Contribution of Dense Array Analysis to the Identification and Quantification of Basin-Edge-Induced Waves,
Part II: Application to Grenoble Basin (French Alps)

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Abstract Settled on a deep sediment-filled valley, the city of Grenoble (French Alps) faces important site effects: large amplification and significant duration increase of ground motion, even for moderate-size events. In order to study multidimensional site effects, a very dense array composed of 29 three-component seismometers over a 1-km aperture was operated during spring 1999 in the center of the city. A total of 18 events (6 local, 4 regional, and 8 teleseismic) with an acceptable signal-to-noise ratio could be recorded over a 4-month period. The complexity of the wave field and in situ seismic noise constraints led us to develop a procedure based on time–frequency coherence and the multiple signal classification algorithm to identify and characterize wave arrivals (Cornou et al., 2003). Applying the procedure to the 18 records, it is clearly indicated that ground motion inside the valley is dominated by basin-edge-induced waves that carry 4 times more energy than the direct wave field, regardless of the type of event considered. In addition, the basin-induced wave field is composed of 60% Rayleigh waves and 40% Love waves when considering energy carried by the three components. If one considers only the energy of horizontal components, this proportion is 50% Rayleigh waves and 50% Love waves. The diffraction phenomena are mostly constrained by the 3D structure of the basin, regardless of the azimuth of the event. A study of the relative contribution of 1D and 2D/3D effects on recorded ground motion suggests, at least at frequencies below 1 Hz, that the difference between the standard spectral ratio and 1D transfer function, or possibly the horizontal-to-vertical ratio (receiver function and Nakamura estimates) might be due mainly to laterally propagating waves.

Introduction

While the importance of site effects in earthquake damage has been recognized for a long time now, the increase in ground-motion amplification related to peculiar 2D (valley) or 3D (basin) geometries is still only very poorly quantified. After the long (and not yet resolved) debate about the relative importance importance of 1D and 3D effects in Mexico City, the damage belt observed in Kobe and its explanation in relation to basin-edge effects (Kawase, 1996) were among the first (dramatic) experimental evidence of the engineering importance of such multidimensional effects. Some additional work has thus been conducted to investigate this issue and provide quantitative numbers. By comparing site to reference spectral ratios computed over increasing window lengths, Field (1996) observed that 2D effects increase Fourier spectral amplification by a factor of 2 in the Coachella valley. Later, Chavez-Garcia and Faccioli (2000) proposed a way to include complex site effects in response spectra for seismic regulations, but they had too few examples to really propose statistically meaningful numbers. Therefore, much work remains to be done before such basin effects can be explicitly taken into account in building codes.

Numerical modeling had indeed shown for a long time that such phenomena were very likely due to the generation of local surface waves by thickness variations and their trapping inside the surficial sediments, which result in larger spectral amplifications and lengthened durations. However, direct observations of such basin-edge-induced waves are still relatively rare and have been performed only in large-size structures (pluridecakilometric scale), where the late edge-generated surface-wave phase can be easily seen sev-
eral tens of seconds after the S-wave train (Kagawa et al., 1992; Kawase and Sato, 1992; Kinoshita et al., 1992; Phillips et al., 1993; Frankel, 1994; Hatayama et al., 1995; Field, 1996; Malagnini et al., 1996).

In smaller structures (i.e., those having lateral dimensions smaller than around 10 km), the shorter travel times for the laterally propagating surface waves, the complexity in the structure geometry, and the possibility of multiple reverberations between the borders result in very complex wave fields that consist of multiple waves mixed in the early S-wave portion (and in the P-wave part as well). Basin-induced waves are thus much more difficult to observe directly, and the first examples of such effects reported in the literature for small-size structures are most often indirect evidence, based on the fact that 1D modeling cannot capture the essential feature of the measured response (amplification, duration). This was the case in the Marina District in San Francisco (Liu et al., 1992; Graves, 1993), in the Ubaye valley in the French Alps (Jongmans and Campillo, 1993), in the city of Grenoble in the French Alps (Lebrun, 1997), and in the Volvi test site (Beauval et al., 2003). Recently, however, such waves could be observed directly in several small basins: the Caille valley in the French Alps (Gaffet et al., 1998), Colfiorito in central Italy (Caserta et al., 1998; Rovelli et al., 2001), and Parkway in New Zealand (Chavez-Garcia et al., 1999). In each of these three cases, direct observation was made possible through dense array recordings; array analysis allows indeed a deeper insight into the characteristics of the wave field, by giving access to the azimuth, phase velocity, and possibly the polarization of the various waves propagating through the array. None of these studies, however, drew any quantitative conclusions on the respective importance of vertical (1D) and lateral (2D and 3D) reverberations in the amplification of ground motion.

The present study, focusing on the Grenoble basin, is following the same array approach to identify the wave field associated with site amplification, with an attempt to go one step further in order to obtain quantitative information on the actual importance of 3D effects, on the sole basis of experimental observations.

The city of Grenoble is located in a small-size, deep sediment-filled basin, typical of almost all alpine valleys, and it faces important site effects. Observations from temporary networks (Lebrun et al., 2001) and permanent accelerometric stations (RAP, www-rap.obs.ujf-grenoble.fr) showed a large amplification (up to 10 times) in a broad frequency band (0.2–5 Hz), as well as an important increase of ground-motion duration throughout the basin for moderate-size seismic events.

As outlined by Lebrun et al. (2001), amplification values observed on standard site-to-reference spectral ratios (as first proposed by Borcherdt (1970)) are systematically at least 2 times larger than the theoretical 1D values derived from the S-wave transfer function. Figure 1 reproduces some of the spectral ratios computed by Lebrun (1997) after an M 2.5 event occurring 15 km east of the city center and compares them with the amplitude of the horizontal/vertical (H/V) peaks derived either from earthquakes (receiver function technique; Langston, 1979) or from noise (Nakamura, 1996). The four spectral ratios correspond to four different locations in the city center, while the reference station is located on a low-altitude outcrop of Jurassic limestone very close to the city center. Standard spectral ratios are characterized by a flat amplification over a wide range of frequencies that is not observed on other curves. Therefore, Lebrun (1997) proposed 2D/3D basin effects for explaining such wideband amplification. This suggestion was supported by numerical simulations (up to a frequency of 1.6 Hz) of ground motion in the Grenoble basin (Cotton et al., 1998; Bard et al., 1999) that outlined the large complexity of wave propagation with important trapping effects within the basin and multiple diffraction and reverberations between the sharp borders.

In order to improve our understanding of these peculiar wave propagation phenomena, a very dense array of three-component seismological stations was installed in February 1999 within the city and operated during 4 months (hereafter, the Grenoble99 experiment). Despite the moderate seismicity of the area, and the high noise level, a number of teleseismic, regional, and local events could be recorded. A specifically designed array analysis method was then used to investigate the entire wave field of the 18 events presenting the best signal-to-noise ratios. This array analysis technique is a combined procedure based on signal coherence between array sensors and multiple signal classification (MUSIC) analysis (Goldstein and Archuleta, 1987); it is fully described in the accompanying article (Cornou et al., 2003 [article 1]). Its primary goals are first to identify and characterize the basin-edge-induced waves and second to try to quantify the importance of the edge-diffracted wave field. This latter point is one of the entrance doors to a better assessment of the relative importance of 2D/3D effects with respect to the classical 1D approach.

We first briefly describe the main characteristics of the Grenoble99 experiment and present the seismic events that were selected for the study. Next, we present experimental evidence of diffraction of impinging high-frequency waves at the basin border and of the amplification effects of the basin at low frequency. The spatial origin and nature of the most coherent wave trains crossing the array with an azimuth different from the source azimuth are then detailed and discussed; they suggest important basin-induced surface-wave phenomena. The quantitative importance of such surface waves is then assessed through an evaluation of the total energy they carry. A comparison with the energy carried by direct waves finally allows us to propose an interpretation of the difference between standard spectral ratio and 1D estimates (or possibly H/V ratios) in terms of basin-induced surface waves.
The Grenoble99 Experiment

Quick Overview

Geological and geophysical characteristics of the Grenoble basin were described in article 1. We only recall here the most important points. The sedimentary fill consists of late-Quaternary postglacial deposits overlaying Jurassic marls and a marly limestone substratum. Several geophysical surveys have been conducted in recent years using different techniques: gravimetry (Vallon, 1999), active reflection and refraction seismic (Dietrich et al., 2001; Cornou, 2002), and microtremor recordings (Bettig et al., 2001; Lebrun et al., 2001). These methods were recently calibrated and validated by the drilling of a deep borehole in the northeast branch of the valley (Lemeille et al., 2000), which reached the marly substratum at 532-m depth and was complemented by a shallow one nearby (Lemeille, 2002). These measurements provided a reliable view of both the depth of the sediment/bedrock interface and of the $P$- and $S$-wave velocity profile (article 1). In addition, Lebrun et al. (2001) conducted a site effect study within the area and observed a fundamental resonance at 0.3 Hz, corresponding to the whole Quaternary deposit, and another one near 3 Hz that they assigned to a thin surface layer. A temporary dense array composed of 29 three-component seismic sensors was installed in the eastern part of the city (article 1). Figure 2 shows the array geometry, which is mainly composed of a small-aperture array (inner array) equipped with short-period sensors (L22, with a flat response between 2 and 50 Hz) and a larger one (outer array) equipped with wider band sensors (Le3D/CMG40 with a flat response from 0.2/0.05 to 50 Hz).

Data

The experiment lasted 4 months, from February to May 1999. The stations were operated in a continuous recording mode. Although a total of 36 seismic events were recorded, we selected for this study a subset of 18 only, based on two criteria: (a) their satisfactory signal-to-noise ratio and (b) a minimum value of available simultaneous recordings of 8 and 11 when using the outer and inner array, respectively. Event main characteristics are listed in Table 1, where they are arranged in three data sets according to epicentral distance. We thus distinguish local, regional, and teleseismic events. Figure 3 displays examples of L1, R1, and T8 events recorded at the center of the inner array. As the analysis method described in article 1 requires very precise time, instrumental correction was done each time we mixed different types of sensors. This correction consisted of a removal of the instrumental response and then a convolution by the response of the shortest band sensor available in the data set. Thus, a data set composed of only CMG40 was used to study wave propagation below 0.2 Hz, a data set composed of Le3D and CMG40 was used for frequencies between 0.2 Hz and 1 Hz, and L22 sensors and one CMG40 were used for frequencies above 1 Hz.

Figure 1. Example of spectral ratios obtained at four different sites in Grenoble for an $M_l$ 2.5 seismic event that occurred 15 km from Grenoble. The thick line represents the soil/bedrock spectral ratio, and the thin lines indicate spectral ratios computed from seismic noise (Nakamura, 1996) and from recorded earthquakes using the receiver function method (Langston, 1979). Reproduced from Lebrun (1997).
Evidence of Basin Effects: Amplification and Diffraction

Teleseismic Events

Out of the eight teleseismic events, four exhibited clear surface waves propagating at frequencies below 0.2 Hz. Figure 4a presents, for each of the four events, the distribution of the identified backazimuth as a function of frequency for all the time–frequency windows that were processed with MUSIC. The whole seismogram duration is analyzed and the selection of time–frequency windows is described in article 1. In the figure, all components are considered, that is, we simply plot on the same figure results from the up–down, north–south, and east–west components. For frequencies below 0.2 Hz, identified backazimuths are nearly the same as the great-circle backazimuth to the epicenter, while above
0.2 Hz, arrivals are extremely scattered and come from every direction. Dispersion around the theoretical backazimuth observed below 0.2 Hz may come from either the lack of resolution due to the seven-element array (CMG40 sensors) or from scattering of surface waves by regional-scale heterogeneities. Cotte et al. (2000) have shown indeed that in the western Alps this deviation can reach 30° for long-period surface waves. This 0.2-Hz threshold frequency agrees with the mean resonant frequency of the basin evaluated by Lebrun et al. (2001) and outlines that incident waves with a frequency content lower than the resonance frequency are not significantly influenced by the basin.

Another indication of this behavior is obvious when computing spectral ratios between recordings of one CMG40 sensor of the array and the SSB station from the Geoscope network (www.geoscope.ipp.jussieu.fr). The SSB station is located 100 km west of the array center and is equipped with an STS1 sensor. Spectral ratios were computed using the first 300 sec of signal. Time series were bandpass filtered (0.05–0.5 Hz) and tapered using a 10% Hanning window, and the amplitude spectra were smoothed using the Konno and Ohmachi (1998) formulation with a frequency width factor $b = 40$. One can see (Fig. 4b) that below 0.2 Hz no amplification within the basin is observed, whereas above 0.2 Hz amplification is huge, emphasizing once again the amplification effects of the structure.

### Local Events

Four of the six local events are aftershocks of an $M_3$, 3.5 local event that occurred 2 weeks before the beginning of the experiment. The backazimuth of the events is 180° N (180° from north in a clockwise direction). These aftershocks, listed as events L1, L2, L4, and L6 in Table 1, present the same characteristics in terms of backazimuth, linear polarization, and apparent inclination estimates during the very first seconds of ground shaking. The frequency content of first arrivals ranges from 5 to 15 Hz, but the most energetic arrivals are near 7 Hz. We detailed with MUSIC the $P$-wave signal (before onset of the direct $S$ wave) of the L6 event (Fig. 5). As the frequency content of the ground motion is high, we considered, in that case only, a Nyquist wave number of 0.2 rad/m instead of the one considered in article 1 (0.1 rad/m). As a consequence, we checked for each time window processed by MUSIC that the identified wave was not an aliased one (see article 1 for more explanation). Figure 5b shows identified backazimuths as a function of frequency for all components. For the direction of maximum polarization, the apparent inclination from the vertical and apparent linear polarization angles are then computed by using $t-f$ windows that were used to process the north–south and east–west component (Fig. 5c,d).

Until 2.6 sec the motion is vertically polarized with high linear polarization indicating pure body waves. Values of backazimuth show that the very first motion comes from 120° N. Then the main azimuthal trend is a linear deviation of backazimuth up to 220° N. After 3 sec, arrivals exhibit again the same azimuthal trend from 120° up to 220° N. This may suggest that the impinging wavefront coming from the south (180° N) was deviated by the basin edges, as shown on Figure 5e.

### Statistical View of Estimates

From these two examples, it is obvious that the 3D structure of the Grenoble basin significantly affects the wave field inside the valley. As it is impossible to follow the wave field in its whole complexity, we will try in the following to outline the main characteristics of the wave propagation phenomena through a statistical analysis of MUSIC estimates.

### Selection of Identified Waves

When investigating the wave field of seismic events, we first evaluated on each of the three components the time–frequency ($t-f$) windows exhibiting the most spatially coherent signals (see article 1). Values of coherence threshold and number of $t-f$ windows processed with MUSIC are indicated in Table 1. Next, MUSIC array analysis was performed separately on the three components. As mentioned in article 1, body and Rayleigh waves should in principle be simultaneously identified on all three components and Love waves on both horizontal components. After array analysis is performed, we kept only those waves identically identified.
Figure 4. (a) Backazimuth of identified wave trains as a function of frequency for four teleseismic events (events T4, T5, T6, and T7). Arrows indicate the theoretical event backazimuth. (b) Spectral ratios between one CMG40 of the array (MINI station) and the SSB reference station. Signals are bandpass filtered between 0.05 and 0.5 Hz, and spectral ratios are computed on the first 300 sec of the four teleseismic events. The thick line indicates the mean spectral ratio, and the dashed line is the mean plus/minus standard deviation.

Statistical Merging of Estimates

As introduced earlier, we allowed some deviations on frequency and backazimuth when tracking waves identically identified on both horizontal components. We also mentioned in article 1 that a statistical description of final estimates (frequency, velocity, and backazimuth) should enhance the reliability of results. Thus, we aggregated the results by using time/backazimuth/frequency grids. The grid steps used here are 0.2 sec $\times 10^6 \times 0.5$ Hz when studying above 1 Hz and 5 sec $\times 10^6 \times 0.1$ Hz when looking at estimates below 1 Hz. In other words, we considered as identical all waves within the same grid cell. In the following results, we use a backazimuth/frequency (BF) representation, regardless of the time window for which the waves were identified. We computed the wave and energy densities as follows:

- For each cell $(\Theta, F)$, the wave density $b(\Theta, F)$ is defined as

$$b(\Theta, F) = \frac{\sum_{i=1}^{M} N(\theta_i, f_i)}{B},$$

with $M$ the total number of windows processed by MUSIC, $N_i$ the number of waves propagating at a frequency $f_i$ and with a backazimuth $\theta_i$ that fell into grid cell $(\Theta, F)$ of the BF grid. $B$ is a normalization factor, corresponding to the $(\Theta, F)$ cell with the maximum number of identified waves, so that $b(\Theta, F) \leq 1$.

- For each cell $(\Theta, F)$, the energy density $e(\Theta, F)$ is defined as

with respect to backazimuth and frequency content on both horizontal components. For such a selection, backazimuth has to be within a deviation of $10^\circ$ and frequencies within a deviation of 0.05 and 0.25 Hz for studies below and above 1 Hz, respectively. We did not use apparent wave velocity as a criterion for wave selection because this parameter is much more unstable than other attributes (Zerva and Zhang, 1996; article 1). When the final selection of waves is done, the components are rotated along radial and transverse components, and subsequent polarization and energy estimations are performed using the three-component covariance matrix (article 1).
Figure 5. Evidence of diffraction of incident wavefront by the basin border for event L6. (a) First 2.5 sec of the three-component seismogram for the sensor located at the center of the small array; (b) backazimuth as a function of time for the north–south and east–west component all together; (c) apparent inclination angle; (d) apparent linear polarization; (e) simple scheme depicting diffraction of incident wavefront.

\[ e(\Theta, F) = \left[ \sum_{l=1}^{M} E_{\text{total}}^i N_i(\theta, f) \right] / E, \]

\[ E = \max_{\theta, f} \left[ \sum_{l=1}^{M} E_{\text{total}}^i N_i(\theta, f) \right]. \]

with \( E_{\text{total}}^i \) the total energy of wave \( i \) having a backazimuth belonging to \( \Theta \) and frequency to \( F \); the total energy was computed as presented in article 1. \( E \) is a normalization factor, corresponding to the \((\Theta, F)\) cell with the maximum energy.

The proportion of energy within the radial plane \( (E_{\text{radial}} + E_{\text{vertical}}; \text{see article 1}) \) compared to total analyzed energy \( (E_{\text{total}}) \) and apparent inclination of the maximal direction of polarization are displayed too using the BF grid.

In addition, we similarly evaluated the distribution of apparent velocities as a function of frequency by using velocity/frequency grids with grid steps of 40 m/sec \( \times 0.1 \) Hz when studying below 1 Hz and 40 m/sec \( \times 0.5 \) Hz above 1 Hz. The distribution of apparent velocity is simply defined as the number of waves having a propagating frequency \( f \) and velocity \( V \) that fall into a grid cell of the velocity/frequency grid. The frequency step used here allows a detailed distribution of apparent velocity as a function of frequency, whatever the backazimuth.

Furthermore, we separated waves coming directly from the epicenter from waves coming from elsewhere. We assumed that waves having a backazimuth falling within \( \pm 30^\circ \) of the source backazimuth are direct waves coming directly from the source. This \( \pm 30^\circ \) comes from the great-circle deviations observed by Cotte et al. (2000). We simultaneously consider all other waves as “off-source.” By doing so, we may only overestimate the source term, since there may exist...
off-source waves in the epicentral azimuth. This separation criterion was applied for frequencies below 1 Hz for teleseismic events and for all frequencies for regional events. For local events, we used a criterion slightly different to discriminate off-source and source waves, as explained later.

Identification and Stability of Diffractors at Low Frequency (Below 1 Hz)

Teleseismic Events

For frequencies above 0.2 Hz, we computed the backazimuth and energy density of the off-source arrivals. Energy densities for the eight teleseismic events are depicted in Figure 6. Despite a significant event-to-event source location variation, most energetic arrivals come from two main stable directions of propagation: north and east of the array location. The proportion of energy within the radial plane as well as apparent inclination were found to be very similar for all events, too. In Figure 7, we show, as an example, these estimates as a function of frequency for all identified backazimuths of events T6 and T8.

Because of this stability of results and in order to outline the main characteristics of the off-source wave propagation, we plot for all events the average density of wave (Fig. 8a), energy (Fig. 8b), apparent inclination of dominant polarization (Fig. 8c), and proportion of energy carried in the radial plane as a function of backazimuth and frequency (Fig. 8d). BF grids corresponding to each of the eight events (wave and energy density, apparent inclination, and proportion of energy within the radial plane) were simply added and averaged. Figure 8e is another representation of Figure 8b, where contributions from all frequencies are summed up. We have also calculated the distribution of apparent velocities as a function of frequency for all events. Rather than using the average, velocity/frequency grids of each event were simply added (Fig. 8f). The main azimuthal contribution is coming from northwest and east of the array. The associated energy is mostly located around 0.3 Hz.

High values of apparent inclination indicate that waves are mostly horizontally polarized. Apparent velocities are small at high frequencies and larger for low frequencies. This strongly suggests that off-source waves are mainly composed of surface waves, at least Love waves for which apparent inclination has to be 90°, with much less contribution from body waves. In order to check the hypothesis of surface waves, we computed the dispersion curves of the first two modes of Love and Rayleigh waves by using the velocity model derived from borehole measurements from 0 to 532 m. The actual sediment thickness straight below the array (see article 1) was taken into account by extending P- and S-wave velocity profiles down to a depth of 700 m. We simply extrapolated the depth–velocity gradient above 532-m depth down to 700 m. We also considered a thinner surficial layer with an S-wave velocity of 250 m/sec. Lebrun et al. (2001) have observed indeed on H/V ratios computed on noise recordings at various sites within the basin a secondary peak near 3 Hz that they attributed to resonance of a surficial layer in the depth range of 20–40 m. This suggestion was confirmed by Cornou (1998) when computing the H/V ratio on synthetic noise generated for a 1D horizontally stratified medium rather representative of Grenoble’s soil column (Fig. 9). Effects of a topmost soil layer overlaying the sediment fill were also experimentally observed in Quioto and Puji on secondary peaks observed on H/V ratios (Guéguen et al., 1998, 2000). H/V ratios computed on noise recorded within the same time window by the outer array sensors exhibit in some cases a secondary peak between 2.5 and 4.5 Hz (Fig. 10). Thus, assuming an S-wave velocity of 250 m/sec, we consider a 20-m-thick layer in order to get a resonance frequency of about 3 Hz within this surficial layer. The velocity profile finally considered at the array location is indicated in Figure 11a. At low frequency, we are confident in these velocity profiles because the first mode of Rayleigh waves fits rather well between 0.5 and 2 Hz the dispersion curves obtained from dense array measurements of noise with an array located in the same place as ours (Fig. 11b). For frequencies below 0.5 Hz, experimental dispersion curves are not reliable because of limitations due to array size and sensor response problems (Bettig et al., 2001).

Most of the apparent velocities evaluated in this study lie between the Love- and Rayleigh-wave dispersion curves down to 0.4 Hz. Below this frequency apparent velocities are largely dispersed and can reach high values. Three main reasons for such high apparent velocities at 0.3 Hz can be invoked. The first reason could be a correlation phenomenon involving waves having very close propagation velocities. Simulations performed in article 1 showed that in such cases only one of the waves is detected with a quite accurate backazimuth but with an overestimated velocity. Overestimation factors range from 1.2 to 2 on synthetic correlated waves at 0.3 Hz and for a theoretical propagation velocity of 2000 m/sec. Overestimations observed here are of the same order if one assumes that the identified waves are surface waves. The second reason could come from the numerical uncertainty at such frequency. As mentioned in article 1, the uncertainty expected for a wave propagating at a velocity of 2000 m/sec is about 1000 m/sec at 0.3 Hz. Third, the sediment depth varies with azimuths, and the actual dispersion curves might exhibit a similar azimuthal dependence. The mean distribution of apparent velocities fits quite well with dispersion curves of surface waves. This argues that the off-source wave field is mainly composed of surface waves. When plotting energy versus backazimuth on a contour map of the Grenoble basin (Fig. 8e), one can see that these off-source waves are apparently generated by the Belledonne (east) and the Chartreuse Massifs (north). Taking into account the strong impedance contrast, the 3D geometry of the basin, the surface-wave nature of these off-source waves that appear above 0.2 Hz, and previous numerical simulations that outlined at least up to 1.6 Hz important trapping wave effects within the basin (Cotton et al., 1998), we postulate that off-
source waves are actually edge-generated surface waves. However, since we did not have a similar array outside of the basin in order to observe directions of coherent propagation in late portions of the signal, we cannot discriminate from the azimuths alone whether these waves are diffracted by the basin or by the further regional heterogeneities.

Therefore, we have compared the signals observed at the MINI station (see Fig. 2 for MINI location) with those ones recorded at the reference rock SSB station. We show here only one example of that comparison. Figure 12 displays the smoothed pseudo Wigner–Ville time–frequency distribution (Flandrin, 1993) for the north–south component of event T6. Records were aligned to fit their respective first P-wave arrival time. Around 400 sec after the first arrival,
energy is concentrated below 0.2 Hz for the signal at rock. For the site, energy is dispatched in two distinct frequency bands: one below 0.2 Hz, as is the case for the rock site, and one between 0.2 and 0.4 Hz. Since no energy is impinging in the last frequency band, only some wave propagation phenomena within the basin can explain this focus of energy above 0.2 Hz. Furthermore, if we assume effects of regional heterogeneities on ground motion above 0.2 Hz, one should observe a time correlation of energy jumps observed for signals recorded at the site and the rock location. In order to see such a time correlation, we have computed the cumulative energy over time of bandpass-filtered (0.2–0.5 Hz) signals. Cumulative energies have an arbitrary relative scale. Figure 12 (bottom) shows that, most often, energy jumps have an arbitrary relative scale. Cumulative energies have an arbitrary relative scale. Figure 12 (bottom) shows that, most often, energy jumps observed at station MINI do not correspond to the ones observed at station SBB. This provides an additional argument in favor of basin-induced surface-wave effects for the origin of off-source waves. However, some scattering from regional heterogeneities may exist, as observed in the Alps by Gaffet et al. (1998), but it may be overshadowed by the local effects of the basin.

Assuming basin-induced surface waves, the proportion of energy carried in the radial plane should thus provide the proportion of Rayleigh waves of the basin-induced wave field. Such energy proportions for each event are indicated in Table 2. On average the basin-induced wave field is therefore composed of 60% Rayleigh waves and 40% Love waves. However, if one considers only the energy carried by horizontal components, this proportion reduces to 50% Rayleigh waves and 50% Love waves (Table 2).

Regional Events

As is the case for teleseismic events, the off-source arrivals below 1 Hz from regional events exhibit the same general patterns in terms of wave density, apparent inclination, proportion of radial energy, and energy density as a function of frequency and backazimuth. The average values of estimates for all events are shown in Figure 13. Identified arrivals cover a wider range of frequencies (from 0.2 to 0.9 Hz) than for teleseismic events. This difference probably comes from the broader frequency content of the regional events. However, the most energetic coherent waves are generally around 0.3 Hz, except for arrivals coming from 90° N. When reporting energy as a function of backazimuth on the contour map of Grenoble, one can see that off-source waves are coming predominantly from the northwest (Chartreuse Massif), east (Belledonne Massif), and southwest. Since we did not have adequate reference stations for regional events, we could not compare the coda observed at rock and site locations in order to discriminate between regional and basin effects. According to the observations for teleseismic events, we assume that most of the coherent energy is carried by basin-induced waves. In such a case, the southern diffractor corresponds to a small subbasin delimited by a buried bedrock high. This diffractor was not (or was only slightly) observed when studying teleseismic events. Values of apparent inclination and apparent velocities suggest, as for teleseismic events, that surface waves dominate off-source arrivals with a mean proportion of Rayleigh waves of about 60% and Love waves of 40% (Table 2). When considering only the energy of horizontal components, this proportion is 50% Rayleigh waves and 50% Love waves (Table 2).

Identification and Stability of Diffractors at High Frequency (Above 1 Hz)

Local Events

It was previously explained that the impinging wave field of four of the six local events was most probably diffracted at the basin borders, leading to a deviation up to 60° from the theoretical backazimuth. We considered, for these four events, as source waves all arrivals during the first 12 sec coming from backazimuths between 100° and 210° N. After this arbitrary time threshold, all late arrivals are considered as off-source-type ones. Since this time threshold is large compared to the epicentral distance (20 km), source waves identified within the previously defined backazimuth range should include some diffracted off-source waves too.

For events L3 and L5, source waves are the ones that, during the first 15 sec, come from the source area within a backazimuth deviation of ± 30° from the theoretical backazimuth. We disregarded arrivals with a frequency content above 10 Hz because their aliasing domain corresponds to a large range of velocities of interest, from 200 to 1000 m/sec. Below 10 Hz, we get the same distribution of backazimuth and energy regardless of the event. Figure 14 summarizes the results. As is the case for the regional events, two main energetic peaks in the frequency range of 2–4 Hz come from the Belledonne Massif and especially from the southwest.
subbasin. At frequencies above 6 Hz, waves come from everywhere, but with low energy. The high values of apparent inclination and the distribution of apparent velocities that lie between theoretical phase velocities of Rayleigh and Love waves suggest that off-source arrivals are mainly composed of surface waves. These surface waves are composed of 60% Rayleigh waves and 40% Love waves (Table 2). If one considers only the energy of horizontal components, this proportion becomes 43% Rayleigh waves and 57% Love waves (Table 2). As is the case for teleseismic events, we compared records at rock and sediment sites. The reference rock station OGMU is at the same location as the one used by Lebrun et al. (2001), about 3 km from the center of the array. This station is equipped with an accelerometric CMG5 sensor and...
is part of the French permanent accelerometric network (RAP). CMG5 signals were integrated and highpass filtered above 1 Hz. Figure 15 displays the smoothed pseudo Wigner–Ville time–frequency distribution of rock (top) and sediment (middle) records and the cumulative energy over time (bottom) for the north–south component of event L5. At rock, the incident signal is short in duration (5 sec) and the energy is concentrated between 1 up to at least 10 Hz. At the site, the signal duration is much longer and its energy concentrated between 1 and 4 Hz. When comparing cumulative energy over time of bandpass-filtered (from 1 to 10 Hz) signals, a few seconds after the first arrival, no more energy is observed at the rock site, whereas the energy still increases at the sediment site. This argues that off-source waves are generated within the basin. Other studies in other sites (Satoh et al., 2001) have shown that, for local and regional earthquakes, P- and S-wave portions have a relatively short duration inside the basins and that the later part of signals consist mainly of basin-induced surface waves. Therefore, we suggest that off-source waves are basin-induced waves propagating within the topmost surficial layers.

Regional Events

For frequencies above 1 Hz, we could not observe consistent shape in the distribution of backazimuth and energy densities for the three regional events we could analyze (events R1, R2, and R4). The 90°–100° N and 180°–200° N ranges of backazimuth appear in each case, but the corresponding frequency content is different from one event to another. Nevertheless, we plot the average results in the same way as for previous events in Figure 16. Most of the energy comes from the eastern and southwestern directions, with frequencies ranging from 2 to 4 Hz. At 3 Hz, apparent velocity values of about 300 m/sec agree with the S-wave mean velocity of the most surficial layers, whereas below 3 Hz, apparent velocities slightly increase. The high values of apparent inclination and the distribution of apparent velocities that lie between theoretical phase velocities of Rayleigh and Love waves again suggest that off-source arrivals are mainly composed of surface waves. Surface waves are equally composed of Rayleigh and Love waves regardless of the method used to estimate these proportions (Table 2). According to comparison between rock and site records performed for local events, we postulate that these surface waves are generated at basin borders too.

Importance of Basin-Induced Surface Waves

In order to evaluate the relevance of our results, we first try to calculate the proportion of energy we analyzed ($E_{\text{total}}$) compared to the entire energy of the seismogram. Second, we estimate the proportion between off-source and source energy, as an indication of the importance of edge-generated surface-wave phenomena. Cumulative energy over time was used to calculate these proportions. The cumulative energy $E_o(t)$ of the entire seismogram was obtained by summation of the square of the three-component [$x(t)$, $y(t)$, $z(t)$] amplitudes scaled by the time sampling $dt$:

$$E_o(t) = \int_0^t [x(\tau)^2 + y(\tau)^2 + z(\tau)^2]d\tau,$$

with $t$ the time.

For each time $t_o$, we compute the total analyzed energy that is the sum of the off-source and source energy:

$$E_{\text{total}}(t_o) = E_{\text{off-source}}(t_o) + E_{\text{source}}(t_o).$$

Thus energy may then be summed over all the frequency bands where it is coherent, and we can in that way get a curve $E_{\text{total}}(f)$ that may be compared with the total energy $E_o(f)$. At the end of the seismogram defined by the time $T$, the proportion of analyzed energy $E_{\text{total}}$ compared to the total energy of the seismogram $E_o$ is $E_{\text{total}}(T)/E_o(T)$ and the proportion of off-source energy compared to analyzed energy is $E_{\text{off-source}}(T)/E_{\text{total}}(T)$.

We plot in Figure 17 cumulative energies of regional
event R1 for the range of frequencies from 0.1 to 1.0 Hz. As this example is quite representative of what is observed for other events, we will not present all the curves. Off-source waves appear very shortly after the first source arrival. At the end of the analysis, the proportion of analyzed energy compared to the seismogram total energy is 80% and the proportion of off-source energy compared to the total analyzed energy is 79%. Table 2 lists for all events the proportion of final analyzed energy compared to the seismogram energy and the proportion of off-source wave energy compared to the total analyzed energy. Regardless of the class of events, the MUSIC analysis allows us to identify about 40% of the total energy, of which 80% is associated to basin-induced energy.

The variations of $E_{\text{total}}/E_{o}$ and $E_{\text{off-source}}/E_{\text{total}}$ as a function of frequency must specify the range of the frequency content investigated and the frequency range dominated by off-source waves. These proportions are depicted on Figures
Figure 11. (a) Seismic $P$- and $S$-wave velocity profiles considered at the array location; (b) phase velocity of the two first modes of Rayleigh waves (thin lines) and Love waves (dashed lines). The thick line is the surface-wave phase velocity curves obtained by Bettig et al. (2001) from dense array measurements of ambient noise in the area of our experiment.

18, 19, and 20, as are their mean values (Table 2), for teleseismic, regional, and local events, respectively. In the figures, gray shaded zones indicate the distribution of the total energy of seismograms with an arbitrary relative scale as a function of frequency. In addition we described MUSIC estimates by using a BF grid with grid steps of $10^6 \times 0.1$ Hz when studying above 1 Hz and $10^6 \times 0.05$ Hz when looking at estimates below 1 Hz. For any given frequency band, the

Figure 12. Smoothed pseudo Wigner–Ville (SPWV) time–frequency distribution and energy spectral density of the unfiltered north–south component of event T7 recorded at the SBB reference station (top) and at the MINI station (middle). Cumulative energy (with an arbitrary relative scale) over time (bottom) of bandpass filtered (0.2–0.5 Hz) signals. Black bars indicate the center time of the time windows processed with array analysis.
ground motion was performed for local and regional events. Investigation of the full range of frequencies concerned with events below 0.3 Hz (events T2, T3, T5, and T8). A better understanding of the basin (site effects). For teleseismic events (Fig. 18), narrow frequency bands underline the strong filtering effects of frequency. In contrast, the Chartreuse Massif appears as other diffractors in the southwest and northwest directions, and the Belledonne Massif, is involved for all frequencies and classes of events. It can be simply explained by the proximity of the array to this basin border. In contrast, the two other diffractors in the southwest and northwest directions are not seen for all event types. On the one hand, both regional and local events suggest a large region of high-frequency diffraction that is southwest of the array. On the other hand, the Chartreuse Massif seems to play a major role in generating surface waves at frequencies below 1 Hz. Because the Chartreuse Massif border and the southwest sub-basin structure are roughly located at the same distance from the array, we may invoke the shape of the structures, especially surface curvatures of basin edges, to explain such diffraction patterns. From the array location, the southwest basin looks like a lens and should send a large number of edge-generated wave trains toward the array site, regardless of frequency. In contrast, the Chartreuse Massif appears as a linear structure of large extent. One can imagine that only waves of large wavelength, not sensitive to the massif-border surface curvatures, will be sent toward the array location. We are aware that we only investigated wave motion propagation in a small part of the basin; we are quite sure that putting such a dense array in other places would certainly identify the western Vercors Massif as a strong diffractor too. Besides, our previous interpretation is only focused on the local basin structure effects. As was outlined by Gaffet et al. (1998), regional heterogeneities can contribute to reading.

### Table 2

Comparison of Energies

<table>
<thead>
<tr>
<th>Event</th>
<th>Frequency Range (Hz)</th>
<th>$E_{\text{total}}/E_o$ (%)</th>
<th>$E_{\text{off-source}}/E_{\text{total}}$ (%)</th>
<th>$E_{\text{off-source radial + vertical}}/E_{\text{total}}$ (%)</th>
<th>$E_{\text{off-source radial}}/E_{\text{radial + vertical}}$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>L1</td>
<td>56.</td>
<td>47.</td>
<td>60 ± 13</td>
<td>42. ± 12</td>
<td>65. ± 15</td>
</tr>
<tr>
<td>L2</td>
<td>12.</td>
<td>90.</td>
<td>64 ± 7</td>
<td>38. ± 8</td>
<td>65. ± 7</td>
</tr>
<tr>
<td>L3</td>
<td>1–10</td>
<td>56.</td>
<td>96.</td>
<td>54 ± 10</td>
<td>47. ± 10</td>
</tr>
<tr>
<td>L4</td>
<td>22.</td>
<td>77.</td>
<td>59 ± 10</td>
<td>45. ± 9</td>
<td>58. ± 12</td>
</tr>
<tr>
<td>L5</td>
<td>37.</td>
<td>98.</td>
<td>54 ± 13</td>
<td>45. ± 12</td>
<td>72. ± 18</td>
</tr>
<tr>
<td>L6</td>
<td>46.</td>
<td>42.</td>
<td>67 ± 16</td>
<td>43. ± 14</td>
<td>75. ± 14</td>
</tr>
<tr>
<td>Mean</td>
<td>38. ± 18.</td>
<td>75. ± 25</td>
<td>60 ± 12</td>
<td>43. ± 11</td>
<td>67. ± 14</td>
</tr>
<tr>
<td>R1</td>
<td>21.</td>
<td>79.</td>
<td>53 ± 20</td>
<td>58. ± 18</td>
<td>84 ± 17</td>
</tr>
<tr>
<td>R3</td>
<td>12.</td>
<td>72.</td>
<td>55 ± 9</td>
<td>50 ± 10</td>
<td>81. ± 8</td>
</tr>
<tr>
<td>R4</td>
<td>13.</td>
<td>—</td>
<td>58 ± 8</td>
<td>42. ± 9</td>
<td>58. ± 9</td>
</tr>
<tr>
<td>Mean</td>
<td>—</td>
<td>—</td>
<td>55 ± 13</td>
<td>50 ± 13</td>
<td>74 ± 12</td>
</tr>
<tr>
<td>R1</td>
<td>80.</td>
<td>79.</td>
<td>56 ± 12</td>
<td>51 ± 12</td>
<td>78 ± 12</td>
</tr>
<tr>
<td>R2</td>
<td>41.</td>
<td>81.</td>
<td>61 ± 6</td>
<td>50 ± 7</td>
<td>62. ± 9</td>
</tr>
<tr>
<td>R3</td>
<td>47.</td>
<td>82.</td>
<td>63 ± 7</td>
<td>49 ± 9</td>
<td>56. ± 10</td>
</tr>
<tr>
<td>R4</td>
<td>65.</td>
<td>77.</td>
<td>60 ± 7</td>
<td>51 ± 8</td>
<td>68 ± 10</td>
</tr>
<tr>
<td>Mean</td>
<td>58. ± 18.</td>
<td>80. ± 3.</td>
<td>60 ± 8</td>
<td>50 ± 9</td>
<td>66 ± 11</td>
</tr>
<tr>
<td>T1</td>
<td>80.</td>
<td>85.</td>
<td>56 ± 14</td>
<td>50 ± 14</td>
<td>78 ± 12</td>
</tr>
<tr>
<td>T2</td>
<td>27.</td>
<td>91.</td>
<td>55 ± 16</td>
<td>49 ± 17</td>
<td>80 ± 10</td>
</tr>
<tr>
<td>T3</td>
<td>31.</td>
<td>87.</td>
<td>60 ± 14</td>
<td>50 ± 12</td>
<td>70 ± 17</td>
</tr>
<tr>
<td>T4</td>
<td>44.</td>
<td>93.</td>
<td>65 ± 6</td>
<td>52 ± 8</td>
<td>56 ± 7</td>
</tr>
<tr>
<td>T5</td>
<td>23.</td>
<td>91.</td>
<td>61 ± 9</td>
<td>52 ± 12</td>
<td>78 ± 10</td>
</tr>
<tr>
<td>T6</td>
<td>51.</td>
<td>85.</td>
<td>63 ± 10</td>
<td>50 ± 12</td>
<td>55 ± 11</td>
</tr>
<tr>
<td>T7</td>
<td>49.</td>
<td>75.</td>
<td>61 ± 9</td>
<td>51 ± 10</td>
<td>62 ± 12</td>
</tr>
<tr>
<td>T8</td>
<td>39.</td>
<td>96.</td>
<td>62 ± 8</td>
<td>51 ± 12</td>
<td>62 ± 10</td>
</tr>
<tr>
<td>Mean</td>
<td>43. ± 18.</td>
<td>88. ± 7</td>
<td>61 ± 11</td>
<td>51 ± 12</td>
<td>68 ± 11</td>
</tr>
</tbody>
</table>

Proportion of total analyzed energy ($E_{\text{total}}$) compared to total energy of seismograms ($E_o$); proportion of off-source total analyzed energy ($E_{\text{off-source}}/E_{\text{total}}$) compared to total analyzed energy ($E_{\text{total}}$); proportion of off-source energy carried in the radial plane ($E_{\text{radial}}/E_{\text{radial + vertical}}$) compared to off-source total energy; proportion of off-source radial energy compared to off-source horizontal energy ($E_{\text{off-source radial}}/E_{\text{off-source radial + transversal}}$); proportion of radial energy compared to energy carried in the radial plane ($E_{\text{off-source radial}}/E_{\text{radial}}$).
load the basin with seismic energy. Nevertheless, in our study, the properties of off-source wave fields hint at basin structure effects only.

For all classes of events, the main energetic azimuthal contributions do not depend on event backazimuth and on epicentral distance. Our event azimuthal coverage is obviously not the best, but Caserta et al. (1998), Chavez-Garcia et al. (1999), and Rovelli et al. (2001) also observed in other small-size basins a similar stability of directions for basin-edge-induced waves. This suggests that the depth and geometry of the basin strongly constrain the wave motion patterns.

We also observed that basin-edge-induced waves at low and high frequencies are mainly composed of surface waves.
with a mean proportion of Rayleigh waves of about 60% and a mean proportion of about 40% for Love waves. The proportion of Rayleigh waves was estimated comparing radial and vertical energy with total three-component energy. In order to get an idea of the distribution of energy with respect to component type, we calculated the proportion of radial energy compared to horizontal energy and the proportion of radial energy compared to energy carried in the propagation plane. Proportions are indicated in Table 2. As previously indicated, horizontal energy is, on the mean, equally carried by radial and transverse components. The vertical component carries 2–2.5 times less energy than each horizontal component.

When studying teleseismic events we observed largely
Contribution of Dense Array Analysis to the Identification and Quantification of Basin-Edge-Induced Waves, Part II

Figure 15. Smoothed pseudo Wigner–Ville (SPWV) time–frequency distribution and energy spectral density of the north–south component of event L5 recorded at the OGMU reference station (top) and at the MINI station (middle). Signals are highpass filtered above 1 Hz. Unscaled cumulative energy over time (bottom) of bandpass filtered (1–10 Hz) signals.

dispersed waves with high values of apparent velocity at 0.3 Hz. We argued that either a large uncertainty in velocity estimate at that frequency, some lateral variations of the sub-stratum topography, or some very particular correlation phenomena could explain the observed distribution of velocity. From dense array measurements of noise located at the same place as ours, Bettig et al. (2001) compared dispersion curves obtained from the spatial autocorrelation method (SPAC) method with those derived from the $f$-$k$ technique (Fig. 21). From 0.5 to 1 Hz, $f$-$k$ estimates are systematically much higher than the SPAC ones and the expected values (Fig. 11b). Below 0.5 Hz, $f$-$k$ estimates are not reliable. Since the $f$-$k$ method is, like MUSIC, a delay estimation problem (article 1), and since Bettig et al. (2001) did not use the same array configuration as ours, we suggest that the high values of phase velocity around 0.5 Hz simply reveal the existence of several waves propagating at similar velocities in different directions.

Consequences on Amplification Values

Chavez-Garcia et al. (1999) concluded in the case of Parkway valley that it was not possible to separate 1D and 2D effects because of their contribution to the same frequency band. In our study, assuming that direct waves include 1D effects and the off-source waves are related to 2D/3D effects (basin-induced waves), we computed $E_{\text{total}}/E_o$ and $E_{\text{off-source}}/E_{\text{total}}$ as a function of frequency. We observed that 1D and 2D/3D effects contribute to the same amplified frequency band from 0.2 to 5 Hz. One of the main features of site effects in the Grenoble area is the flat amplification observed from 0.2 to 5 Hz of the spectral ratio relative to the reference site. This amplification is much different from that determined by the receiver function and 1D estimates. Because of the lack of a reference station for local and regional events, we could not compute the spectral ratio relative to the reference site. However we used records from the rock SSB station previously mentioned to get the amplification for teleseismic events. Figure 22 displays results for two teleseismic events and underlines once more the important amplification on the standard spectral ratio.

According to the rough approximation about the nature of 1D and 2D/3D effects, the horizontal spectral motion $H$ observed on sediments can be decomposed as

$$H = H^{1D} + H^{2,3D},$$

with $H^{1D}$ corresponding to vertical reverberation of $S$ waves and $H^{2,3D}$ corresponding to 2D/3D basin-induced waves. This may be rewritten as

$$H = (1 + \alpha)H^{1D},$$

with $\alpha = H^{2,3D}/H^{1D}$. The value of $\alpha$ is related to the rate of production of edge-generated surface waves.

In our study, this value can be estimated by computing the ratio $\tilde{\alpha} = H^{\text{off-source}}/H^{\text{source}}$, with $H^{\text{off-source}}$ the Fourier

Log. scale

Energy spectral density

Log. scale

Energy spectral density

Event L5 (NS comp.)

Time [s]

0 10 20 30

SPWV time–frequency distribution

0 1 2 3

Frequency [Hz]

0 2 4 6 8 10

Event L5 NS comp.

ROCK

SEDIMENT

Energy spectral density

Log. scale

Energy spectral density

Time [s]

0 10 20 30

SPWV time–frequency distribution

0 1 2 3

Frequency [Hz]

0 2 4 6 8 10

Event L5 (NS component)

Time [s]

0 10 20 30

Cumulative energy

0 50 100 150

0 200

0 10 20 30

SPWV time–frequency distribution

0 1 2 3

Frequency [Hz]

0 2 4 6 8 10

Event L5 NS comp.

ROCK

SEDIMENT

Energy spectral density

Log. scale

Energy spectral density

Time [s]

0 10 20 30

SPWV time–frequency distribution

0 1 2 3

Frequency [Hz]

0 2 4 6 8 10

Event L5 (NS component)

Time [s]

0 10 20 30

Cumulative energy

0 50 100 150

0 200
amplitude spectrum of the off-source arrivals and $H_{\text{source}}$ the Fourier amplitude spectrum of the source arrivals. We calculated $\tilde{\alpha}$ for six teleseismic events (events T1, T3, T4, T5, T6, and T7). Details about the calculation of $\tilde{\alpha}$ values can be found in Cornou and Bard (2003). Values of $\tilde{\alpha}$ are displayed for the frequency range from 0.1 to 1 Hz in Figure 23. The lack of $\tilde{\alpha}$ values at some frequencies simply indicates that no source waves were identified for these frequencies, and thus $\tilde{\alpha}$ could not be estimated. One can see that values range between 0 and 6, indicating a proportion of off-source waves compared to total ones varying from 0% to 85%.
Using the theoretical 1D transfer function computed for the velocity model depicted in Figure 11 and the \( \tilde{E} \) estimate, Cornou and Bard (2003) have derived an estimate of the site-
to-bedrock spectral ratio including the effects of locally gen-
erated surface waves. When almost all the energy could be
analyzed with the MUSIC technique (\( E_{\text{total}}/E_\circ \approx 100\%)\), they
have observed a very good match between the estimated
standard spectral ratio values and the actual values. When
only a portion of energy is analyzed with array analysis, they
have assumed that, above the resonance frequency of the
basin, the noncoherent energy (i.e., that could not be inves-
tigated with array technique) comes from basin-induced
waves. On the contrary, below 0.25 Hz, they have assumed
that the noncoherent energy comes mainly from direct
waves. By doing so, upper (above 0.25 Hz) and lower (below
0.25 Hz) standard spectral ratio estimates can be provided.
We present in Figure 22 results for two teleseismic events.
Open circles correspond to the standard spectral ratio esti-
mated values based on the 1D response shown by thin lines,
crosses to upper estimated values (above 0.25 Hz), and tri-
angles to lower estimated values (below 0.25 Hz) when the
whole energy could not be explained. Spectral ratio esti-
mates fit rather well the actual values and, above 0.25 Hz,
strongly suggest a major role for basin-induced surface
waves.

Besides, the receiver function is very similar here to the
1D transfer function (Fig. 22). It suggests that the non-
reference-site H/V ratio is not sensitive or is much less sen-
titive to basin-induced waves than the standard spectral ratio
technique. Thereafter, for site amplification evaluation pur-
poses in such small-size structures, we strongly recommend
the site-to-bedrock ratio over the whole duration and draw
attention to the nonrelevance of non-reference-site techniques.

Conclusions

In this article, we have presented the systematic analysis
of 18 events recorded by a temporary three-component seis-
mometer array located in the city of Grenoble. A combined
procedure based on coherence of signals between array sen-
sors and MUSIC analysis (article 1) allowed us to precisely
investigate the whole record motion and to isolate wave
fields not coming from the source backazimuth (off-source
wave field) and its associated energy. The study leads to
the following main conclusions:

1. Locally generated surface waves are seen only above the
fundamental frequency of the basin and comprise almost
all off-source wave fields.

2. The spatial stability of diffractors outlines the importance
of structure geometry and frequency input motion in
shaping wave fields, regardless of source epicenter and
azimuth.

3. Almost 40\% of the seismogram energy was coherent
enough to be analyzed. Among this coherent part, 80\%
is associated to laterally propagating waves; in other
words, the basin-induced surface waves carry 4 times
more energy than the waves coming from the source
backazimuth. On average, the surface wave field is
composed of around 60\% Rayleigh waves and 40\% Love
waves. If one considers only the energy of horizontal
components, this proportion becomes 50\% Rayleigh
waves and 50\% Love waves.

4. Evaluation of energy carried by each identified wave train
allows us to compare the relative contribution of 1D and
2D/3D effects in the ground motion: at low frequencies
(below 0.5 Hz) for teleseismic events, the discrepancy
between the standard spectral ratio and 1D transfer func-
tion (or possibly H/V ratios) could be completely ex-
plained by basin-induced surface-wave effects.

Additional observational studies are needed in different
types of valleys to specify whether the steady response of
the basin is peculiar to the Grenoble basin (or more generally
to alpine valleys) or not. Numerical simulations should be
done to elucidate some basic features of our observations.
Our feeling about the way to evaluate the relative contri-
bution of 1D and 2D/3D effects echoes in a sense the work
of Field (1996), who suggested applying factors to 1D pre-
dictions at some frequencies to take into account overam-
plification induced by multidimensional site effects.

Acknowledgments

We thank R. Guiguet, B. Bettig, M. Bouchon, H. Havenith, and D.
Hatzfeld for their help in the field and people who kindly hosted some
seismological stations in their yards. We are also very thankful to H.
Kawase, K. Kudo, M. Horike, and an anonymous reviewer for their thorough
and fruitful comments on the manuscript. Most of the computations were
performed at the Service Commun de Calcul Intensif de l’Observatoire de
Grenoble (SCCI). This work was supported by the Pôle Grenoblois des
Risques Naturels and the Programme National de Recherche sur la Prévi-
Figure 18. Energy ratios for teleseismic events. For each of the eight analyzed events, several ratios are displayed: the proportion of analyzed energy compared to the whole energy of the seismogram (top, dots) and the proportion of off-source energy compared to analyzed energy (bottom, bars) as a function of frequency. The gray shaded area indicates the variation over frequency of the total energy of the seismogram. Dashed lines indicate overall values of $\frac{E_{\text{total}}}{E_o}$ and $\frac{E_{\text{off-source}}}{E_{\text{total}}}$ as listed in Table 2.
Figure 19. Energy ratios for regional events. For each of the four analyzed events, several ratios are displayed: the proportion of analyzed energy compared to the whole energy of the seismogram (top, dots) and the proportion of off-source energy compared to the analyzed energy (bottom, bars) as a function of frequency. The gray shaded area indicates the variation over frequency of the total energy of the seismogram. Dashed lines indicate overall values of $E_{\text{total}}/E_o$ and $E_{\text{off-source}}/E_{\text{total}}$ as listed in Table 2.
Figure 20. Energy ratios for local events. For each of the six analyzed events, several ratios are displayed: the proportion of analyzed energy compared to the whole energy of the seismogram (top, dots) and the proportion of off-source energy compared to the analyzed energy (bottom, bars) as a function of frequency. The gray shaded area indicates the variation over frequency of the total energy of the seismogram. Dashed lines indicate overall values of $E_{\text{total}}/E_o$ and $E_{\text{off-source}}/E_{\text{total}}$ as listed in Table 2.
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Figure 21. Surface-wave dispersion curves obtained by Bettig et al. (2001) on array measurements of microtremors. The array was located at the same place as ours, and the methods used are the SPAC modified technique and a semblance-based $f-k$ analysis.

Figure 22. Estimate of amplification evaluated using the standard spectral ratio technique (SSR), the receiver function method (HVSR), and the theoretical 1D transfer function of $SH$-incident plane waves (1D). The quadratic mean horizontal component was used to evaluate the average standard spectral ratio and HVSR. See text for explanation of open circles, crosses, and triangles. The SSB station is the reference station.

Figure 23. $\tilde{\alpha}$ estimates for six teleseismic events within the frequency range from 0.1 to 0.5 Hz.

References


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Manuscript received 21 June 2002.