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Subsoil Structure Using SPAC Measurements along a Line

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Abstract The SPAC (SPatial AutoCorrelation) method was proposed almost 50 years ago by Aki (1957). This method allows a phase-velocity dispersion curve to be obtained from microtremor measurements using an array of stations arranged in a circle. The subsoil structure is subsequently derived from the inversion of that dispersion curve. In this article we show that it is possible to get similar results using microtremors recorded along a line. We use microtremor records obtained by using four broadband seismographs disposed along a line, with different interstation spacings (5, 10, 20, and 40 m). Our data are precessed by using the standard SPAC procedure, with the exception of the azimuthal average. The final subsoil structure as determined from the inversion of the phase-velocity dispersion curve shows excellent agreement with previous results at the site of our measurements. Our results suggest that the use of the SPAC method is not restricted to a particular geometry of the array, provided that the basic requirement of stationarity is fulfilled.

Introduction

The specification of site effects frequently relies on the determination of the soil profile at a site, followed by the forward computation of the local transfer function. The reliability of the result depends to a large extent on the reliability of the determined soil profile. The more reliable techniques are usually too expensive (exploration using boreholes) or unsuitable (e.g., explosion seismology is out of the question in an urban environment) to be of general application. For these reasons, alternative methods have been devised. Currently, the most popular technique involves the computation of horizontal to vertical ratios of Fourier spectra of microtremor amplitude (Lermo and Chávez-García, 1994). Although there is a general agreement that this technique allows the resonant frequency at a given site to be identified with confidence, the debate continues regarding the reliability of the maximum amplitude obtained from those ratios. Moreover, even if we accept that the resulting transfer function is reliable, we do not get information on subsoil structure.

Other techniques are based on surface-wave analysis. Surface waves are popular because of the close link between their phase or group velocity dispersion and the properties of the subsoil materials. The inversion of dispersion curves has been used to investigate geological structure at the crustal scale (e.g., Mokhtar *et al.*, 1988) or at the very shallow scale for geotechnical applications (e.g., Chávez-García, 1995). A recent proposal (Louie, 2001) analyzes phase-velocity dispersion curves by using microtremor measurements obtained with seismic exploration gear.

Almost 50 years ago, Aki (1957) proposed what was then an innovative technique to obtain information on the subsoil structure at a site, the Spatial Autocorrelation, or SPAC technique. Aki (1957) showed that the azimuthal average of the cross-correlation function at a fixed interstation distance allowed the phase-velocity dispersion curve at a site to be obtained. In applications of the SPAC method it is usual to arrange the seismographs on a half-circle, with a central station. This geometry allows the sampling of different azimuths between pairs of stations at the same distance to compute the azimuthal average (e.g., Ferrazzini *et al.*, 1991; Chouet *et al.*, 1998). In a previous article (Chávez-García *et al.*, 2005), we presented an extension of SPAC, where phase-velocity dispersion curves were obtained from data recorded using a temporary seismic array with a very irregular geometry. (The use of the SPAC method with data from an array with irregular geometry has also been presented in DeLuca *et al.*, 1997, and Ohori *et al.*, 2002, and discussed in Chávez-García *et al.*, 2005.) The good results obtained led us to perform an additional experiment using a line of stations at the same location. Thus, the geometry of the array is as different as possible from a circle. In this article we present the results of that analysis, which support the use of the SPAC method without constraints in the geometry of the array.

The SPAC Method

Aki (1957, 1965) proposed a technique to deduce a phase-velocity dispersion curve from the microtremors recorded by a seismic array. He established that the spatial cross-correlation coefficient as a function of frequency for a given interstation distance, r, and angular frequency, ω,

$\rho(r, \omega)$, averaged over many different azimuths, τ, can be written as

$$\rho(r,\omega) = \frac{1}{2\pi\phi(r = 0,\omega)} \int_0^{2\pi} \phi(r,\theta,\omega) d\theta = J_0\left(\frac{r\omega}{c}\right), \qquad (1)$$

where $\phi(r = 0, \omega)$ is the average autocorrelation function at the center of the array, $\phi(r, \theta, \omega)$ is the cross-correlation function between the record at a site at coordinates (r, θ), and the record obtained at the station at the origin, c, is the phase velocity at frequency ω at the site, and J_0 is the Bessel function of first kind and order zero. The wave field is assumed to consist of surface waves propagating with equal power in all directions. The only unknown in the preceding equation is the phase velocity for each frequency, which can be obtained from the inversion of the observed correlation coefficients. In turn, it is possible to invert that phase-velocity dispersion curve to obtain a shear-wave velocity profile with standard techniques (e.g., Herrmann, 1987). The details of the method have been published several times (e.g., Asten, 1976; Chouet *et al.*, 1998).

The approach tested in this article is similar to that described with the exception of the azimuthal averaging of the cross-correlation coefficients. A hypothesis in the method is that microtremors consist of surface waves propagating along different directions with comparable power. Thus, if long-enough records are obtained at a single station pair, the recorded motion can be considered to have sampled waves propagating along many different directions. Under this hypothesis, the equations and results that Aki (1957) obtained using the azimuthal average of the spatial cross-correlation coefficients can be validly applied, substituting a temporal average of those same coefficients computed for a single station pair. However, a possible difficulty with equation (1), especially when only two signals are processed, is that the normalization only uses information from the record obtained at one of the intervening stations, $\phi(r = 0, \omega) = \phi_1(\omega)$. When the power is strictly the same at the two stations, this is no problem and the cross-correlation coefficients are bounded at unity. If, however, small variations in the power of the signals between the two stations appear, then equation (1) may lead to values slightly larger than unity, as was the case in Chávez-García *et al.* (2005). We have corrected this behavior by using $\sqrt{\phi_1(\omega)}\ \sqrt{\phi_2(\omega)}$, in the denominator, instead of $\phi_1(\omega)$. Here, $\phi_1(\omega)$ and $\phi_2(\omega)$ are the average autocorrelation functions of the two signals involved. Of course, even if all cross-correlation coefficient values are bounded at 1, we could still have average plus one standard deviation values larger than unity. We have not imposed an additional restriction on the standard deviation values because we will use those only as an estimate of the scatter of the mean values. They will allow us to estimate the error of the estimated phase-velocity value from the inversion scheme. In the next paragraphs we will describe the setup for the measurements and the obtained results. The details of the procedure followed in this article are the same as were used in Chávez-García *et al.* (2005), of which this article is a follow-up. The interested reader is invited to consult that article and the references cited therein.

Data Acquisition and Processing

Parkway basin is a small (400 m wide), shallow alluvial valley located in Wainuiomata, North Island, New Zealand (Fig. 1). Very soft clays overlay a greywacke, giving as a result a large velocity contrast between the sediments and the bedrock, inferred to be at 70 m below the ground surface by Duggan (1997). The shear-wave velocity varies from

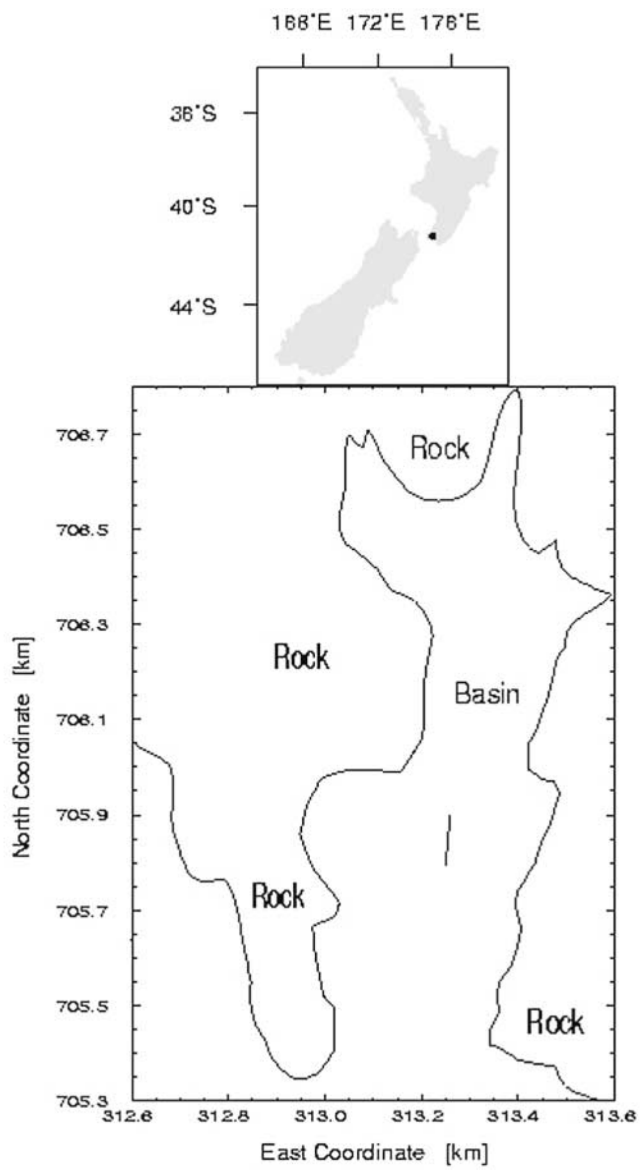

Figure 1. Location map of Parkway valley, Wainuiomata, New Zealand. The solid line shows the limits between soft soil sediments and underlying bedrock. The short line within the basin indicates the location of the linear array measurements.

about 80 m/sec at the surface to 200 m/sec at 14 m depth. Stephenson and Barker (2000) found a low-velocity layer (133 m/sec) between 25.5 m and 31.5 m. Site response at this basin was studied with earthquake data by Chávez-García *et al.* (1999).

We used data recorded along a line at the center of Parkway basin (Fig. 1) in February 2003. Five broadband, Guralp, CMG40 seismometers were used, coupled to Orion recorders (manufactured by Nanometrics). These data loggers have 24-bit A/D converters and were synchronized by using a Global Positioning System (GPS) receiver at each station. We chose, as our recording site, a grassy stretch along the stream at Parkway. During the recording time, very little traffic was traveling in the nearby streets. All five stations were placed along a line, with a spacing of 5 m, and ground vibration was recorded for 30 min, with a sampling rate of 100 Hz. Then, the spacing between stations was increased to 10 m, recording again for 30 min. The spacing between stations was subsequently increased to 20 and 40 m, recording seismic noise for 30 min in each configuration. One of the stations at the end of the array failed during the experiment. For this reason, we will only use data from four stations. The interstation distances span the range from 5 to 120 m. We have considered all possible pairs of stations to compute the cross-correlation between records. Thus, data from the very first array, for example, provided data for interstation spacing of 5, 10, and 15 m. Although three-component data were recorded, we use only the vertical component records in this article, because we can then be sure that we are measuring Rayleigh wave dispersion.

From the 30 min of data recorded by each array, we selected 60-sec windows for the analysis. Overlapping between neighboring windows was 20 sec. For each simultaneous pair of 60-sec windows, the records were baseline corrected and cosine tapered. They were then filtered by a series of 8-pole Butterworth, bandpass filters (0.4 Hz bandwidth), and the average cross-correlation was computed. The central frequency of the filters varied from 0.3 Hz up to 10.9 Hz, with a 0.2-Hz step. The number of 60-sec windows that could be analyzed for each array was not constant. We could analyze 34 simultaneous windows for the first array (with 5-m spacing between stations), 56 for the second (10-m spacing), 77 for the third (20-m spacing), and 79 for the last one (40-m spacing).

Analysis

Our first objective is to verify whether our measurements satisfy our main hypotheses regarding the stationarity in time of the microtremor wave field and the presence of waves in many different directions. The first, obvious approach was to compute Fourier amplitude spectra for all the microtremor time windows used for the analysis, including the data obtained from all the linear arrays. The average Fourier amplitude spectra did not change throughout the experiment, neither did the scatter change for any of the four stations used. In terms of our cross-correaltion coefficients, Figure 2 shows those coefficients computed for all possible station pairs separated by 15 m. Each solid line shows the correlation coefficients computed for a single 60-sec window. Open circles with vertical bars indicate the average and standard deviation values at each frequency. Following our previous paper (Chávez-García *et al.*, 2005), we have assumed a Gaussian probability distribution, given that we average many estimates of the correlation coefficients. We observe the quite good agreement among all 34 windows, suggesting that the microtremor wave field is stationary in time. The average values describe a curve that resembles a zero order, first kind Bessel function, at least for frequencies larger than 2 Hz. This suggests that it may be possible to substitute temporal averaging for azimuthal averaging in the computation of correlation coefficients.

Given that our stations were arranged along a line, it is not possible to determine directions of propagation. However, we can use beamforming to evaluate whether a single propagation direction dominates our microtremor records by measuring apparent propagation velocities along our linear array. We have computed the beam for several of our 60-sec windows, in the direction of the line. We first filtered the traces with a series of bandpass filters (0.6 Hz wide) with central frequencies between 0.3 and 10.9 Hz. For each frequency we computed the beam for 51 apparent velocity values, between 5 and 1000 m/sec. Even if apparent velocity could be infinite (for propagation perpendicular to our line of stations), we did not feel it necessary to go to higher apparent velocity values. The reason is we do not intend to measure phase velocities, and the aim of this computation is only to identify whether individual directions of propagation

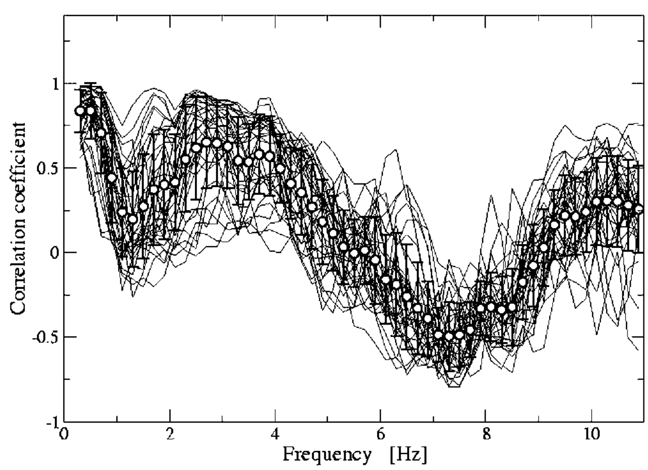

Figure 2. Correlation coefficients computed for all simultaneous windows recorded by the single station pair separated 15 m (from the 5-m spacing array). Each solid line corresponds to a single 60-sec window (a total of 34 windows are shown). Open circles with their error bars show the average plus or minus one standard deviation computed from the 34 individual measurements.

dominate our records. In addition, an apparent velocity of 1000 m/sec means a time delay of only half our sampling interval for the smaller interstation distance. Beamforming for each apparent velocity value produced a resulting trace for each bandpass filter. From those traces, we selected the one that had the largest peak amplitude and kept the corresponding value of apparent velocity for that frequency This computation was done twice, one for propagation down our line of stations and one for propagation up our line of stations. If, for some frequency band, a single direction of propagation dominates our records, we expect some coherency among the phase velocities obtained for the maximum beam in that frequency range. An example of the results is shown in Figure 3, for one of the microtremor windows recorded with the 40-m interstation spacing array. The values of velocity that produce the largest beam show a large scatter as a function of frequency, without coherency among different frequencies. This suggests that no single direction dominates the recorded microtremor wave field.

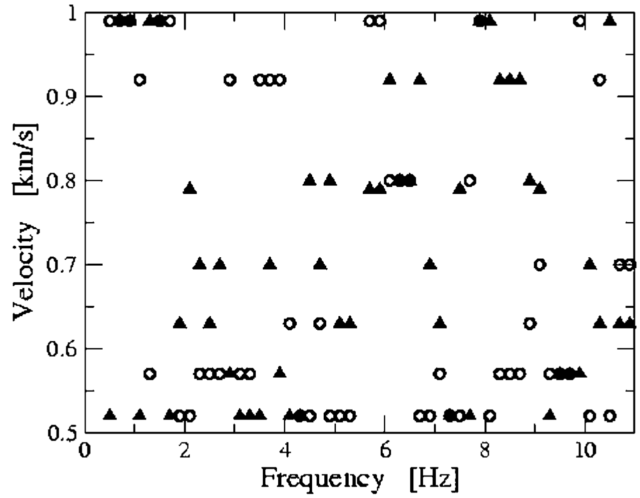

Figure 3. Apparent velocity as a function of frequency. Each symbol shows the apparent velocity for which the beam computed for a single microtremor window, bandpass filtered about the corresponding frequency, has the largest amplitude. We include results computed for beamforming down and up our line of stations (open circles and solid triangles). The results shown correspond to a sample window recorded by the four stations of the array with 40-m spacing between stations.

The final estimate of correlation coefficients for each distance is shown in Figure 4, computed using all possible station pairs at each distance. Figure 4 shows a steady decrease in frequency of the first zero crossing with increasing distance, from more than 9 Hz at 5-m interstation distance to about 2 Hz for the results at 120-m interstation distance. Our results show the expected shape, with the exception of the values at low frequencies, where a J_0 function should tend smoothly to unity at 0 Hz while our average correlation coefficients show a conspicuous hole between 1 and 2. We inverted the correlation coefficients shown in Figure 4 to obtain the argument of a J_0 function that best fits these data. The only unknown in the argument is the phase-velocity dispersion curve, $c(\omega)$. The inversion procedure was described in detail in Chávez-García *et al.* (2005). The theory of the SPAC method does not constrain the range of frequencies for which the information in the cross-correlation curves is useful. However, as shown by Henstridge (1979), the results are reliable in a region about the first zero crossing of the Bessel function. Henstridge (1979) proposed 0.4–3.2 as the limiting values of the Bessel function argument. Given that J_0 has its first zero crossing at a value of the argument of 2.4, these limits correspond to 0.17 and 1.3 times the value of the argument at the first zero crossing. Because of the problems shown by our results at low frequencies, we raised the lower limit and used only those correlation coefficients for values of the argument larger than 1.2, that is 0.5 times the value of the argument at the first zero crossing of J_0. As upper limit for the argument of the Bessel function we used 1.5, slightly larger than the value of 1.3 suggested by Henstridge, because our data do seem to contribute useful information at higher frequencies.

The phase-velocity dispersion curve obtained is shown in Figure 5, with open circles. A measure of the reliability of the results is that a single phase-velocity dispersion curve was able to satisfy the measurements for all our distances, from 5 to 120 m. The solid circles in Figure 5 show the phase-velocity values determined by Chávez-García *et al.* (2005). The range of validity of the dispersion curve is not easy to establish with the criterion described earlier, given that we used data from nine different distances. If we average our distances, weighting by the number of windows available at each distance, we obtain 36.7. If we use the value of the first zero crossing at 40 m to compute a limit at low frequencies, we obtain a limiting value of wavelength (λ) of 147 m, shown by a dotted line in Figure 5. Using the same value of first zero crossing to compute the lower wavelength limit gives 49 m, which coincides with the open circles in Figure 5 at 5 Hz. The dispersion curve shown with open circles, however, is the same as the one with solid circles for higher frequencies (smaller wavelengths), at least down to the line corresponding to a λ of 18 m, shown in Figure 5. That wavelength value corresponds to 1.5 times the argument at the first zero crossing observed for the correlation coefficients at 20 m distance.

Figure 5 allows us to compare the dispersion curve that is obtained, by jointly inverting the correlation coefficients shown in Figure 4 with those obtained by Chávez-García *et al.* (2005). The lower wavelength limit of validity obtained in that article was 80 m, also shown in Figure 5 with a dotted line. The result obtained joining the data from 1995 with the data from 2003 shows greater phase velocities (about 100 m/sec) in the frequency range 2.3–4 Hz than those obtained for the linear array only. This small difference could result from the spatial distribution of the measurements. Our measurements were obtained only along the

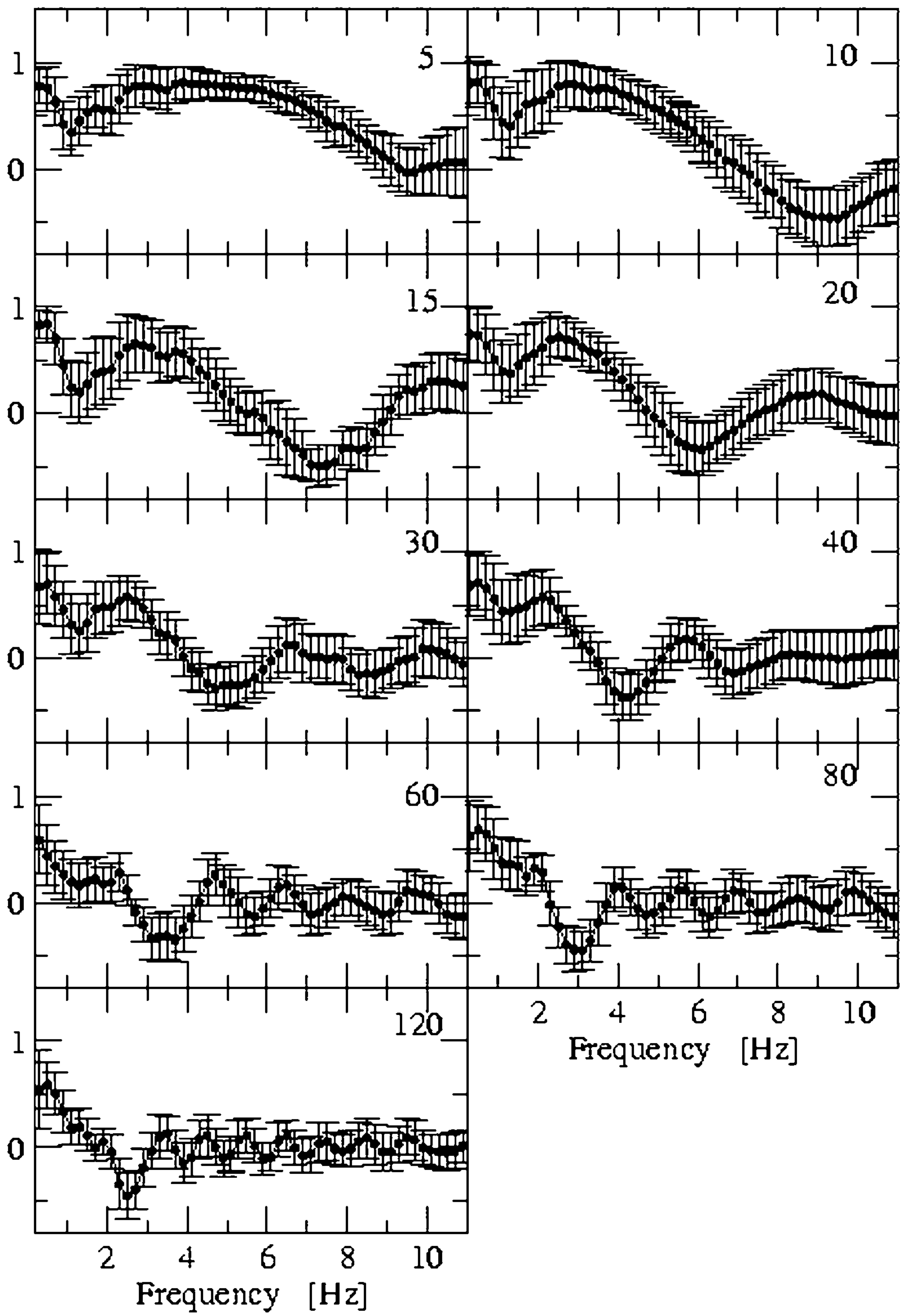

Figure 4. Final, average cross-correlation coefficients computed for all available interstation distances from our arrays. The bars for each symbol indicate the average value plus or minus one standard deviation. The number of windows averaged was 102 for 5 m, 236 for 10 m, 34 for 15 m, 343 for 20 m, 56 for 30 m, 391 for 40 m, 77 for 60 m, 158 for 80 m, and 79 for 120 m.

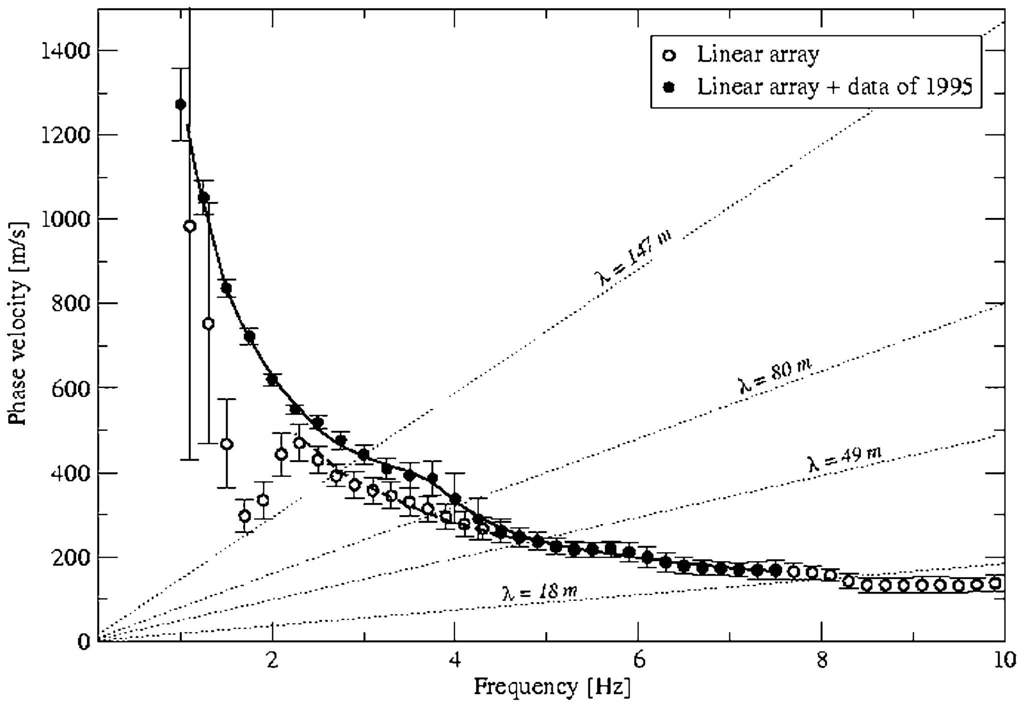

Figure 5. Phase-velocity dispersion curves obtained from the correlation coefficients. Open circles, estimates obtained from the microtremor measurements along our line at Parkway. The error bars show the uncertainty of the estimate. Solid circles, estimates obtained from the joint inversion of the results from the linear array and the microtremor measurements obtained with a weak-motion temporary seismograph network analyzed using SPAC by Chávez-García *et al.* (2005). The straight lines starting from the origin of the plot show points of constant wavelength (λ). The thick dashed line is the phase-velocity dispersion curve computed for the soil profile inverted using only the data from the linear array (profile "Linear array" in Fig. 6). The thick solid line shows the dispersion curve computed for soil profile inverted using together the data from the linear array and the 1995 measurements (profile "Linear array + data of 1995" in Fig. 6).

small line shown in Figure 1, whereas those of Chávez-García *et al.* (2005) were obtained using data recorded over a much larger portion of the basin. In addition, in that article correlation coefficients were obtained for 70 different distances spanning 40 to 396 m, as compared with the nine distances analyzed here.

The final subsoil models are shown in Figure 6. We compare the model obtained from the inversion of the linear array data only (solid line) with that obtained from the combination of the data from the linear array with those from the 1995 experiment (dot-dashed line). We observe that the differences between the phase-velocity dispersion curves in these two cases relate mainly to the layer between 35 and 78 m depth. This layer has a shear-wave velocity of 530 m/sec in the former case and of 720 m/sec in the latter. The computed amplification, however, is very similar for the two profiles, as it is governed by the large impedance contrast at about 80 m depth. The computed phase-velocity dispersion for these two models is shown in Figure 5 with thick solid line (composite data) and thick dashed line (linear array data only). Figure 6 shows two additional soil profiles. The profile identified as Spectral Analysis of Surface Waves (SASW) was obtained by Sutherland and Logan (1998) using the SASW technique (Nazarian and Stokoe, 1984). An additional profile (identified with SCPT), was obtained by Stephenson and Barker (2000) by using the seismic cone penetration test (SCPT). The agreement among all soil profiles is good at shallow depths. However, neither SASW nor SCPT were able to retrieve information for depths greater than 20 and 30 m, respectively. SASW indicates that a high-velocity contrast exists, but is unable to constrain its depth, whereas SCPT met with probe refusal. Our results are able to provide a reliable estimate of the velocity profile at this basin down to bedrock, which we find at slightly less than 80 m depth. We note that a gravity survey and a seismic refraction survey by Duggan (1997) yielded bedrock depths of 70 and 75 m respectively.

Conclusions

We have presented the results obtained using the SPAC method to analyze microtremors recorded along a line of

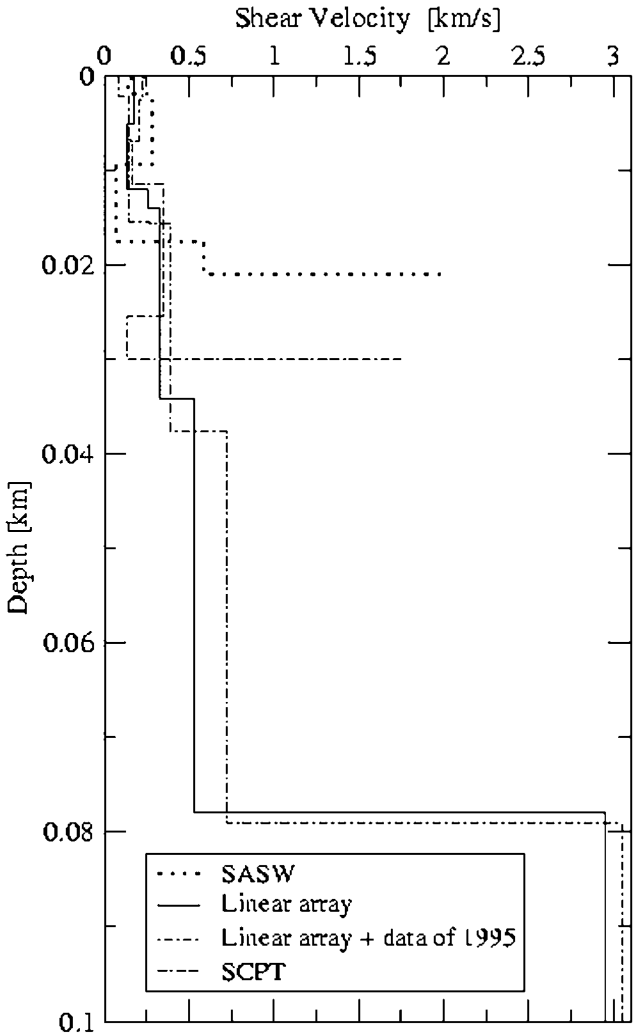

Figure 6. Velocity models for Parkway basin. The models determined in this study are the ones identified as "Linear array," using data from the linear array at Parkway, and "Linear array + data of 1995," adding the results of the analysis of data recorded in 1995 to the data analyzed in this article. The line identified as "SASW" was determined by Sutherland and Logan (1998). The line identified as "SCPT" was determined by Stephenson and Barker (2000).

stations. Four broadband seismographs were used to record ambient vibration using four different linear arrays. The first had interstation spacing of 5 m, the second of 10 m, the third and fourth of 20 and 40 m. These data were analyzed using the SPAC method, following a previous, more detailed article published previously (Chávez-García *et al.*, 2005), where the antecedents, tests, and results are described in more detail. Our results suggest that it is possible to substitute temporal averaging for the azimuthal averaging required by the SPAC method. Even if these results are not backed by a theoretical demonstration, the results presented here and in Chávez-García *et al.* (2005) strongly suggest that the SPAC method does not require a particular geometry, thereby expanding greatly the possible applications for this method. However, using many distances to compute correlation coefficients makes it difficult to select a range of validity (in terms of wavelength) for the results.

The resulting phase-velocity dispersion curve was inverted to obtain the subsoil profile. The comparison with previous results from a temporary seismic network showed that the results from the linear array are reliable, especially at frequencies higher than 4 Hz. For frequencies between 2.3 and 4 Hz, the results of a previous experiment suggest that phase velocities are underestimated by about 100 m/sec. In terms of computed amplification, however, this difference is not significant. Finally, our results indicate that the maximum depth of the basin is close to 80 m.

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