Seismic Ambient Noise Analysis in Southern Italy
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Abstract We studied the ambient noise recorded at Irpinia Seismic Network (ISNet), a seismic network installed along the Campania–Lucania Apenninic chain (southern Italy), with the aim of characterizing the noise spectrum for each station as a function of time and the detection threshold of the network. For the latter purpose, we proposed a mixed indirect approach based on the signal-to-noise ratio (SNR) in the time domain, with parameterization in the frequency domain. The source signature is represented by the convolution of the Brune source time function with the Azimi attenuation curve. We found that 1.3 is the minimum magnitude an event should have to be detected at least at five stations with an SNR larger than five, wherever it occurs. We observed a space variability of the detection threshold as large as 0.3 units, ascribed to both the geometrical configuration of the network and the differences in the noise levels at the different stations. A sensitivity study indicates that the estimation of the detection threshold is robust for changes in the focal depths and stress drop, while it is strongly affected by the anelastic attenuation. In our case, changes of the reduced time $t^*$ in the range 0.015–0.035 s generate changes in the completeness threshold of 0.5 units.

Noise levels were obtained by a statistical analysis on the power spectral density curves along almost three years of continuous data from 22 stations. We found that, at short periods, major time variations are generated by diurnal changes in the wind intensity and other meteorological factors. At longer periods, we retrieved the microseismic peak, resulting from the constructive interference of oceanic waves. We also found an additional peak between 2 and 4 s, correlated with the sea wave height along the Tyrrhenian coast.

Introduction

Local networks are generally installed in the vicinity of a fault or a fault system that may produce a moderate to large magnitude earthquake in the near future. They are aimed at monitoring the seismicity during the seismic cycle to detect anomalies and changes in the rate of occurrence of small earthquakes. With the installation of digital networks, the transmission and processing of data in real time, the accurate location of small earthquakes, and their characterization in terms of source properties have become fundamental to image active structures (Waldhauser and Ellsworth, 2000; Fukuyama et al., 2003), to understand the fault mechanics and the forcing mechanisms (Carena et al., 2002; Crampin et al., 2002), to estimate the state of the stress at depth via the focal mechanisms computation (Hardebeck and Shearer, 2003; Hardebeck, 2006), and to characterize the statistical parameters associated with a seismic cycle (Sammis and Smith, 1999; Helmstetter and Sornette, 2002). In such a framework, interpreting data from smaller and smaller events may allow researchers to better constrain the geometry and the activity state of a fault system.

To investigate the mechanical and statistical properties of a seismically active area, we need to decouple the recorded seismicity by the geometrical configuration of the network. We should rule out the events that are below the completeness threshold of the catalog, which is directly linked to the detection capability of the network itself. We assume the detection threshold of a network as the minimum magnitude of an event that has a 90% of probability of being identified and localized with an accuracy comparable to the standard location errors of the network, the choice of the probability level being a standard for seismology (Ringdal, 1975). The importance of the detection threshold historically arose from the analysis of signals induced by nuclear tests (Ringdal, 1975; Ringdal and Kvaerna, 1989); at that point, the techniques developed to investigate the signal-to-noise ratio (SNR) for data recorded at teleseismic and regional distances were extended to smaller-scale networks (Gomberg, 1991).
Two strategies are generally followed for the estimation of the detection threshold. The recurrence curve method uses the seismicity actually recorded by a seismic network to compute the Gutenberg–Richter exponential decay function and compares the data recorded by each single station with the ones recorded by the whole network. Plotting the frequency–magnitude curve for a given station, we expect the same slope at large magnitudes and a smaller one (or even a level or a positive slope) at small magnitudes because some of the small events detected by the network do not come out of the noise at that station. The magnitude at which the curve changes its slope is the detection threshold for such a station. Because this magnitude is a function of the distance, data need to be grouped by distance or an attenuation law should be used to reduce the data at the same distance (Cao and Gao, 2002; Wiemer and Wiss, 2002). An alternative procedure, avoiding the explicit assumption of a Gutenberg–Richter scaling law, consists of building up a station-dependent probability map of detecting an event of given magnitude recorded at a given distance, based on a graph of detected/missed events from such a station (Schorlemmer and Woessner, 2008; Schorlemmer et al., 2010). This latter method requires a large catalog populated in magnitude and distance, which, for instance, is not available for the Irpinia region. Since the installation of the seismic network, we recorded events of magnitude ranging between 0.5 and 3.0, and we are not even able to accurately estimate the occurrence rate (the $b$ parameter of the Gutenberg–Richter relationship) associated with the area.

In southern Italy we prefer to use an indirect technique based on the noise analysis. This approach, already implemented for other seismic networks such as the Advanced National Seismic System (ANSS) backbone, the Global Seismic Network, and the northern Italy networks (McNamara et al., 2004; McNamara et al., 2005; Marzorati and Bindi, 2006), is based on the comparison of the noise level recorded at the single stations with theoretical spectra associated with rupture models of small earthquakes. The detection of an event occurring at a point inside the area of interest is then declared when the earthquake spectrum associated with the $P$ and/or $S$ wave sufficiently exceeds the noise level at a sufficiently large number of stations to accurately locate the event itself (McNamara et al., 2004). For this technique, analysis and parameterization in the frequency domain are motivated by the fact that some of the anomalies in the noise generally are due to very localized peaks in the frequency domain, which can be easily identified and possibly removed from the records. Additionally, when working with velocity records, time-domain analysis is not directly linked with the local magnitude because the noise level generally increases after integration because of baselines. In the frequency domain, detection based on comparison of theoretical source models and noise power spectral densities is directly linked to the moment magnitude and the corner frequency when we are able to properly correct for the anelastic attenuation.

As concerns the seismicity analysis, the noise is considered as an almost uncontrollable effect reducing the quality of the data. Hence, strategies are applied to reduce the noise level by installing instrumentation at depth in wells or by stacking data from very dense arrays. However, seismic noise has recently become a huge resource for the knowledge of the structure underlying a seismic network and its variation with time through the stacking of cross-correlated records at pairs of stations. At regional and teleseismic distances, low-frequency sources of noise will propagate across the networks, allowing the surface waves to be enhanced in the stacking (Shapiro et al., 2005; Bensen et al., 2008). At local scale and in the high-frequency range, the extraction of the Green’s function may be prevented by the spatial localization of sources. The characterization of noise spectrum and of its sources then becomes fundamental to extract the elastic properties of the propagation medium below a seismic network.

In this paper we characterize the noise level at the stations of the Irpinia Seismic Network (ISNet), a seismic network installed along the Campano–Lucano Apennine (southern Italy), with the aim of studying the main features of the noise itself as a function of the frequency, its space and time variations, and the detection threshold of the network. In the first section, we present the data used in the analysis. Then we show the power spectral density curves for the stations of ISNet and the behavior of the noise as a function of frequency and time. In the last section, we analyze the detection threshold of the network using a statistical description of the noise levels.

Data

We studied the seismic noise using continuous data recorded by ISNet. ISNet is a dense, local network of strong-motion, short-period, and broadband seismic stations displaced over an area of approximately $100 \times 70$ km$^2$ along the southern Apennines chain (southern Italy; Weber et al., 2007; Iannaccone, Zollo, et al., 2010) around the fault system that generated the 1980 M 6.9 Irpinia earthquake. The location of the network and its geometrical configuration is shown in Figure 1. ISNet is composed of 22 stations and is subdivided into four subnets, with each station being connected to a local control center (LCC) in real time. The LCCs are then linked to the network center, gathering parametric information (first arrival picks, velocity and acceleration peaks, maximum Wood–Anderson amplitude, etc.) and computing earthquake source parameters within a short time after the event.

To ensure a high dynamic recording range, each seismic station is equipped with two three-component instruments: an accelerometer (Gürälp CMG-5T) and a short-period velocimeter (Geotech S-13J). At five stations, short-period instruments are replaced by broadband sensors (Nanometrics Trillian 40s) to better record regional and teleseismic events and to broaden the spectrum of the low-frequency noise in the area up to 40 s. The data acquisition is performed by the
Osiris-6 data logger (Agecodagis), equipped with a $\Sigma - \Delta$ 24-bit A/D converter. Input signals entering the data logger are not amplified and are acquired by the analog-to-digital converter (ADC) at 125 samples/second (sps). The seismic stations are housed in shelters and equipped with solar panels and batteries. The sensors are installed on an internal reinforced concrete block and isolated by the external walls of the shelter. No special care was taken to reduce the thermal effects, such as the temperature changes, and to suppress any possible air movement around the instruments.

The data sent by the stations to the control centers are managed by the software Earthworm (Johnson et al., 1995) that automatically picks the first arrivals and locates the events inside ISNet or on its outskirts. For the analysis, we selected three-component velocity data, acquired by the broadband and short-period sensors from January 2008 to September 2010.

**Power Spectral Density Curves**

Seismic noise was here synthesized by power spectral density (PSD) curves. In the hypothesis of a stationary noise, the PSD curves represent the energy density per unit time as a function of the frequency. Hence the integral of the PSD around a given frequency is the power associated with the signal in such a frequency band. For a stationary ergodic time series $n(t)$, we consider the autocorrelation function as

$$R(t) = \lim_{T \to \infty} \frac{1}{T} \int_{0}^{T} n(\tau)n(t + \tau) d\tau.$$  

Its Fourier transform is the PSD

$$PSD(f) = \int_{-\infty}^{\infty} R(t) e^{-2\pi if t} dt.$$  

For a discrete signal spanning the time interval $[0, t_0]$, the PSD is computed as the squared modulus of $N(f)$, the fast Fourier transform of $n(t)$, normalized by half the window length $t_0/2$. The unit of the PSD is the unit of $n^2$ per Hz.

It is a common practice to represent the PSD in a logarithmic scale, normalized by the unit amplitude of the PSD at 1 Hz. The quantity 10 log (PSD) is commonly referred to as the decibel scale of the PSD.

We computed the PSD curves from velocity signals because the self-noise of the velocimeters installed in ISNet is more suitable to record the ambient noise within the frequency band of interest. For this purpose, the continuous record from a single station was subdivided into nonoverlapping one-hour-long segments, discarding traces along which gaps had occurred. It is worth noting, however, that the presence of gaps is only due to anomalies in the station working—not to a communication breakdown. Data loggers have a local backup of the data collected by the stations; and, when the communication with the LCCs breaks down, the data are automatically recovered at the local control centers after the transmission has been restored. To minimize the long-period contamination, the mean and the linear trend were removed from the seismic traces. Before computing the spectrum, a 10% cosine taper was applied to the ends of each time series to suppress side lobe leakage. Finally, the Welch method (Welch, 1967) was used to compute the PSD curves to avoid localization of peaks, with a 50% of overlapping on about one-minute-long time windows.

Figure 2 shows the PSD curves collected for one day at the vertical components of both the broadband station RDM3 (Fig. 2a) and the short-period station SNR3 (Fig. 2b), which are representative of the noise levels recorded in ISNet. As common in seismology, the PSD curves are here represented as a function of the period instead of the frequency, and they are compared with the new low-noise (NLNM) and the new high-noise (NHNM) models of Peterson (1993). These models represent the lower- and upper-bound envelopes of a cumulative compilation of representative PSD curves determined at 75 continental worldwide seismic stations. They are currently assumed as the acceptable limits for the seismic noise at permanent inland stations (Bormann, 2002). For RDM3 (Fig. 2a), all the PSD curves fall within the NLNM and NHNM levels; for station SNR3 (Fig. 2b), the PSD curves lie within the Peterson levels only for periods smaller than 15 s. At larger periods, the signal overcomes the NHNM level. Moreover the increasing trend above 5–6 s is not representative of the seismic noise as compared to the levels provided by the broadband stations. Such a limit indicates that at large periods the ambient noise goes below the instrumental self-noise, and its features cannot be retrieved by the short-period sensors even when removing the instrumental response.
To study the variability of the seismic noise with time, we plot the PSD levels as a function of the period versus time in one-month-long spectrograms. Figure 3 shows some examples of spectrograms for four broadband stations (RSF3, TEO3, RDM3, and COL3), covering the month of November 2008. For the sake of completeness, we show the data of the vertical and north–south components.

The spectrograms enhance several interesting patterns with different characteristics, on the basis of which we can identify two bands of investigation: below 1 s (short periods) and between 1 and 10 s (long periods). At long periods, the noise shows a strong variation up to 30 dB. Amplifications are mostly limited to the 2–9-s band and correspond to two peaks on the PSD curves clearly identifiable in Figure 2, the first one between 2 and 4 s and the second one between 5 and 9 s. This latter is generally associated with the secondary microseismic peak caused by the superposition of oceanic waves of equal period traveling in opposite directions, thus generating gravity standing waves, each with a period half of the standard water wave period (Longuet-Higgins, 1950). In order to investigate the time variation of such peaks, we plot the PSD curves at station RSF3, averaged between 2–4 s and 5–9 s, as a function of time (Fig. 4a). In the figure, we plot both the short- and long-term variability of the two peaks, with the latter obtained by smoothing the data with an average moving window of 3 months duration. Specifically, we represent the long-term trend of the noise in the ranges 2–4 s and 5–9 s with dashed and dotted thick lines, respectively and their short-term counterparts with dark and light solid lines. Both peaks show a long-term variability with a decrease of the amplitude during the summer and an increase during the winter. The seasonal difference is as large as 5 dB, indicating an average energy in the noise at these frequencies that is seven times larger in the winter as compared to the summer. The short-term variability of the two peaks with time shows a more complex pattern and large amplitude changes with a characteristic time of hours to weeks. We recognize, however, that the temporal changes associated with the 5–9-s peak are less pronounced with respect to the 2–4-s peak especially in the summer when the largest amplitude variations are on average as large as 15 dB. The changes associated with the 5–9-s peak during the same period have an average amplitude of 5 dB.

To further investigate the origin of the 2–4-s peak, we plot the noise level averaged in this period range at the broadband stations RSF3, COL3, and RDM3 (Fig. 4b). We also show the amplitude of the sea motion measured at Cetraro, a village along the coast about 150-km south of station CMP3 (the location of the site with respect to ISNet is shown in the upper panel of Fig. 1). The sea wave height was measured by the directional wave buoy TRIAXY (Axys Technologies Inc.) of the Italian wave measurement network (Rete Ondametrica Nazionale, RON) managed by the Istituto Superiore per la Protezione e la Ricerca Ambientale (ISPRA). As compared to the time series in Figure 4a, the time window is now reduced to the period 1/3/2008–5/4/2008 because of the availability of the sea motion data. We see a strong correlation between peaks observed along the PSD curves and peaks associated with the sea motion, indicating that the variability in the motion of the Tyrrhenian Sea is influencing the noise amplitude in the range of 2 to 4 s.

At short periods, spectrograms show several thin bands below 0.1 s, corresponding to the oscillation of the shelter hosting the instrumentation caused by the wind. Although very localized, such peaks may influence the detection and the location of small events ($M < 1.5$) inside ISNet, with errors in the $P$-wave picking as large as 0.2 s. However, because the peaks are known and very localized in frequency, picking procedures can be improved with band-reject filters.

To investigate the time variation of the noise at short periods, in Figure 5 we plot the PSD curves averaged between 0.1–1 s for five stations of ISNet (data from three components are superimposed on the same panel), using the data recorded during the month of June 2008. In this band, most of the variations with time are diurnal, with the noise undergoing a rapid decrease during the night and reaching its minimum around midnight. Such variations occur at all the stations of ISNet. We do not believe that such changes are
related to human activity, as is generally suggested for
stations located in or close to urban areas (McNamara and
Buland, 2004). In fact, ISNet stations are generally located
far from the urbanized areas and the observed diurnal vari-
tions do not change their intensity during the weekend when
the energy of the cultural noise is expected to decrease. In
Figure 6 we compare the wind intensity recorded at the
meteorological station Conza (Fig. 1; data provided by Cen-
tro Irpino per l’Innovazione ed il Monitoraggio Ambientale
[CIMA]), with the averaged PSD curve related to the vertical
component of the closest station, CLT3. We observe a strong
correlation between the peaks in the PSD and the peaks in the
wind intensity. Such a correlation is stronger when the wind
intensity is high for time windows longer than one day. For
diurnal variations, the occurrence time of peaks in the PSD is
still correlated to the time at which the wind intensity is high,
but we do not always observe a strong increase in the wind intensity that corresponds with a strong increase in the amplitude of the noise, indicating that other meteorological factors, such as the wind direction and the temperature, may play a role on the noise recorded in ISNet.

**Statistical Analysis of the PSD Curves**

Noise levels change with time because the sources of noise are not stationary over long time scales. To study the noise levels at the ISNet stations, we carried out a statistical analysis on the PSD curves. As proposed by McNamara and Buland (2004), the PSD curves were used to define a probability density function (PDF) of the seismic noise for each period of investigation. For the computation of the PDF, the single PSD curves were smoothed using a centered averaging moving window with a size of 3 months: the dashed line corresponds to the 2–4 s range and the dotted line to the 5–9 s range. In both cases, we observe a seasonal effect with differences in the noise of 5 dB between the winter and the summer. Both nonsmoothed traces show variations as large as 15 dB for the 2–4 s signal and 10 dB for the 5–9 s bands. (b) Time variation of PSD curves, averaged between 2 and 4 s, for the vertical components of the broadband stations RSF3 (dot-dashed line), COL3 (dotted line), and RDM3 (dashed line). The solid line indicates the sea wave height, measured at the Cetraro buoy. A strong correlation is observed between the noise recorded in this frequency band and sea wave motion along the Tyrrhenian coast. The color version of this figure is available only in the electronic edition.

**Figure 4.** (a) Time variation of PSD curves, averaged between 2 and 4 s (solid dark line) and between 5 and 9 s (solid light line) are represented for the vertical component of station RSF3. We superimpose their smoothed versions, obtained using an average moving window with a size of 3 months: the dashed line corresponds to the 2–4 s range and the dotted line to the 5–9 s range. In both cases, we observe a seasonal effect with differences in the noise of 5 dB between the winter and the summer. Both nonsmoothed traces show variations as large as 15 dB for the 2–4 s signal and 10 dB for the 5–9 s bands.

**Detection Threshold of ISNet**

We propose an indirect method for the computation of the detection threshold of ISNet based on the SNR at the stations of the network. An SNR value larger than 5 on the horizontal components of the seismic records guarantees a correct estimation of the local magnitude for events recorded by ISNet (Bobbio et al., 2009). We hence assumed such a value as the minimum required for an S wave to be detected on the horizontal components. By similarity, we expect that the direct P wave can be correctly individuated when the
SNR associated with this phase is larger than 5 on the vertical component. The use of direct waves is here motivated by the fact that they generally dominate the seismograms at local distances. Because we defined the detection threshold as the minimum magnitude for which 90% of the events will be identified and localized, we compared a theoretical signature of the direct $P$ and $S$ waves with the ninetieth percentile of the PDFs at the stations of ISNet. We used the Brune source model (Brune, 1970; Brune, 1971) in a homogeneous medium to represent the $P$ and $S$ amplitude spectra of the recorded velocity associated with an earthquake of seismic moment $M_0$, recorded at hypocentral distance $R$.

Figure 5. Time variation of the PSD level averaged in the period band [0.1 s, 1 s], related to the three components of five stations, representative of ISNet. Each panel shows the noise levels of a single station as a function of time. Most of the variations are diurnal with differences of 20–40 dB from the daylight to the night time.
Finally the corner frequency was computed using a constant stress drop (Margheriti and Zollo, 2010). We chose the following for the Irpinia region: $\rho$ the density, $c_s$ the P/S wave velocity, $f_c$ the corresponding corner frequency, $\rho$ the density, and $\tau^*$ the travel time reduced by the noise. Following Aki and Richards (1980), we defined the SNR as

$$S(f) = M_0 \frac{R_{\phi z} F_s f_c}{2 \rho R c^3} \frac{f}{1 + (f / f_c)^2} e^{-\pi f f_c},$$

where the positive frequency $f$ is the reciprocal of the period $T$, $R_{\phi z}$ the radiation pattern, $F_s$ the free-surface correction, $c$ the P/S wave velocity, $f_c$ the corresponding corner frequency, $\rho$ the density, and $\tau^*$ the travel time reduced by the quality factor. Such a model was shown to hold for small earthquakes occurring in southern Italy (Margheriti and Zollo, 2010). We chose the following for the Irpinia region: $R_{\phi z} = 0.6$ (Boore and Boatwright, 1984), $F_s = 2$, $c_s = 3$ km/s, $c_p / c_s = 1.85$ (De Matteis et al., 2010), $\rho = 2.5$ g/cm$^3$ (Improta et al., 2003), and $\tau^* = 0.025\ s$ as a mean value for both $P$ and $S$ waves, averaged on the distance (Margheriti and Zollo, 2010). Finally the corner frequency was computed using a constant stress drop $\Delta \sigma = 2$ MPa (Margheriti and Zollo, 2010) and the formula

$$f_c = \frac{2.34}{\pi c} \left( \frac{16 \Delta \sigma}{M_0} \right)^{1/3}.$$

It is worth noting that the amplitude spectrum is not directly comparable with the power spectral density associated with the noise. Following Aki and Richards (1980), we defined the detection threshold as a function of the SNR in the time domain. First, we converted the Brune spectrum into a time-domain function. For the goal, we separated the source effect from the anelastic attenuation. The time-domain representation of the velocity spectrum is the time derivative of the Brune time function:

$$s_1(t) = M_0 \frac{\pi R_{\phi z} F_s f_c^2}{\rho R c^3} (1 - 2 \pi f_c t) e^{-2 \pi f_c t}, \quad t > 0,$$

while the function $s_1$ is zero for negative time. Using the analogy between pairs of Fourier transforms and defining the attenuation factor as negative for negative frequencies (Aki and Richards, 1980), the time-domain representation of such a kernel is

$$s_2(t) = \frac{2}{\pi} \frac{t^*}{t^* + 4 t^2}.$$

The amplitude of the time signal is

$$S(t) = s_1(t) \ast s_2(t)$$

$$= \frac{2 R_{\phi z} F_s f_c^2}{\rho R c^3} \int_0^{\infty} \left(1 - 2 \pi f_c t\right) e^{-2 \pi f_c t} t^* dt \times e^{-2 \pi f_c t^*} \frac{t^*}{t^* + 4 (t^*)^2}$$

Using the symmetry of the functions $s_1$ and $s_2$, $S(t)$ is still an odd function, with a maximum at a positive time. The numerator of the SNR was defined as the maximum of the function $|S(t)|$. Such a quantity is

$$S = \max |S(t)| = M_0 C(f_c, t^*) \frac{2 R_{\phi z} F_s f_c^2}{\rho R c^3 t^*},$$

where $C$ is the maximum of the integral as a function of time. It was numerically computed as a function of the moment magnitude, fixing $t^*$ to the value chosen for the area.

The mean square amplitude of the noise in the time domain was computed using the Parseval theorem and the PSD as the Fourier transform of the autocorrelation function:

$$\langle |n(t)|^2 \rangle = 2 \int_{f_{min}}^{f_{max}} P(f) df = 2 (P(f))(f_{max} - f_{min}).$$

$f_{max}$ was chosen to be 40 Hz on the basis of the acquisition sampling rate (125 sps) and $f_{min}$ as 1 Hz to reduce the baseline effect generated by the long-period noise in the microseismic band. This latter frequency is also the cut-off frequency of the high-pass filter used by the automatic picker of ISNet. We used the square root of the mean square amplitude as representative of the noise level in time domain. The SNR is now written as

$$\text{SNR} = \frac{M_0 C(f_c, t^*) \frac{2 R_{\phi z} F_s f_c^2}{\rho R c^3 t^* \sqrt{2 (P(f))(f_{max} - f_{min})}}}{\rho R c^3 t^* \sqrt{2 (P(f))(f_{max} - f_{min})}}.$$

The theoretical amplitude of the $P$ and $S$ waves depends on the hypocentral distance and hence on the specific location of the earthquake. To investigate such a dependence, we defined a regular grid on the investigation area, with cell size of $1 \times 1$ km$^2$. We moved the epicenter of the earthquake on each node of the grid while the earthquake depth is fixed at 10 km, at which most of the seismicity in the area is concentrated (Margheriti and Zollo, 2010). For each node, we then computed the smallest seismic moment associated with a seismic event recorded by at least five stations with an SNR $> 5$. These threshold levels insure an accurate estimation of both location and magnitude. For each node, we converted the seismic moment into the moment magnitude according to the formula $M_0 = 0.67 \log M_0 - 6.07$, such a value representing the detection threshold associated with the single node. The detection threshold for the investigated area and for $P$ and $S$ waves was finally obtained by computing the magnitude threshold for each node of the grid.

Figure 6. Time variation of PSD curves at the station CLT3 (vertical component) averaged between 0.1 and 1 s (solid line) and wind velocity (dashed line) recorded by the meteorological station of Conza, located about 7 km from the CLT3 station. The two curves are correlated over time windows larger than one day. Nevertheless, for daily variations, we do not necessarily observe a strong variation in the wind intensity corresponding with a strong variation in the noise level. The color version of this figure is available only in the electronic edition.
In Figure 8a,b, we plot the detection threshold maps for $P$ and $S$ waves. The smallest magnitudes detectable at all points inside the network are 1.3 for $P$ waves and 1.1 for $S$ waves. However, the maps show significant spatial variations of about 0.2–0.3 units in $M_w$, reflecting the differences in the noise levels at the stations of ISNet. The smallest magnitude values on these maps are 1.0 and 0.9 for $P$ and $S$ waves, respectively, occurring in the middle and in the west regions of the network for the $P$ map and eastward for the $S$ map. To stress the effect of the space variation of the noise on the detection capability of ISNet, we show the detection thresholds for $P$ and $S$ waves (Figs. 8c and 8d, respectively) that we would have if the noise levels at all stations were the same as that measured at LIO3, where the noise is minimum almost in all the short-period range. For these cases, the smallest magnitude detectable at all points inside ISNet is 0.9 for $P$ waves and 0.8 for $S$ waves; and, specifically for $P$ waves, the magnitude distribution reflects the geometrical configuration of the network.

Several of the parameters used in this study are assumed to be constant inside ISNet and are fixed when comparing the Brune model to the noise levels. However, such parameters
are affected by uncertainties that may influence the detection threshold. We hence studied the variability of the detection threshold as a function of the reduced time $t^*/0.0003$, the stress drop, and the depth of the earthquakes, which are the most critical parameters in the definition of the Brune spectrum. We synthesized the results by evaluating the minimum magnitude of an event, wherever it occurs, that will be detected by the network is $1.3$. The maps also show a variability as large as 0.3 units in magnitude. In (c) and (d), the detection thresholds refer to the hypothetical case in which all stations would have the same noise level as the lowest noise station, LIO3, only reflecting the geometrical configuration of the network. The minimum magnitude would decrease to 0.8 for $S$-wave detection in all the investigated area.

Changes of $t^*$ between 0.015 and 0.035 s result in changes in the threshold of 0.5 units in $M_w$, from 0.8 to 1.3 for $S$ waves and from 1.0 to 1.5 for $P$ waves. Hence a correct estimation of $t^*$ and eventually its dependence with the distance may be relevant for the evaluation of the detection threshold.

In Figure 10 we finally compare the completeness threshold obtained through the analysis of the noise with the one predicted by the Gutenberg–Richter (GR) law. The law was computed using all the seismic events acquired by the ISNet stations since January 2008 and located inside the network. The cumulative frequency–magnitude distribution was built by grouping the earthquakes in classes of magnitude having width of 0.3. Finally, the coefficients of the GR law were retrieved by a linear fit performed on the associated cumulative distribution. To investigate the completeness threshold, several linear fits were performed, including points at smaller and smaller magnitude, and the minimum magnitude to be included in the fit was defined as the point

Figure 8. Detection thresholds maps for the Campania–Lucania Apennines: (a) and (c) are related to the $P$ waves and (b) and (d) to $S$ waves, respectively. In (a) and (b), the detection thresholds are represented for the effective noise levels measured at the single stations of ISNet. The minimum magnitude of an event, wherever it occurs, that will be detected by the network is 1.3. The maps also show a variability as large as 0.3 units in magnitude. In (c) and (d), the detection thresholds refer to the hypothetical case in which all stations would have the same noise level as the lowest noise station, LIO3, only reflecting the geometrical configuration of the network. The minimum magnitude would decrease to 0.8 for $S$-wave detection in all the investigated area.
beyond which the quality of the fit started to degrade as compared to the previous curves. The estimated completeness threshold of the seismic catalog is 1.1, which is in good agreement with the values retrieved by our analysis (Fig. 8) in the middle of the network where the recorded seismicity of the area is mainly concentrated.

**Conclusions**

We studied the seismic noise acquired by 22 velocimetric stations of ISNet, a seismic network installed along the Apenninic chain in southern Italy. The analysis was carried out by computing the velocity power spectral density curves on 1-hr-long time windows, covering a period of almost three years.

We found that the noise levels fall between the NLNM and NHNM of Peterson (1993) for most of the stations. Discrepancies were observed for the horizontal components of broadband stations at periods larger than 10 s, where the average noise level does not decrease beyond the microseismic peak as expected and overcomes the NHNM curve at very large periods \( T > 10 \text{s} \). Such a high level is likely to be associated with such installation problems as bad insulation, air circulation, or other external forces acting on the sensor cover.

In the discussion of the noise behavior, we distinguished two period ranges. At short periods \( 0.1 \text{s} < T < 1 \text{s} \), the seismic noise shows significant diurnal variations, with maximum values generally occurring in the daylight hours and the minimum reached during the night. Because the stations are located far from the urban environment and we do not observe any significant decrease in the noise during the weekend, we believe that meteorological agents are mainly responsible for the noise variation at the time scale of days to weeks. Peaks in the noise are well correlated with peaks in the wind intensity, but additional meteorological factors are likely to contribute to the noise behavior as a function of time.

At longer periods we retrieved the microseismic peak between 5 and 9 s generated by oceanic waves. In addition we also found a second peak between 2 and 4 s, which undergoes larger short-term variations and is well correlated with the sea wave height measured along the Tyrrhenian coasts. In this band the noise increases during heavy sea conditions, and it shows clear seasonal variations. We believe that the noise is generated by the interaction of the sea near the shore, probably in regions where the swell hits against the rocky coast at an almost normal incidence. Similar results were also found by several authors in the Mediterranean area (Marzorati and Bindi, 2006; Chevrot et al., 2007; Vassallo et al., 2008; Frontera et al., 2010; Iannaccone, Vassallo, et al., 2010). Such sparse observations could indicate that the effect of the noise generated by waves in the Mediterranean Sea is slightly different from the one coming from oceanic gravity waves. Alternatively, because similar effects have been observed in other coastal zones outside the Mediterranean region (Bungum et al., 1971; Bormann, 2002), such a peak could be simply related to the complex pattern of sea waves impinging against the coast, for which the effect rapidly decreases as the distance from the coast increases. However, this latter hypothesis is not supported by the fact that the peak, located in the Mediterranean Sea, is also present in the ambient noise recorded in Bucharest, Romania.
(Groos and Ritter, 2009) several hundred kilometers from the Mediterranean coast.

We computed the detection threshold of the network and its spatial variability by comparing station by station the amplitude levels of the seismic noise with the theoretical signature of small earthquakes based on the Brune model. For the comparison, we proposed a new approach based on a condition on the SNR in the time domain, but maintaining a parameterization in the frequency domain. Convolving the effect of the small earthquake source, in the framework of the Brune model, with the anelastic attenuation, we came out with a signal-to-noise expression that is a function of the moment magnitude, of the average rheological and attenuation properties of the medium, and of the hypocentral distance. This criterion was recursively used to define for each possible hypocenter the minimum magnitude of an event that will be identified at least at five stations with an SNR larger than 5. We separately computed the detection threshold maps for P and S waves. Theoretical P waves were compared with the noise recorded on the vertical component whilst S waves were confronted with the noise on the horizontal components. The correct identification of the direct S phase is here linked to the capability of estimating the magnitude of the event, while identification of the direct P wave is associated with the possibility of locating the event through the picking of the first arrivals. It is worth noting that the correct individuation of the direct P wave does not necessarily imply that we are able to perform a correct picking of the P wave. P first arrivals could be characterized by an emergent phase owing to the focal mechanism, wave path, and local heterogeneities. However, in most cases, we expect that the magnitude order of the error on the P choice is the same as the dominant period of the P phase. This hypothesis is supported by the good agreement that we found between the completeness threshold of the ISNet seismic catalog and the one predicted by this study. A strong correlation between the P detection threshold and the minimum magnitude of the events was also found for ANSS stations in the United States, where the events were detected by the NEIC automatic picker (McNamara et al., 2004).

We found that the minimum magnitude of an event detectable by the network, wherever it occurs, is 1.3 for P and 1.1 for S waves. Space variations of the detection threshold may be as large as 0.3 and are due to a combination of the geometrical configuration of the network and the differences in the noise levels at the stations of ISNet. These maps could be additionally used to improve the quality of the data at the single stations: if all stations were to have the same noise as the lowest noise station LIO3, the S-wave detection threshold would decrease to 0.8 for all earthquakes occurring inside ISNet.

We finally investigated the variability of the completeness threshold as a function of the reduced time $t^*$, the stress drop, and the hypocentral depth. Changes in the stress drop and in the focal depth do not significantly affect the magnitude completeness of the network while changes in the $t^*$ parameter are responsible for changes in the threshold as large as 0.5 for both P and S waves, when $t^*$ ranges between 0.015 and 0.035 s. For this reason, the characterization of anelastic attenuation, here represented by $t^*$ parameter, is crucial in the determination of the detection threshold of a seismic network.

Data and Resources

Seismic data used in this study were collected by Irpinia Seismic Network (ISNet) managed by Amra Scarl (Analisi e Monitoraggio del Rischio Ambientale). The computed power spectral densities can be obtained upon request to vassallo@fisica.unina.it. The meteorological data were provided by Centro Irpino per l’Innovazione ed il Monitoraggio Ambientale (CIMA). The sea wave height data were provided by the Italian wave measurement network (Rete Ondametrica Nazionale, RON) managed by the Istituto Superiore per la Protezione e la Ricerca Ambientale (ISPRA), and the data availability is subject to registration at web site: http://www.idromare.it/ (last accessed October 2011). Spectral analysis and some of the figures were made using the Generic Mapping Tools (www.soest.hawaii.edu/gmt; last accessed October 2011).

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