Chapter 3
Seismic Noise

Abstract Seismic signals recorded by analog or digital instruments always contain noise. This can have two origins: Noise generated in the instrumentation and 'real' seismic noise from the ground. Here only the latter is described. Noise has traditionally been measured as ground displacement at different frequencies using analog seismograms and a typical ground noise at 1 Hz is in the range 1–100 nm while at 0.2 Hz it is between 10 and 10,000 nm. This measure is dependent on the filter band used. In addition, the natural noise (or microseismic noise) is dependent on frequency. This has lead to the modern way of presenting the noise as noise spectra. For practical reasons, this is calculated as acceleration noise power density spectra, which is now made routinely for most seismic stations. These spectra are a good measure of the quality of the station as well being used to spot potential problems.

The origin of the microseismic noise below 1 Hz is dominated by the wave motion in the oceans. The dominating noise is generated by the superposition of ocean waves of equal period traveling in opposite directions, thus generating standing gravity waves of half the period, typically 5–8 s. This noise is seen globally. Other sources of noise, typically above 1 Hz, are caused by water motion, wind and cultural movements, particularly traffic. The cultural noise is often the limiting factor when looking for a site for a new seismic station. Peterson noise curves give the average maximum and minimum noise spectra to be expected a the best and worst seismic stations, respectively, and are used as references to qualify instrumental noise and sites for seismic stations.

Recorded seismic signals always contain noise and it is important to be aware of both the source of the noise and how to measure it. Noise can have two origins: Noise generated in the instrumentation and 'real' seismic noise from earth vibrations. Normally, the instrument noise is well below the seismic noise although most sensors will have some frequency band where the instrumental noise is dominating (e.g. an accelerometer at low frequencies). The instrumental noise is dealt with in more detail in Chap. 2 and how to measure it in Chap. 10. So from now on in this section, it is assumed that noise is ground noise.

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3.1 Observation of Noise

All seismograms show some kind of noise when the gain is turned up and at most places in the world, harmonic-like noise (called microseismic noise) in the 0.1–1.0 Hz band is observed in the raw seismogram (Fig. 3.1), unless obscured by a high local noise level. From Fig. 3.1, it is also seen that, although the microseismic noise dominates (see noise sources later), there is also significant seismic noise in other frequency bands. So, obviously, the noise level must be specified at different frequencies.

Intuitively, the simplest way should be to measure the earth displacement in different frequency bands and plot the amplitude as a function of frequency or period. This was in fact the way it was done before the use of digital recording and an example of measurements from the old manual of Seismological Observatory Practice (Willmore 1979) is shown in Fig. 3.3.

The use of filtering and measurements in the time domain presents two problems:

(1) Bandwidth used is often an arbitrary choice, (2) Getting average values over long time intervals is not possible. Both of these problems are solved by presenting the noise in the spectral domain, (see Chap. 6 for more on spectral analysis). Figure 3.2 illustrates the problem of the filter bandwidth. The figure shows the same signal filtered with an increasingly narrow filter. The amplitude of the signal decreases as the filter becomes narrower, since less and less energy gets into the filter band. In the example in Fig. 3.2, the maximum amplitude at around 1 Hz varies from 23 to 6 nm depending on filter width. This decrease in amplitude is mainly due to making the filter narrower. However, for the widest frequency band (0.6–1.7 Hz), relatively more low frequency energy is also present in the original

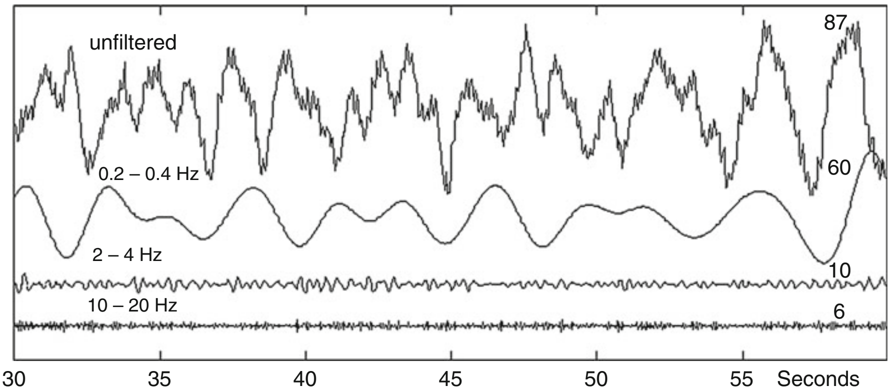

Fig. 3.1 Seismic noise in different filter bands at station MOL in the Norwegian National Seismic Network The short period station (1 Hz) is situated about 40 km from the North Sea and the unfiltered trace clearly shows the high level of low frequency noise (~0.3 Hz) generated by the sea. All traces are plotted with the same scale and the numbers to the *right above* the traces are the maximum amplitudes in counts

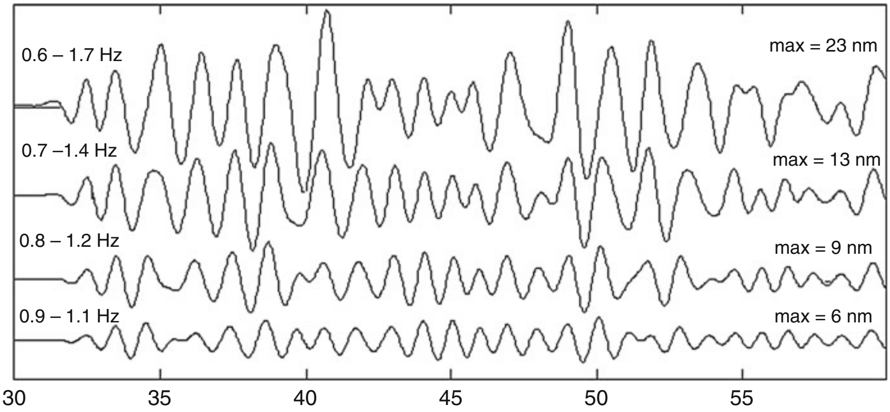

Fig. 3.2 The signal from Fig. 3.1 band pass filtered with different filter widths. The signals have been corrected for instrument response to show displacement. The maximum amplitude in nm is shown to the *right on top* of the traces

signal. Comparing the amplitudes to Fig. 3.3, the noise level can be considered worse or better than average depending on which filter band is used. So in order to do measurements in time domain, the noise values can only be compared if the same bandwidth of the filter is used.

3.2 Noise Spectra

With digital data, it is possible to make spectral analysis, and thereby easily get the noise level at all frequencies in one simple operation. It has become the convention to represent the noise spectra as the noise power density acceleration spectrum $P_a(\omega)$, commonly in units of dB referred to 1 $(\mathrm{m/s^2})^2/\mathrm{Hz}$. Noise Level is thus calculated as

$$\text{Noise Level} = 10 \log \left[\mathrm{P_a}(\omega)/\left(\mathrm{m/s^2}\right)^2/\mathrm{Hz}\right] \tag{3.1}$$

Figure 3.4 shows the new global high (NHNM) and low noise models (NLNM) (Peterson 1993) and an example of noise spectra at a seismic station. The curves represent upper and lower bounds of a cumulative compilation of representative ground acceleration power spectral densities determined for noisy and quiet periods at 75 worldwide distributed digital stations. These so-called Peterson curves have become the standard, by which the noise level at seismic stations is evaluated. *A power density spectrum can be defined in different ways and it is important that the definition used is identical to the one used originally by Peterson (Fig. 3.4) in order to compare the spectra. The exact definition is given in Chap. 6.*

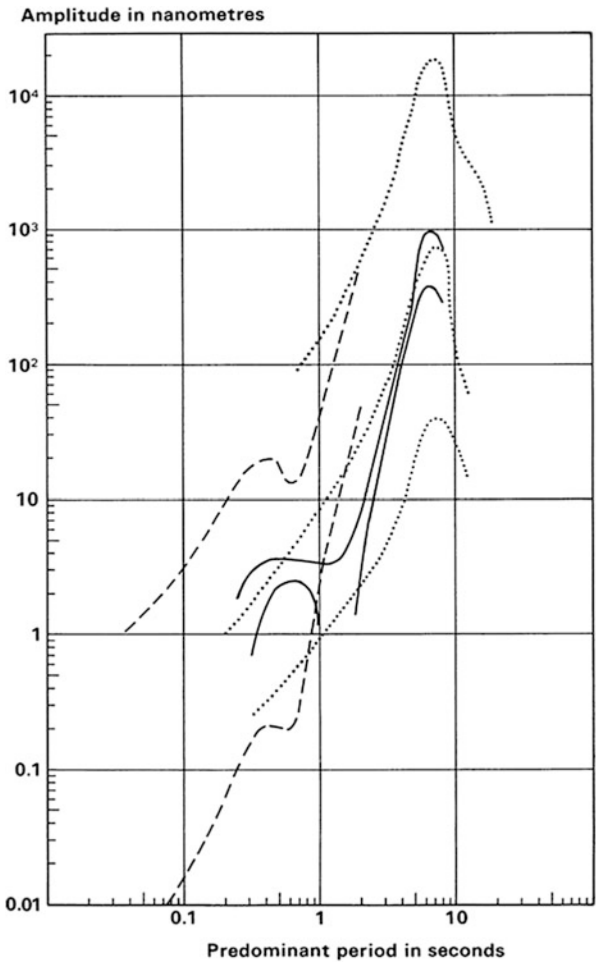

Fig. 3.3 Noise curves in a rural environment. The three *dotted lines* correspond to the maximum, mean and minimum levels published by Brune and Oliver (1959), the dashed lines give two extreme examples observed in the US and the full line curves give the limits of fluctuation of seismic noise at a European station on bedrock in a populated area 15 km away from heavy traffic (Figure from Willmore 1979)

3.3 Relating Power Spectra to Amplitude Measurements

The Peterson noise curves and the way of representing them have standardized the way of representing seismic noise. However, looking at such a curve, it is difficult to relate them to something physical as seen in Figs. 3.1, 3.2, and 3.3 and a standard question is often: So what is the physical meaning of the Peterson curve? The old noise curve in Fig. 3.3 can be directly related to the seismogram in Fig. 3.2, while the Peterson curve cannot since one is a frequency domain measure and the other a time domain measure. However, as will be shown below, under certain conditions it is actually possible to go from one to the other (the following text largely follows Bormann in Chapter 4 of the NMSOP, Bormann 2012).

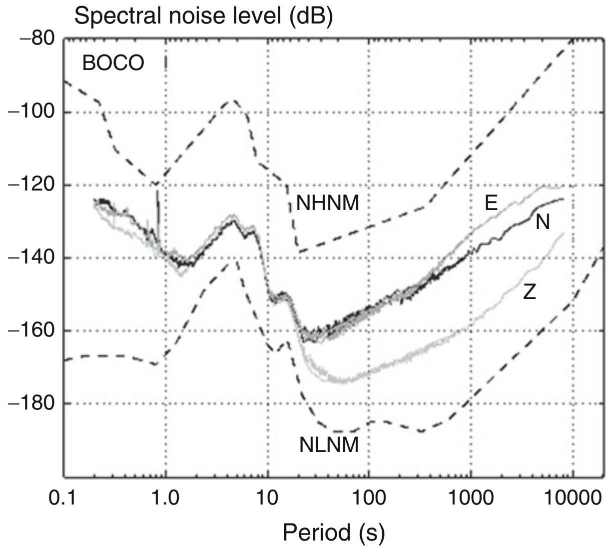

Fig. 3.4 The Peterson noise curves and noise spectral level for the IRIS station BOCO. The noise level is in dB relative to 1 $(ms^{-2})^2$/Hz. The Peterson high and low noise models are shown with *dashed lines*. The noise spectra are shown for all three components. Note the lower noise level for the vertical (Z) component (Figure modified from FDSN station book found at www.fdsn.org/station_book)

The problem is how to relate a spectral amplitude at a given frequency to a time domain amplitude in a given frequency band. The root mean squared amplitude, a_{RMS} of a signal in the time interval *0-T*, is defined as

$$a_{RMS}^2 = \frac{1}{T}\int_0^T a(t)^2 dt \tag{3.2}$$

The average power of the signal in the time interval is then equal to a_{RMS}^2. The average power can also be calculated (Parseval's Theorem) from the power density spectrum as

$$a_{RMS}^2 = \int_{f_1}^{f_2} P(\omega)df \approx P \cdot (f_2 - f_1) \tag{3.3}$$

under the assumption that the power spectrum is nearly a constant P in the frequency range f_1 to f_2, which is not unreasonable if the filter is narrow. In the general case, P would represent the average value of $P(\omega)$ in this frequency band. It is important that $P(\omega)$ is the normalized power spectral density as defined in Chap. 6 and that the power represents the total contribution at ω from both positive and negative frequencies. If, as is the usual practice, the power spectrum is calculated using only the positive frequencies as defined in the standard complex spectral analysis, a^2_{RMS} will have to be calculated as $2P(f_2 - f_1)$. *The power values given by* *the New Global Noise Model by Peterson (1993) do already contain this factor of two, or said in other words, represent the total power.*

Under these assumptions, we then have a relationship between the power spectral density and the RMS amplitude within a narrow frequency band:

$$a_{RMS} = \sqrt{P \cdot (f_2 - f_1)} \tag{3.4}$$

There is thus a simple way of relating the power spectral densities to amplitudes, as seen on a seismogram; however, note that relation (3.4) gives the RMS amplitude. There is statistically a 95 % probability that the instantaneous peak amplitude of a random wavelet with Gaussian amplitude distribution lie within a range of $2a_{RMS}$. Peterson (1993) showed that both broadband and long period noise amplitudes closely follow a Gaussian probability distribution. In the case of narrowband-filtered envelopes, the average peak amplitudes are 1.25 a_{RMS} (Bormann, Chapter 4 in NMSOP, Bormann 2012). From measurements of noise using narrowband filtered broadband data, values of 1.19–1.28 were found (Peterson 1993). Thus, in order to get the true average peak amplitude on the seismogram, a factor of about 1.25 can be used. (NB: For a pure sine wave, $a = a_{RMS}\sqrt{2}$, not so very different). We can now set up the relation between the power spectral values and the average peak amplitudes

$$a = a_{RMS} \cdot 1.25 = 1.25\sqrt{P \cdot (f_2 - f_1)} \tag{3.5}$$

The frequency band depends on instrument (mainly if analog), while for digital data the user can select the filter. A common way of specifying filter bands is to use the term octave filter. An n-octave filter has filter limits f_1 and f_2 such that

$$\frac{f_2}{f_1} = 2^n \tag{3.6}$$

For example, a half-octave filter has the limits f_1 and $f_2 = f_1 \cdot 2^{1/2}$ (e.g. 1–1.41 Hz). Many of the classical analog seismographs have bandwidths of 1–3 octaves and digital seismographs might have a bandwidth of 6–12 octaves. However, the signal bandwidth of many dominating components of seismic background noise might be less than 1 octave. Thus, the frequency of measurement is really a frequency range. However, for practical reasons, the average frequency will be used to represent the measurement. For the average frequency, the geometric center frequency f_0 must be used

$$f_0 = \sqrt{f_1 f_2} = \sqrt{f_1 f_1 \cdot 2^n} = f_1 \cdot 2^{n/2} \tag{3.7}$$

and

$$f_1 = f_0 2^{-n/2} \text{ and } f_2 = f_0 2^{n/2} \tag{3.8}$$

For narrow filters, the geometric center frequency is almost the same as the average frequency. The filters used in Fig. 3.2 have average frequencies of 1.2, 1.1, 1.0 and

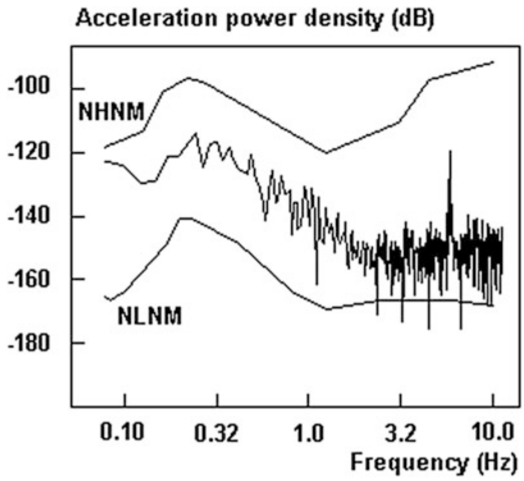

Fig. 3.5 Noise power density spectrum of the raw signal seen in Fig. 3.1, *top trace*. The acceleration power is in dB relative to 1 $(\mathrm{m/s^2})^2/\mathrm{Hz}$. The limits of the Peterson noise model are indicated (NHNM and NLNM). The spectrum has not been smoothed

1.0 Hz while the geometric center frequencies are all 1.0. For the filter 10–20 Hz, the geometric and average frequencies are 14 and 15 Hz, respectively.

In comparing time and frequency domain signals, there was no mention of units since this has no importance on the relations. However, the Peterson curves are in acceleration, so unless the time domain signal is in acceleration too, the power spectrum must first be transformed to acceleration. The most common unit for the original seismogram is velocity, sometimes acceleration and rarely displacement. If the power spectra of acceleration, velocity and displacement are called P_a, P_v, and P_d respectively, the relations are:

$$P_v(\omega) = P_d(\omega) \cdot \omega^2 \tag{3.9}$$

$$P_a(\omega) = P_v(\omega) \cdot \omega^2 = P_d(\omega) \cdot \omega^4 \tag{3.10}$$

Now, some examples. Figure 3.2 has a peak amplitude of 13 nm for the frequency band 0.7–1.4 Hz. This can be converted to P_d

$$P_d = \left(13 \cdot 10^{-9}/1.25\right)^2/(1.4-0.7) = 1.55 \cdot 10^{-16}\mathrm{m}^2/\mathrm{Hz} \tag{3.11}$$

The center frequency is $\sqrt{0.7 \times 1.4} = 1.0\,\mathrm{Hz}$ and the acceleration power density is

$$P_a = 1.55 \cdot 10^{-16} \quad 16 \cdot \pi^4 \cdot 1^4 = 2.4 \cdot 10^{-13}\left(\mathrm{m/s^2}\right)^2/\mathrm{Hz} \tag{3.12}$$

or -126 dB relative to 1 $(\mathrm{m/s^2})^2/\mathrm{Hz}$. Figure 3.5 shows the complete noise power density spectrum for the time window used in Figs. 3.1 and 3.2. We can see that the level (-126 dB) is high compared to the spectral level which is around -136 dB. How can that be explained? We have to remember that (3.5) is based on the assumption that the amplitude is the *average peak amplitude,* while what has

Table 3.1 Maximum amplitudes of signals (Fig. 3.2) in the different filter bands and corresponding noise levels

Filter bands (Hz)	Amplitude (nm)	Noise power level (dB)
0.6–1.7	23	–123
0.7–1.4	13	–126
0.8–1.2	9	–127
0.9–1.1	6	–127

been used is the largest amplitude in the window. If the average peak amplitude is 3 times smaller than the maximum amplitude, the level would be corrected by $-10 \log(3^2) = -9.5$ dB and the time domain and frequency domain measures would be similar.

Assuming a one octave filter, an approximate relationship can be calculated (Havskov and Ottemöller 2010) between the noise power density *N(dB)* given in dB (Fig. 3.4) and the ground displacement *d* in m

$$d = \left(1/38f^{1.5}\right)10^{N(dB)/20} \tag{3.13}$$

or going the other way

$$N(dB) = 20log(d) + 30log(f) + 32 \tag{3.14}$$

where *f* is the average frequency of the filter.

This discussion should demonstrate that, by far, the most objective way to report the noise level at a given site is to use the power density spectrum, although it is nice to be able to relate it, at least approximately, to some amplitude measure.

The spectrum in Fig. 3.5 shows a relatively high noise level at lower frequencies relative to high frequencies. This is not surprising, considering the general high microseismic noise level along the Norwegian West coast. The noise level above 3 Hz is quite low since the station is on granite in a rural area (20 m from nearest house).

The noise spectral level can be calculated for all four cases in Fig. 3.2, see Table 3.1. It is seen that, although the amplitudes were quite different in the four filter bands, power spectral levels are nearly equal except for the filter band 0.6–1.7 Hz, where the amplitudes in the relatively wide filter is influenced by the stronger background noise at lower frequencies.

From Figs. 3.4 or 3.5, we see that the NLNM has a level of -166 dB at 1 Hz. Assuming a 2-octave filter and using (3.5) and (3.8), the corresponding average peak displacement is 0.3 nm. From Fig. 3.3 it is seen that the lowest displacement is 1 nm at 1 Hz so there is a reasonable agreement considering the uncertainty of whether RMS values or average peak values have been used. At 10 Hz, the NLNM gives 0.01 nm, which also agrees well with the values on Fig. 3.3. So a rule of thumb (and easy to remember) is: *A peak displacement of 1 nm at 1 Hz means a good site in terms of ambient noise*.

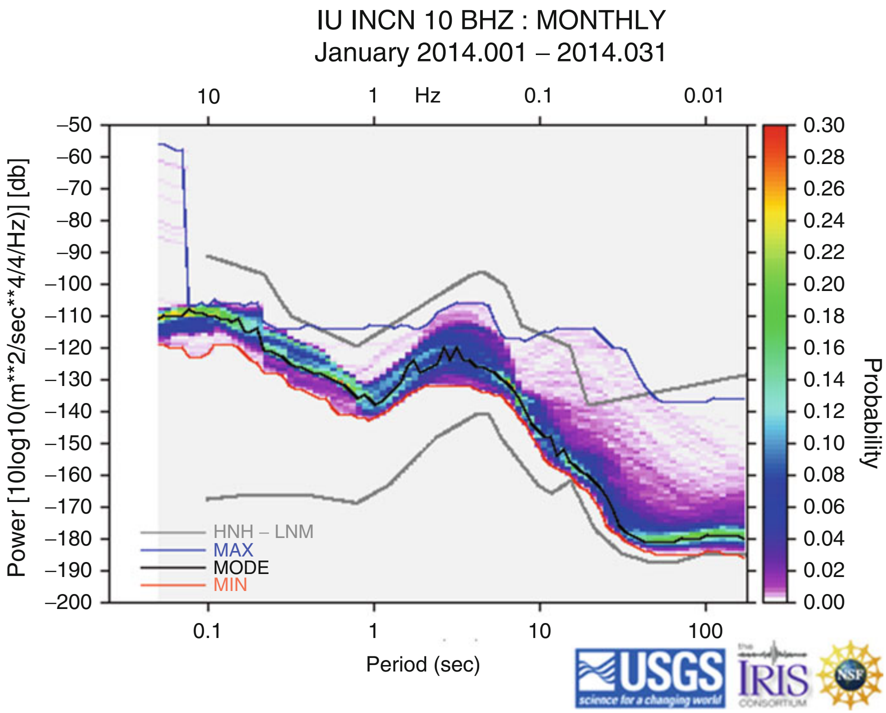

Fig. 3.6 PSD function of seismic noise recorded at station INCN, vertical component. The analysis is made with 1 month of data using overlapping time windows. HNM and LNM are the high and low noise models, respectively (The figure is from www.iris.edu/servlet/quackquery)

One point to be underlined is that noise is of random nature (any "predictable" disturbance is not noise as defined here!). The power spectrum estimated by the simple Fourier transform of a sample window of raw data is also random. The standard error of such an estimate is very high (see, e.g. Blackman and Tukey 1966; Press et al. 1995). Actually, when we express the noise level as the power spectrum density or RMS within a certain frequency band, we are implicitly assuming that noise is a stationary process, which means that its statistical characteristics are not time-dependent or at least vary slowly enough to be considered constant within a certain time interval. Therefore, if a reliable estimation of the noise level, at a given site, is needed, an average of power density estimation on several overlapping sample time-windows or some spectral smoothing will decrease the estimate variance (e.g. Blackman and Tukey 1966).

It is also useful to look at the distribution of the noise levels as function of frequency, which can be done by computing probabilities of the observations. A method for this has been suggested by McNamara and Buland (2004) in which probability density functions (PDF) are computed from the power spectral density (PSD) data. This approach was implemented in the PQLX software (McNamara and Boaz 2005) (earthquake.usgs.gov/research/software/pqlx.php) that is now widely used (Fig. 3.6).

3.4 Origin of Seismic Noise

Man Made Noise This is often referred to as "cultural" noise. It originates from traffic and machinery, has high frequencies (>2–4 Hz) and die out rather quickly (m to km), when moving away from the noise source. It propagates mainly as high-frequency surface waves, which attenuate fast with distance and decrease strongly in amplitude with depth, so it may become almost negligible in boreholes, deep caves or tunnel sites. This kind of noise usually has a large difference between day and night and can have characteristic frequencies depending on the source of the disturbance. The noise level can be very high. Figure 3.7 shows the spectrogram of 10 days of a seismic record, including two distant earthquakes, from an urban seismic station, where the daily cycle of human activity is clearly seen.

Wind Noise Wind will make any object move so it will always generate ground noise. It is usually high frequency like man-made noise; however large swinging objects like masts and towers can generate lower frequency signals. Trees also transmit wind vibrations to ground and therefore seismic stations should be installed away from them. In general, wind turbulences around topography irregularities such as scarps or rocks generate local noise and their proximity must be avoided.

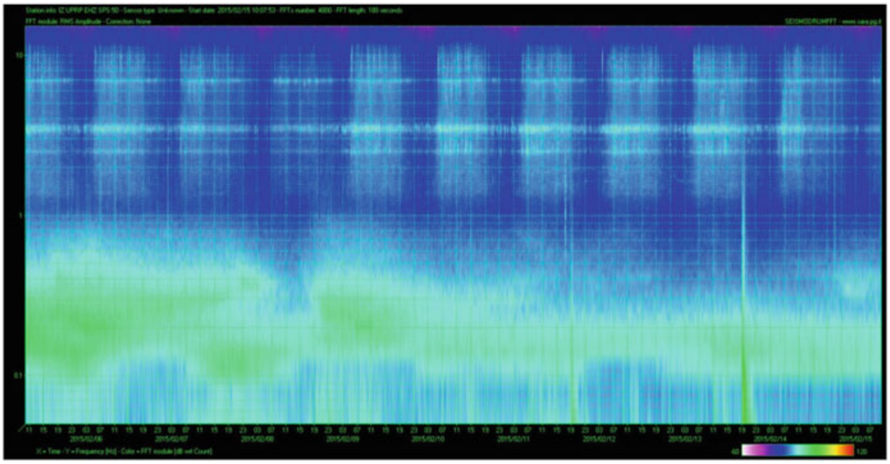

Fig. 3.7 Spectrogram of a horizontal component of a seismic record from an urban station, spanning 10 days, from Feb 05 (Thursday) to Feb 15 (Sunday) of 2015. The horizontal axis is time and the vertical axis is frequency. The *colors* represent the spectral amplitude in dB relative to 1 count/Hz (color scale on the *bottom-right*). A 10 Hz low-pass filter has been applied. Note the cycle of human-generated noise every day and the lower level on the week ends (Feb 7–8 and 14–15). A variable microseismic noise under 0.5 Hz and probable resonant modes at 2.5, 3.5 and 7 Hz –likely corresponding to vibration modes of a building or some structure- are also visible. Two teleseisms appear on Feb 11 and Feb 13, both at 19 h, with a characteristic wide-band spectral content from low frequencies (Figure courtesy of Mauro Mariotti)

Ocean Generated Noise This is the most widespread noise (called microseisms or microseismic noise), and it is seen globally, although the interior of continents has less noise than coastal regions. Long period ocean microseisms are generated only in shallow waters in coastal regions (NMSOP, Bormann 2012), where the wave energy is converted directly into seismic energy either through vertical pressure variations, or smashing surf on the shores. They therefore have the same period as the water waves ($T \approx 10$–16 s). Shorter period microseisms can be explained as being generated by the superposition of ocean waves of equal period traveling in opposite directions, thus generating standing gravity waves of half the period. These standing waves cause perturbations which propagate without attenuation to the ocean bottom. The higher frequency microseisms have larger amplitude than the lower frequency microseisms (Figs. 3.3, 3.4, and 3.5). During large storms, the amplitudes can reach 20,000 nm at stations near the coast and make analog seismograms useless.

Other Sources Running water, surf and volcanic tremor (an almost harmonic noise associated to fluids motion, often lasting hours or days) are other local sources of seismic noise. Man made noise and wind noise are usually the main source at high frequencies and since the lower limit is about 0.01 nm at 10 Hz, very small disturbances will quickly get the noise level above this value. For more details on seismic noise generation, see NMSOP, Chapter 4, Bormann (2012).

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