CHAPTER V

Near-Surface Measurements of P- and S-Wave Velocities from the Dense Three-Dimensional Array near Garni, Armenia

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ABSTRACT

P- and S-wave arrivals from local earthquakes were studied using an array of ten three-component instruments in and around a tunnel at Garni Observatory, Armenia. The array has a three-dimensional configuration with lateral dimensions of 300 to 500 m and a depth extent of 100 m. Estimates of the horizontal and vertical components of slowness for P and S wavefronts were used to determine the angles of approach and the propagation velocity. The results show that the region around the array has low average velocities for both P (1.43 km/sec) and S (0.61 km/sec) waves, so wavefronts approach the array at steep angles of incidence. Waveforms from one event show clear reflections from the free surface for both P and S waves. The timing of these reflections gives the velocity variation with depth within the array. We estimate a P-wave velocity of 0.33 km/sec within a few meters of the surface increasing to 6.0 km/sec for the deepest portion of the array. Local site variations can greatly complicate the high-frequency waveforms even for tunnel stations in bedrock. The S waves exhibit stronger site-dependent waveforms and time delays than do the P waves.

INTRODUCTION

An array of ten three-component instruments was installed in and around a 200 meter horizontal tunnel (Figure 5-1) at Garni Observatory, Armenia (40.136^{o}N, 44.724^{o}E). The Garni array is located on the lower southwestern slope of Mt. Azhdahak in the Gegham Range of central Armenia. Mt. Azhdahak is a Cenozoic volcanic center with layered tuffs dominating the geology near the tunnel. With horizontal spacings of 60 to 480 meters and vertical spacings of 14 to 100 meters, this dense array provides one of the few opportunities to record three-components of ground motion in a three-dimensional array. We have used these data to make measurements of the three-dimensional slowness for incoming P and S arrivals from eleven earthquakes at distances of 10 to 60 km. These three-dimensional slowness estimates are useful because they show the angle of incidence with which the wavefront approaches the array and also directly give the velocity of the material in the region of the array. The predominant frequencies of the data used in this study are from 3 to 20 Hz which correspond to wavelengths of 90 to 570 meters for the P wave and 30 to 170 meters for the S wave. The instrument spacing is comparable to these wavelengths, so we are able to accurately map wavefronts as they pass through the upper 100 meters of the near-surface material.

The first part of this paper treats the data as single incoming wavefronts and we estimate the directions of approach and the average propagation velocity across the array. The second part of the paper looks at one specific detail of the waveforms, the free-surface reflections. The relative arrival time of the reflection as a function of station depth gives more information about the depth dependent velocity variation within the array.

DATA

The data for this study were recorded by 1 Hz (Mark Products L4C) velocity transducers both on a triggered IBM-compatible personal computer (PC) system at 200 samples/second/channel with 12-bit resolution (Tottingham and Lee, 1989) and on the General Earthquake Observations System

(GEOS) recorders (Borcherdt *et al*., 1985) at 200 samples/second/channel with 16-bit resolution. The PC system recorded all ten vertical and four horizontal components with a common time base. The five GEOS instruments recorded the complete three-component data for all ten sites. Comparisons between the two systems show spectral coherence above 0.9 in the frequency range of 1 to 50 Hz. The P-wave data used in this study were primarily taken from the PC system. The S-wave data were taken from horizontal components recorded by the GEOS system.

We studied eleven earthquakes recorded by the array from September through November 1990 (Table 5-1). Approximate hypocentral distances (km) were determined by multiplying the S-P time (seconds) by 7.8. Magnitudes were determined by convolving a Wood-Anderson instrument response with the data to produce equivalent Wood-Anderson amplitudes (Kanamori and Jennings, 1978). Amplitudes from all the available horizontal components (four to twenty per event) were used with the Richter (1937) distance correction to obtain the magnitudes. Based on the similarity of the waveforms and the relative arrival times, three of the earthquakes were located closely together at a hypocentral distance of about 12 km, nearly under the array (events 4, 5, and 8 in Table 5-1). Another two events (9,10), located 20 km to the east, also have similar hypocenters. The remaining 6 earthquakes arrived from a wide spread of azimuths (Table 5-1).

Figure 5-2 shows an example of the data for event 11. The P waves show similar impulsive waveforms across the array that are well suited for timing arrivals. The S waves on the horizontal components are also fairly clear, although some procedure of waveform matching is needed to accurately estimate arrival times. There are significant differences both in waveform and amplitude among these sites, which are located within a few hundred meters of each other. Differences in site response for closely spaced stations have also been noted at other arrays, such as in New York state (Menke *et al.*, 1990) and southern California (Vernon *et al.*, 1991; Mori, 1993). These site-response characteristics, especially for the S wave, are an indication of the problems that will be encountered in studying coherent wavefronts across the array.

THREE-DIMENSIONAL SLOWNESS MEASUREMENTS

For a plane wave traveling across a three-dimensional array of stations, the arrival time (t_i) at station *i* can be written as,

$$t_i = S_x x_i + S_y y_i + S_z z_i ,$$

where S_x, S_y, S_z are the three components of slowness and x_i, y_i, z_i are the spatial coordinates of the *i* th station. With arrival times measured at four or more sites distributed in space, the slowness in three dimensions can be determined. For the data used in this study, we determine the slowness vector using a grid search method that tests the correlation of the waveform data for each value of the three-component slowness. This is a three-dimensional extension of the two-dimensional method used by Frankel *et al.* (1991) to study apparent velocities with a small aperture array. The procedure used in this study consists of:

1. Calculating the relative arrival times at each station for the given value of slowness.
2. Shifting the waveforms by these time lags.
3. Calculating the cross correlation of each pair of stations and summing the values. For 10 arrivals, this involved 45 cross correlations.

The combined cross correlation for all the station pairs is a measure of how well the relative arrival times fit the given value of slowness. In this manner we searched for the three components of slowness that that give the highest correlation of the waveforms. After determining the slowness vector, the back azimuth (ϕ) to the event is,

$$\phi = \arctan(S_x/S_y) \ .$$

The incoming angle of incidence (θ) from vertical is,

$$\theta = \arctan((S_x^2 + S_y^2)^{1/2} / S_z) \ .$$

The apparent velocity (V_a) is,

$$V_a = 1/(S_x^2 + S_y^2)^{1/2} \ .$$

The material velocity (V_o) is,

$$V_o = 1/(S_x^2 + S_y^2 + S_z^2)^{1/2} \ .$$

We used time windows of 0.5 sec for the P wave and 0.7 sec for the S wave to compute cross correlations. The slowness was tested in increments of 0.03 sec/km. An example of the grid search for the P-wave window of the event recorded at 1500 on 11/16/90 is shown in Figure 5-3. The contoured values of the combined correlation are shown for horizontal and vertical slices out of the three-dimensional slowness space that was searched. The contour interval is 0.05. The highest values of the correlation for the P wave (shaded areas) show an incoming wavefront with a horizontal slowness of 0.12 sec/km (corresponding to an apparent velocity of 8.3 km/sec) moving toward the southwest in the horizontal slice. In the vertical slice the P wave is approaching the surface at an incidence angle of 11 degrees with a total slowness of 0.64 sec/km (P-wave velocity of 1.56 km/sec). The S-wave window (Figure 5-4) shows a wavefront of similar orientation moving with a horizontal slowness of 0.30 sec/km (apparent velocity of 3.3 km/sec) in the horizontal slice and total slowness of 1.71 sec/km (S-wave velocity of 0.58 km/sec) in the vertical slice.

In the horizontal slices, the contours are elongated in the east-west direction, indicating that there is better resolution for the north-south slowness than for the east-west slowness. This is due to the geometry of the array. The stations tend to be lined up in the direction along the axis of the tunnel, so that there is good station spacing in the north-south direction and more limited distribution of stations in the east-west direction. The contours in the vertical slice show contours that are elongated in the vertical direction, which is also probably due to station geometry. Since the vertical extent of the array is only 100 meters compared to the larger horizontal extent of 300-500 meters, there is a better constraint on the horizontal slowness compared to the vertical slowness. Another reason for poorer resolution in the vertical direction could be due to strong velocity variations as a function of depth.

In addition to the grid search method described above, we checked our results by picking the onset times and directly determined the slowness using a least-squares fit to the arrival times. In general the two methods give consistent results. The advantage of the more tedious grid search method is that the cross correlations look at a larger portion of the waveform and are particularly useful for emergent P and S waves, where it is difficult to accurately pick the onset of the arrival. Also, searching through the whole slowness space shows the range of values that is consistent with the data and is useful for identifying directions from which secondary energy is arriving across the array.

SLOWNESS RESULTS

Using the grid search method, we determined the P-wave slowness for the eleven earthquakes and also the S-wave slowness for six earthquakes. There were incomplete horizontal data to determine S-wave slowness for the other five events. The cross correlation values generally have well-defined maxima, giving good estimates of the horizontal and vertical slowness.

For the P waves, the results presented in Table 5-1 show a large spread of incoming azimuths with steep incidence angles. 6 of the events (1, 4, 5, 7, 8, and 11) have large apparent velocities (>10 km/sec), indicating relatively deep earthquakes with raypaths to the array that take-off upward from the source. The other events, even to a distance of greater than 60 km, also have steep incidence angles (< 20 degrees) which is an indication of low-velocity material in the near-surface region. The low-velocity material will bend the ray paths toward the vertical as they approach the surface. One consistent result for all events is the P-wave velocity for the region near the array. For the wide range of incoming azimuths the estimates for the P-wave velocity had a standard deviation of 15%. The average velocity was determined to be 1.43 $\pm$ 0.22 km/sec. The stated uncertainty is one standard deviation.

For the S waves, the north and east horizontal components were rotated into a transverse direction assuming the back-azimuth obtained from the P wave, before the cross correlation procedure was run on the waveforms. The estimates of the incoming S wavefronts are generally in agreement with the P wavefonts. However, there is more variation among the S waveforms compared to the P, thus the correlation values were lower indicating more uncertainty in the results. The determinations for the back azimuth to the event and the vertical incidence angle are within 14 degrees of the values from the P waves. This indicates that the P and S wavefronts are approaching from approximately the same angle. The average S-wave velocity is 0.61 $\pm$ 0.06 km/sec. This gives a high P to S velocity ratio of 2.3 $\pm$ 0.2.

As mentioned above there tends to be more uncertainty in the S-wave estimates because of the variation in waveforms across the array. At a given frequency, the S waves might be expected to show more variations than the P waves, because the S-wave velocity is slower than the P-wave velocity and the wavelength is correspondingly shorter. For this reason, the S waves might also be more sensitive to site-dependent time delays. Figure 5-5 shows the P- and S-wave arrivals for the event at 1839 on 10/15/90. Note that the pattern of relative arrival times across the array is quite similar for the P and S, with the exception of G5A, where the S wave arrives late. This significant S-wave site delay appears to be azimuthally dependent and is strong for events 5 and 8, which approach the array steeply from the northwest. For these two events, G5A was not used in the S-wave slowness analyses. There is also some evidence that the delay could be due to phase response problems in the instrumentation at G5A.

FREE-SURFACE REFLECTIONS

Figure 5-6 shows the P- and S- wave arrival from the event at 1839 on 10/15/90 as recorded on the tunnel stations and one surface station directly above, plotted as a function of station depth. The instrument response was removed to convert the data to ground displacement and the waveforms were aligned on the first arrival. Note that there is a secondary arrival (dashed line) following both the initial P and S arrivals on the tunnel stations, but not on the surface station. This arrival is interpreted to be the downward reflection from the free surface, since it moves out with station depth. The timing and polarity of this arrival is also consistent with the direct arrivals at the surface station G4A and the tunnel station G1A located 60 meters directly below. Between these two stations there is a 0.04 sec time difference in the direct P arrival which corresponds to the 0.08 sec two-way travel time of the surface reflection seen on G1A (Figure 5-6). This type of arrival has been previously observed from borehole data (Hauksson *et al.*, 1987). The amplitude of the first arrival at the tunnel stations is about half that of the surface station. This difference can be partly explained because the energy is divided between the direct and reflected arrivals for the sub-surface stations, while for the stations on the surface all the energy is contained in the direct arrival. Also, there is an amplitude difference because the surface station is sited on lower-velocity material. At these high frequencies, the tunnel sites show more complicated waveforms than the surface sites. These differences in waveforms may affect the estimated slowness, if the waveform correlations are dominated by these high-frequency characteristics. To avoid this, we use relatively long (0.5 to 0.7 sec) time windows to give more weight to the lower-frequency components.

We can use this secondary arrival to estimate the P-wave velocity in the material between the tunnel stations and the surface. Assuming that the ray is traveling at a near-vertical angle of incidence as indicated by the slowness results, the time difference between the direct and reflected arrivals is the two-way travel time from the station to the surface. The measured times give average velocities of 0.33, 0.70, 1.1, and 1.5 km/sec for the material between the surface and stations G3A, G2A, G2B, and G1A, respectively, as shown in the top portion of Figure 5-7. If we assume that the velocity structure is parallel to the topography, the data can also be interpreted as a layered structure shown in the bottom portion of Figure 5-7, with velocities ranging from 0.33 km/sec at the surface to 6.0 km/sec at a depth of 60 m. The layer thicknesses in this model were arbitrarily set to match the station spacings. The actual velocity structure may be quite different from either of the simple models shown in Figure 5-7, especially given the various layers of volcanic tuff that are observed in the local geology near the tunnel. However, at present we do not have data to constrain the velocity of these more complicated structures.

The S waveforms in Figure 5-6 are very similar to the P waveforms, therefore the timing of the reflection phases gives an S-wave velocity structure similar to the the P-wave velocity structure but with 0.5 times the velocity. This P to S velocity ratio of 2.0 is the same as obtained from the slowness analysis for this event. The ratio of 2.0 measured from the reflected arrivals is probably the more reliable estimate because it is a direct measurement of the time differences of P and S waves arriving along similar paths.

DISCUSSION

We do not have independent determinations for the locations of these events, so it is difficult to judge the accuracy in the estimates for back azimuth and incidence angles. The estimates of the P- and S-wave velocities are much better constrained. The largest time differences are between the tunnel and surface stations, indicating that most of the arrivals have steep angles of incidence. Therefore, small errors in the back azimuth and incidence angle do not affect the velocity estimates significantly. This is reflected in our consistent estimates of the material velocity. Determination of apparent velocity for steep incidence angles are strongly dependent on the angle of incidence, so there is much more uncertainty in these values.

The slowness analysis used in this study assumes a plane wave incident on the array. This assumption breaks down for any vertical or lateral variations in velocity near the array, as was clearly inferred from analysis of the reflected phases. However, the values from the slowness analyses are still meaningful as velocities for the near-surface region averaged over the dimensions of the array. In particular, since the largest time differences are measured between the tunnel and surface stations, the velocity estimates in our study are dominated by the near-surface material above the tunnel. The consistent results we obtain for a wide range of P-wave azimuths indicate that lateral variations are not strong enough to affect the P-wave velocity estimates. However, there appear to be stronger lateral variations in the S-wave velocity structure to the extent that incoming wavefronts are not as well approximated by a plane wave.

The velocities estimated by the slowness method are also frequency dependent, since lower frequencies, corresponding to longer wavelengths, will sample material properties deeper in the crust. We see this effect in the higher P-wave velocity for event 2, which was the most distant event (S-P time > 9 seconds) and had P waves with the lowest frequencies. This is further evidence that the low velocities are associated with the near-surface material. In additon, the most direct evidence for the very low velocities near the surface are the time delays from the free-surface reflections at tunnel stations. Using four tunnel stations at various depths below the surface, we see a strong increase of both the P- and S-wave velocities with depth (Figure 5-7). The range of values we obtain are similar to the results from a borehole experiment at Anza, California (Fletcher *et al.*, 1990) and show P-wave velocities near the surface of a few hundred meters per second and increasing to several kilometers per second at 50 to 100 meters depth. This strong depth dependence of the velocity is some of the detailed velocity structure that is blurred together in our average P-wave velocity of 1.43 km/sec obtained from the slowness method.

This three-dimensional array experiment has been useful for making direct estimates of the P- and S-wave velocities and providing approximate locations of local earthquakes (Figure 5-8). Knowledge of the near-surface velocities is important for evaluating the site effects associated with strong ground-motions. However, in the slowness analysis we assume that the region around the tunnel was a uniform medium, and thus obtain only average near-surface velocities. Denser instrumentation would be useful for studying the lateral variations that exist on the scale of tens to hundreds of meters. These site conditions affect the ampltudes of the high-frequency waves and also affect our direction estimates of the incoming wavefronts. Further, for the purposes of more accurately locating earthquakes, the aperture of the array should be expanded to several kilometers. Even in that case, it would be necessary to have some indepently known source locations to detemine the individual site delays.

CONCLUSIONS

Estimates of slowness for incoming P and S waves recorded on the Garni three-dimensional array show steep incidence angles from local earthquakes. These arrivals give average P- and S-wave velocities of 1.43 $\pm$ 0.22 and 0.61 $\pm$ 0.06 km/sec, respectively, for the region around the array. Since the data used for slowness correlations have relatively high predominant frequencies (3-20 Hz), these values reflect the material close to the ground surface. This result was confirmed by measurements of the time delays from free-surface reflections observed on the tunnel stations. The P-wave velocity profile inferred from the surface reflections varied from 0.33 km/sec near the surface to 6.0 km/sec at 60 meters depth. At the higher frequencies used in this study, the tunnel sites have more complicated waveforms than the surface sites, because of the reflection off the ground surface. In addition to differences in waveform and amplitude between the closely spaced instruments, there are also arrival-time delays which are particularly strong for the S wave and affect the ability of the array to resolve the incoming azimuth of seismic waves.

Despite the high frequency complexities observed in the incoming wavefronts, the array was still useful for giving approximate locations of small earthquakes in the immediate region, which may be important for seismic hazard assessments. Garni lies on the northern edge of a narrow, east-west trending valley, a strong topographic feature that may be due to active faulting. This feature extends west to the outskirts of Yerevan, a city of 1.2 million people.

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Table 5-1. Parameters of incident plane waves from events recorded on the Garni array. Distances were estimated from S-P times. Magnitudes were determined by convolving a Wood-Anderson instrument response with the data.

					P wave				S wave			
No.	Date	Time	Distance (km)	Magnitude	Back Azimuth (deg)	Apparent Velocity (km/s)	Velocity (km/s)	Incidence Angle (deg)	Back Azimuth (deg)	Apparent Velocity (km/s)	Velocity (km/s)	Incidence Angle (deg)
1.	09/21/90	0715	16	1.2	282	16.7	1.28	4				
2.	10/05/90	0755	>60		294	8.3	2.02	14				
3.	10/09/90	2054	12	0.8	176	4.2	1.27	18				
4.	10/15/90	0746	12	0.9	338	11.1	1.44	7				
5.	10/15/90	1839	12	1.7	338	11.1	1.44	7	326	6.7	0.71	6
6.	10/18/90	1118	30	2.3	82	6.7	1.21	11	90	3.3	0.55	10
7.	10/27/90	0544	42	3.0	204	11.1	1.44	7	198	10.0	0.62	4
8.	11/08/90	2037	12	1.6	354	16.7	1.33	5	8	5.0	0.62	4
9.	11/16/90	0653	20	1.8	74	6.7	1.36	12	80	3.3	0.58	10
10.	11/16/90	0703	20	1.8	74	6.7	1.36	12				
11.	11/16/90	1500	14	2.2	32	8.3	1.56	11	34	3.3	0.58	10

Figure 5-1. Approximate station configuration for the three-dimensional Garni Array. G5B which appears to be above the ground is on a hillside slope east of the array.

Figure 5-2. Example of the three-component velocity data for the event at 1500 on 11/16/90.

Figure 5-3. Contoured plots of combined cross correlation values for tested values of slowness from the P-wave window of the event at 1500 on 11/16/90. Vertical (top) and horizontal (bottom) slices of the three-dimensional slowness space are shown. High correlation values (shaded areas) correspond to values of slowness which are consistent with a plane-wave propagating across the array.

Figure 5-4. Contoured plots of combined correlation values from the S-wave window of the event at 1500 on 11/16/90.

Figure 5-5. P- and S-wave arrivals for the event at 0746 on 10/15/90. Note that the relative arrivals for the P and S are similar, except at G5A where the S wave arrives significantly late.

Figure 5-6. Ground displacements of the P and S waves for the event at 1839 on 10/15/90. Reflections from the free surface (dashed line) can be seen following both the P- and S-wave arrivals at the tunnel stations.

Figure 5-7. (Top) Average P-wave velocities between the tunnel stations and the ground surface from time delays of surface reflections.

(Bottom) Velocity model consistent with the time delays of the surface relections, assuming a layered structure parallel to the topography.

Figure 5-8. Rough locations of the numbered events from Table 5-1 shown on a map of northern Armenia.