**Geothermal Ground Noise
Measurements at
Roosevelt Hot Springs
and
Cove Fort, Utah**

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1. INTRODUCTION

As part of a number of studies sponsored by the National Science Foundation under the Research Applied to National Needs (RANN) program and by the Department of Energy, a research program was conducted using seismometer arrays to study the background noise field at Roosevelt Hot Springs and Cove Fort in Utah.

The objective of the program was not to conduct a geothermal ground-noise survey in the usual sense, i.e. measuring noise levels at discrete stations. This approach has been discussed in the literature by a number of authors (Clacy, 1968, Douze and Sorrells, 1972, Iyer and Hitchcock, 1974).

There is considerable controversy about the usefulness of the geothermal ground-noise method in exploring for geothermal reservoirs. Most surveys to date indicate high noise levels associated with geysers and hot springs (Iyer and Hitchcock, 1974). The results over geothermal reservoirs with no surface manifestations are more ambiguous (see for example Douze and Sorrells, 1972, and Iyer, 1974).

One of the principal reasons for the controversy is a lack of understanding of the source of the noise anomalies. Either the geothermal reservoir, because of a convective process, acts as a source of noise, or the presence of a reservoir somehow amplifies the normal background noise level. A third possibility is that the noise anomalies, while real, are not directly related to the geothermal

reservoir. The noise could be generated by weather, cultural sources, etc., and be locally amplified by the local near-surface geology.

The present study was basically aimed at determining the type of noise found around a geothermal reservoir. For example, if the reservoir acts as a noise source, one would expect to find body waves at stations located directly above a reservoir. Cultural or weather related noise would probably occur as surface waves.

Six element arrays were used to determine the structure of the noise field and the types of waves present. Arrays were used at both the Roosevelt Hot Springs and Cove Fort areas. The arrays consisted of six seismometers, one in the center and the others evenly spaced on a circle. The interelement array spacing varied from 50 m to 300 m depending on the experiment. The instrumentation and the arrays are discussed in more detail in section 2.

2. FIELD EXPERIMENTS AND INSTRUMENTATION

Two field experiments were performed under this grant.

The first field experiment took place at Roosevelt Hot Springs (Beaver County, Utah) between June 25 and July 21, 1976. A total of 18 data samples were recorded, each recording site using a six element pentagonal array. Each element was a single HS-10 seismometer. All seismometer locations were surveyed, and the seismometers were buried about 18 inches. Data was recorded by telemetry on 7 track analog magnetic tape. The recording crew consisted of a two man field party from Senturion Sciences Inc. of Tulsa, Oklahoma, and Professors Laster and Douze of The University of Tulsa. All equipment was provided by Senturion. The nominal overall response characteristics of the recording system are shown in Figure 1.

The initial recording site was near a section marker approximately 1/2 mile north of a Phillips Petroleum steam well. This well is on the line between Sections 9 and 10, Township R9W, T27S. The initial pentagonal array had a spacing between elements of 200 feet, and was recorded for about 48 hours (June 25-26). The third day the spacing was increased to 400 feet (at the same site) and an additional 48 hours of data were recorded. Thereafter the 400 feet spacing was used. The final day of recording consisted of a cluster test, in which all six seismometers were placed in a small group a few feet in diameter. The purpose of this test was to determine the consistency of recording response from channel to channel.

The initial recording site was on alluvium in a gently rolling terrain. The surface dips down to the west toward the valley floor. To the north and east lie the foothills of the Mineral mountains, and the terrain is very rugged.

Some general comments can be made concerning the data. The wind was generally blowing about 15-25 miles per hour during the day, making the data very noisy even though the seismometers were buried. The wind generally decreased considerably at night. Also, several other groups were conducting geophysical surveys in the area at the same time, leading to some traffic noise. For this reason most of our samples for detailed analysis were selected from night time recordings. A difficulty not originally anticipated was seismic energy generated by railroad traffic on the major North-South track several miles to the west of Roosevelt KGRA. This noise was of very high amplitude and occurred at intervals of a few hours both day and night. This energy was useful in determining near surface seismic velocities, but it also made selection of data samples difficult. Unfortunately, we did not have a log of train traffic in the area.

Because of the large amount of data recorded, it is obvious that only selected samples from each site could be digitally analyzed. The samples to be analyzed were selected from analog playbacks used in conjunction with the data logs. Generally we chose four to six, ten second samples from the night recording period. Sinusoidal calibration samples for each seismometer were also selected. The desired samples were located by use of the WWVB channel on the analog tape.

The analog tapes were shipped to Geotechnical Corporation in Dallas where they were digitized at 100 samples per second. All six channels were digitized.

The second experiment was performed near Cove Fort (Millard County) Utah between November 11 and November 15, 1977. This field work was performed by Ensco Inc. of Springfield, Virginia. Three recording sites were occupied - two in Dog Valley and the third in the foothills below Dog Valley Peak. At each recording site two types of arrays were used. A small array (170 foot radius) consisting of six elements, each a single HS-10 seismometer, was used to study surface wave noise. A large array (1,000 foot diameter) was used to study possible body waves. These large arrays had six elements, each element consisting of a cluster of 72 geophones (Mark Products L-25B) with 10 foot spacing. The data were recorded on 14 track analog magnetic tape. Unfortunately, because of electrical problems, it was not possible to record the large and small array simultaneously at any site.

The data, particularly at the two valley sites, was very noisy due to traffic on a major north-south highway to the west. An east-west highway and possible mining activity, both south of the area, also contributed to the noise. The quietest period of the day at all three sites was around sunrise. Ensco selected appropriate quiet data samples and converted these to digital form with 120 samples per second. Tapes containing this data were transmitted to Tulsa University. In addition, Ensco independently processed the data using their Seismic Activity Map algorithm. Their results are described in a separate report accompanying this report.

3. THEORY OF ARRAY PROCESSING

3.1 Frequency Domain Processing

The noise of interest in geothermal ground-noise surveys is a random time series; therefore, it is appropriate to use power spectra, coherences and frequency - wave number (f-k) spectra to analyze the data obtained in the field experiment. In the following section, the theory required to gain an understanding of the results is briefly discussed, emphasizing those features that are not readily available in the literature.

To compute power spectra the method discussed in detail by Welch (1967) was used. It consists of dividing the time series in segments and transforming each segment into frequency domain. Spectral estimates are obtained by multiplying in the frequency domain and averaging. The method is well known and need not be discussed further. In the computations, 12 to 20 segments of data together with a Hanning smoothing window were used, resulting in approximately 30 to 50 degrees of freedom (Blackman and Tukey, 1958).

In order to efficiently compute f-k spectra from the array data, a combination of the algorithm described by Smart and Flinn (1971) and the averaging procedure of Welch (1967) were employed. The basic approach is described below, for details the reader is referred to the above mentioned references.

The array is composed of N elements whose locations are specified by vectors $\vec{r}_i$ ($i = 1, N$). The Fourier transform of a

chosen time interval of the noise of seismometer j is given by

$A_j(\omega) \exp [i \alpha_j(\omega)]$. The f - k spectrum is defined by:

$$P(\omega, \bar{k}) = \sum_{j=1}^N \sum_{m=1}^N \Psi_{jm}(\omega) \exp [i \bar{k} (\bar{r}_j - \bar{r}_m)] \quad (1)$$

where Ψ_{jm} is the cross spectrum between the j^{th} and m^{th} seismometers; if $j = m$ the autospectrum is obtained.

$$\Psi_{jm} = A_j(\omega) \exp [-i \alpha_j(\omega)] \cdot A_m(\omega) \exp [i \alpha_m(\omega)]$$

Equation 1 can be rearranged to give

$$P(\omega, \bar{k}) = \left| \sum_{n=1}^N \{A_n(\omega) \exp [i \alpha_n(\omega)]\} \cdot \exp (i \bar{k} \bar{r}_n) \right|^2 \quad (2)$$

which involves a single summation. A major reduction in computation time can be achieved by using the following relationship:

$$A_n(\omega) \exp \{i[\alpha_n(\omega) - (\bar{k} + \Delta \bar{k}) \bar{r}_n]\} = A_n(\omega) \exp \{i[\alpha_n(\omega) - \bar{k} \bar{r}_n]\} \cdot \exp (-i \Delta \bar{k} \bar{r}_n) \quad (3)$$

By using this relationship (Equation 3), each of the products in the summation in Equation 2, which defines the f - k spectrum, can be derived from the previous product by multiplying by $\exp (-i \Delta \bar{k} \bar{r}_n)$. Since $\Delta \bar{k}$ can be taken as a constant (the wave number grid), the generation of the kernels $\exp (-i \bar{k} \bar{r}_n)$ is eliminated.

This computational approach is extremely efficient on a computer. However, because only one time segment from each array element is used, the number of degrees of freedom of the resulting f-k spectrum is small. A simple way of increasing the degrees of freedom, and thus the confidence in the results, is by averaging f-k spectra. This procedure is entirely analogous to averaging Fourier transforms to obtain power spectra (Welch, 1967). In the case of f-k spectra it is necessary to assume that the time series is both time and space stationary.

If L time segments from each seismometer are used, the final result is simply:

$$P(\omega, \bar{k}) = \frac{1}{L} \sum_{j=1}^L P_1(\omega, \bar{k}) \quad (4)$$

The f-k spectra discussed in this report were computed using this approach. The method is efficient and flexible; it allows one to examine the f-k spectra of individual segments and the spectra resulting from the sum at any point in the computations.

In order to gain some insight into the significance of a peak in the f-k spectra, the Fisher statistic is computed. The Fisher statistic is defined as (Schumway, 1971):

$$F(\omega, \bar{k}) = \frac{E(\omega, \bar{k})}{E(\omega) - E(\omega, \bar{k})} (N - 1)$$

where

$$E(\omega, \bar{k}) = \frac{1}{N} \sum_{j=1}^N \left| A_j(\omega) \exp(-i \bar{k} \bar{r}_j) \right|^2$$

and

$$E(\omega) = \frac{1}{N} \sum_{j=1}^N |A_j(\omega)|^2$$

It is simply the ratio of the power of the peak in f-k space to the difference between that power and the total power at that frequency, all weighted by N-1. The theoretically expected Fisher statistic is difficult to derive for our case where a large number of non-independent f-k estimates are computed. From the theory (see Section 3.2) and some experimentation with synthetic data the following guidelines have been used when interpreting the results. For an average of L=4, normally used in the computations, an F-statistic above 10 is significant and indicates the presence of coherent energy recorded by the array; values below 5 are considered statistically insignificant, while values between 5 and 10 are questionable.

When operating an array very close to the source of the seismic noise, one of the basic assumptions used in the computation of frequency-wave number is not valid. It is assumed that plane waves, i.e. waves from a distant source, are being recorded by the array. If an extended source is close to the array, waves will arrive from a multiplicity of directions with circular (not plane) wave fronts.

In this section we derive some theoretical consideration intended to increase our understanding of the results being obtained from the experiment conducted with an array in close proximity to

the source. The theory is an adaptation of results previously published by Douze (1967) and a more detailed explanation of the mathematical treatment can be found there.

The two basic types of waves, body waves and surface waves, are treated separately; the methods used are very similar.

Assume surface waves arriving at two seismometers located some distance apart (Δx) from a number of directions (N), each of which has the same power level. The power spectra at each seismometer would be the same, simply the sum of the individual spectra, $\Psi_n(\omega)$.

$$\Psi_{11}(\omega) = \Psi_{22}(\omega) = \frac{1}{N} \sum_{n=1}^N \Psi_n(\omega) \quad (6)$$

where we have normalized by N .

The cross-spectra between the recordings of the two seismometers is:

$$\Psi_{12}(\omega) = \frac{1}{N} \sum_{n=1}^N \Psi_n(\omega) \exp \left[-i \omega \left(\frac{\Delta x}{V} \cos \beta_n \right) \right] \quad (7)$$

where

$V(\omega)$ = phase velocity of the surface waves

β_n = angle between the direction connecting the two seismometers and the direction of arrival of the wave.

If we desire the surface wave noise to be isotropic, i.e. equal energy from all directions, we let N approach infinity. The spectra are still identical, $\Psi_n(\omega)$, and the cross-spectra becomes

$$\psi_{12}(\omega) = \frac{1}{2\pi} \int_0^\pi \psi(\omega) \exp[-j\omega(\frac{\Delta x}{V} \cos \beta)] d\beta = \psi(\omega) J_0(\frac{\omega \Delta x}{V}) \quad (8)$$

where J_0 is the zero order Bessel function.

The cross-spectra becomes a purely real quantity, instead of the usual complex one, it can be positive or negative. Thus, no apparent phase velocities can be obtained from the data.

If instead of generating surface waves, the geothermal source generates predominantly body waves, the mathematics becomes somewhat more complex. Consider the case of compressional body waves arriving from N random directions and angles of incidence. The time series for a vertical motion seismograph becomes

$$x_1(t) = \frac{1}{N} \sum_{n=1}^N f_n(t) \cos \theta_n$$

where θ_n = angle of incidence.

Each wave has the same energy content and the conversion from P to S is ignored. This last restriction is not necessary and the results have been published (Douze, 1967), but the increased complexity does not change the results appreciably. At seismometer 2, located some distance away, the time series recorded becomes:

$$x_2(t) = \frac{1}{N} \sum_{n=1}^N f_n(t - \frac{\Delta x}{V_p} \sin \theta_n \cos \beta_n) \cos \theta_n$$

where β_n = the direction

Δ_x = seismometer separation

V_p = body wave velocity.

If, again, we are interested in obtaining a solution for omnidirectional P waves of all angles of incidence and all directions at the surface of the half space, we let N approach infinity:

$$\Psi_{11}(\omega) = \Psi(\omega) \frac{1}{\pi} \int_{-\pi/2}^{\pi/2} \cos^2 \theta d\theta = \frac{\Psi(\omega)}{2} \quad (9)$$

$$\Psi_{22}(\omega) = \Psi_{11}(\omega)$$

$\Psi(\omega)$ is the power spectrum of $f(t)$.

The cross-spectrum becomes

$$\Psi_{12}(\omega) = \Psi(\omega) \frac{1}{2\pi} \cdot \frac{1}{\pi} \int_0^{2\pi} \int_{-\pi/2}^{\pi/2} \exp -i \omega \left(\frac{\Delta_x}{V_p} \sin \theta \cos \beta \right) \cos^2 \theta d\theta d\beta$$

By performing the last integration, the expression reduces to

$$\Psi_{12}(\omega) = \Psi(\omega) \frac{1}{\pi} \cdot \frac{1}{2\pi} \int_0^{2\pi} \frac{J_1 \left(\frac{\omega \Delta_x}{V_p} \cos \beta \right)}{\frac{\omega \Delta_x}{V_p} \cos \beta} d\beta \quad (10)$$

where J_1 is the Bessel function of order 1. Again the cross-spectra has no imaginary part, however, examination of the formula indicates that it will remain positive at all times. In the limit of Δ_x or ω approaching zero, the result approaches 1/2 because $J_1(x)/x \rightarrow 1/2$.

The formula (10) was solved numerically as the authors could not find a closed form solution.

The same approach can be used to obtain a solution for shear waves arriving at all angles of incidence and direction.

$$\Psi_{11}(\omega) = \Psi_{22}(\omega) = \frac{\Psi(\omega)}{2} \quad (11)$$

and

$$\Psi_{12}(\omega) = \Psi(\omega) \frac{1}{2\pi} \int_0^{2\pi} \left[J_0 \left(\frac{\omega \Delta x \cos \beta}{V_s} \right) - \frac{J_1 \left(\frac{\omega \Delta x}{V_s} \cos \beta \right)}{\frac{\omega \Delta x}{V_s} \cos \beta} \right] d\beta \quad (12)$$

where V_s is the shear wave velocity. As Δx or ω approach zero, the J_0 factor approaches 1.0 and the J_1 factor, 0.5.

The theory discussed above for plane waves arriving from all directions is an extreme case; the other extreme being a single plane wave travelling across the array. Neither case is correct for an array above a geothermal source, however, we suspect that waves arriving from a multiplicity of directions is closer to the true situation than a single plane wave. One can, of course, numerically integrate the formulas between any limits to obtain intermediary situations between the two extremes.

One measurement often used when computing spectra of multiple time series is the coherence between time series usually defined as

$$C_{12}(\omega) = \frac{|\Psi_{12}(\omega)|^2}{\Psi_{11}(\omega) \cdot \Psi_{22}(\omega)} \quad (13)$$

For a single plane wave the coherence will be unity between two seismographs. If the noise is isotropic, the coherence will decrease rapidly from unity at zero frequency or zero separation to smaller values.

The theoretical results discussed above are used in the interpretation of the experimental results in Section 4 1.2.

3.2 Time-Domain Processing

The frequency-wave number spectra discussed in the previous section is one of the most effective techniques of array processing; it allows examination of the velocity and direction of waves frequency by frequency. If the waves of interest are dispersed wave trains, such as surface waves, this approach is almost a necessity.

If the waves to be detected by the array consist of non-dispersive waves such as body waves, one has the option of using either time-domain or frequency-domain techniques.

As the time-domain methods use all the frequencies at the same time, they can be expected to be effective if the signal contains a broad band of frequencies travelling at the same velocity.

One of the main objectives of the present study was to determine if P-waves from geothermal sources were present in the noise. The f-k spectra indicate that, with the exception of array 3 at Roosevelt Hot Springs, the noise consisted of surface waves.

To test the possibility that broad-band P wave noise might be present at low levels, it was decided to try a time-domain technique, the Fisher detector, a method based on power levels.

The F-statistic as a measure of the presence of coherent energy has already been briefly mentioned in the section on f-k computations (Section 3.1).

The F-statistic as a detector can also be used in the time domain; most of the experimental studies have been done by the Seismic Data Laboratory (Blanford, 1974).

The statistical theory underlying the use of the F-statistic is quite complex and has been described in the literature by Shumway (1970). The approach is briefly summarized below.

Suppose that a collection of N independent normal time series is regarded as composed of a signal $s(t)$ and a zero mean stationary normal noise process, $n(t)$ in discrete time.

$$\begin{aligned} y_J(t) &= s(t) + n_J(t) \quad J = 1, N \\ t &= 0, 1, 2, \dots, L-1 \end{aligned} \quad (14)$$

For the case of $N = 1$, a large number of solutions in terms of orthogonal expansions are available in the literature.

In order to test the hypothesis that there is no signal present, i.e. $s(t) = 0$, the data could be transformed into the frequency domain as was done in the f-k computations.

If the noise is band limited, with bandwidth B, over which the noise spectrum is constant, an approximation to the F-statistic can be used which is given by

$$F(N_1, N_2, \lambda) = \frac{(N-1) \langle \bar{\gamma}^2 \rangle}{\frac{1}{N} \langle \sum_{J=1}^N \gamma_J^2 \rangle - \langle \bar{\gamma} \rangle^2} \quad (15)$$

where

$$\bar{\gamma} = \frac{1}{N} \sum_{J=1}^N \gamma(t - \tau_J)$$

τ_J are the time delays required to form a beam of the desired velocity and direction of approach

and

the angle brackets indicate a time average over L data points

$$\langle \gamma_J \rangle = \frac{1}{L} \sum_{t=0}^{L-1} \gamma_J(t)$$

The F-detector indicates the presence of a signal if $F(N_1, N_2, \lambda) > F_0$, where $F(N_1, N_2, \lambda)$ is the computed statistic, which belong to a population distributed as a non-central F with N_1 and N_2 degrees of freedom and non-centrality parameter λ , and F_0 is the theoretical deduction threshold.

The number of degrees of freedom are given by

$$N_1 = 2 BT$$

$$N_2 = 2 BT (N-1)$$

where B = the band width in Hz

T = the time window in sec

The noncentrality parameter is defined as a function of frequency

$$\lambda(\omega) = \frac{2 \text{ BTM [signal power]}}{[\text{noise power}]} \quad (16)$$

As the signal is not known in the problem under consideration here, all comparisons between experimental and theoretical F-statistic were made with a noncentrality parameter of zero. When a signal is present in the beam, the F-statistic increases for two reasons: first the beam power in the numerator increases, and second the denominator decreases to the residual noise.

In order to obtain a value for F_0 , the computed F-statistic that indicates the presence of a signal, a theoretical F-distribution needs to be calculated. Estimates of the degrees of freedom are required. Power spectra of the different elements of the arrays were examined to determine the band width. This examination showed that the different array stations have, in general, neither the same amplitudes nor a flat spectrum as required by the theory (see for example, Figures 4 and 5). Therefore, it is not possible to accurately determine the degrees of freedom of the experimental computations; however, we can still expect to receive some guidance in determining an F-statistic level that would indicate the presence of a signal. Figure 2 shows some theoretical probability distributions with a noncentrality parameter of zero.

The results given in the figure indicates the probability (in percent) that random, uncorrelated noise will result in a given F-statistic value. For example, for $F(8,40,0)$, random noise will

result in an F-statistic larger than 2.18 for 5 percent of the values computed. (These degrees of freedom are the ones applicable to the F-statistic from the f-k spectra).

However, in both the time and frequency domain a large number of estimates are computed, although they are not all independent estimates.

Thus, the F_0 , or threshold at which we declare a detection, must be set high. The actual F_0 used in the experiments depended on the length of the time window used in the computations.

Several other time domain detectors have also been used in the past, particularly in the geophysical exploration industry. It might seem worthwhile to evaluate these for the detection of body waves around geothermal reservoirs. One of the most popular detection statistics is the Semblance Coefficient introduced by Neidell and Taner (1971). They define this to be the normalized output to input energy ratio, or in the notation used here,

$$S_c = \frac{N \langle \bar{Y}^2 \rangle}{\langle \sum_{J=1}^N Y_J^2 \rangle}$$

Thus there is a simple relationship between the Semblance Coefficient and the time domain F-statistic,

$$F = \frac{(N - 1)S_c}{1 - S_c} \quad \text{or}$$

$$S_c = \frac{F}{(N - 1 + F)}$$

Computations of F will also yield S_c and vice versa, but more important, the theory outlined earlier in this section can be used to establish tests of significance for the Semblance Coefficient.

4. EXPERIMENTAL RESULTS

The data from the two survey areas, Roosevelt Hot Springs and Cove Fort were obtained in two separate surveys with different instrumentation. Therefore, the interpretations are discussed separately. However, the format for the results is the same in both cases. The frequency-wavenumber spectra are given in the form of computer letter plots. The origin, representing vertically upward traveling waves, is at the center of the diagram. Decreasing velocities are represented farther from the center. The figure captions will often refer to "edge" velocities. These refer to the largest circles which can be inscribed on the diagram. The corners of the rectangular region will of course represent lower velocities. The relative amplitudes of the spectra are coded in the following manner:

Q = 0 to 1 decibels down from the total energy in the
frequency band

0 = 1 to 3 db down

6 = 6 to 9 db down

2 = 12 to 15 db down

8 = 18 to 21 db down

. = 24 or more db down

Each of the above ranges is separated by a blank character to facilitate reading the diagram. Arrows have been drawn from the origin to the largest spectral peak.

4.1 Roosevelt Hot Springs

At Roosevelt Hot Springs only the small arrays were used, with a maximum interelement spacing of 200 feet. These arrays do not attempt to discriminate against surface waves as do the large arrays used at Cove Fort. Thus, the principal analysis technique was the frequency-wavenumber computation. The time-domain F-statistic was tested on some of the Roosevelt Hot Springs data.

Three arrays are discussed in detail as they are representative of the different noise fields encountered in the area. The locations are shown in Figure 3. The first array (referred to as Array 1 in the test) was located between two steam wells, thus it is above the known geothermal reservoir. The other two arrays are outside the area of the geothermal reservoir, one in the valley (Array 2) and one in the Mineral Mountains (Array 3).

4.1.1 Power Spectra

Power spectra of the noise samples to be used in array processing were computed to determine the frequency bands of high power and to determine the coherences. As mentioned previously, with the exception of a sample purposely chosen to study the cultural noise, the rest of the time segments were from windless periods with a minimum of cultural noise.

Figures 4 and 5 show some typical examples of power spectra and coherences. Figure 4 is from the larger of the two arrays at Array Site 1, above the known geothermal reservoir. Figure 5 is data from Array 3 on the Mineral Mountains.

The spectra from Array 1 show two distinct peaks, one at a frequency around 30 Hz. This peak has usually been found to be associated with cultural activities (Douze, 1967). As will be determined in the section on f-k spectra, the noise appears to consist of surface waves from a number of directions, some of which are clearly of cultural origin.

Both spectra show an additional peak on the spectra at higher frequencies. It should be noted that the peak is at different frequencies for the two locations only a 160 m apart. This behavior makes it extremely unlikely that this peak is caused by travelling elastic waves on the ground; wind-generated noise together with different ground-to-seismometer coupling may be the cause of these peaks on the noise spectra.

The coherences between the two seismometer traces are statistically insignificant except for a few values below 2.0 Hz. This is typical of the results obtained, although a few selected samples show higher coherences.

The spectra from Array 3 shows a similar behavior at a lower power level. The first peak, despite its similarity to the one discussed above, was found to be caused by body waves and not surface waves. These examples clearly point out that it is not possible to draw reliable conclusions from single-station spectra. The second peak showed no spatial organization from f-k spectra or coherences. Therefore, it is not at present possible to determine if the energy in this peak is of any significance.

4.1.2. Frequency Wave Number Spectra

As mentioned in the introduction, the discussion here is confined to three arrays. Array 1 was located on top of the geothermal reservoir. The other two were located outside the geothermal area, one (Array 2) in the valley, a somewhat noisy environment, and the third on the Mineral Mountains further away from the cultural noise source.

Frequency wave number spectra were computed for frequencies between 0.5 and 10.0 Hz. Even for these small arrays, the results of the f-k spectra usually showed no organization in the noise field for frequencies greater than 5.0 Hz, and often for frequencies greater than 3.0 Hz.

The f-k plots were computed for a number of velocity ranges; in order to orient the reader to the dimensions of the plot, an arrow giving the velocity is drawn to the largest peak in the f-k spectra. The main peak has been shaded; the figure heading gives the F-statistic which is a measure of the power in the peak.

Experiments have shown that, as is the case for most sites, the noise field is, in general, neither time-nor space-stationary. The variations are mostly associated with changes in cultural activities and wind generated noise. Wind-generated noise is not coherent between seismometers at the spacing used here. Cultural noise is well known to consist mainly of surface waves (Douze, 1967).

In order to obtain an estimate of the velocities of the noise, f-k spectra were computed for a noisy interval at Array 1 (see Figure 2). The results are shown in Figure 6; only f-k spectra with an F-statistic greater than 10.0 were plotted, this is the reason for the gaps in the data points. The resulting dispersion curve clearly shows that the phase velocities are appropriate to surface waves, i.e., Rayleigh waves as recorded by the vertical seismometers, throughout the frequency range of interest. The f-k spectra do not provide sufficient resolution to determine if more than one mode of propagation is present, and are probably the cause of the decrease in velocity for frequencies less than 1.0 Hz.

Array 1 is located between two steam wells, thus it can be assumed that it is directly above the reservoir. If the reservoir is a source of body waves, one would expect that the results could be explained by using the theory of isotropic noise discussed in the section on theory.

Figure 7 shows a typical example of the experimental coherence (Equation 13) between two seismometers separated by 177 feet during a quiet interval. Also included is the theoretically expected coherence for isotropic P-waves using a half-space velocity of 2.0 km/sec. and the isotropic surface waves using the velocities determined from the cultural noise (Figure 6).

The isotropic P-wave model does not fit the experimental data. The isotropic Rayleigh wave model is a better fit to the experimental coherence, however, the theoretical values are still considerably higher than the experimental ones. The most likely solution is that

surface waves, coming from numerous directions together with some wind-generated noise, are the main components of the noise field predominate.

One important conclusion that can be drawn is that the geothermal noise field cannot be explained by P-waves generated by a geothermal convection cell. If instead of isotropic body waves, the angles of incidence were restricted to smaller values than 90 degrees, the theoretical coherence would be even larger. Mixing P-and surface-wave energy also results in coherences for higher than the experimental results.

The conclusions obtained from the coherences are reinforced by the results obtained from computing f-k spectra from Array 1. In Figures 8 and 9, two f-k spectra are shown at a frequency of approximately 6.0 Hz. Figure 8 shows the experimental results, the arrow indicates the largest peak at 402 m/sec. with an F-statistic of 5.0; the shaded areas are all the peaks greater than 12 dB below maximum power. Figure 9 shows a theoretical plot using the theory for isotropic surface waves to compute the spectral matrix (Equations 6 and 8) and then the f-k spectrum (Equation 1). The velocity of the Rayleigh waves was taken as 300 m/sec. (from Figure 6.) Again the arrow shows the maximum peak, at 380 m/sec., and the shaded areas are peaks greater than 18dB of a maximum power. The computed F-statistic is 4.2.

The theoretical f-k spectrum shows clearly that, as one would expect, a six-element array cannot determine velocities if isotropic surface waves are present. There is no peak at 300 m/sec. to be found in the spectrum, instead close examination shows a number of

peaks of about the same power level scattered throughout the plot. The symmetry of the peaks is a result of the symmetrical array configuration. This result is not surprising if one refers back to Equation 8, the phase angle can only be 0 or 180 degrees, the result is a spectral matrix with purely real values, the f-k spectra computations cannot cope with this situation.

Thus, allowing for the fact that the noise will not be completely isotropic, the experimental f-k spectrum shows the same type of behavior as found in the theoretical one. In the first place there is no hint of a high-velocity peak in the center of the plot, strongly suggesting that there are no body waves present in the noise field. Also the largest peak in the spectrum (shown by the arrow) does not agree with the expected surface wave velocity at this frequency, 300 m/sec. from the results shown in Figure 6. There are a number of peaks in the spectrum of almost the same level as the main peak, a very similar behavior as that found in the theoretical results (Figure 9). The F-statistic of the result is low, a value of 5.0 is obtained from the main peak.

It should be noted that there is no resemblance in the location of the peaks in either spectra to the impulse response of the array, thus the peaks in the spectra are not the sidelobes of the array response.

The resemblance between the theoretical and experimental results strongly suggest that the noise field is composed of surface waves arriving from a large number of directions, and that body waves are not a major contribution to the noise.

While only one f-k spectra has been shown, it is typical of the results obtained for frequencies from 2.0 to 10.0 Hz, except that the F-statistic is somewhat larger than average.

For the lower frequencies, with these small arrays it is difficult to distinguish between surface waves and body waves at a higher velocity. In most cases a peak will be found close to the center of the f-k plot, however, measured velocities and directions will show random differences from frequency to frequency. In no cases did the maximum velocity fall in the center of the plot as would be the case for either isotropic waves or body waves coming from the reservoir below the array site.

Figure 10 shows an example of the results obtained at a frequency of 1.56 Hz. The peak value is indicated by an arrow, it shows waves with a velocity of 960 m/sec. arriving from S 5° E with an F-statistic of 5.2. This particular example suggests the presence of Rayleigh waves, however, the results do not always allow the same conclusions to be drawn, sometimes the velocities are more representative of body waves.

For comparison, the theoretically computed f-k spectrum at 1.56 Hz for isotropic P-waves of 2.0 km/sec. is shown in Figure 11. As expected from the phase angles (all 0 degrees), it shows a peak at infinite velocity, with the computed F-statistic of 28.0. A plot of isotropic surface waves shows the same general features with a lower F-statistic. No general conclusion to be drawn from the results at these lower frequencies.

The array (Array 2) in the valley was located approximately 4.0 km west of the reservoir in a culturally noisy environment.

Figures 12 and 13 shows typical examples of the f-k spectra from noise samples during quiet periods. The main point to note is that the results at this site are not significantly different from those at Array 1; the noise level was somewhat higher, but the structure of the noise field is the same.

Figures 12 and 13 show the results at two frequencies, 1.17 Hz and 2.7 Hz. Figure 12 shows a velocity of 570 m/sec. (see arrow), and a F-statistic of 6.0. This velocity might be appropriate for Rayleigh waves in the low-velocity alluvium of the valley. The direction is not towards the geothermal reservoir, but towards the cultural noise sources. Figure 13 shows a velocity of 1.0 km/sec. with an F-statistic of 8.1. In both cases, as in all the f-k spectra from this site, the F-statistics are low, indicating that the noise field is poorly organized. Velocities vary widely and are not easily interpreted in terms of wave types.

The third array used in this study was located in the Mineral Mountains approximately 7.0 km from the geothermal reservoir (see Figure 3). The results of f-k computations were surprising and completely different from those obtained at the other sites.

Figures 14 and 15 illustrate the results obtained at frequencies of 1.96 Hz and 2.3 Hz respectively. The arrow in Figure 14 gives the maximum peak in the f-k spectra with a velocity of 6.4 km/sec. and an azimuth of N 31° E, the F-statistic is 37. In Figure 15 the arrow indicates a velocity of 4.29 km/sec. and an azimuth of N 37° E with an F-statistic of 37. Other frequencies showed similar results for the frequency range between 0.5 to 4 Hz.

The velocities are too high to be explained by surface waves and tests were run to insure that the peaks shown are not sidelobes in the f-k spectra from low velocity surface waves.

Thus, considering the high F-statistics, the noise must consist of organized P-waves coming from approximately N35° E. The same result persisted for several noise samples spread over the two days the array was operated. Thus, it is not an isolated case such as would occur if a low-amplitude microearthquake was present in the chosen sample. To test the hypothesis that recurring microearthquakes were the cause of the noise, the time traces for each seismometer were plotted. A typical example is shown in Figure 16. Approximately 10 seconds of data is shown, and it is obvious that no impulsive events of significant amplitude are present.

Because of the uniqueness of these results a complete set of f-k spectra from a long data sample is presented in Figures 17 through 21. Probably because of the longer than average sample and the fact that the noise is not stationary in space, the F-statistics are lower than for the shorter samples. In these examples, several frequencies showed the highest peak arriving from the south-west, this may be caused by minor phase shifts in the seismograph responses. It is also possible that the source is very close and below the array.

These results are not readily explained, the direction does not point towards the known geothermal reservoir, even allowing for some refraction caused by horizontal velocity changes. There is a source of microearthquakes known to exist some 30 km away in the Mineral Mountains in the general direction obtained from the f-k spectra (personal communication, Senturion Sciences).

Two possible explanations for the P-wave noise are:

- (a) that a microearthquake source is also a continuous generator of body-waves,
- or (b) that an unknown source of energy, possibly connected with a deeper geothermal convection cell is present to the East of the known geothermal reservoir.

At present, the only reasonable explanation why the other arrays did not record this noise lies in the relative noise levels, with the site in the mountains less noisy because it is further removed from the cultural noise sources.

4.1.3 F-Statistic

The time domain techniques discussed in Section 3.2 were intended primarily to analyze the data recorded by the large arrays at Cove Fort. These arrays were designed to emphasize the detection of high velocity P waves, and to attenuate the surface waves.

In order to check the computer program that computes the experimental F-statistic it was tested with the data from Array 3 at Roosevelt Hot Springs. As discussed in Section 4.1.2 this array showed a relatively broad band of P-waves in the noise.

Figure 22 shows a typical example of the results obtained. The edges of the plot start at a velocity of 1.0 km/sec. with the center representing infinite velocity. The values on the plot are a coded computer printout of the experimental F-statistics. The output shows an area of F-statistics above 2.0 (shaded area) with a peak of $F = 2.6$ at a velocity between 25.0 and 12.5 km/sec from N28E.

The time window used was 600 sec. Taking the band width as a few Hz, and using Figure 2, it is clear that an F-statistic of 2.6 has a high probability of being significant.

The velocity obtained from the time-domain computations is higher than most of the values obtained from the f-k spectra, at least suggesting that the source is close to and below the array. It should, however, be noted that the array size is too small to accurately determine the velocities of these waves.

4.2 Cove Fort

4.2.1 Frequency Wavenumber Results for Cove Fort

The Cove Fort area in Utah (Figure 23) has been designated a KGRA (Known Geothermal Resources Area) by the U. S. Department of the Interior. It currently is being explored for geothermal resources (Rowley and Anderson, 1975). The locations of three array sites used in this study are marked as 1, 2 and 3 in the figure.

A large and a small array were used at each site. Although the same general configuration was used at each site, the spacing differed slightly. Details are shown in Figures 24 through 26. The small arrays consisted of single seismometers, while elements of the large array consisted of groups of six strings of 12 seismometers, the pattern taking the form of a six-armed star. Distance between seismometers on a string was 10 feet. These groups were used to cut down on surface wave energy, which would be aliased if single seismometers were used. Unfortunately, while such groups work very well at frequencies above 20 Hz, they were found to offer little surface wave attenuation below 5 Hz. Figure 27 shows the group response at 5 Hz. Waves with velocities are low as 500 meter/sec. (which

corresponds to the edge of this diagram) are attenuated by 3 decibels or less. Nevertheless, the groups did cut down on wind noise and helped minimize the differences in geophone plant, etc.

The recorded data is summarized in Table 1. As mentioned previously, the large and small arrays at each site could not be recorded simultaneously because of electrical problems. In addition, all six array channels were simultaneously operational only on the large array at Site 3. Considering the small number of channels used, a bad channel produced pronounced effects on the array response. As an example of this, Figure 28 shows the array response at 2.34 Hz for the large array 3 using all six channels. Figure 29 shows the analogous response with channels 4 and 5 deleted. The pattern becomes much more directional and large sidelobes or aliases become apparent. The latter makes interpretation of the results extremely difficult in some instances. (It should be noted that the responses in the last two figures do not include the effect of the 72 seismometer groups, but as we have discussed above this should have little effect).

For completeness, the array response of small array 2 is shown in Figure 30 for 2.34 Hz. Again, the edge of the diagram corresponds to 500 meters/sec. The small arrays show little directionality, but have no significant side lobes in the frequency band of interest.

Two channel coherences were computed for all arrays. Typical examples are shown for small array 3 in Figure 31 and for large array 3 in Figure 32. The vertical scales for the power spectra are relative, as no absolute calibrations were attempted during this experiment. The vertical scales for the coherence run from 0 to 1.

In both figures Channel 1 has been used as the reference. It would be more desirable to use the center channel, Number 6, as reference, but this was bad on the small array. Coherences on the small array were less than .45, and were typically much lower. Apparent high coherences between some pairs of channels above 24 Hz are artifacts, since there is no significant energy in those channels at those frequencies. On the large array coherences were very small; this is to be expected because of the larger spacing.

The coherence results can be summarized as follows, based on processing all of the digitized data. The maximum observed coherence always occurred at 4 Hz or below. For frequencies below 1 Hz the coherences were often high, but this is expected because of the long wave lengths involved (assuming propagating waves). We will, therefore, ignore those values. For large array 1 the maximum observed coherence was .31. At large array 2 it was .41, and at large array 3 it was .65. For the small arrays, the maxima at Site 2 was .61 and at Site 3 was .89. These results came from computations on successive 32 second lengths of data. The coherences changed significantly from one data segment to the next, and the high values persisted for only a short time.

Using the coherence results as a guide, data samples and specific frequencies were chosen for frequency wavenumber analysis, using the methods described previously. It was found desirable to normalize the power in the various channels. The power in the frequency band of interest (usually 1 to 5 Hz) was computed and used as a weighting factor to make the average powers the same. The

standard frequency wavenumber computation used here is relatively insensitive to normalization errors. However, certain of the high resolution techniques, such as the Maximum Likelihood method (LaCoss, 1971), are very sensitive to errors of this kind.

Usually, because it is representative of surface wave velocities in this area, .5 km/sec. was used as a starting velocity in two-dimensional wavenumber space. Four blocks of data, each block 256 data points long ($0.0083 \times 256 = 2.13$ sec.) were used in the computation of the f - k spectra. The f - k spectral technique was utilized to process four or five different time segments at each array for analysis. Besides computing and displaying the spectrum in the desired region of wavenumber space, the program determined the largest spectral peak, evaluated its F-statistic, and computed azimuth and phase velocity associated with the peak. Subsidiary peaks of interest were picked visually by the geophysicists.

Before showing examples of the spectra we will summarize the results. Because of cost and time involved it was not feasible to compute spectra for all the digitized data. However, a total of 49 time segments were analyzed for the three sites and two types of arrays. In each instance wavenumber spectra were computed at five different frequencies. The phase velocities obtained for the maximum peak varied somewhat from sample to sample, probably because there is more than one noise source which gives rise to a time varying interference pattern. However, 74% of the peaks measured had phase velocities between .4 km/sec. and .55 km/sec.

The highest velocity (3.6 km/sec.) was measured on small array 2, which had the least precision due to its small size and use

of only four channels. The largest secondary peaks in the spectra also tended to have velocities near .5 km/sec., although these results were not tabulated. No peaks that could be construed as body waves were observed.

The spectral peaks showed a fairly consistent directionality. A peak was usually present indicating a source in the approximate direction N50°W. This was assumed to be due to the north-south highway which has its closest approach west and northwest of Dog Valley. The direction varied somewhat between the three array sites. A second peak was usually present to the south, but its direction varied from S40°E to S30°W. It may be due to the east-west highway to the south, or possibly to mining activity. Finally, a third peak indicating a source N50°E was usually present at all sites. There are no known cultural sources in that direction. The area to the northeast of Dog Valley is faulted, so this might represent true geological activity. Spectral peaks from all three directions typically have velocities near .5 km/sec.

While the above discussion summarizes the results of this study, we feel that it is worthwhile to show examples of the frequency wavenumber spectra. The following series of figures show examples from each recording site. We have chosen examples that have the highest F-statistic. We have also attempted to show results for at least three contiguous frequencies. The data from large array 1 is very poor (only two frequencies are shown). The maximum F-statistic is 5., which is very marginal. The direction of the largest peak changes from frequency to frequency, but the directions are northeast, northwest-west, as observed above.

The data from the large array 2 was again not very good. The maximum F-statistic was 6. Figures 35 to 37 suggest noise sources to the east, northeast, and southeast.

The small arrays of course offer much less precision in direction. The F-statistic is much higher here (18.) and the apparent direction of the noise source is northeast or northwest. The data from small array 2 is shown in Figures 38 to 40.

The best data recorded in this experiment came from the third recording site. The data from the small array consistently showed a noise source in the southeast, although the F-statistic was low (about 5.). Data from the large array showed a noise source to the northeast which cannot be identified with any cultural source.

4.2.2 F-Statistic

The computer program used to compute experimental F-statistics was applied to the data from the three large arrays at Cove Fort. This has been discussed before in this report; the large arrays were designed specifically to detect P-waves.

The results obtained from the computations did not indicate the presence of high velocity noise at any of the three arrays. The experimental F-statistics were too low to be considered statistically significant. The time window used was 500 sec., while the bandwidth, from the spectra, is at least several Hz. Figure 47 shows an example of the results from Array 3, which because of its location at a less noisy site, was considered to be most likely to detect P-waves.

The edges of the plot are at 1 km/sec. chosen because the lower velocities characteristic of surface waves, are not of primary

interest in this experiment.

There are two peaks in the plot, with F-statistics of 1.85 and 1.89 and velocities of 1.1 and 1.7 km/sec. Neither of these velocities is indicative of P-waves.

Furthermore, it is necessary to decide if the F-statistics are significant. Because the spectra are not flat nor are the seismometers recording the same noise level. It is difficult to decide how many degrees of freedom to use in the computations.

Therefore, the following approach was used: an experimental probability distribution was computed using all the values in the plot.

Figure 48 shows the results obtained, a theoretical distribution for $F(40,200,0)$ is included for comparison. The theoretical and experimental values are in close agreement, suggesting that the number of degrees of freedom one would obtain from using the band width and time window are far too large. Examination of the results shown in this figure suggest clearly that an F-statistic of 1.89 does not have a high statistical significance, and that the peaks are random noise.

5. CONCLUSIONS AND RECOMMENDATIONS

The noise field was investigated using arrays at two geothermal sites, Roosevelt Hot Springs and Cove Fort, both in Utah.

Frequency wavenumber spectra, together with the associated spectra and coherences was the main analyses technique used. Time domain detection methods were also employed in an attempt to find P-waves in the noise.

The time domain methods used here were inefficient in terms of computer time. However, the data was extremely over-sampled. If the data were filtered and resampled these techniques might prove competitive with the frequency wavenumber technique.

The range of analyses covered the frequency range of 0.5 to 10.0 Hz. The instrumentation and digitizing rate allows higher frequencies to be analyzed. However, at higher frequencies no organization in the noise field could be found in the processing.

At frequencies higher than 10.0 Hz the power levels of different seismometers in an array varied considerably, furthermore spectral peaks did not often occur at the same frequency. It appears quite likely that, because of the presently used installations, true ground motion is not being recorded. The most likely source for at least part of this noise is wind generated noise. Thus, if there is any geothermal noise at the higher frequencies we are not at present in a position to detect it. If, for example, geothermal noise is caused by small, noisy microearthquakes, the peak frequencies

would be on the band not being utilized. We suggest that this problem be investigated. Two possible approaches are: first, attempt to improve the installation of the seismometers; second, it might be possible to predict the wind generated noise using a microbarograph and then subtract it, leaving a trace without wind noise for analysis. This approach has been used at the lower frequencies (Douze and Sorrells, 1975).

At Roosevelt Hot Springs, both the arrays located on top of the known geothermal field, and the one in the valley, showed that the noise consisted largely of Rayleigh waves at frequencies from 0.5 to 5.0 Hz. There was no clear indication of body waves in the noise as would be expected close to a buried source. The array located in the mountains to the southeast of the reservoir did show body waves predominating at frequencies from 0.5 to 5.0 Hz. However, the direction of arrival did not point towards the known geothermal reservoir. The presence of body waves at one array indicates that they should be present at the other arrays only a few miles away. The reason that they are not detected must be the overwhelming amount of cultural noise in the valley.

This strong body wave source should be investigated. As mentioned above the direction does not agree with the known geothermal reservoir. However, it is such an unusual phenomenon that the possibility exists that it may be connected with stronger geothermal activity in the Mineral Mountains.

REFERENCES

- Blackman, R. B. and Tukey, J. W., 1958, The measurement of power spectra: Dover Publ. Inc., New York.
- Blanford, R. R., 1974, An automatic event detector at the Tonto Forest Seismic Observatory: *Geophysics*, 39, 633-643.
- Clacy, G. R. T., 1968, Geothermal ground-noise amplitude and frequency spectra in New Zealand volcanic area: *J. Geoph. Res.*, 73, 5377-5383.
- Douze, E. J., 1967, Short-period seismic noise: *Bull. Seism Soc. Am.*, 57, 55-81.
- Douze, E. M. and Sorrells, G. G., 1972, Geothermal ground-noise surveys: *Geophysics*, 37, 813-824.
- Douze, E. J. and Sorrells, G. G., 1975, Prediction of pressure-generated earth motion using optimum filters: *Bull. Seism. Soc. Am.*, 65, 637-650.
- Goforth, T. T., Douze, E. J. and Sorrells, G. G., Seismic noise measurements in a geothermal area: *Geophysical Prospecting*, 20, 76-82.
- Iyer, H. M., 1974, Search for geothermal seismic noise in the East Mesa area, Imperial Valley, California: U.S.G.S. open-file report no. 74-96, 52 p.
- Iyer, H. M. and Hitchcock, T., 1974, Seismic noise measurements in Yellowstone National Park: *Geophysics*, 39, 389-400.
- LaCoss, R. T., 1971, Data Adaptive Spectral Analysis Methods, *Geophysics*, Vol. 34, 661.
- Neidell, N. S. and Taner, M. T., Semblance and other coherency measures for multichannel data, *Geophysics*, 36, 482-497.
- Rowley, P. D. and Anderson, J. J., 1975, Guidebook to the Cenozoic structural and volcanic evolution of the southern Marysvale volcanic center, Iron Springs Mining District and adjacent areas, Southwestern Utah: *Geol. Soc. of Am., Annual Meeting Salt Lake City, Utah.*

References (Cont'd)

- Schumway, R. H., 1971, On detecting a signal in N stationarily correlated noise series: *Technometrics*, 13, 499-520.
- Smart, E. and Flinn, E. A., 1971, Fast frequency-wavenumber analyses and Fisher detection on real-time infrasonic array data processing: *Geoph. J.* 26, 279-284.
- Welch, P. D., 1967, The use of the fast Fourier transform for the estimation of power spectra: *IEEE Trans. Audio and Electroacous.* AU-15.

7. Table and Figures

TABLE 1 : Seismic ground noise data chart

DATA	START TIME	ARRAY	TOTAL TIME DIGITIZED (MIN)	PIECES PLOTTED	CHANNELS WITH GOOD DATA	WIND	TRAFFIC	TU TAPE	FILE
11/15/77	11:45 AM	L3 L3*	65.4	21-90 193-228	1,2,3,4,5,6	Light	Medium	L100 L100,L500	1 2,3
11/15/77	2:00 PM	S3	32.1	21-90	1,2,3,4,5,-	Light	Medium	L300	1
11/13/77	9:00 AM	L2A	27.6	21-74	1,2,3,4,-,6	Light	Medium/ Light	L400,L500	1,2
11/13/77	10:30 AM	L2B	21.0	3-74	1,2,3,4,-,6	Very Light	Medium/ Light	L400	2
11/14/77	3:00 PM	S2	33.3	41-76	1,2,-,-,5,6	Light	Medium	L500	1
11/12/77	8:20 AM	L1A	26.2	3-34 60-91	1,2,-,4,-,6	None	Very Light	L300	3
11/11/77	1:45 PM	L1B	37.3	48-83	1,2,-,4,5,6	Light	Medium	L300	4

Note; Small Array 1 (S1) was deployed but no useable data was obtained.

Figure 1. Velocity magnification versus frequency for the Senturion Sciences radio telemetry, seismometer system under normal field operating conditions.

Seismometer: HS-10, 2000 ohm, .492 damping.

Field Amplifier Attenuation: 18 dB.

Playback Amplifier Gain: 0.1

Figure 2. Theoretical probability distributions of F-statistics with centrality parameter of zero.

Figure 3. Map of the Roosevelt Hot Springs geothermal area with location of steam wells and arrays.

Figure 4. Power spectra and coherence of the noise recorded by two seismometers at Array 1: Roosevelt Hot Springs.

Figure 5. Power spectra and coherence of the noise recorded by two seismometers at Array 3, Roosevelt Hot Springs.

Figure 6. Measured velocities during a noisy period, Array 1, Roosevelt Hot Springs.

Figure 7. Theoretical and experimental coherences for a seismometer separation of 177 ft., Array 1, Roosevelt Hot Springs.

Figure 8. Experimental frequency-wavenumber spectrum at 6.25 Hz, Array 1. F-statistic 5.0, Roosevelt Hot Springs.

Figure 9. Theoretical frequency-wavenumber spectrum at 6.0 Hz for surface waves at 300/m pec, Array 1. F-statistic, 4.2, Roosevelt Hot Springs.

Figure 10. Experimental frequency-wavenumber spectrum at 1.56 Hz,
Array 1. F-statistic 5.2, Roosevelt Hot Springs.

Figure 11. Theoretical frequency-wavenumber spectrum at 1.56 Hz for P-wave velocity of 2.0 km/sec, Array 1. F-statistic 28.0, Roosevelt Hot Springs.

Figure 12. Experimental frequency-wavenumber spectrum at 1.17 Hz, Array 2. F-statistic 6.0., Roosevelt Hot Springs.

Figure 13. Experimental frequency-wavenumber spectrum at 2.7 Hz, Array 2. F-statistic 8.1, Roosevelt Hot Springs.

Figure 14. Experimental frequency-wavenumber spectrum at 1.96 Hz, Array 3. F-statistic 37.0, Roosevelt Hot Springs.

Figure 15. Experimental frequency-wavenumber spectrum at 1.7 Hz, Array 3. F-statistic 12.1, Roosevelt Hot Springs.

Figure 16. Plot of time domain traces for Array 3. Each trace represents 10 seconds of data.

Figure 18. Experimental frequency-wavenumber spectra at 1.17 Hz, Array 3. F-statistic 12.0, Roosevelt Hot Springs.

Figure 19. Experimental frequency-wavenumber spectra at 1.56 Hz, Array 3. F-statistic 11.1, Roosevelt Hot Springs.

Figure 20. Experimental frequency-wavenumber spectra at 2.0 Hz,
 Array 3. F-ststistic 8.2, Roosevelt Hot Springs.

Figure 22. Experimental F-statistic at Array 3, Roosevelt Hot Springs.

Figure 23. Cove Fort area, Utah. Dot points (1 or 2 or 3) represents array locations

Figure 24. Array 1 Configuration. Circles are Large Array elements, squares are Small Array elements.

Figure 25. Array 2 Configuration. Circles are Large Array elements, squares are Small Array elements.

Figure 26. Array 3 Configuration. Circles are Large Array elements, squares are Small Array elements.

Figure 27. Array response for group of 72 geophones (10 foot spacing) deployed as a six arm star. Edge velocity is 500 m/sec.

Figure 28. Impulse response in wavenumber space (edges = 500 m/sec.)
for Array 3
Channels: 1, 2, 3, 4, 5, 6

Figure 29. Impulse response in wavenumber space (edges = 500 m/sec.)
for Array 3
Channels: 1, 2, 3, 6

Figure 30. Impulse response in wavenumber space (edges = 500 m/sec.)
for Small Array 2
Channels: 1, 2, 5, 6

Figure 31. Auto-power spectra, coherence vs. frequency for Small Array 3 (S3) (records: 1-24).

Figure 32. Auto-power spectra, coherence vs. frequency for Large Array 3 (L3) (records: 1-24).

Figure 33. Frequency-wavenumber spectrum for Large Array 1:
 Frequency = 1.41 Hz, Velocity of maximum peak = 542 m/sec,
 Direction of maximum peak = N77°E,
 F-statistic = 5.

Figure 34. Frequency-wavenumber spectrum for Large Array 1.
 Frequency = 1.88 Hz, Velocity of maximum peak = 612 m/sec,
 Direction of maximum peak = N23°W,
 F-statistic = 5.

Figure 35. Frequency-wavenumber spectrum for Large Array 2.
Frequency = 1.41 Hz, Velocity of maximum peak = 624 m/sec.
Direction of maximum peak = N92°E,
F-statistic = 4.

Figure 36. Frequency-wavenumber spectrum for Large Array 2.
 Frequency = 1.88 Hz, Velocity of maximum peak = 416 m/sec.
 Direction of maximum peak = N560E,
 F-statistic = 5.

Figure 37. Frequency-wavenumber spectrum for Large Array 2.
 Frequency = 2.34 Hz, Velocity of maximum peak = 494 m/sec.
 Direction of maximum peak = S170E,
 F-statistic = 3.

Figure 38. Frequency-wavenumber spectrum for Small Array 2.
Frequency = 2.81 Hz, Velocity of maximum peak = 2.4 km/sec.
Direction of maximum peak = S50°W,
F-statistic = 4.

Figure 40. Frequency-wavenumber spectrum for Small Array 2.
Frequency = 3.75 Hz, Velocity of maximum peak = 1.22 km/sec.
Direction of maximum peak = S26°E,
F-statistic = 18.

Figure 41. Frequency-wavenumber spectrum for Large Array 3.
Frequency = 1.88 Hz, Velocity of maximum peak = 445 m/sec.
Direction of maximum peak = N51°E,
F-statistic = 4.

Figure 42. Frequency-wavenumber spectrum for Large Array 3.
Frequency = 2.34 Hz, Velocity of maximum peak = 453 m/sec.
Direction of maximum peak = N65°E,
F-statistic = 6.

Figure 43. Frequency-wavenumber spectrum for Large Array 3.
 Frequency = 2.81 Hz, Velocity of maximum peak = 542 m/sec.
 Direction of maximum peak = N49°E,
 F-statistic = 3.

Figure 47. Experimental F-statistics at Array 3, Cove Fort.

Figure 46. Frequency-wavenumber spectrum for Small Array 3.
Frequency = 2.81 Hz, Velocity of maximum peak = 444 m/sec.
Direction of maximum peak = S31°E,
F-statistic = 11.

Figure 48. Experimental and theoretical probability distributions at Array 3, Cove Fort.