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1 Variations of crustal elastic properties during the 2009
2 L'Aquila earthquake inferred from cross-correlations of
3 ambient seismic noise

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4 We retrieve seismic velocity variations within the Earth's crust in the re-
5 gion of L'Aquila (central Italy) by analyzing cross-correlations of more than
6 two years of continuous seismic records. The studied period includes the April
7 6, 2009, M_W 6.1 L'Aquila earthquake. We observe a decrease of seismic ve-
8 locities as a result of the earthquake's main shock. After performing the anal-
9 ysis in different frequency bands between 0.1 and 1 Hz, we conclude that the
10 velocity variations are strongest at relatively high frequencies (0.5-1 Hz) sug-
11 gesting that they are mostly related to the damage in the shallow soft lay-
12 ers resulting from the co-seismic shaking.

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1. Introduction

13 On April 6, 2009 a M_W 6.1 earthquake struck the central Apennines region near L'Aquila
14 (Italy) causing severe damage and more than 300 fatalities [*Scognamiglio et al.*, 2010].
15 This area had been long recognized as seismically active [see the official seismic hazard
16 map of Italy, *MPS Working Group*, 2004] and an occurrence of a strong earthquake in
17 the central Apennines could not be considered as totally unexpected. Before the main
18 shock, an increase in the rate of seismicity started on September 2008 and many small size
19 events (about 300 with $M_L \leq 2.5$) occurred beneath the L'Aquila city area. This foreshock
20 sequence culminated with a $M_L = 4.1$ earthquake on March 30, 2009. In the following
21 days, the seismicity decreased until two earthquakes ($M_L = 3.9$ and $M_L = 3.5$) occurred
22 just a few hours before the L'Aquila main shock. In agreement with the extensional
23 tectonics of the central Apennines, the focal mechanism of the L'Aquila earthquake has
24 been determined to be a normal fault on a South-West dipping plane with the rupture
25 area of $\sim 20 \times 15$ km² and the dipping angle of about 50 degrees [*Cirella et al.*, 2009]. The
26 main shock was followed by an aftershock sequence that included 33 earthquakes greater
27 than $M_L = 4$.

28 In this study, we use a recently proposed monitoring technique based on ambient seis-
29 mic noise. The idea of this method is to use signals reconstructed from repeated cross-
30 correlations of continuous seismic records as virtual seismograms generated by highly
31 repeatable sources. In case of well distributed noise, the reconstructed virtual sources
32 are close to point forces and the cross-correlations functions can be considered as Green
33 functions [e.g., *Weaver and Lobkis*, 2001; *Shapiro and Campillo*, 2004; *Sabra et al.*, 2005;

34 *Shapiro et al.*, 2005]. Highly accurate temporal monitoring can be also performed even
35 with inhomogeneous noise sources distributions when a perfect reconstruction of the Green
36 function is not achieved [e.g., *Hadziioannou et al.*, 2009]. The changes of the travel times
37 measured from the noise cross-correlations reflect variations of the elastic properties in
38 the propagating media, i.e., in the Earth's crust. This approach has been recently applied
39 to monitor active volcanoes [e.g., *Sens-Schönfelder and Wegler*, 2006; *Brenguier et al.*,
40 2008a; *Duputel et al.*, 2009; *Mordret et al.*, 2010] and large seismogenic faults [e.g., *We-*
41 *gler and Sens-Schönfelder*, 2007; *Brenguier et al.*, 2008b; *Chen et al.*, 2010] and to detect
42 seasonal changes in the Earth's crust resulting from thermoelastic variations [e.g., *Meier*
43 *et al.*, 2010].

44 In a seismogram or in a correlation function, the delay accumulates linearly with the
45 lapse time when the wave speed changes homogeneously within the medium. As a con-
46 sequence, a small change can be detected more easily when considering late arrivals.
47 This makes the use of coda waves particularly suited to measure temporal variations.
48 This can be done either by using the so-called stretching technique [e.g., *Wegler and*
49 *Sens-Schönfelder*, 2007] or with a method that was initially developed for repeated earth-
50 quakes (doublets) by *Poupinet et al.* [1984]. Here, we use this latter approach that has
51 **been** specifically adopted to make measurements from the noise cross-correlations [e.g.,
52 *Clarke et al.*, 2011]. We apply this method to two years of continuous recordings by three
53 seismic stations located in the vicinity of the L'Aquila main shock fault (Figure 1) to
54 measure variations of crustal seismic velocities caused by this earthquake.

2. Selecting and pre-processing the data and computing cross-correlations

55 Istituto Nazionale di Geofisica e Vulcanologia (INGV) operates two large seismologi-
56 cal networks: the Italian National Seismic Network (INSN) and the Mediterranean Very
57 Broadband Seismographic Network MedNet. The INSN consists of more than 250 stations
58 with various characteristics [Amato and Mele, 2008]. MedNet consists of 22 very broad-
59 band stations distributed over the Euro-Mediterranean area with 13 of them located in
60 Italy [Mazza *et al.*, 2008]. During period of interest for our study, four broadband stations
61 operated in continuous mode within a radius of 25 km from the main shock epicenter.
62 However, records of one of these stations contained too many gaps and we finally decided
63 to use three stations: CAMP and FIAM from INSN and AQU from MedNet (Figure 1).
64 The longest period of data availability at these three stations was between March 27, 2008
65 and April 18, 2010.

66 We re-sampled time series recorded at the three stations in order to get a perfect time
67 synchronization and filled existing small gaps via a linear interpolation. Then, we pre-
68 processed the vertical component seismograms by whitening their spectra between 0.1
69 and 1 Hz and by normalizing their amplitude through a one-bit normalization. In this
70 way, the contributions arising from strong transient phenomena were reduced in both
71 time and frequency domains [e.g., Bensen *et al.*, 2005; Brenguier *et al.*, 2008b]. Finally,
72 we computed cross-correlations between the three pairs of stations for every hour of the
73 available recordings.

3. Measurement of seismic velocity variations

74 We adopted the Multi Window Cross-Spectrum (MWCS) analysis [e.g., *Clarke et al.*,
75 2011]. This technique was first proposed by *Poupinet et al.* [1984] for retrieving the relative
76 velocity variation between earthquake doublets. *Brenquier et al.* [2008a, b] applied this
77 technique to the cross-correlations of the seismic noise. The main idea of the method is
78 that the noise cross-correlations computed from subsequent time windows can be analysed
79 similar to records from earthquake doublets. When analyzing long time series, we compare
80 a single reference cross-correlation with many subsequent current functions. The reference
81 cross-correlation CC^R for a particular station pair is obtained from stacking all available
82 cross-correlations for this pair and, therefore, is representative of the background crustal
83 state. The current cross-correlations CC^C are obtained from stacking a small sub-set of
84 cross-correlations representative of a state of the crust for a given short period of time.
85 There is a trade-off between the length of the stack required to have stable estimates of
86 the CC^C and the time resolution for detecting the variations. To find an optimal stacking
87 duration for the current function we tested different lengths between 10 and 100 days. For
88 each tested stacking length, we computed all possible functions CC^C by applying moving
89 windows shifted by two days. Then, we computed the correlation coefficient r between the
90 reference function CC^R and every CC^C . The distribution of r characterizes the similarity
91 between CC^R and CC^C for a given stacking length. We represent the overall degree of
92 similarity by the mean and the standard deviation of this distribution. Figure 2 shows
93 these parameters for the three station pairs. We observe that the degree of similarity

94 increases rapidly for short stacking durations and then it tends to stabilize. We selected
 95 a value of 50 days as stacking length for computing the current correlation functions.

96 The MWCS analysis consists of two computational steps [e.g., *Clarke et al.*, 2011]. In
 97 the first step, we estimate for a station pair k delay times δt_i^k between CC^R and CC^C
 98 within a set of time windows centered at t_i . In case of uniform velocity perturbations,
 99 the measured delays δt_i^k are expected to be a linear function of time t_i with a slope
 100 corresponding to the relative time perturbation:

$$\frac{\Delta t}{t} = -\frac{\Delta v}{v} \quad (1)$$

101 where $\frac{\Delta v}{v}$ is the relative uniform seismic velocity perturbation that can be estimated in
 102 the second step from a single station pair k via linear fitting of the following equation:

$$\delta t_i^k = -\left(\frac{\Delta v}{v}\right)_k \cdot t_i \quad (2)$$

103 In order to obtain one estimates representative of the entire crustal volume, we merged
 104 together the delays δt_i^k measured from the three station pairs before proceeding with the
 105 second step of the analysis. We computed the median value $\widetilde{\delta t}_i$ of the delays δt_i^k for every i -
 106 th window, and we inserted it into (2) to estimate of $\frac{\Delta v}{v}$ for the entire region encompassed
 107 by the three stations. When performing this analysis, we removed the central part of
 108 the cross-correlations containing direct waves (group velocities faster than 2.5 km/s; see
 109 Table 1) because they may be sensitive to the changing position of the noise sources [e.g.,
 110 *Froment et al.*, 2010]. Relative velocity variations were then computed by taking into

111 account the coda of the cross-correlation up to a length of 60 s where the signal decreases
112 to values close to the noise level.

113 To estimate uncertainties of our measurements, we followed the method proposed by
114 *Clarke et al.* [2011] and performed a synthetic test on the L'Aquila noise cross-correlations.
115 We perturbed the reference cross-correlation function by stretching its waveform and
116 simulating different values of velocity variations (from 0.01% to 0.5%). Then, we added
117 a random noise with a signal-to-noise ratio of 5 (that is the mean value measured from
118 the observed cross-correlations). Finally, we applied the MWCS technique to measure the
119 apparent velocity variations $\frac{\Delta v}{v}$ between the perturbed cross-correlations and the original
120 CC^R . The RMS deviations between the estimated velocity variations and those introduced
121 through stretching characterize the uncertainties of our measurements.

122 To investigate the depth extent of the measured crustal velocity perturbations, we
123 performed the MWCS analysis inside three different frequency bands: [0.1–1], [0.1–0.6],
124 and [0.5–1] Hz. It has been shown both theoretically and observationally that at these
125 frequencies the coda of seismograms and correlation functions are mainly composed of
126 surface waves [e.g., *Hennino et al.*, 2001; *Margerin et al.*, 2009]. We therefore expect that
127 the sensitivity of the coda waves to a velocity change at depth depends on their spectral
128 content with shorter periods sensitive to shallower structures and longer periods sampling
129 deeper parts of the crust. The measurement results for the three frequency bands are
130 presented in **Figure 3** and show a sudden velocity decrease at the time of occurrence of
131 the L'Aquila main shock. The amplitude of this velocity drop is largest at frequencies

132 higher than 0.5 Hz and decreases at lower frequencies. This indicates that a large part of
133 the observed variations have their likely origin within the shallow crustal layers.

4. Discussion

134 A limited number of available stations (only three) and the fact that only
135 one of them is located in the immediate vicinity of the earthquake fault did
136 not allow us to identify exact regions that produced the observed velocity
137 variations. Also, the dataset used in this study did not allow us to make
138 robust measurements with refined time resolution. A denser network covering
139 the source area would be required to obtain better space and time resolutions
140 [e.g., *Brenquier et al.*, 2008a]. Therefore, we interpret here only the most
141 robust average features.

142 The results presented in our study show that the L'Aquila main shock caused a de-
143 tectable reduction of seismic velocities within the surrounding crust. We observe that
144 the velocity dropped by 0.3%, which is more than 3 times larger than the fluctuations
145 observed before the main shock. Co-seismic velocity reductions can be attributed to
146 increasing crack and void densities in the shallow crustal structure and/or to reduced
147 compaction of the near-surface granular material. The presence and migration of fluids
148 can further contribute to modification of the seismic properties in the shallow crust. Our
149 results can be compared with other studies that have addressed changes of the crustal
150 parameters prior and after the L'Aquila earthquake. *Amoruso and Crescentini* [2010]
151 used strain measurements obtained in the Gran Sasso laboratory during the two years
152 prior to the main shock to infer that no anomalous signal was observed. They concluded

153 that the possible earthquake nucleation zone was confined to a volume less than 100 km^3 .
154 In contrast, v_p/v_s anomalies have been reported by *Di Luccio et al.* [2010] in the weeks
155 prior to the main shock with an abrupt variation after the $M_L = 4.1$ foreshock occurred
156 on March 30. Similar results were obtained by *Lucente et al.* [2010] who used shear wave
157 splitting in addition to v_p/v_s ratios. They attribute the velocity anomalies occurring in
158 the week prior to the main shock to a complex sequence of dilatancy-diffusion processes in
159 which fluids play a key role. *Terakawa et al.* [2010] inverted the stress field obtained from
160 the aftershock sequence focal mechanisms to determine the fluid pressure and to conclude
161 that the spatial pattern of the sequence is driven mainly by fluid migration.

162 Our results are based on current cross-correlation functions stacked over a 50 day period
163 and, therefore, do not have the time resolution required to identify possible short-term
164 precursory variations and to separate them from the co-seismic effect. On the other hand,
165 with stacking large data volumes our estimation of the co-seismic velocity reduction is
166 inherently very robust. The observed velocity reduction is larger at higher frequencies.
167 Therefore, we prefer the hypothesis the perturbation is mainly due to damaging of shallow
168 soft sedimentary layers by the co-seismic strong ground motion [e.g., *Wu et al.*, 2009]. This
169 effect may be also enhanced by the presence of fluids.

170 We compare the co-seismic perturbation observed during the L'Aquila earthquakes with
171 other cases when the co-seismic crustal velocity variations were measured from noise cross-
172 correlations (Table 2). The co-seismic velocity drop measured for the L'Aquila earthquake
173 ($\sim 0.3\%$) is significantly larger than the values measured within a similar frequency band
174 for the M_W 6.0 Parkfield and the M_W 7.9 Wenchuan events ($\Delta v/v \sim 0.08\%$ as reported by

175 *Brenquier et al.* [2008a] and *Chen et al.* [2010], respectively). At the same time, a stronger
176 variation ($\sim 0.6\%$) has been observed **with** the stretching technique and frequencies higher
177 than 2 Hz during the M_W 6.6 Mid-Niigata earthquake. The results of this comparison
178 suggest that the level of measured co-seismic velocity variation is not a simple function of
179 the total moment release during an earthquake but is controlled by different factors such
180 as local geological conditions and, possibly, focal mechanism and source depth. Also, the
181 frequency range used in the analysis controls the depth extent of the measured anomaly.
182 Finally, the aperture of the used seismic network (i.e., the distance between the station
183 pairs) can play an important role. So far, the velocity variations reported in this study
184 were measured over a relatively large area. Therefore, they may be less sensitive to the
185 processes occurring in the immediate vicinity of the fault, where stress-induced velocity
186 perturbations are expected to be most important.

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Table 1. Parameters of the three inter-stations paths used in the study. The Rayleigh wave arrival times are roughly estimated considering a group velocity of 3 km/s [*Chiarabba et al.*, 2009]). Parts of the correlation functions with group velocities faster than 2.5 km/s we excluded from the analysis to avoid the influence of the noise source variability in direct arrivals.

stations	distance km	Rayleigh arrival s	cutoff s
AQU_CAMP	20	6.67	± 7.5
AQU_FIAM	26	8.67	± 10
CAMP_FIAM	38	12.67	± 15

Table 2. Comparison between the L'Aquila event and other earthquakes where co-seismic velocity variations were measured from noise cross-correlations. Values of velocity variations are from *Brenquier et al.* [2008a], *Wegler and Sens-Schönfelder* [2007], and *Chen et al.* [2010], for the Parkfield, the Mid-Niigata, and the Wenchuan earthquakes, respectively.

Earthquake	M_w	depth km	focal mechanism	$\Delta v/v$ %	frequency Hz	stations
L'Aquila	6.1	8.8	normal	0.15 0.3 0.4	0.1–0.6 0.1–1 0.5–1	3
Parkfield	6.0	7.9	strike-slip	0.08	0.1–0.9	13
Mid-Niigata	6.6	5	thrust	0.6	> 2	1
Wenchuan	7.9	19	thrust	0.08	0.3–1	> 30

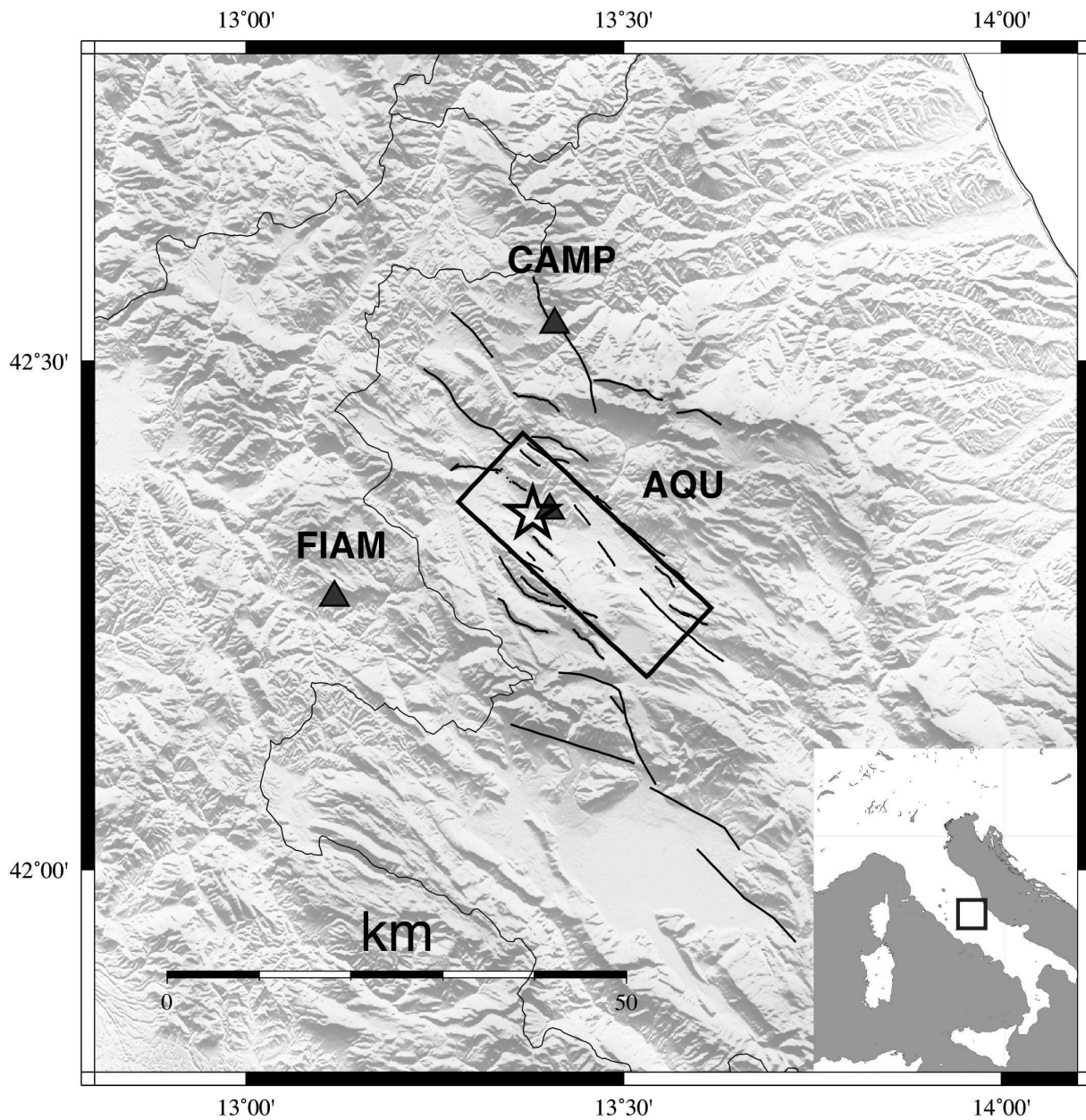


Figure 1. Map of the central Apennines showing the location of the L'Aquila epicenter (black open star) and of the fault plane projection (black rectangle) from *Cirella et al.* [2009]. The gray triangles are the three stations considered in this study. Black thin lines indicate main tectonic faults from *Emergeo Working Group* [2010]. Light gray lines show the regional boundaries.

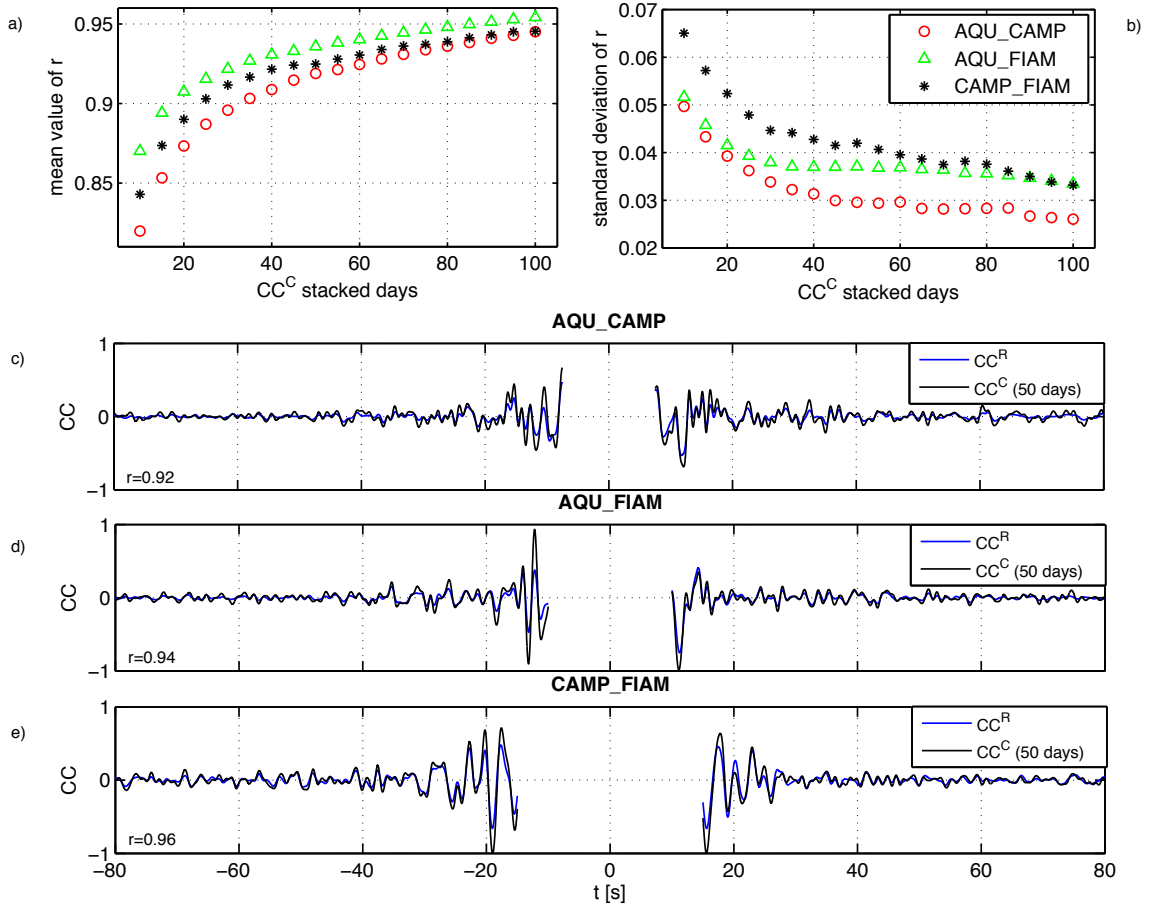


Figure 2. Mean (a) and standard deviation (b) values of the correlation coefficients r between CC^C and CC^R as a function of number of days used to construct the current correlation functions CC^C . Mean and standard deviations were computed after a Fisher transformation that returns an almost normally distributed variable [VanDecar and Crosson, 1990]. Panels (c),(d), and (e) show the reference cross-correlation functions CC^R (blue) together with an example of 50 day current function CC^C (black) for the three couples of stations. Only portions of the the signal considered in the analysis are plotted (Table 1). Numbers in the bottom left corners are the respective correlation coefficients r .

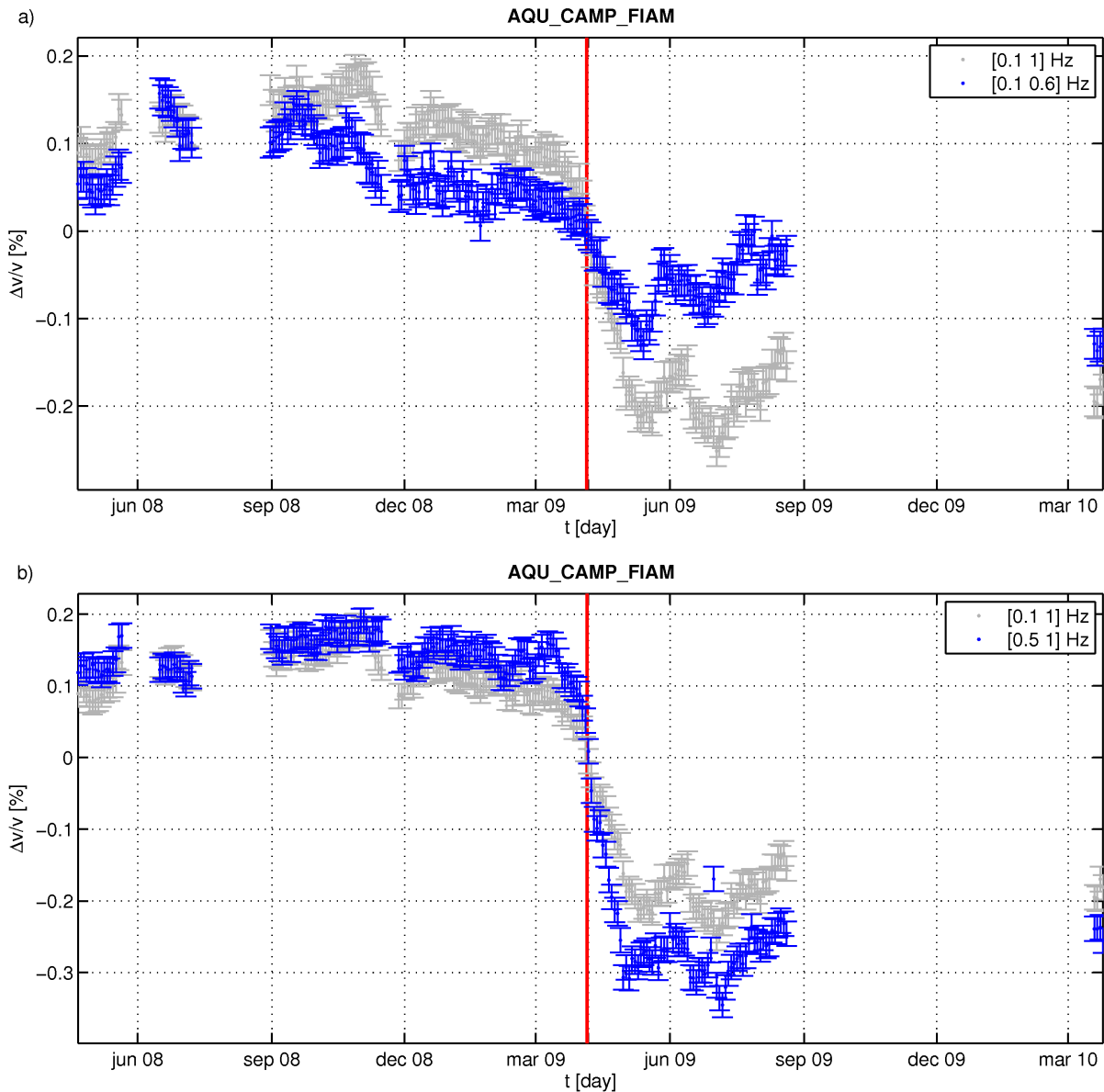


Figure 3. Relative velocity variations measured from cross-correlations of seismic noise recorded at the three stations (gaps correspond to periods when the stations were not operating simultaneously). Results obtained by analyzing the whole frequency range [0.1 1] Hz are shown with a gray color. Blue color shows the results from narrower frequency ranges: (a) [0.1 0.6] Hz and (b) [0.5 1] Hz). Vertical bars indicate the uncertainties of the measurements. The vertical red line highlights the time of occurrence of the L'Aquila main shock.