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The nature of noise wavefield and its applications for site effects studies
A literature review

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Abstract

The aim of this paper is to discuss the existing scientific literature in order to gather all the available information dealing with the origin and the nature of the ambient seismic noise wavefield. This issue is essential as the use of seismic noise is more and more popular for seismic hazard purposes with a growing number of processing techniques based on the assumption that the noise wavefield is predominantly consisting of fundamental mode Rayleigh waves. This survey reveals an overall agreement about the origin of seismic noise and its frequency dependence. At frequencies higher than 1 Hz, seismic noise systematically exhibits daily and weekly variations linked to human activities, whereas at lower frequencies (between 0.005 and 0.3 Hz) the variation of seismic noise is correlated to natural activities (oceanic, meteorological...). Such a surface origin clearly supports the interpretation of seismic noise wavefield consisting primarily of surface waves. However, the further, very common (though hidden) assumption according which almost all the noise energy would be carried by fundamental mode Rayleigh waves is not supported by the few available data: no "average" number can though be given concerning the actual proportion between surface and body waves, Love and Rayleigh waves (horizontal components), fundamental and higher modes (vertical components), since the few available investigations report a significant variability, which might be related with site conditions and noise source properties.

Keywords: seismic noise wavefield; microseisms; microtremors; surface waves; seismic hazard

1. Introduction

Over the last two decades, many cities have grown considerably and demographists expect a similar trend for at least the two next decades. These urban areas are very often built on soft sediments, and a large number is unfortunately located in seismic areas, emphasizing the need for a careful and reliable assessment of site amplification phenomena. This issue has been addressed for a long time by scientists and engineers who developed many techniques to identify the main characteristics of site responses for soft deposits (i.e., resonance frequencies and amplification factor). These techniques may be grouped into three main categories. The first is based on a numerical simulation approach (see Panza et al., 2001 for a review of numerical simulation methods), and is coupling with classical geophysical and geotechnical tools (such as seismic refraction, seismic reflection, boreholes, penetrometers, etc.) in order to provide reliable estimates of the required input parameters (sediment depth, S and P wave

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velocities, etc.). However, many such classical geophysical tools suffer severe limitations in urbanized areas, mainly because of their cost (seemingly prohibitive especially in developing countries which face many other priorities), and their environmental impact that is less and less accepted by the community (use of explosives, drillings, etc.). The second category of technique consists in directly measuring the site response on the basis of earthquake recordings on specific stations located on carefully chosen sites. Although this technique provides an unbiased experimental estimation of the site transfer amplification factor, its use in areas of low to moderate seismicity is limited by the time required to gather a significant number of recordings with satisfactory signal to noise ratio. Finally, the latest category of methods, based on ambient noise recordings, became more and more popular over the last decades as it offers a convenient, practical and low cost tool to be used in urbanized areas. Two techniques are predominantly used to determine site response parameters: the simple horizontal to vertical Fourier amplitude spectral ratio (HVSR), and the more advanced array technique.

The ability of the HVSR technique to provide a reliable information related to site response has been repeatedly shown in the past (Nakamura, 1989; Lachet and Bard, 1994; Kudo, 1995; Bard, 1998). However, its theoretical basis is still unclear as two opposite explanations have been proposed. Nakamura (1989, 2000) claims that the horizontal to vertical spectral ratio mainly reflects the S-wave resonance in soft surface layer (removing effects of surface waves), and hence that HVSR curves provide a consistent estimate of the site amplification function. This "body wave" interpretation has been contradicted in several papers highlighting the relationship between the HVSR and the ellipticity of fundamental mode Rayleigh waves (Lachet and Bard, 1994; Kudo, 1995; Bard, 1998), and thus seriously questioning the existence of any simple direct correlation between HVSR peak value and the actual site amplification factor (the HVSR peak would then be associated with the vanishing of the Rayleigh waves vertical component, instead of amplification of S-wave on horizontal components). This brief summary about the two hypothetical origins of the HVSR peak shows the close link between the composition of the seismic noise wavefield (body or surface waves) and the interpretation of the HVSR curve.

To go one step further, when one assumes that the seismic noise wavefield is mainly constituted by surface waves, then new questions arise concerning the type of surface wave (Rayleigh or Love waves) and propagation mode (fundamental or higher modes). Answering these questions is very important for the processing and interpretation of array microtremor recordings, which mainly focus on the derivation of surface waves dispersion curve. For instance, processing of horizontal components should definitively take into account the simultaneous existence of both Rayleigh and Love waves (Arai and Tokimatsu, 1998, 2000; Yamamoto, 2000; Bonnefoy-Claudet, 2004), while improper identification of surface waves type (Rayleigh or Love) or order (fundamental or harmonics) might severely bias the inversion of S-wave velocity profile (Tokimatsu et al., 1992; Tokimatsu, 1997; Beatty et al., 2002; Zhang and Chan, 2003; Bonnefoy-Claudet, 2004; Wathelet, 2005).

Despite the lack of theoretical agreement about the nature of the ambient seismic noise wavefield, the number of site-specific studies based on noise recordings has increased dramatically in recent years. Time has come thus for a clear assessment of the HVSR and array methods, in order to better identify their actual possibilities and limitations. This was one of the main goals of the *SESAME* European project (Site EffectS using AMbient Excitations, EESD project n° EVG1-CT-2000-00026). Within this framework, a wide set of numerical and experimental researches work has been conducted over the last 4 years to investigate the nature of the ambient noise wavefield in the frequency range of interest for site effects estimation purposes (i.e., from 0.2 to 10 Hz). The first step of this *SESAME* project has been to update a survey of the scientific literature dealing with seismic noise, in order to establish a state of the art about the knowledge of the nature of the ambient noise wavefield. The present paper is devoted to a presentation of the main outcomes of this comprehensive survey. After reviewing the evolution of concerns about seismic noise over the year (Section 2) we will focus on its origin (Section 3) and the composition of the noise wavefield (Section 4), with special attention to the respective proportion between a) surface and body waves, b) Rayleigh and Love waves, c) fundamental and higher modes.

2. The use of seismic noise: main historical periods

Seismic noise is observed very early on, from the very beginning of instrumental seismology in the nineteenth century. A pendulum led Bertelli (1872), in Italy, to observe that it was continuously moving for hours or days with regional weather conditions. He noticed a correlation between the long period noise and disturbed air pressure. Since this date many studies about seismic noise have been carried out. We can distinguish three predominant time periods in subsequent studies of ambient ground motion.

2.1. Before 1950

Until the middle of the twentieth century, studies were more qualitative than quantitative: progress in knowledge was limited by the frontiers of instrumental and processing techniques. The extensive work carried out by Gutenberg (1911) into the nature and the origin of microseisms is probably the first major review concerning seismic noise. Gutenberg (1958) quoted a bibliography containing 600 references relating to microseisms. Unfortunately, most of these references are in Russian, German, Italian, etc. and are published in local scientific journals.

Following the pioneer work of Gutenberg (1911, 1958) some authors highlighted the relations between microseisms, meteorological conditions and oceanic waves (Bonnefoy-Claudet et al., 2004). Banerji (1924, 1925) observed microseisms associated with Indian monsoons in south Asia, and suggested that they were due to Rayleigh waves generated at the bottom of the sea by the train of water waves maintained by the monsoon currents. Bernard (1941a,b) and Longuet-Higgins (1950) showed the relation between the microseism periods and oceanic swells (the predominant period of microseisms is equal to half the natural period of swell height).

2.2. From 1950 to 1970

During the 1950–1970, the expansion of seismology and technical improvements in equipment allowed significant advances to be made in the understanding of noise phenomena. Several authors felt the interest of using seismic noise for different applications, and investigated its origin and nature. Several techniques based on noise recordings were then developed. Array techniques, originally developed to detect and localize nuclear explosions, were adapted by seismologists to derive the surface wave dispersion curve from ambient noise array measurements, in view of inverting the soil shear wave velocity profiles. Two main noise based array techniques were developed: the spatial auto-correlation analysis of signal (*SPAC*) (Aki, 1957, 1965), and the frequency–wave number analysis (*f–k*) (Capon et al., 1967; Capon, 1969; Lacoss et al., 1969).

Other methods were also used to investigate the noise wavefield, such as particle motion analysis (Toksöz, 1964), or borehole techniques sometimes coupled with array analysis (Douze, 1964; Gupta, 1965; Douze, 1967).

The emergence of these techniques has been useful for improving understanding of the origin of the noise (oceanic, meteorological, human, etc.) and the nature of the noise wavefield nature, as discuss below.

2.3. From 1970 up to now

Since the 1970 up to now, the number of publications dealing with seismic noise has accelerated dramatically. It is not easy to consult all the publications (especially those written in Japanese), but we estimate their number to exceed 500. Some of them are devoted to the nature of the noise wavefield, but the overwhelming majority (about 95%) deal with the applicability of seismic noise, and/or their direct applications to some specific case studies.

Analyzing seismic noise for seismic city microzonation is probably one of the most important and noteworthy applications. Two major techniques have been developed for this purpose: the site-to-reference spectral ratio and the HVSR ratio (the spectral ratio between horizontal and vertical components). The HVSR ratio technique was proposed first by Nogoshi and Igarashi (1971), and then strongly emphasized by Nakamura (1989, 1996, 2000). Since that date, many authors have published papers about the use of the HVSR ratio for microzonation purposes (Ansary et al., 1995; Field et al., 1995; Gaull et al., 1995; Theodulidis and Bard, 1995; Abeki et al., 1996; Konno, 1996; Teves-Costa et al., 1996; Wakamatsu and Yasui, 1996; Alfaro et al., 1997; Fäh, 1997; Abeki et al., 1998; Bour et al., 1998; Duval et al., 1998; Guéguen et al., 1998; Ishida et al., 1998; Konno and Ohmachi, 1998; Mucciarelli, 1998; Ogawa et al., 1998; Ibs-Von Seht and Wohlenberg, 1999; Al Yuncha and Luzon, 2000; Maruyama et al., 2000; Tobita et al., 2000; Alfaro et al., 2001; Ansal et al., 2001; Bindi et al., 2001; Duval et al., 2001a,b; Giampiccolo et al., 2001; Lebrun et al., 2001; Lombardo et al., 2001; Delgado et al., 2002; Huang et al., 2002; Parolai et al., 2002; Cara et al., 2003; Maresca et al., 2003; Uebayashi, 2003; Al Yuncha et al., 2004; Nguyen et al., 2004; Talhaoui et al., 2004; Tuladhar et al., 2004; Dikmen and Mirzaoglu, 2005; Panou et al., 2005; Raptakis et al., 2005). In most of these studies the authors assume that the nature and origin of the microtremor are known, and they do not discuss this issue. It is usually assumed that a microtremor consists of surface waves without any further discussion about the nature of the noise wavefield. Although a few authors (Lermo and Chavez-Garcia, 1993; Lachet and Bard, 1994; Kudo, 1995; Delgado et al., 2000; Luzon et al., 2001;

Rodriguez and Midorikawa, 2003; Al Yuncha et al., 2004) have attempted to find qualitative explanations for Nakamura's technique, most authors assume that the method's basic assumptions are right.

Another application of noise background vibration is the use of the array technique. By inverting the Rayleigh waves dispersion curve derived from measurement, array processing leads to subsurface S-wave velocity profiles. Applications are mainly dedicated to site hazard studies and geotechnical engineering purpose. Such array studies started in the late 1950, but improvements in computers and instrumental techniques (3-component seismometers, signal processing of numeric data) in the last three decades have resulted in a drastic increase in the quantity and quality of array recording (Asten and Henstridge, 1984; Bache et al., 1986; Tokimatsu et al., 1992; Malagnini et al., 1993; Arai et al., 1996; Horike, 1996; Kagawa, 1996; Milana et al., 1996; Miyakoshi and Okada, 1996; Tokimatsu et al., 1996; Friedrich et al., 1998; Miyakoshi et al., 1998; Maresca et al., 1999; Scherbaum et al., 1999; Kanno et al., 2000; Liu et al., 2000; Bettig et al., 2001; Louie, 2001; Satoh et al., 2001; Kudo et al., 2002; Ohori et al., 2002; Flores Estrella and Aguirre Gonzalez, 2003; Scherbaum et al., 2003; Morikawa et al., 2004; Roberts and Asten, 2004; Chavez-Garcia et al., 2005; Roberts and Asten, 2005; Yamanaka et al., 2005).

In the late 1990 a 'hybrid' technique based on noise ambient vibrations was pointed out. This technique consists in inverting the HVSR ratio curve, interpreted as a proxy to the Rayleigh wave ellipticity curve, in order to obtain the S-wave velocity profile estimate (Tokimatsu et al., 1998; Fäh et al., 2001; Arai and Tokimatsu, 2004). Recently Arai and Tokimatsu (2005) and Parolai et al. (2005) propose new techniques to derive the S-wave velocity profile by joint inversion of dispersion curve and HVSR. Wathelet et al. (2005) show how to get the S-wave velocity profile by direct inversion of spatial auto-correlation curves.

Fig. 1 summarizes the development of ambient noise literature. While the number of papers devoted to microzonation and array processing is increasing every year, the emphasis on understanding the nature of the noise wavefield has decreased dramatically. This phenomenon would be logical if the physical basis of the noise-based methods was known. The following sections will stress that this is not the case.

Note that current applications of seismic noise go much beyond earthquake hazard studies. Lognonné et al. (1998), Suda et al. (1998), Tanimoto (1999), Nishida et al. (2002), Rhie and Romanowicz (2004), Kedar and Weeb (2005), Stehly et al. (in press) and Tanimoto (2005) investigate in-depth the structure of noise wavefield for very low frequency (lower than 10 mHz). However this frequency range is outside the scope of the present paper dealing with 0.1–10 Hz. Recently Shapiro and Campillo (2004), Sabra et al. (2005a,b) and Shapiro et al. (2005) have also attempted to give an estimate of Green's functions for a crustal structure, applying a spatial correlation method on very long noise windows. This method, based on the multiple scattering theory, showed that it is possible to investigate the crustal structure with very low-frequency microseisms. Moreover Larose et al. (2005) show that

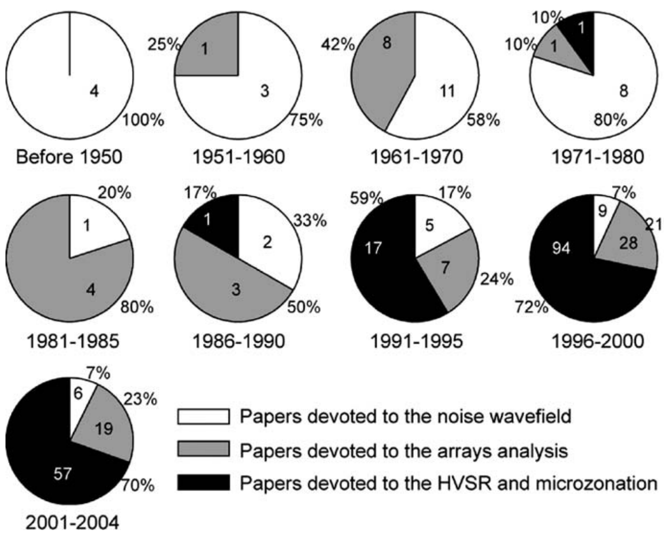

Fig. 1. Growth in the number of papers (in terms of percentage and number), from 1911 to 2004, devoted to the nature of noise wavefield analysis (white), and devoted to noise-based methods (array analysis (grey) and H/V (black)).

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the extraction of the Rayleigh wave Green function by cross-correlating seismic noise can be extended to extraterrestrial planets, with an example for the Moon.

The use of ambient vibrations for civil engineering applications is also widespread in some scientific and technical communities (wind and vibration engineering) (Cunha and Caetano, 2005), and receives a renewed interest for earthquake engineering applications especially after the work done by Trifunac (1970) concerning building ambient vibrations. More details on this latter application can be found in Dunand (2005).

3. Origin of the seismic noise

Noise is the generic term used to denote ambient vibrations of the ground caused by sources such as tide, water waves striking the coast, turbulent wind, effects of wind on trees or buildings, industrial machinery, cars and trains, or human footsteps, etc. Clearly classifying all noise sources is not an easy task. Gutenberg (1958) established a list of the different types of sources according to their frequency. Asten (1978) and Asten and Henstridge (1984) reached the same conclusions in a noise review (Table 1). These analyses show that noise has basically two different origins: natural or cultural, and differ in frequency content. This difference led these authors to distinguish between microseisms and microtremors, corresponding respectively to natural and cultural sources, and relatively low and high frequency. Based on the summaries of Gutenberg (1958) and Asten (1978) we may conclude, as a first approximation, that at low frequency (below 1 Hz) the sources are natural (ocean, large-scale meteorological conditions); at intermediate frequency (1 to 5 Hz) the sources are either natural (local meteorological conditions) or cultural (urban); and at higher frequencies the sources are essentially cultural. At this stage a warning must be issued that the term microtremor largely overlaps the

Table 1
Summary of ambient noise sources according to frequency

	Gutenberg (1958)	Asten (1978, 1984)
Oceanic waves striking along the coasts	0.05–0.1 Hz	0.5–1.2 Hz
Monsoon/Large scale meteorological perturbations	0.1–0.25 Hz	0.16–0.5 Hz
Cyclones over the oceans	0.3–1 Hz	0.5–3 Hz
Local scale meteorological conditions	1.4–5 Hz	
Volcanic tremor	2–10 Hz	
Urban	1–100 Hz	1.4–30 Hz

Summary established after studies by Gutenberg (1958), Asten (1978) and Asten and Henstridge (1984) studies.

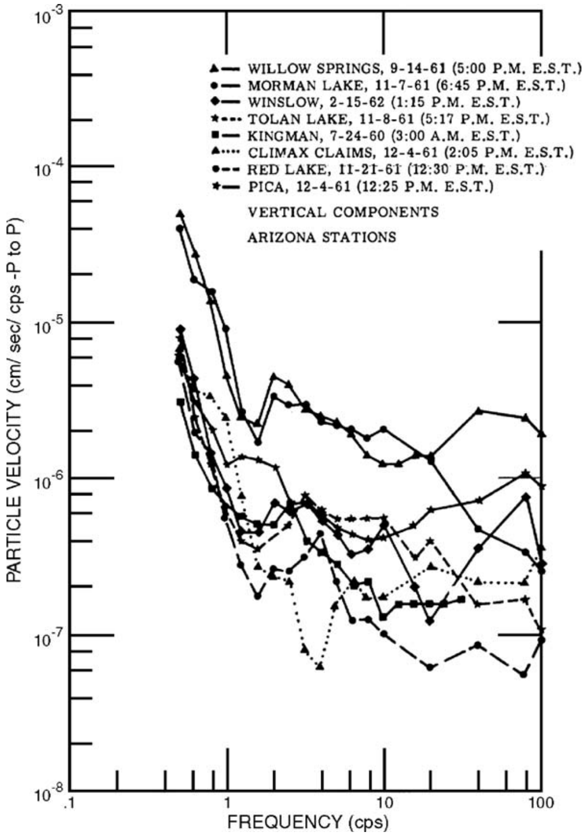

Fig. 2. The vertical component of noise particle velocities recorded at 8 distinct sites located in Arizona after Frantti (1963), show the typical seismic noise level between 0.6 and 100 Hz.

term microseism in most of the recent publications. One has to be aware that both terms may have the same meaning in other work. Here we will strictly follow the generic names as defined previously.

Extensive investigations performed by Frantti et al. (1962) and Frantti (1963) suggest a change in noise behavior around 1 Hz. These authors measure the ground particle velocity of seismic noise at 48 sites in the United States and other countries. These noise measurements are performed at sites with different geology (rock or sediment), different geographical locations (close to the ocean, mountains or cities), and different times (varying hours and seasons). For processing purposes, a two-minute recording interval was selected, which was free from any known disturbances nearby. The vertical component of the noise particle velocities observed on only 8 different sites in Arizona (Frantti, 1963) is plotted here in Fig. 2. A consistent decrease in seismic noise level from 0.6 up to 1 Hz can be observed. At higher frequency the noise levels show relatively stationary trends over frequencies. While the absolute noise level

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exhibits large changes from site to site, the general trend showed in Fig. 2 is similar for all sites presented in Frantti et al. (1962) and Frantti (1963). Peterson (1983), Stutzmann et al. (2000), Okada (2003), Berger and Davis (2004) and McNamara and Buland (2004) show similar results in the frequency band of interest for site effect studies. However, their works are extended toward lower frequencies (down to 1 mHz in some cases).

Yamanaka et al. (1993) performed continuous noise measurements on the campus of the University of Southern California in Los Angeles. Fig. 3 displays the time-varying characteristics of the spectral amplitude at periods of 0.3 s and 6.5 s (the geometric means of the two horizontal spectra are shown). Also shown is the swell height observed at Begg Rock, an oceanic buoy station located about 100 km southwest of the Los Angeles coast (Fig. 3c). Short period microtremors are clearly related to cultural activities revealed by a regular daily spectral amplitude variation with a minimum at midnight and a maximum at midday, and a spectral amplitude decrease on the weekends (Fig. 3a). The spectral amplitude around a period of 6.5 s also exhibits a variation over time and shows a rather good correlation with the swell height amplitude (at period equal two times the microseism period) (Fig. 3b and c). This similarity in the time-varying characteristics indicates that long-period microseisms in the Los Angeles basin are closely related to ocean disturbances. Similar correlation between microseism amplitude and swell height, and more generally meteorological activities over ocean, was consistently observed by Bernard (1941a,b), Longuet-Higgins (1950), Haubrich et al. (1963), Akamatsu et al. (1992), Friedrich et al. (1998), Tindle and Murphy (1999), Grevemeyer et al. (2000), Okeke and Asor (2000), Bromirski (2001), Bromirski and Duennebier (2002), Bowen et al. (2003), Essen et al. (2003), Dolenc and Dreger (2005).

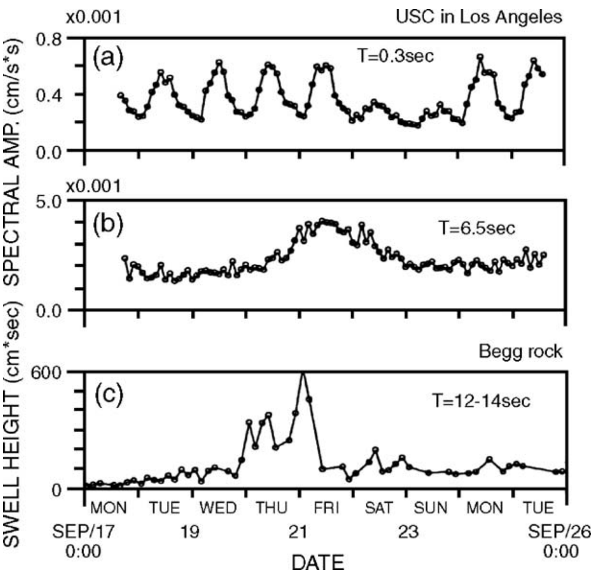

Fig. 3. Fourier spectra amplitude variations of seismic noise (Los Angeles) over time (horizontal components) at (a) 0.3 and (b) 6.5 s. (c) Recordings of swell height variations over time about a hundred kilometers from Los Angeles, after Yamanaka et al. (1993). At 0.3 s a good correlation between microtremor and cultural activities is shown, while at 6.5 s microseism variation is linked to ocean disturbances.

Such daily and weekly variations in the spectral amplitudes of microtremors were also reported in the earlier stage of research on noise. Kanai and Tanaka (1961) present results for continuous noise measurements performed in Tokyo, Japan. The maximum daytime amplitude is around 0.4 to 0.5 µm, and drops to 0.1–0.2 µm at nighttime (Kanai and Tanaka, 1961). The spectral amplitude average over time shows that the predominant microtremor period is different according to the hour (longer period during nighttime than during daytime). In addition, measurements during daytime and nighttime were performed at thirty sites in Japan, taking into account various kinds of subsoil. Kanai and Tanaka (1961) suggest the following empirical relation (1):

$$\text{Amplitude}(\text{nightime}) = 0.3 \times [\text{Amplitude}(\text{daytime})]^{1.5} \quad (1)$$

More recent studies lead to the same conclusion. Okada (2003, pp. 9–16) very clearly establishes the temporal variation of the noise Fourier spectra according to natural or cultural sources. Bonnefoy-Claudet (2004) also observed similar distinctions between microseism and microtremor spectra after analyzing continuous seismic noise measurements conducted over 2 months in the Grenoble basin (French Alps). Fig. 4 displays the vertical and horizontal normalized Fourier spectra amplitude for a four week period. Once again, high frequency microtremors ($F>1$ Hz) exhibit clear daily and weekly variations linked to cultural activities, while variations over time of low frequency microseisms are not linked with cultural activities. To go beyond earthquake hazard studies it is amazing to compare noise Fourier spectra spectrogram observed in Grenoble basin (Bonnefoy-Claudet, 2004) (Fig. 4a) with those observed in Lanzarote island (Canary, Spain) by Gorbatikov et al. (2004) (Fig. 5). These figures outline the differences between Spanish and French rhythm of life. According to the variation of the noise Fourier spectra amplitude, French and Spanish people start to work at 8 am, have lunch break at mid-time and 2 pm,

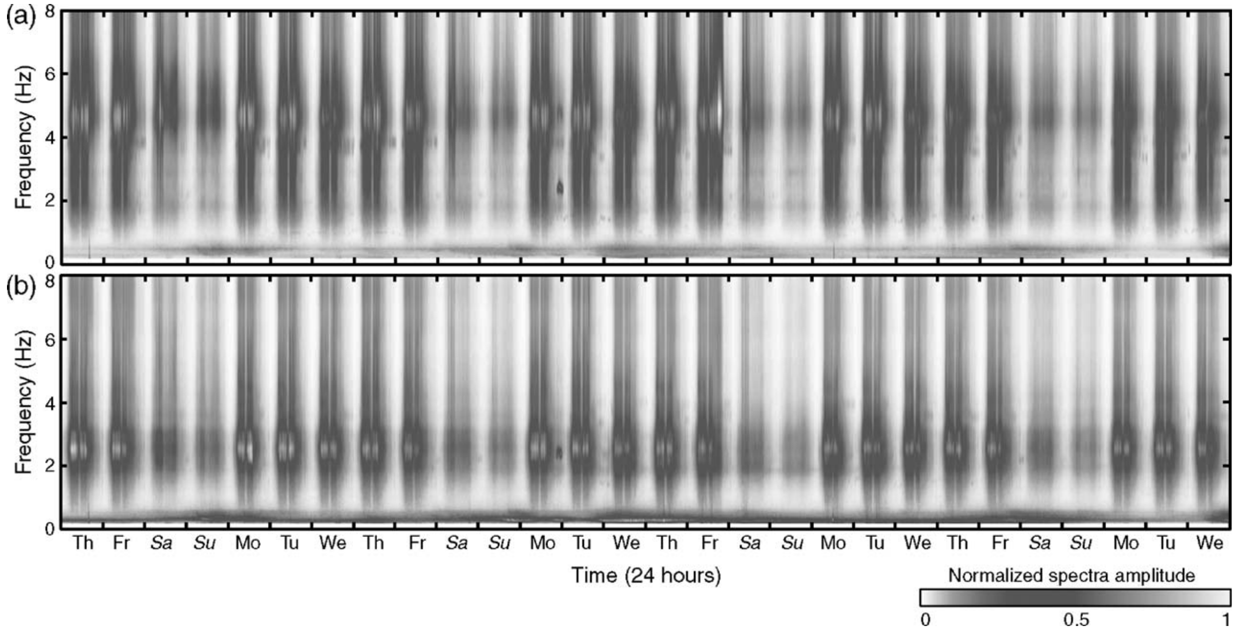

Fig. 4. Normalized Fourier spectra amplitude continuously recorded in urban area (Grenoble, French Alps) from 10.06.2004 midnight to 22.07.2004 midnight, a) vertical component, and b) north-east component. Modified from Bonnefoy-Claudet (2004). At frequency higher than 1 Hz microtremor exhibit clearly daily and weekly variations linked with cultural activities, and at low frequency microseism shows variations over time not linked to cultural activities.

respectively, and stop activities around 8–9 pm and 12 pm, respectively.

Since the fifties, improvements in array techniques have also brought some useful light on the origin of seismic noise. Toksöz and Lacoss (1968) use the frequency–wave number technique (*f*–*k*) on *LASA* (Large Aperture Seismic Array) data in Montana, United States, to investigate noise propagation directions. At low frequencies (0.2 to 0.6 Hz), two distinct noise source directions are identified, both linked with oceanic activity (Pacific Ocean and Labrador Sea). Using the same array analysis method (*f*–*k*), Horike

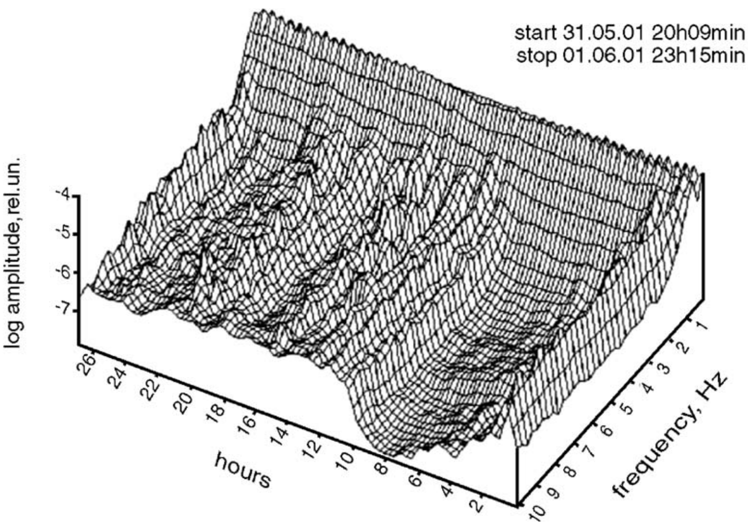

Fig. 5. Continuous seismic noise recorded at Lanzarote island (Canary, Spain). The resulting spectra amplitude are displayed according to frequency (*X* axis) and time (*Y* axis). Experiment started on 31.05.2001 at 20:09 local time and concluded late evening the next day. After Gorbatikov et al. (2004).

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(1985) observe similar results in the Osaka area. At intermediate frequencies (0.68 to 0.8 Hz) microseism sources are distributed along Osaka Bay and are mainly due to water waves striking the coast. At higher frequencies (1.4 to 1.7 Hz), microtremors sources are located inland and produced by Osaka city traffic. Satoh et al. (2001) provides another illustration of this point. Array measurements are conducted for one night in the Sendai basin, Japan. Noise propagation directions are estimated by *f–k* analysis on the vertical components. At frequencies close to 2 Hz, the directions of the microtremor propagation at this site come from the Sendai city. At lower frequencies (from 0.55 to 1.2 Hz) the microseism directions are relatively scattered in the direction of the coastal line of the Pacific Ocean. Once more, these results support the idea that the sources of long-period noise (microseisms) are mainly oceanic waves, and those of short-period noise (microseisms) are due to human activities such as traffic or machine vibrations.

On the basis of these results, it may be concluded that there are two distinct origins of noise: natural and cultural. Depending on the origin the noise behavior is different (in time and spectral domains). Microtremor amplitudes have daily and weekly variations, while natural noise exhibits amplitude variations linked with

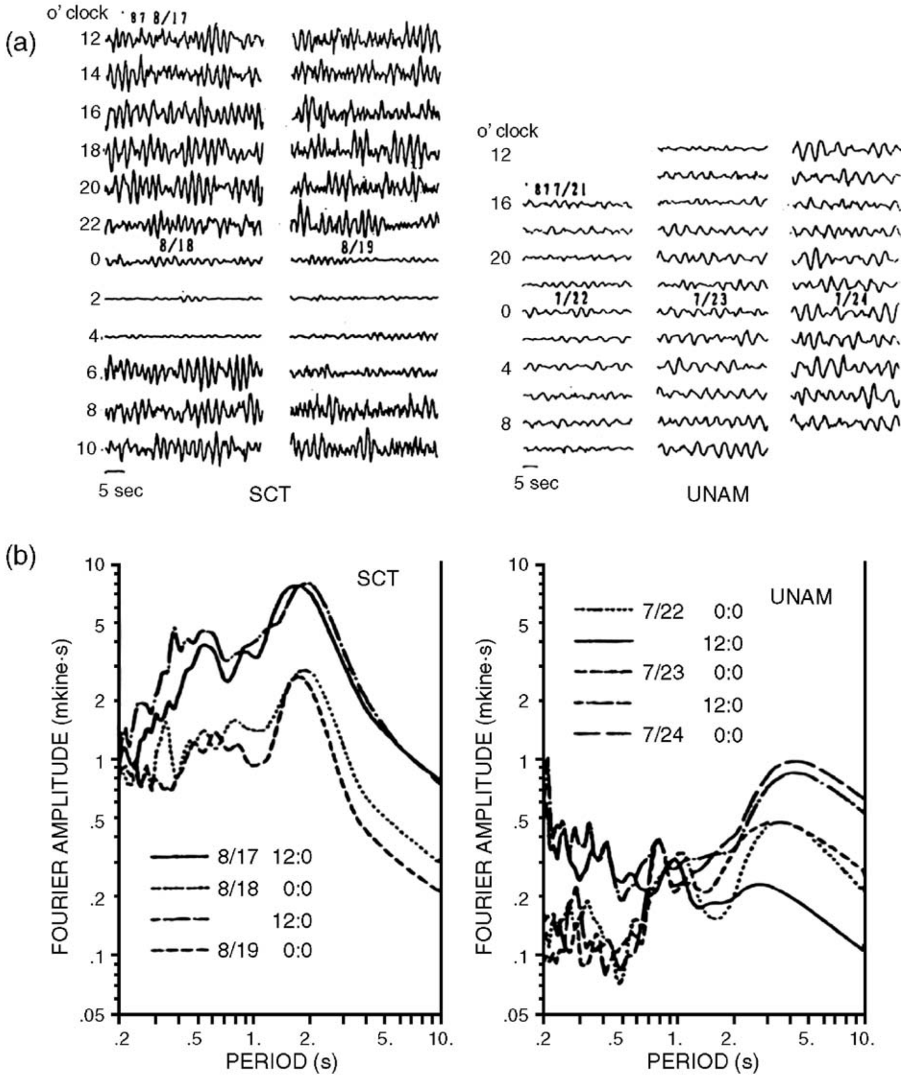

Fig. 6. Continuous noise measurements in Mexico. (a) Seismic noise amplitude variations over time for sediment site SCT and rock site UNAM. (b) Fourier spectra amplitudes for SCT and UNAM at mid and night time, after Seo (1997). The daily variations in noise Fourier spectra amplitude over the entire frequency range at sediment site suggest that microseism can also exhibit night and day variation.

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natural phenomena, most often oceanic activity for long period noise. Thus spectral boundary between microseisms and microtremors is around 1 Hz. We can summarize all the results concerning the origin of noise in order to draw up the following scheme: 1) seismic noise below 0.5 Hz is generated by oceanic and large-scale meteorological conditions; 2) noise at frequencies around 1 Hz is due to wind effects and local meteorological conditions; 3) above 1 Hz, seismic noise is due to human activities.

The 1 Hz boundary, however, is not a universal limit. According to Seo (1997) depending on geology, the limit between microseisms and microtremors can be shifted to a lower frequency. In deep soft basin, such as the Mexico basin, microtremor behavior (i.e., daily variation) can be observed at frequencies lower than 1 Hz. Fig. 6a displays two horizontal seismic noise traces recorded observed within Mexico city. The first was located in the lake-bed zone (SCT) and the second at a rock site (UNAM). In the time domain, diurnal variations are apparent at the sediment site, but not at the rock site. In the spectral domain, diurnal variations in noise spectral amplitude are obvious over the whole frequency range at the sedimentary site, while daily variations appear only for frequencies higher than 1 Hz at the rock site (Fig. 6b). This shows that a daily variation of seismic noise could be observed also at low frequency (i.e., below the 1 Hz limit) in case of deep soft basin. This observation is also noticed by Bonnefoy-Claudet (2004), who shows a daily variation on noise

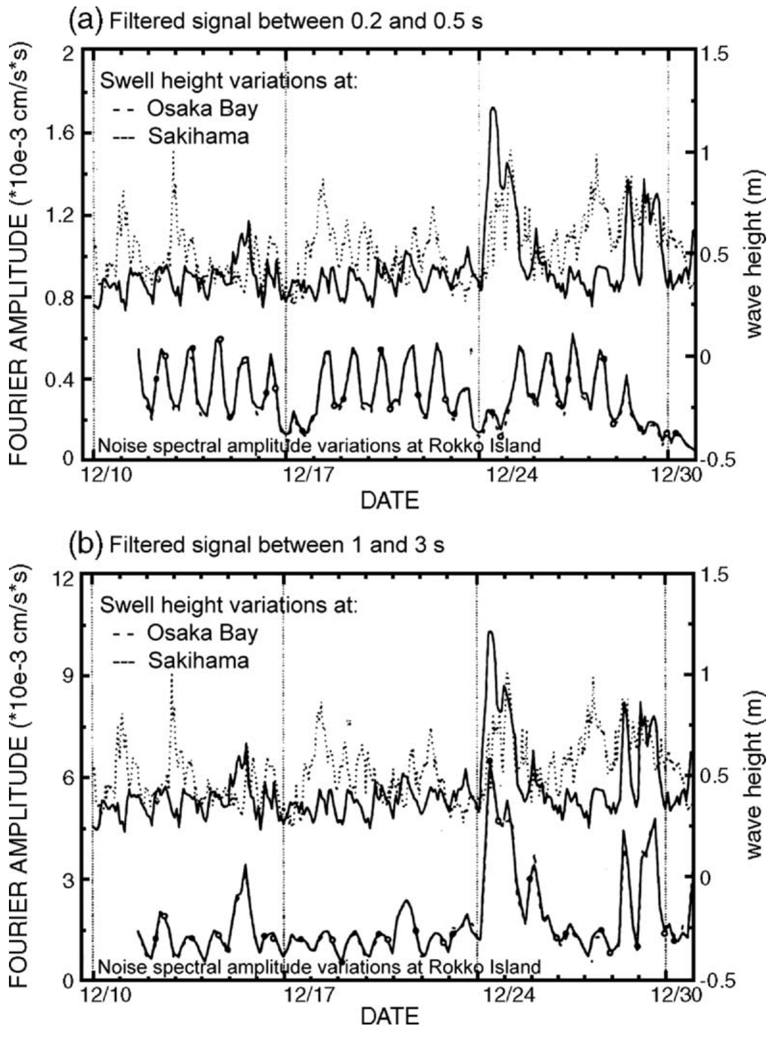

Fig. 7. Comparison between recordings of noise spectral amplitude at Rokko Island (bottom lines) and swell height amplitude variations at Osaka Bay (plain line) and at Sakihama on the Pacific coast (dashed line). The results are presented for two period ranges: (a) 0.2–0.5 s, (b) 1–3 s, after Seo (1997). The daily variation of microseism at low frequency can be explained by the night and day ocean activity (due to tide effect?).

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spectral amplitude for frequencies lower than 1 Hz (Fig. 4), in the deep soft Grenoble basin. However, Seo (1997) also shows results from other continuous noise measurements from Rokko Island (Kobe, Japan). Temporal variations in the Fourier spectrum amplitudes of microtremors for horizontal components are shown Fig. 7 for both the shorter period range (0.2– 0.5 s) and the longer period range (1–3 s). The variations in sea height measured at Kobe (in Osaka Bay) and at Sakihama (Pacific coast) sites are also shown in the same figure. For both period ranges the daily variation in the spectral amplitude of the microtremors are observed. As a daily variation is simultaneously observed in the swell height amplitude (mainly at the Sakihama site), the shift of the limit between microseisms and microtremors towards a lower frequency may be partly due to the night and day alternations of swell height amplitude (possibly in relation with tide effects?), at least in coastal areas.

4. Composition of the seismic noise wavefield

Ideally the goal would be to split up the noise wavefield into body waves (P, SV, SH) and surface waves (Rayleigh and Love waves), and quantify the proportion of each type of waves. The literature survey does prove that achieving that goal is not an easy task, and that many issues are still wide open. This section will thus summarize the relatively few results that have been obtained concerning 1) the relative proportion of surface waves and body waves, 2) the relative proportion of Rayleigh waves and Love waves, 3) for the Rayleigh part, the relative proportion of fundamental and higher modes.

4.1. Relative contribution of surface waves

Toksöz and Lacoss (1968) use the *f–k* method on noise measurements observed from the *LASA* array (Large Aperture Seismic Array), Montana, USA, to estimate the phase velocities of waves crossing the array. Note that the *LASA* array were constituted by 546 seismometers distributed over an area about 200 km in diameter, thus their analysis were focused mainly on crust structure. Their comparison with the theoretical value for body waves and Rayleigh waves (Fig. 8) lead them to draw the following conclusions about the nature of low frequency noise: 1) at 0.2 Hz the observed phase velocity is about 3.5 km/s corresponding to the theoretical phase velocity of the 1st or 2nd higher Rayleigh waves mode; 2) at 0.3 Hz two distinct velocities are identified, namely a 3.5 km/s velocity corresponding to Rayleigh waves, and another equal to 13.5 km/s. This second velocity is attributed to compressional body waves (P waves) since so high velocities on the vertical component can be reached only for slightly oblique P waves; 3) between 0.4 and 0.6 Hz, only the fast P waves are detected with a velocity around

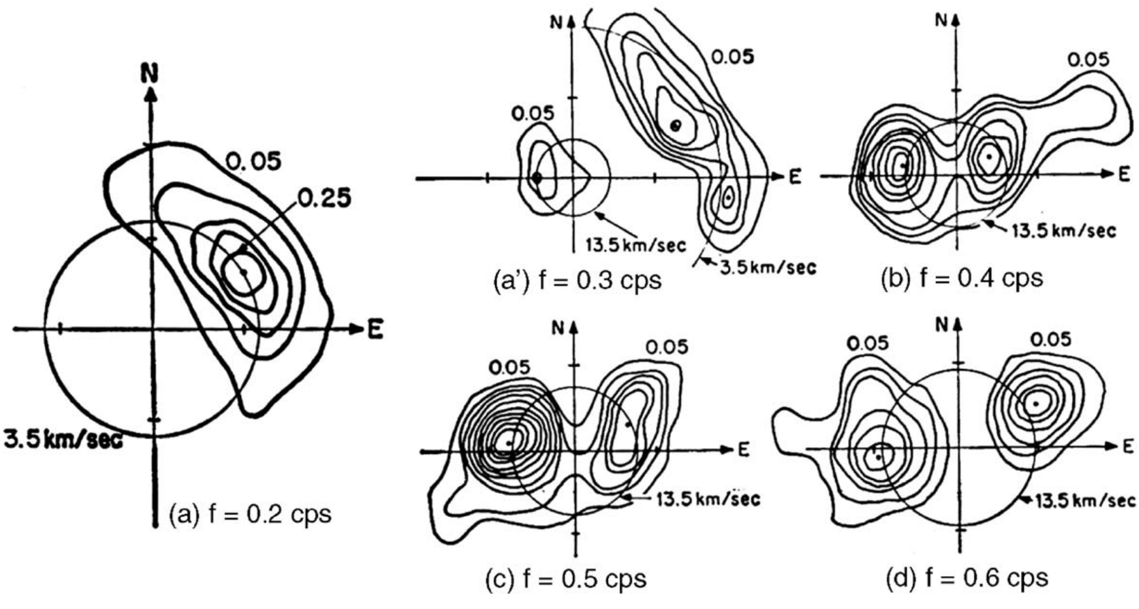

Fig. 8. *f–k* analysis performed on seismic noise recorded at the LASA antenna for different periods, after Toksöz and Lacoss (1968). Depending on the frequency, body wave (13.5 km/s) and/or surface waves (3.5 km/s) phase velocities are observed.

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13.5 km/s; 4) at frequencies lower than 0.15 Hz, these authors assume without further discussion that the seismic waves are fundamental Rayleigh waves.

During the sixties, Douze (1964, 1967) performed several noise measurements in a deep borehole and compared the depth dependence of the noise Fourier spectral amplitude with theoretical values for Rayleigh and P waves. Two examples are depicted in Fig. 9, for the Eniwetok Island borehole (1288 m deep, Marshall Islands, Fig. 9a), and for the Apache borehole (2917 m deep, Oklahoma, Fig. 9b). For the former example, the very significant decrease of the depth-to-surface noise Fourier spectral ratio at periods shorter than 2–3 s, suggests that the noise energy is mainly carried by surface waves. Beyond 3 s however, this ratio gets close to 1, suggesting that these waves are mainly body waves (no decrease in Fourier spectra amplitude with depth). The depth-to-surface ratios observed at the second site (Fig. 9b) for 3 distinct periods (0.5, 1 and 2 s), do not allow definite conclusions. According to Douze (1967), 1) at 0.5 s (2 Hz), noise is a mixture between P waves and 3rd Rayleigh waves mode; 2) at 1 s (1 Hz), it is still a mixture between P waves and 1st Rayleigh waves mode; 3) at 2 s (0.5 Hz), no clear conclusion can be drawn. However, with regard to Fig. 9b, the noise seems to be a mix between P waves and 1st Rayleigh waves mode. These pioneering studies (Douze, 1964, 1967; Toksöz and Lacoss, 1968) addressed the issue of noise wavefield in relation with crustal waves for rocky sites, while more recent studies (see below) mainly considered the noise structure for soft sites in relation with local amplification issues. While the former suffered from severe limitations in instrumental quality, computer and processing capacities in the sixties (compared to the latter), one may regret that a systematic comparison of noise levels at surface and large depth in relation with the crustal waves has not been yet performed with most recent data.

During the 1980, Li et al. (1984) and Horike (1985) used array analysis to determine the nature of the seismic noise on sedimentary sites. Phase velocities derived from array observations with *f–k* processing are compared with the theoretical expected wave velocities corresponding to the soil structure. At Lajitas (southwest Texas), Li et al. (1984) concluded that microtremors in the 1–20 Hz frequency range are a mixture between higher Rayleigh modes and P waves. Horike (1985) investigated the noise within the Osaka basin (Japan) in lower frequency ranges using a *f–k* processing on an eleven station array. This analysis shows that while noise consists of fundamental Rayleigh waves, between 0.5 and 0.9 Hz, higher mode could be detected in addition to the fundamental Rayleigh mode for frequencies between 0.9 and 3 Hz.

Recently, using array analysis, Bonnefoy-Claudet et al. (in press) investigated the composition of synthetic noise wavefield computed at the surface of a simple horizontally layered medium consisting of one soft layer over a half-space. They conclude that the relative proportion of Rayleigh waves to body waves is linked to the spatial noise source distribution (in terms of source–receiver distance and source depth). If the sources are below the sedimentary layer (i.e., inside the bedrock) then only non-dispersive body waves are present in the noise wavefield. Such a situation may be representative only for microseism with distant origin, since cultural noise sources are located at the surface. The results for surface sources depend on their distance from the receiver: if they are distant (i.e., more than twenty times the sedimentary layer thickness), the vertical noise wavefield is constituted by a mixture of Rayleigh and body waves (head waves propagating along the sedimentary/bedrock interface); if the sources are close, the vertical noise wavefield mainly consists of fundamental mode Rayleigh waves. Finally if all the sources are considered then near surface sources are dominant and the vertical noise wavefield is mainly constituted by fundamental mode Rayleigh waves. Note that the effect of source properties has been investigating only for a given high sediment/bedrock impedance contrast media. Further work has to be done in order to generalize Bonnefoy-Claudet et al. (in press) conclusions to larger data set. Nevertheless this shows the sensitivity of the noise wavefield composition to the sources properties.

Yamanaka et al. (1994) perform continuous noise measurements in the northwest part of Kanto plain (Japan) and show comparison between observed HVSR curves and theoretical ellipticity curves of the fundamental mode of Rayleigh waves for sedimentary sites. HVSR curves have been observed for different time during day and night. The good fit between the two curves leads Yamanaka et al. (1994) to suggest that microseisms over the frequency range 0.1 to 1 Hz consist mainly of fundamental mode of Rayleigh waves. Bonnefoy-Claudet (2004) also compares the HVSR curve to the ellipticity curve of the fundamental mode of Rayleigh waves, but unlike Yamanaka et al. (1994), for seismic noise synthetics computed at the surface of various horizontally layered media excited by random surface sources. This kind of simulation is thus valid only for cultural noise resulting from local surfaces sources, and cannot yet be extended to low-frequency microseisms caused by distant oceanic sources. Her

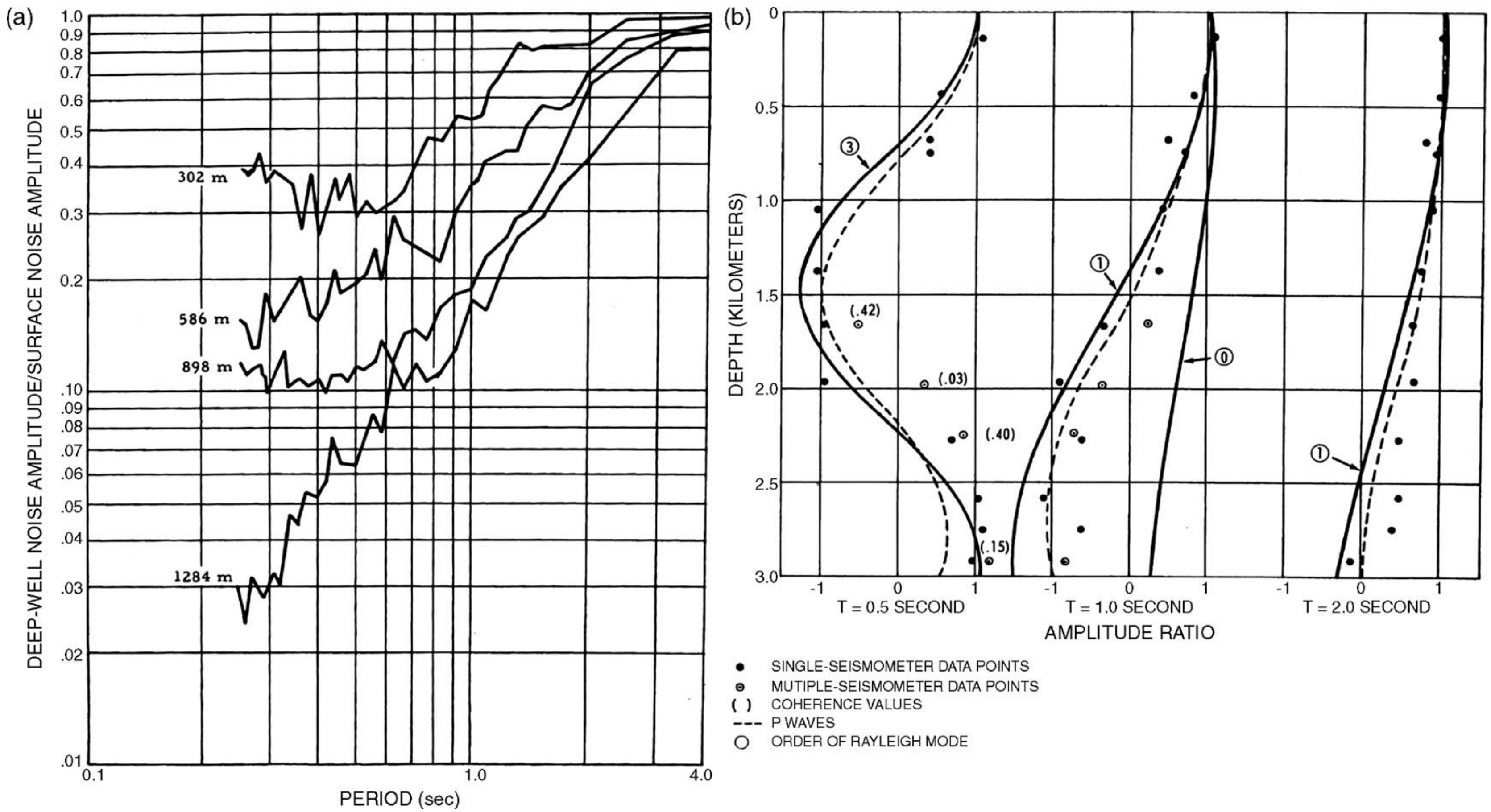

Fig. 9. Deep-to-surface noise Fourier spectra amplitude ratio (vertical component) as a function of period for (a) Eniwetok Island borehole (1288 m deep, Marshall islands), after Douze (1964), (b) for Apache borehole (2917 m deep, Oklahoma), after Douze (1967). Dots represent observed values, dashed lines theoretical curve of P waves, and plain lines Rayleigh waves.

main conclusion based on analysis of the HVSR curve is that the noise wavefield is dominated by surface waves only in the case of large impedance contrast (i.e., for a velocity contrast between the surface and the underlying bedrock above 3–4), while for smaller velocity contrast (below 3), the wavefield also includes body waves generated by close surface sources.

Fig. 10 summarizes conclusions about the composition of the noise wavefield in terms of body and surface waves. At lower frequencies (below 1 Hz), microseisms consist mainly of the fundamental mode of Rayleigh waves (Li et al., 1984; Horike, 1985; Yamanaka et al., 1994), as far as only the surficial structure is concerned (top kilometers to 100 m). At higher frequencies there is no agreement between the authors to define the nature of the seismic noise, and the results are very few. The noise wavefield is suspected to be a mix of body waves and Rayleigh waves (fundamental mode and/or higher modes). As suggested by Bonnefoy-Claudet (2004) and Bonnefoy-Claudet et al. (in press) from a thorough analysis of noise synthetics, this absence of any definite conclusion on relative proportion of surface waves may be simply a consequence of its larger sensitivity to soil and noise source properties.

4.2. Relative contribution of Rayleigh waves

The authors of all the studies presented in this section assume that seismic noise consists mainly of surface waves. The relative proportion of Rayleigh and Loves waves in the seismic noise wavefield has been seldom investigated in the literature, especially in the past, simply because array measurement have historically been done with vertical geophone only.

However, some interesting, though incomplete results could be obtained with numerical simulation using random, surface and local sources with arbitrary directions in horizontally layered half-spaces. The latest results were obtained by Bonnefoy-Claudet (2004, pp. 147–172), who concluded that Love waves are always present in the synthetic noise wavefield as soon as some noise sources have a horizontal component.

Similarly, Ohmachi and Umezono (1998) simulated noise propagation in simple 1D models (one layer over a half space) with the closed form equations established by Harkrider (1964). For such a soil model and considering different excitation types (number, position and direction of forces), the horizontal to vertical ratio (HVSR ratio) and the coherence between radial and vertical synthetic noise components are analyzed. They claim that this coherence value should be a good estimate of the Rayleigh to Love waves ratio (and a good estimate of the site amplification factor, as well). Nevertheless these results show a strong variation in the Rayleigh to Love waves ratio in the synthetic noise, as it depends on the type of excitation (vertical, transverse or radial forces), on the observation direction, and on the impedance soil contrast. This ratio varies between 10% and 90% with an average around 30%, and it seems difficult to establish reliable conclusions concerning the proportion of Rayleigh waves in actual noise. This numerical approach thus does not give quantitative estimate of the relative proportion of Rayleigh to Love waves in real noise, since it is highly dependent in the orientation of the random sources.

A few results however could be achieved with array measurements, most often with SPAC processing. Chouet et al. (1998) perform array measurements to

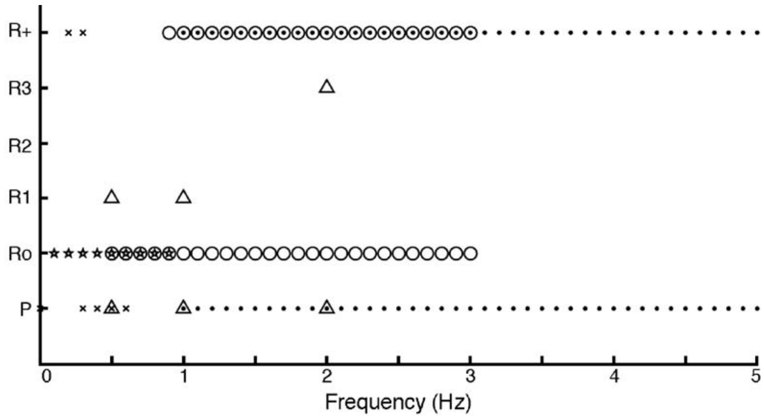

Fig. 10. Synthesis of the type of waves (body waves or Rayleigh waves), according to frequency, contained in the seismic noise wavefield. Letter P refers to body waves and letter R to Rayleigh waves (subscript indicate the order of Rayleigh mode: 0 for fundamental mode, 1 for first mode, 2 for second mode, 3 for third mode, and + when there is no order precision), after Douze (1964; 1967) (triangles), Toksöz and Lacoss (1968) (crosses), Li et al. (1984) (dots), Horike (1985) (circles) and Yamanaka et al. (1994) (stars).

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investigate the properties of the Stromboli volcano tremor. The spatial auto-correlation method developed by Aki (1957, 1965) is used to compute the spatial auto-correlation coefficients. The Rayleigh and Love waves phase velocities are obtained by inverting the spatial auto-correlation curves. In a first step, the Rayleigh waves phase velocity is estimated by azimuthally averaging the correlation coefficients obtained for the vertical component of motion. Then, the Love waves dispersion curve and the fraction of power carried by both types of waves is estimated by minimizing the errors between the azimuthally-averaged correlation coefficients calculated on both radial and transverse components, and the predicted correlation coefficients (see Eqs. 4 and 7 in Chouet et al., 1998). These authors found that 77% of the total energy contained in noise is carried by Love waves, and 23% by Rayleigh waves. However the percentage of energy ratio given is for a volcanic tremor, and there is no physical reason allowing to assume that wavefield composition is similar for volcanic tremors and microtremors. The interest of Chouet et al. (1998) for the present review relies much more in the method it proposes, than in the quantitative values they obtain.

Along the same direction, Okada (2003, chapter 3) describes how the SPAC method can detect Rayleigh and Love waves from three-component array measurements, and allow to derive an average contribution of each wave type. An example application to long-period seismic noise data results in a frequency-dependent energy power fraction for Love waves (Okada, 2003, fig. 5.25). This proportion increases from about 50% at 1 s to larger values (up to 90%) at longer periods (up to 3 s). Using a modified 3-component SPAC technique, Köhler et al. (submitted for publication) also observe a preponderance of Love waves (65–90%) at intermediate frequencies (0.5 to 1.3 Hz) in the seismic noise wavefield recorded at the Pulheim site (Germany).

Yamamoto (2000) also used the SPAC technique to estimate the energy ratio between Rayleigh and Love waves for 3 sites in the city of Morioka, Japan. Fig. 11 presents the results obtained for one site (Nioh): in the 3–8 Hz frequency range, 60% to 85% of the noise energy is carried by Love waves. For the two other sites (not shown here) although different ratios were observed, their values were exceeding 50%. The author concludes that for urban sites, in the frequency range 3 to 8 Hz, more than 50% energy is carried by Love waves than by Rayleigh waves in microtremors.

Similarly Arai and Tokimatsu (1998, 2000) estimate the relative proportion of Rayleigh and Love waves and its frequency dependence from 3-component array

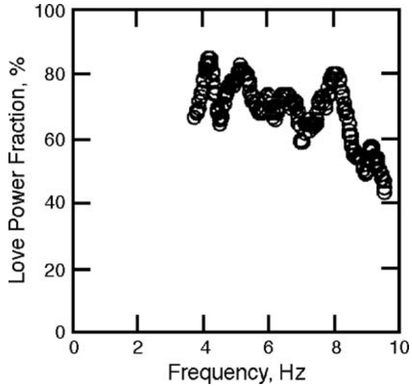

Fig. 11. Proportion of Love waves contained in noise wavefield for the 3 to 10 Hz frequency range at Nioh site (Morioka, Japan), after Yamamoto (2000).

measurements obtained at four sites in Japan (A: Yumenoshima (Tokyo), B: Rokko Island (Kobe), C: Asahi (Kushiro), D: Kotobuki (Kushiro)). They use both an f–k analysis on radial and transverse motion, and a three-dimensional SPAC approach (Fig. 12), and obtain quite similar results whatever the method. For the four sites, in the 1 to 12 Hz frequency range (extended down 0.3 Hz for site B, Rokko Island), the Rayleigh to Love waves energy ratio varies around an average value of 0.7 (independently of the method used). Hence for frequencies higher than 1 Hz (cultural noise), the noise consists of around 60% Love waves and 40% Rayleigh waves in this area.

Slightly different results were obtained in the deep Grenoble basin (France), where continuous noise measurements were also obtained over a four month period with an array consisting of thirteen 3-component seismometers. The low frequency seismic noise (lower than 1 Hz) was investigated with the *MUSIC* high-resolution array analysis (Schmidt, 1981). Cornou (2002) and Cornou et al. (2003a,b) observed a good agreement between the measured phase velocities and the theoretical surface wave velocities, and estimated the proportion of energy carried by Rayleigh waves through the ratio of the energy carried by radial and transverse components. Table 2 presents the results obtained for the different days and hours analyzed. While the meteorological conditions changed over the period of the study, Cornou (2002) points out a constant proportion of around 50% of Rayleigh waves contained in microseisms over the frequency range 0.2 to 1 Hz.

Table 3 summarizes the results obtained by all the authors concerning the relative contribution of Rayleigh waves in the seismic noise wavefield. Given the very few case studies, it is not straightforward to deduce

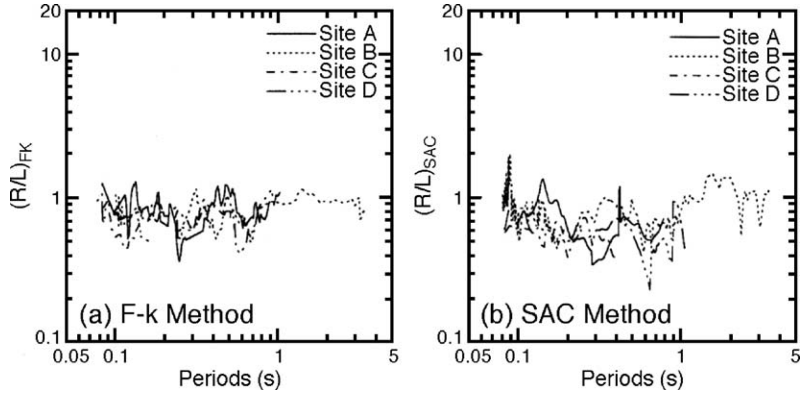

Fig. 12. Energy ratio between Rayleigh and Love waves estimated at four different sites and from 2 array techniques: (a) *f–k* analysis, (b) spatial auto-correlation method, after Arai and Tokimatsu (1998).

general conclusions about this proportion. For more or less the same high frequency range (higher than 1 Hz) and the same method (spatial auto-correlation method) Arai and Tokimatsu (1998) and Yamamoto (2000) obtain different quantitative Rayleigh to Love waves ratios. Their results, however, agree on highlighting the predominance of Love waves in the noise wavefield. At lower frequencies (below 1 Hz) it is also difficult to reach an agreement between different authors (Cornou, 2002; Okada, 2003; Köhler et al., submitted for publication). Note that all studies presented here were conducted at sites characterized by different sediment thickness (see Table 3). This suggests again that the site geology and geometry (sediment thickness, slope, geometry of the sediment-to-bedrock interface) as well as the source properties (see Ohmachi and Umezono, 1998) may influence the relative contribution of Rayleigh to Love waves in the seismic noise wavefield.

Although the results concerning the Rayleigh to Love waves ratio in seismic noise are not conclusive there is a common point in each study: there are all based on array analysis. 3-component array analysis, especially SPAC method (see Okada (2003)) seems to be the best way to determine the relative proportion of Love waves and Rayleigh waves in seismic noise.

Table 2
Rayleigh waves proportions (as %, ± standard deviation) contained in noise wavefield, for different days and hours (*MUSIC* high-resolution array analysis), after Cornou (2002) and Cornou et al. (2003a,b)

Julian days	Hours (TU)	Rayleigh waves proportions (%)
076	01–02	55±6
	15–16	50±6
085	01–02	52±12
	15–16	50±8
090	01–02	52±7
	05–06	50±8
	15–16	57±10
105	01–02	51±9
	05–06	50±7
	15–16	49±6

4.3. Relative contribution of the fundamental mode of Rayleigh waves

The actual energy distribution for various modes of Rayleigh waves is almost impossible to predict due to the heterogeneous ground conditions. Therefore the existence of Rayleigh higher modes in seismic noise has been seldom investigated in scientific the literature.

Indirect information can be obtained from the analysis of the HVSR. For instance, Stephenson (2003) investigates the link between HVSR and ellipticity of particle orbits of Rayleigh waves, and states that if the seismic noise wavefield is constituted only by the fundamental mode of Rayleigh waves then the HVSR curves should have a peak/trough structure. This statement is also supported by Konno and Ohmachi (1998), who show peak/trough structure on actual HVSR curves. Some other field examples can be quoted: at site ACAZ in Acapulco, Mexico (Lermo and Chavez-Garcia, 1994), in the Po Valley in Italy (Mucciarelli, 1998), at Baguio site BG2 in the Philippines (Ohmachi and Nakamura, 1992), and at the Embarcadero Freeway in San Francisco (Seekins et al., 1996). Besides, the simulations of Bonnefoy-Claudet (2004), already mentioned above, never exhibit such a peak/trough structure in HVSR curves. Careful array analysis performed on the noise synthetics indicates clearly the co-existence of the fundamental mode and higher modes of Rayleigh waves in the seismic noise wavefield (in most of the media considered). However this study does not provide information about the relative energy between the

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Table 3
Summary of Rayleigh to Love waves ratios contained in seismic noise wavefield

	Frequency range	Rayleigh waves proportion	Love waves proportion	Sites
Chouet et al. (1998)	>2 Hz	23%	77%	Volcanoes
Yamamoto (2000)	3–8 Hz	<50%	>50%	Sedimentary (thickness < 100 m)
Arai and Tokimatsu (1998)	1–12 Hz	40%	60%	Sedimentary (thickness < 100 m)
Cornou (2002)	0.1–1 Hz	50%	50%	Sedimentary (thickness ~ 500 m)
Okada (2003)	0.4–1 Hz	<50%	>= 50%	Sedimentary (thickness ~ 50 m)
Köhler et al. (submitted for publication)	0.5–1.3 Hz	10–35%	65–90%	Sedimentary (thickness ~ 200 m)

Summary established after studies by Arai and Tokimatsu (1998), Yamamoto (2000), Cornou et al. (2003a,b), Okada (2003), Köhler et al. (submitted for publication) and Chouet et al. (1998) for volcanic tremors.

fundamental and higher modes in the noise. The existence of higher modes at the frequency corresponding to the vertical polarization of the fundamental mode of Rayleigh waves (the "trough frequency") explains why in most of cases a peak/trough structure is not observed in HVSR curves (Bonnefoy-Claudet, 2004).

Bodin et al. (2001) observe a consistent second peak on HVSR curves from the Memphis basin. As no clear

(a)

Layer No. (1)	Thick–ness H (m) (2)	Density ρ ($Mg.m^3$) (3)	V_p (m/s) (4)	V_s (m/s) Case 1 (5)	Case 2 (6)	Case 3 (7)
1	2	1.8	360	80	180	80
2	4	1.8	1000	120	120	180
3	8	1.8	1400	180	180	120
4		1.8	1400	360	360	360

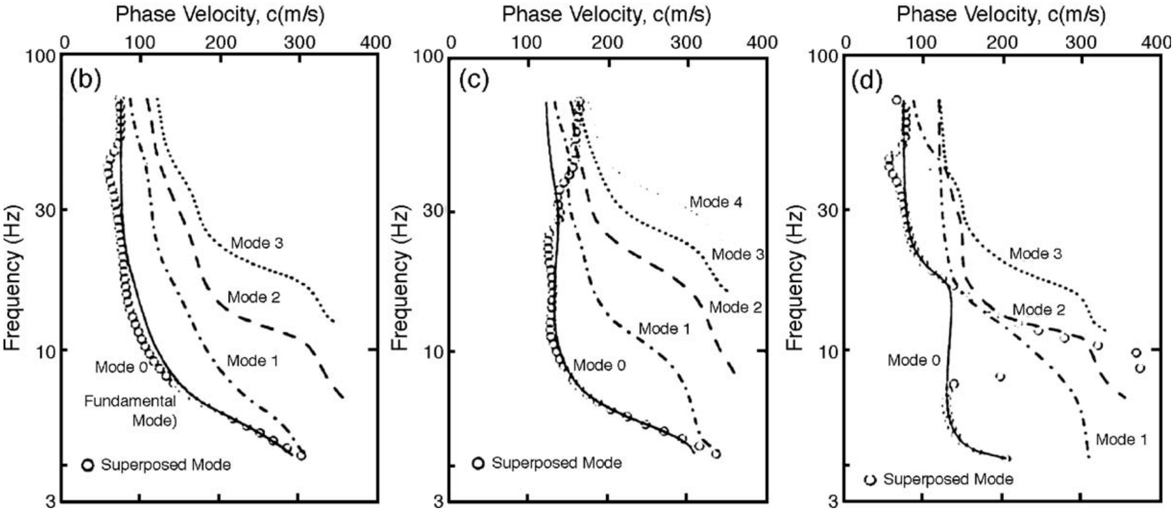

Fig. 13. (a) Velocity profiles used by Tokimatsu (1997) to compute Rayleigh waves dispersion curves numerically. The corresponding dispersion curves (dots) computed from noise synthetics in case 1 (b), case 2 (c) and case 3 (d) are compared with theoretical dispersion curves of higher mode Rayleigh waves, after Tokimatsu (1997). Depending on the soil properties, evidence of higher modes of Rayleigh waves is shown on the dispersion curves.

correlation between the frequency of this HVSR peak and the resonance frequency of surface unconsolidated sediment layers has been found, they interpret this second peak as due to the first higher harmonic of Rayleigh waves. For realistic models of the Memphis embayment, ellipticity curves of higher modes Rayleigh waves do exhibit peaks at higher frequencies that may be associated with the second peak observed on HVSR curves (Asten, 2004). This interpretation is also strongly supported by Asten and Dhu (2002). Another interpretation is however possible, given the simulations results obtained by Bonnefoy-Claudet (2004) and Bonnefoy-Claudet et al. (in press), which do exhibit higher peaks either in case of distant predominant noise sources, explained as body waves resonance. These conflicting conclusions show that the question of the origin of second (or third) peak on HVSR curves is still in debate within the scientific community.

Tokimatsu (1997) also shows the possible existence of higher modes of Rayleigh waves in the noise wavefield and investigate how the soil structure may emphasize the excitation of higher modes. Seismic noise wavefield is modeled at the surface of three different thin soil models (Fig. 13a). In the first case, the soil layer stiffness increases with depth, while it varies irregularly with depth in cases 2 and 3. In each soil case, phase velocities of the noise synthetics were estimated by array analysis (Fig. 13b to d). For comparison, theoretical dispersion curves of the fundamental and higher Rayleigh waves modes were computed for each soil model. In the first case with Vs increasing with depth, the computed dispersion curve follows the theoretical dispersion curve of the fundamental Rayleigh mode (Fig. 13b). By contrast, in the second and third cases, where velocity inversion at depth is present, higher Rayleigh modes could be observed at higher frequencies. In case 2 (Fig. 13c), the fundamental mode dominates up to 30 Hz and then the order of Rayleigh higher modes increases with frequency. In case 3 (Fig. 13d), the fundamental mode dominates except in the frequency range 9 to 16 Hz. Hence depending on the geological characteristics of the soil (velocity inversion at depth) higher modes may be excited. Although Tokimatsu (1997) does not give any quantitative answer concerning the proportion of Rayleigh higher modes in noise, he suggests that: 1) higher modes of Rayleigh waves do exist in noise; 2) soil stratification (shear wave velocity profile) plays an important role in higher mode excitation. This issue should be investigated in more detail, given the results obtained by Zhang and Chan (2003), Feng et al. (2005) and Wathelet (2005) from numerical investigation on the effects of higher modes of Rayleigh waves on the inverted S-wave velocity profile. These authors lead to the conclusion that mixing of data from different dispersion modes can severely affect the accuracy of the inverted ground velocity profile, especially in case of a low velocity zone.

5. Conclusions

This review outlines that, while an overall agreement is reached concerning the origin and gross characteristics of seismic noise, this is not the case for the composition of seismic noise wavefield, mainly because of the lack or scarcity of data.

It seems well established now that the spatial and temporal characteristics of seismic noise are closely related with its natural or cultural origin (low frequency microseisms or higher frequency microtremors, respectively). The amplitude variations of microseisms are well correlated with oceanic and large-scale meteorological conditions. In contrast, the daily and weekly variations of microtremors amplitude are clearly correlated with human activities (machineries, traffic, etc.). The boundary between these two types of noise (microseisms and microtremors) is close to 1 Hz, but may vary from site to site depending on the soil structure – and, may be, on the characteristics of human activity –. More advanced investigations have to be carried out to precisely identify a site-specific frequency limit between microseisms and microtremors. For example, continuous seismic noise measurement should be performed in different urban areas presenting different soil fundamental resonance frequencies (i.e., lower than, equal to and higher than 1 Hz) and lead to a better definition of the role of noise sources and soil geology in noise spectral amplification.

Concerning the composition of the seismic noise wavefield, this literature overview also highlights the scarcity and variability of relevant data which results in the absence of any scientific consensus. It would be however of particular interest for array and HVRS techniques to have reliable estimates of the average proportion of energy carried by Rayleigh waves, and more especially by the fundamental mode, in a given seismic noise recording. This is not possible today, nor even to determine whether it is legitimate to look for such average value. This variability may be due to differences from site to site (soil characteristics and/or source properties) as much as to methodological biases and/or author interpretation issues. Moreover this literature review outlines that in general higher modes of Rayleigh waves may exist in the noise wavefield at high frequency. Seismic noise based methods need to take this observation into account in order to reach the right interpretation, since, for instance, the interpretation of second or higher peaks observed on HVSR curves may depend on the noise wavefield composition (body waves resonance or higher mode of Rayleigh waves). Until now this is still an open question. This review also shows that the relative proportion of Love waves contained in the seismic noise is significant (at least 40%, probably higher than 50% in most cases). This result is important in view of the recent attempts to invert HVSR curves interpreted as a proxy to ellipticity curves, and there is a clear need to use and/or develop array methods and inversion schemes able to take it into account.

This study also provides indications as to possible ways to investigate the composition of seismic ambient noise wavefield. 1) Deep measurements of noise in boreholes may be an appropriate way to estimate the ratio of body waves to surface waves in seismic noise wavefield. A comparison between theoretical and observed eigen-functions of surface waves at known sites may lead to estimate the proportion of surface waves in noise. However, due to technical conditions (3-component broadband seismometer with sufficiently small aperture for non-tubed borehole) and experimental conditions (sufficiently well-known site structure), such experimentations are indeed difficult (and expensive) to achieve. 2) Three-component array measurements are a key element in characterizing the seismic noise wavefield and estimating the relative proportion of each wave type (body or surface waves) and propagation mode. Okada (2003) shows that frequency–wave number (*f*–*k*) based techniques are more efficient than spatial autocorrelation methods (MSPAC) to separate and characterize vertical higher modes propagation. By contrast, 3-component MSPAC methods are more efficient for defining the seismic noise propagation on horizontal components (Okada, 2003). 3) In addition, a numerical modeling approach constitute an informative complementary tool for detailed investigations on the structure of the noise wavefield, see Panza et al. (2001) for a review of the methods mostly used to determine the theoretical site response and their applications to seismic microzonation of urban areas. Investigations of the noise wavefield through numerical simulations under well-controlled conditions in terms of source characteristics and propagation structure do bring helpful and enlightening results. Simulated noise wavefield analysis for horizontally stratified structures (Bonnefoy-Claudet, 2004; Bonnefoy-Claudet et al., in press) and more complex 3D structures (Cornou et al., 2004; Roten et al., 2006) indicate that the composition of the noise wavefield is highly dependent on the soil characteristics (especially the impedance contrast value). Noise simulations still need, however, careful calibration on well-known sites, and are presently available only at frequencies corresponding to cultural (local) noise.

The key outcome from this review, and a challenge for future studies, is the need for further efforts to better assess the physical composition of the seismic noise wavefield, especially as the practical use of noise will certainly get a new boom given the cross-correlation techniques recently proposed by Shapiro and Campillo (2004). A physical understanding of the nature of noise is essential in defining the capabilities of ambient noise-based methods (HVSR, array and cross-correlation methods) and in making a significant contribution to effective seismic risk mitigation, in particular in urban areas.

It is worth however to conclude by emphasizing several facts: the hidden (or implicit) assumption according which seismic noise wavefield would consist only in pure fundamental mode Rayleigh waves (for the vertical component) is not supported by the data. The proportion of different waves (body/surface waves, Rayleigh/Love, fundamental/higher modes) is dependent on site conditions and source properties. However whatever the predominant type of waves, HVSR curves always indicate the resonance frequency (Bonnefoy-Claudet, 2004; Bonnefoy-Claudet et al., in press), at least for 1D structures with large enough impedance contrast.

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