction) distributions as a function of distance and magnitude, and the results of those tests did not show any trend with all the considered magnitudes and distance bins. The ray-paths and magnitude range of Data set-1 and Data set-2 fall within those of PAC16, and this legitimates the adoption of their attenuation term in our inversion. A consistency check is made a posteriori comparing results from different data and analysis methods.

4 SPECTRAL ANALYSIS AND RESULTS

4.1 Separation of source and site terms

The contributions of individual-event source spectra and individual-station site functions are computed after dividing observed ground motion spectra for the empirical propagation term A(f, R<sub>i</sub>) as estimated by PAC16

\[
\log_{10} V_{ij} (f, M_i) = \log_{10} S_j (f, M_i) + \log_{10} G_j (f),
\]

where V<sub>ij</sub>(f, M<sub>i</sub>) = U<sub>i</sub>(f, M<sub>i</sub>, R<sub>i</sub>) / A(f, R<sub>i</sub>) are the original spectra corrected for the propagation term of PAC16.

Eq. (3) is solved imposing a lower frequency bound of 0.3 and 0.5 Hz for Data set-1 and Data set-2, respectively, our data set being composed mainly of small earthquakes. The inversion was performed by constraining the logarithm of the site amplification at a reference site to be zero. Taking into consideration the geological setting of the sites and the shape of the horizontal-to-vertical (H/V) spectral ratios of stations (calculated directly from the amplitude spectra used in the inversion), we selected stations AQ01 and MNT1 as reference for Data set-1 and Data set-2, respectively. Details about the choice of these reference stations will be provided in the following where the station response functions are discussed in terms of local geology.

4.2 Seismic source

Before computing source parameters from the inverted source spectra S<sub>j</sub>(f, M<sub>i</sub>), a check is made on a possible high-frequency bias due to a not refined propagation correction. The need for such a preliminary check on the source terms derived from the inversion was remarked by Oth et al. (2011). These authors compared inversion results from borehole and surface reference stations in Japan and demonstrated that, when surface stations are used as reference, estimated source spectra deviate from the omega square model at high frequency whereas the use of borehole references yields a flat high-frequency asymptote in source acceleration spectra, as expected on the basis of the omega square model. They ascribed that discrepancy to the lack of a high-frequency decay term in the form exp(−πfκ) needed for surface stations (Anderson & Hough 1984). Both our reference stations are surface stations, therefore source spectra assessed from our inversion could be affected by such a bias, similarly to Oth et al. (2011). We have found that the inversion results really suffer this bias that is made evident by the high-frequency trend of source spectra estimated from the linear fit

\[
\log_{10} S_j (f > f_E) = S_{0,j} - \log_{10} e \cdot \pi \kappa \cdot f.
\]

where f<sub>E</sub> is the lower frequency bound of the fit above the corner frequency of the events (Anderson & Hough 1984). In view of the magnitude range of the events considered in this study, we fixed f<sub>E</sub> = 10 Hz for both data sets. The distribution of residual high-frequency decay parameter κ in the source spectra is shown in Fig. 4 for the two data sets. The resulting values of κ are both positive and negative, suggesting random fluctuations of this parameter in the individual-event source spectra. However, positive values predominate showing an approximately normal distribution around an average of 0.011 and 0.013 s for Data set-1 and Data set-2, respectively. Although smaller than the typical values of hard rock sites at the at the Earth’s surface (see Hartzell et al. 1996 or Boore 2003), the results of Fig. 4 indicate that a further correction by κ is needed to achieve a more refined propagation correction.
After this correction for the residual $\kappa$, the acceleration source spectra are written as:

$$S(f) = \left(2\pi f\right)^2 \frac{R^\text{rms} F V}{4\pi \rho v^3 R_0} M(f)$$

(5)

according to Brune (1970, 1971). Eq. (5) represents the source spectrum (see Fig. 3 for some specimens) at the reference source distance $R_0$ that was set to 7 km. In eq. (5), $R^\text{rms}$ represents the average radiation pattern of $S$-waves set to 0.55 (Boore & Boatwright 1984), $V = 1/\sqrt{2}$ accounts for the separation of $S$-wave energy into two horizontal components, $F = 2$ is the free surface amplification, $\rho = 2.88 \text{ g cm}^{-3}$ is the density, and $v_s = 3.1 \text{ km s}^{-1}$ is the shear wave velocity (according to Herrmann et al. 2011). $M(f)$ is the moment-rate spectrum defined as

$$M(f) = \frac{M_0}{1 + (f/f_0)^2}.$$ 

(6)

Using a nonlinear least-squares procedure, we determined the seismic moment $M_0$ and corner frequency $f_c$ for each of the individual acceleration source spectra. Moment magnitude $M_W$ is then derived from the Hanks & Kanamori (1979) relationship. For events with magnitude greater than 4.5, $f_c$ is likely smaller than the lowest frequencies of our analysis bandwidth. This can limit the possibility to determine reliable source parameters of largest earthquakes. To overcome this problem, moment magnitude of these events was constrained to the value given in the regional moment tensor catalogue over this problem, moment magnitude of these events was constrained to the value given in the regional moment tensor catalogue of Herrmann et al. (2011). Fig. 5 shows the pattern of acceleration source spectra of Data set-1 and Data set-2, scaled to the reference source and site terms. Our method differs because we invert only for source and site term, the attenuation term being taken from PAC16. Stress drops of Data set-2, which is characterized by magnitudes smaller than those of Data set-1, vary between 0.01 and 2.8 MPa, with a unique value of 22.5 MPa for a $M_W$ 4.9 event. The stress drop distribution results in a median value of 2.0 MPa, which differs by few tens of per cent from the previous estimates by Ameri et al. (2011), indicated as AME11 hereinafter, and by PAC16. The former found a median value of 3.0 MPa, the latter 2.6 MPa. Both these papers applied the GIT method and inverted for source, propagation and site terms. Our method differs because we invert only for source and site term, the attenuation term being taken from PAC16. Stress drops of Data set-2, which is characterized by magnitudes smaller than those of Data set-1, vary between 0.01 and 2.8 MPa, with only one outlier (11.1 MPa for a $M_W$ 4.2 event), with a median value of 0.3 MPa (Fig. 6a). The estimates of Data set-1 are approximately log-normally distributed, while in the Data set-2 case, the distribution is broader and slightly asymmetric with a prevalence of smaller values. The asymmetric distribution is likely due to the lack of events with magnitudes larger than 4 in Data set-2.

The availability of two subsets clearly separated in time offers the opportunity of investigating a time variation in the stress drops during the seismic sequence evolution. Unfortunately, the magnitude distribution of events is different in the two data sets and there is only a small range ($3.0 < M_L < 3.5$) where there are enough samples. Fig. 6(b) shows the stress drop behaviour versus time in this limited magnitude range. The comparison shows that Data set-1 has higher stress drops than Data set-2, on the average, with a significant difference of the ±1 standard deviation range ($-0.19 < \log \Delta \sigma < 0.51$ for Data set-1 and $-0.90 < \log \Delta \sigma < -0.06$ for Data set-2) in the common magnitude interval. The application of the Kolmogorov–Smirnov (not parametric) T-test (Press et al. 1989) results in a 95 per cent probability that the difference between the two subsets is not random leading to the conclusion that their stress drop values, at least in the magnitude interval $3.0 < M_L < 3.5$, represent two statistically different populations. Although limited by the small size of the two subsets, this result is significant because of its physical meaning: it could indicate a decrease of strength in the

Deviation of the model parameters. To measure the uncertainty of corner frequency, seismic moment and stress drop estimates we determined the bounds of the range of values with 95 per cent confidence intervals, according to the Viegas et al. (2010) approach.

### 4.2.1 Stress drops

The values of $\Delta \sigma$ derived from eq. (7) are shown in Fig. 6(a). The histograms of Fig. 6(a) show that individual-earthquake values of Data set-1 are mostly comprised between 0.3 and 12.0 MPa, with a unique value of 22.5 MPa for a $M_W$ 4.9 event. The stress drop distribution results in a median value of 2.0 MPa, which differs by few tens of per cent from the previous estimates by Ameri et al. (2011), indicated as AME11 hereinafter, and by PAC16. The former found a median value of 3.0 MPa, the latter 2.6 MPa. Both these papers applied the GIT method and inverted for source, propagation and site terms. Our method differs because we invert only for source and site term, the attenuation term being taken from PAC16. Stress drops of Data set-2, which is characterized by magnitudes smaller than those of Data set-1, vary between 0.01 and 2.8 MPa, with only one outlier (11.1 MPa for a $M_W$ 4.2 event), with a median value of 0.3 MPa (Fig. 6a). The estimates of Data set-1 are approximately log-normally distributed, while in the Data set-2 case, the distribution is broader and slightly asymmetric with a prevalence of smaller values. The asymmetric distribution is likely due to the lack of events with magnitudes larger than 4 in Data set-2.
fault likely due to the increase of rock damage versus time caused by repeating ruptures.

### 4.2.2 Source scaling

The general trend inferred from both data sets indicates a mean tendency toward lower values at smaller magnitudes with stress drop variability decreasing when magnitude increases. Fig. 6(a) indicates stress drops of small magnitude events span over three orders of magnitude (0.01 MPa < \( \Delta\sigma \) < 10 MPa). The lack of similarity between larger and smaller size earthquakes shown in Fig. 6(a) agrees with findings of Malagnini et al. (2011) and Calderoni et al. (2013). The former (indicated as MAL11 hereinafter) carried out a combined study on source scaling and crustal attenuation applying the Empirical Green’s Function (EGF) technique to \( S \) waves of 170 foreshocks and aftershocks of the L’Aquila sequence \( (2.8 \leq M_W \leq 6.1) \) at distances up to 200 km. The latter (indicated as CAL13 hereinafter) applied the EGF approach to broad-band seismograms of 64 earthquakes in the magnitude range \( 3.3 \leq M_W \leq 6.1 \) and used a longer time window including \( S \) waves and early codas. Estimates by MAL11 and CAL13 are both based on EGF deconvolution and, although they applied a different procedure with a different choice of EGF events and time windows, their results are very consistent, showing a decreasing stress drop scaling at small magnitudes. Other authors applied the GIT method to L’Aquila earthquakes. Among them, PAC16 found a clear evidence of dependence on the magnitude of the stress drop for magnitude larger than 3.7, with an off-set controlled by the earthquake depth.

Stress drop scaling with earthquake size is of great importance for ground motion prediction and seismic hazard assessment. Under the assumption of constant stress drop, the well-known scaling relation \( M_0 \propto f_c^{-3} \) holds and the slope of the straight line fitting data should be close to \(-1/3\) in a logarithm scale (Aki 1967). A significant number of studies on individual sequences provide evidence for a breakdown of self-similarity with increasing stress drop versus magnitude (e.g. Mayeda & Malagnini 2010). Contrasting these findings, other studies based on large earthquake populations, show evidence for a large variability and a weak stress drop dependence on earthquake size (e.g. Allmann & Shearer 2009). Oth et al. (2010) conclude that variability is smaller in the case of local and regional data sets (i.e. within individual earthquake sequences), depending on a rather uniform tectonic stress regime. Rovelli & Calderoni (2014) observe that stress drops of normal faulting earthquakes in the Apennines show small variability above \( M_w \approx 4.5 \) where stress drop is around 10 MPa; variability increases at smaller magnitudes where stress drops can be as low as 0.01 MPa. Fig. 6 indicates a similar trend in the results of this study.

Kanamori & Rivera (2004) proposed to quantify the deviation from self-similarity through a parameter \( \epsilon \) such that \( M_0 \propto f_c^{-(3+\epsilon)} \) where \( \epsilon \) measures to what extent stress drop changes with varying seismic moment. Thus \( \epsilon = 0 \) represents the self-similar stress drop scaling, while \( \epsilon > 0 \) indicates increasing stress drop with earthquake size and vice versa with \( \epsilon < 0 \) (Oth & Kaiser 2014). Results of Data set-1 and Data set-2 are summarized in the log \( f_c - \log M_0 \) plot of Fig. 7, where the expected straight lines of the constant stress drop scaling are drawn for reference. We determined \( \epsilon \) from the slope of the linear fit, and standard deviations were estimated through a bootstrap analysis using 200 re-sampled data sets. We...
found $\varepsilon = 1.0 \pm 0.2$ for Data set-1 and $\varepsilon = 1.3 \pm 0.1$ for Data set-2. Considering the uncertainties obtained for $\varepsilon$, we found a significant deviation from self-similarity for both data sets at a 95 per cent confidence level. The analysis of PAC16 yields $\varepsilon$ close to 1.5 with an uncertainty of about 0.5 over the entire data set, while CAL13 found $\varepsilon = 1.1$. These results indicate that, at a local scale (for example, within the individual sequences of earthquakes) the release of stress depends on the size of the earthquake, in disagreement with the self-similar scaling obtained in some studies derived from large populations earthquake (e.g. Allmann & Shearer 2009; Oth et al. 2010). As noted by Ben-Zion & Zhu (2002), Ben-Zion (2008) and Oth (2013), large earthquakes occur when the stress field is relatively smooth and the strong localized scale-dependency tends to be averaged out over large rupture areas. In contrast, small events (mostly the aftershock sequences) are correlated to the presence of strong heterogeneities in the focal volume, implying larger variability of stress drop for smaller size ruptures. Interestingly, the varying stress drop model has already been found in other Apenines areas as shown in papers using EGF approach. For instance, Rovelli & Calderoni (2014) estimated a scale-dependent trend with $\varepsilon = 1.4$ in the Umbria-Marche region in central Italy, for the 1997–1998 Colfiorito earthquakes. The similar trend of the L’Aquila and Colfiorito earthquakes suggests a common mechanism of stress release for the shallow normal faults in the Apennines.

4.3 Site functions

The site functions derived from the data inversion (i.e. $G_i(f)$ in eq. 1) are controlled by the local geological conditions that vary from Mesozoic limestone to alluvial fan deposits of the Aterno river. We recall that stations used in this study were installed in a small area a ten kilometres around L’Aquila. Many of them were concentrated in the villages of Onna and Monticchio, with six stations in Onna and four in Monticchio, two in the ancient settlement of the village and other two stations in disused quarries a few kilometres from the village.

As mentioned above, stations AQ01 and MNT1 were chosen as reference for Data set-1 and Data set-2, respectively. The criterion for this choice was based on the requirement of a stiff outcropping rock and an $H/V$ spectral ratio as close as possible to unity (Fig. 8a). AQ01 and MNT1 honour both these constraints, the former being installed on surface gravel deposits over the Paganica alluvial fan in proximity of the industrial area of Bazzano, and the latter on the outcropping limestone of a quarry near Monticchio. Their $H/V$ spectral ratios are close to unity within a $\pm 1$ standard deviation uncertainty (Fig. 8a).

The site response functions that result from the inversion of the horizontal components (indicated as GIT H hereinafter) are shown in Figs 9 and 10 for Data set-1 and Data set-2, respectively. In general, the $\pm 1$ standard deviation band is small suggesting a significant stability of results. Only one station with a low number of records (AQ07, which recorded only four events because of the short recording interval caused by technical problems) displayed a larger uncertainty in its site function. For this reason we excluded station AQ07 from the inversion. Stations with the flattest site functions and the lowest amplitudes are AQ02, MNT1, MNT2 and MNT3, installed on more competent rocks in and around Monticchio. Stations MI03, ONN1, ONN2, ONN3, ONN4 and ONN5, on the fluvial-lacustrine and lacustrine deposits of Onna, show larger amplitudes.

As mentioned above, stations MI03 of Data set-1 and ONN1 of Data set-2 occupied the same site in well separated time intervals. This offers the opportunity to investigate possible time variations in the soil properties near the epicentre. Fig. 8(b) compares the seismic response of MI03 (one week after the main shock) with ONN1 (July–October 2009). This comparison indicates a satisfactory overlap at low and high frequency. In contrast, in the intermediate frequency band, the site function of ONN1 exceeds the one of MI03. Before searching for possible physical causes of this observation, it has to be considered that the reference site used in the inversion was different for the two data sets. Fig. 8(a) shows that the $H/V$ spectral ratios of the reference stations AQ01 and MNT1 differ in the frequency band 2–4 Hz that coincides with the range where MI03 and ONN1 show the largest mismatch in their site functions. This suggests that the differences found versus time in the site response could derive from the use of different reference sites in the inversion, precluding any physical conclusion in terms of time variations in the near-surface soil properties.

In Fig. 10, all of the stations installed in Onna show a common feature: a prominent amplification effects attaining a factor up to 5 and a spectral peak between 2.5 and 3 Hz. As discussed in previous papers (e.g. Di Giulio et al. 2011), the main spectral peak can be ascribed to the fundamental resonant frequency of the soft upper layers. At higher frequencies, up to 10 Hz, the GIT H site functions of most of stations in Onna maintain an amplitude higher than 2 indicating a generalized broad band amplification above the fundamental resonant frequency. As a matter of fact, a broad-band effect larger than 2 at high frequencies could have duplicated the ground

Figure 8. (a) Average $H/V$ spectral ratio of reference stations (AQ01 and MNT1 for Data set-1 and Data set-2, respectively). The grey shaded areas indicate the $\pm 1$ standard deviation band. (b) Comparison between empirical transfer functions of the same site in Onna at different times: MI03 and ONN1 are relative to one week and four months after the main shock, respectively.
Figure 9. Horizontal-component site functions obtained from the inversion (GIT H) using stations of Data set-1. The grey-shaded area is the average ±1 standard deviation. The results of this study are compared with those by AME11 (blue lines): the trend of the two curves is similar but there is a systematic bias (a factor of 2 or less) likely due to the different reference site and the different crustal propagation function used in the inversion.
acceleration with dramatic implications for damage, although a simple extension to ground motion during the main shock is not allowed because of possible nonlinearities in the soil response.

Among the other stations in the Aterno river valley, those installed on alluvial fan deposits with different grain size (gravels and sands), show amplification levels varying from 2 (e.g. AQ04, AQ05 and AQ06) up to more than 4 (e.g. MI01, MI02, MI04 and MI05, all in Fig. 9), especially at high frequencies.

Figs 9 and 10 compare our results with site functions estimated by other authors: in Fig. 9 the comparison deals with GIT H results by AME11 (for Data set-1) whereas Fig. 10 compares our GIT H results of Data set-2 with empirical transfer functions estimated through standard spectral ratios of horizontal components (Giuliana Mele, private communication, 2016). Note that AME11 applied the GIT method to about 100 aftershocks (3.0 ≤ \( M_L \) ≤ 5.3) using many of stations of Data set-1 but with a different reference site (station GFZ05) that is not used in this study. Moreover, the crustal propagation function used in this study is taken from PAC16 and differs from the one of AME11. These differences can explain the mismatch (by a factor of 2 or less) that is evident between the curves of different studies in Fig. 9. However, the trend of the two curves of each panel is fairly well reproduced, and the discrepancy is substantially systematic. Even more significant is the agreement in Fig. 10, favoured by the use of the same reference station (MNT1) both in the GIT inversion and in the spectral ratio computation.

The same GIT inversion applied to the horizontal component of motion is repeated for the vertical component: in this way site response of the vertical motion (GIT Z) is computed as well. In Figs 11 and 12, the estimated vertical site response is shown for stations of Data set-1 and Data set-2, respectively. Similarly to Figs 9 and 10, a comparison is shown with results by other authors (Figs 11 and 12, respectively). Again, the agreement is much better when the same reference station (MNT1) is used both for the GIT inversion and for the standard spectral ratios, which is the case of Data set-2: the discrepancy between the two curves in each panel is negligible in Fig. 12. In contrast, for Data set-1 a systematic mismatch is observed between GIT Z of this study and GIT Z by AME11 (Fig. 11), similarly to Fig. 9. For the vertical component, the discrepancy is even larger exceeding a factor of 2 at some frequencies for a couple of stations (AQ06 and MI03). For the remaining stations, the
Separation of source and site effects in Onna

Figure 11. Vertical-component site functions obtained from the inversion (GIT Z) using stations of Data set-1. The grey-shaded area is the average ± 1 standard deviation. The results of this study are compared with those by AME11 (blue lines). Similarly to Fig. 9, the trend of the two curves is similar but the systematic bias is even larger.

mismatch is a factor of 2 or less (Fig. 11). Interestingly, the mismatch observed for stations of Data set-1 decreases significantly when the horizontal-to-vertical ratios between site functions are considered, they are indicated as GIT H/Z in Fig. 13: this confirms the important role of the different reference station used by AME11. For Data set-2, the $H/V$ spectral ratios (Giuliana Mele, private communication, 2016) agree fairly well with the GIT H/Z curves of this study (Fig. 14).

5 DISCUSSION

A key issue for a correct interpretation of the site response in Onna and inferences on the severe damage would be the identification of a particular signature layer responsible for the strong local effects. Unfortunately, this is possible only partially. In the literature, the resonant frequency around 3 Hz is accepted as due to a 40 m thick soft upper layer with $V_s$ between 300 and 400 m s$^{-1}$, according to shallow (<30 m) in situ borehole measurements (e.g. see
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Figure 12. Vertical-component site functions obtained from the inversion (GIT Z) using stations of Data set-2. The grey-shaded area is the average ± 1 standard deviation. The results of this study are compared with vertical-component standard spectral ratios (SSR Z) of the same stations using MNT1 as reference (courtesy by Giuliana Mele). There is an acceptable agreement between the two curves.

Di Giulio et al. (2011). Conversely, the transfer function amplitude of 4 to 5 found for the resonant frequency is not straightforward. According to the inversion of Data set-2, the amplification at the Onna stations installed on fluvial-lacustrine sediments is relative to the nearby outcropping limestone of Monticchio. Nevertheless, the large amplitude values of Onna are not consistent with the lack of shallow impedance contrasts inferred from geophysical surveys (G. Mele, personal communication, 2016).

Based on the lack of a fully satisfactory geological model for Onna, in this paper we have followed a different approach in trying to solve the ambiguity between vulnerability and site response in the destructive effects of the main shock at Onna. This is particularly appropriate for stations in Onna and Monticchio where damage was so different but vulnerability conditions were similar. We apply a numerical prediction method to estimate a ground acceleration scenario for the main shock at least as an upper bound in linear approximation. The scenario is made in terms of peak ground acceleration (PGA) and is realized through co-located aftershock accelerations scaled to the main shock magnitude through the source scaling law previously determined in this paper. The method dates back to Joyner & Boore (1986), it generates synthetic seismograms of a target earthquake at a generic station (STAT) through eq. (8)

$$GM_{\text{Main}}^{\text{STAT}}(t) = \sum_{j=1}^{N} w_j GM_{\text{After}}^{\text{STAT}}(\tau_j - t)$$

where $GM_{\text{Main}}^{\text{STAT}}(t)$ is the ground motion time-history of a co-located subevent recorded at the same station. The subevent time-history is randomly added with a shift time $\tau_j$ uniformly distributed between 0 and source duration $T$. According to Joyner & Boore (1986) and later procedure variations by Ordaz et al. (1995) and Calderoni et al. (2005), the scaling factor $w$ and the total number $N$ in the summation (8) are obtained through ratios of seismic moment and stress drop of the two events

$$w = \left( \frac{M_{\text{l},\text{Main}}}{M_{\text{l},\text{After}}} \right)^{\frac{1}{3}} \left( \frac{\Delta \sigma_{\text{Main}}}{\Delta \sigma_{\text{After}}} \right)^{\frac{2}{3}}$$

$$N = \left( \frac{M_{\text{l},\text{Main}}}{M_{\text{l},\text{After}}} \right)^{\frac{1}{3}} \left( \frac{\Delta \sigma_{\text{Main}}}{\Delta \sigma_{\text{After}}} \right)^{\frac{2}{3}}$$

under the assumption of the omega-square source model. Source duration $T$ is derived as the inverse of corner frequency. The
Figure 13. Ratio of horizontal-to-vertical site functions (GIT H/Z) for stations of Data set-1 (black curves). The blue curves are the results by AME11.

In order to test the reliability of the procedure, we have used the records of an accelerometric station (AQU) that recorded both the main shock and the selected subevent. The PGA of the two events was as large as 0.31 and 0.01 g, respectively, taken as the largest peak value of the two horizontal components. We applied eq. (8)
separately for the two horizontal components, generating synthetic seismograms of the main shock whose amplitude is compared to the observed one. \(N\) and \(w\) were estimated through seismic moments and stress drops of the main shock (taken from CAL13) and aftershock (from this study). The average over 100 stochastic representations of the main shock predicted ground motions yields a PGA of 0.33 g for AQU, which is very close to the real value recorded during the main shock. In spite of the strong assumptions of the method (site response linearity, adjacent source of aftershock and main shock, absence of near-field terms), the result is quite realistic for the used frequency band and station distance.

We have then applied eq. (8) to the aftershock seismograms of stations in Onna (MI03) and Monticchio (AQ02) available in the Data set-1, values of \(N\) and \(w\) were the same used for AQU. The simulated PGA of the main shock, averaged over 100 stochastic representations, resulted in 0.58 and 0.20 g for MI03 and AQ02, respectively (see Table 3). Applying the regression between MCS intensity and PGA determined by Faenza & Michelini (2010) yields simulated macroseismic intensities varying from VII-VIII to IX between Onna and Monticchio, in satisfactory agreement with observations. The main shock acceleration was estimated also for the nearby sites of MI01 (Paganica) and AQ01 (Bazzano, which is the reference site of Data set-1) that are characterized by outcropping soft sediments and gravel deposits, respectively (Fig. 15). Their PGAs result in 0.42 and 0.20 g, and the corresponding simulated MCS intensities are VIII-IX and VII-VIII (see Table 3) in agreement with values of the local macroseismic intensity, as reported by Tertulliani et al. (2009).

Based on the MCS intensity (IX) derived from the regression of Faenza & Michelini (2010), an extra contribution due to vulnerability in Onna seems to be not necessary since the shaking strength estimated for the main shock is sufficiently large to yield the largest damage in Onna among villages around L’Aquila (Fig. 15). The wide frequency band of amplification could have interfered with the small size typology of the constructions justifying the larger damage. Vulnerability was important but uniformly present in the

![Figure 14. Ratio of horizontal-to-vertical site functions (GIT H/Z) for stations of Data set-2 (black curves). The blue curves are the H/V spectral ratios as estimated by Giuliana Mele (written communication, 2016).](image-url)
Figure 15. Geological map of the study area (redrawn from *Gruppo di lavoro MS-AQ* 2010). The insets show site functions and simulated PGA of representative sites on soft deposits of the Aterno river valley (Paganica and Onna) compared to firm sites (Monticchio and Bazzano). Bazzano (AQ01) is the reference station used in the inversion of Data set-1, therefore the amplitude of its site function is equal to 1 by definition. The larger and the smaller star indicate the epicentres of main shock and subevent, respectively, used to generate synthetics of the main shock accelerations. Legend: 1–Alluvial, lacustrine, alluvial fan, eluvial-colluvial deposits (Holocene-upper Pliocene); 2–terrigenous sediments (Messinian); 3–facies of internal carbonate platform (Trias-Lias and Jurassic); 4 –‘Formazione della Terratta’ Auctt. Cretaceous, Mesozoic internal carbonate platform and Miocene limestones; 5–facies of internal carbonate platform (Cretaceous) with Miocene deposits; 6, 7, 8–Marls and marly limestones (Early Miocene) (6: area of Filetto-Pescomaggiore; 7: area of M. Pettino; 8: area of Roio-Tornimparte); a- normal fault; b- buried fault; c- strike-slip and thrusting along a transpressive fault.

Table 3. Estimates of peak ground accelerations (PGA) for stations in the town of L’Aquila and nearby villages. Macroseismic intensities I inferred from estimated PGAs reproduce satisfactorily the trend of damage-based observations, confirming the importance of the site effects in the study area.

<table>
<thead>
<tr>
<th>LOCALITY</th>
<th>PGA (g) Observed</th>
<th>PGA (g) Simulated</th>
<th>I (MSC) Observed</th>
<th>I (MSC) Simulated</th>
</tr>
</thead>
<tbody>
<tr>
<td>L’Aquila</td>
<td>0.31</td>
<td>0.33</td>
<td>VIII–IX</td>
<td>VIII</td>
</tr>
<tr>
<td>Monticchio</td>
<td>0.18</td>
<td>VI–VII</td>
<td>VII–VIII</td>
<td></td>
</tr>
<tr>
<td>Onna</td>
<td>0.58</td>
<td>IX–X</td>
<td>IX</td>
<td></td>
</tr>
<tr>
<td>Paganica</td>
<td>0.42</td>
<td>VIII–IX</td>
<td>VIII–IX</td>
<td></td>
</tr>
<tr>
<td>Bazzano</td>
<td>0.20</td>
<td>VIII</td>
<td>VII–VIII</td>
<td></td>
</tr>
</tbody>
</table>

villages around L’Aquila because of the similar typology of predominant ancient edifices. Of course, the shortcoming of this approach is the inability of precisely assess the main shock ground motion including possible nonlinear soil effects in Onna: if theoretical amplitudes are overestimated in linear regime, the contribution of vulnerability could be important. However, soft sites of the Aterno river valley recorded accelerations similar (0.36 g at AQK) and even larger (0.66 g at AQV) compared to those estimated for Onna and this suggests that nonlinear effects, if any, should have a limited extent.
6 CONCLUDING REMARKS

We analysed a data set composed of more than 1000 seismograms of 202 aftershocks recorded at 20 stations in the middle Aterno river valley at source distances shorter than 50 km and in the local magnitude range 2.0–5.4. The observed Fourier amplitude spectra were corrected for a regional propagation function as computed through GIT by Pacor et al. (2016), then we applied a non-parametric inversion that provided the complete set of site functions for individual stations and source spectra for individual events. Site functions indicate a large variability depending on the different site conditions, and source spectra were interpreted using an omega-square model yielding seismic moment and stress drop of individual events. The results were finally validated through a comparison with previous studies based on generalized inversion or other independent techniques (e.g. EGF and spectral ratios). The main conclusions can be summarized as follows:

(i) The omega-square model adequately represents the source spectra. Using this theoretical model, source parameters (seismic moment, corner frequency and stress drop) of individual events are assessed down to small magnitudes at stations with satisfactory signal-to-noise ratio;

(ii) Stress drop variability within the 2009 L'Aquila sequence spans three orders of magnitude and is comparable with the range observed for seismic sequences in other Italian and worldwide regions (0.01 MPa < Δσ < 10 MPa). A generalized tendency to decrease at smaller magnitude for earthquakes of the L'Aquila seismic sequence is confirmed in this study;

(iii) The horizontal-motion site functions derived from the GIT inversion show a large variability between stations according to their geological conditions. The important role of the soft resonating alluvial deposits of Onna is confirmed, with a concomitant broad-band amplification by more than a factor of 2 up to high frequencies. Although a realistic velocity profile beneath Onna is not available and nonlinearities in the soil response cannot be excluded, simulations of the main shock ground accelerations though aftershocks in a linear approximation yield large spatial variations, with estimated PGA values that are consistent with macroseismic intensities of the main shock, and likely confirm the role of the local geology on the destruction of Onna;

(iv) The vertical-motion site functions indicate that stations characterized by amplification of the horizontal motions suffer amplification of the vertical component as well. The ratio between horizontal and vertical site functions match fairly well independent estimates of H/V spectral ratios.

ACKNOWLEDGEMENTS

This work was financially supported by ‘FIRB—Abruzzo’ Project (UR7) funded by Ministero dell’Istruzione dell’Università e della Ricerca (MIUR): ‘Indagini ad alta risoluzione per la stima della pericolosità e del rischio sismico nelle aree colpite dal terremoto del 6 aprile 2009’ (http://progettoabruzzo.rm.ingv.it/it/home). We wish to thank Maurizio Vassallo and Giuseppe Di Giulio for their fruitful discussions and Giuliano Milan for his encouragement in this work. We also thank Adrien Oth for sharing his code for spectral analysis and Francesca Pacor that gave us the crustal attenuation functions derived for the same region. Gabriele Ameri made available to us his data set, and Giuliana Mele provided horizontal and vertical standard spectral ratios and H/V spectral ratios of stations in Onna and Monticchio before their publication. Data and unpublished information on Data set-2 were kindly provided by Rocco Cogliano, Antonio Fodarella, Stefania Pucillo and Gaetano Riccio.

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