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INVESTIGATION OF THE INFLUENCE OF DEPTH
ON A SEISMIC SPECTRAL DISCRIMINANT
FOR RESERVOIR INDUCED EARTHQUAKES

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Many reservoirs after or during filling have triggered seismic activity which, in a few cases, has included earthquakes large enough to cause damage. This research is a continuation of an evaluation of a seismic spectral discriminant which may be used to identify areas in which reservoirs will tend to induce earthquakes. The discriminant is based on the high-frequency slope of the displacement spectra. A cubic decay implies a susceptibility to induced seismicity, whereas a frequency square decay indicates tectonic stresses. A major remaining uncertainty in the application of the discriminant is the uncertainty of the influence of depth-of-focus on displacement spectra.

The specific objective of our study of reservoir induced earthquakes is to evaluate the influence of depth-of-focus on spectral slope above the corner frequency. The data set consists of 51 events recorded at close range on digital event recorders at Monticello Reservoir, South Carolina.
motions are used to determine the angle of incidence, and spectra are computed from trace displacement in the direction of propagation. Depth of focus was estimated either by standard location procedures or by using S-P time to project distance along the ray path. The evaluation of particle motion indicates that the arrivals following the first cycle of P-wave motion are highly scattered, and the window for spectral estimation must be chosen carefully to be sensitive to focal mechanism. The events studied to date show no depth variation in slope of the high-frequency spectra, although they do cluster in a narrow depth range. For all 51 events, changes in corner frequency and high-frequency slope appear to be positively correlated. The increase in high-frequency slope with a decrease in fault radius may be an indication that earthquakes with a smaller fault plane area have fewer asperities on the fault plane.
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1. Introduction

In a general sense, all earthquakes are triggered. The problem of discriminating reservoir induced earthquakes from natural events is that of determining whether the triggering mechanism(s) are directly related to changes caused by reservoir impoundment. A study by Castle et al. (1980) found that tectonic environments characterized by normal or strike slip faulting were correlated with reservoir induced seismicity. In previous work (Long and Johnston, 1979) it was found that a cubic high-frequency spectral decay of displacement was a discriminant for earthquakes triggered by reservoir impoundment. Earthquakes occurring near Oroville (Calif.), Clark Hill (Ga.), Jocassee (S.C.), and Monticello (S.C.) showed similar spectral (ω-cube) properties and satisfy the spectral discriminant for reservoir induced earthquakes. Also, theoretical considerations imply that spectra with cubic decay are derived from earthquakes on pre-existing lubricated surfaces slipping under low tractional resistance. A common factor in these discriminants based on tectonic environment and spectra is the implication that earthquakes occur under conditions where the vertical stress is the minimum or intermediate deviatoric stress axis. Such conditions would predominantly be observed in areas where a horizontal stress is less than the vertical and in areas subjected to tensional deviatoric stress conditions. Also, in these areas, existing
cracks, joints or faults could retain significant porosity and water would be expected to penetrate greater depths. In contrast, areas of compressive tectonic stresses would have tight faults or joints which would exhibit high tractional resistance. Earthquakes on these surfaces would exhibit W-square spectral decay. Hence, the existence of open or porous faults and joints in a tensional environment provides easy access for ground water to penetrate the rock and provides lubricated or weak surfaces for earthquakes. These environments are conducive to the occurrence of triggered earthquakes and the penetration of ground water to do the triggering. Areas of tight fractures, probably under compressive stresses, do not allow easy ground water penetration, and some other triggering mechanism must explain the occurrence of earthquakes. If one then accepts pore pressure changes as a triggering agent for reservoir induced earthquakes, the discrimination between natural and induced earthquakes is then the problem of defining the stress conditions or relative porosity. The successful tectonic environment discriminant utilizes largely the geological data, and the spectral discriminant uses earthquake displacement spectra and source mechanism theory to determine the relative ambient stress conditions (extensional or compressive tectonics) and the existence of surfaces which can facilitate slip.
Having shown the feasibility of identifying or, perhaps if data allow, predicting reservoir induced seismicity, the next question to be considered is: what is the largest earthquake a given reservoir might trigger. Since it is known that the maximum magnitude of an event is limited by the maximum area of possible faulting and since most explanations of reservoir induced earthquakes require existing lubricated faults or joints, a logical hypothesis would be that the maximum induced earthquake could be computed (though not necessarily directly) from the maximum depth at which smooth or lubricated faults or joints will allow earthquakes. Areas exhibiting high residual or tectonic tensilal stress will have reduced tractional friction on potential fault surfaces extending to greater depths, and reservoirs in such areas could induce larger earthquakes. Reservoirs in areas of moderate tectonic or residual stress would not have available lubricated planes to as great a depth and only smaller events could be triggered. The transition with increased depth from low to high hydrostatic stress should be accompanied by a parallel increase in tractional friction on the slip surfaces. Theoretical considerations in source spectra imply that the high-frequency content of the deeper focus events should increase because of the increased contact between the two sides of the fault plane. Such an increase should be reflected in the displacement spectra by a change from \( n \)-cube to \( n \)-square decay. Data presented by Fletcher (1974) are suggestive of such a
transition. The specific problem of this proposed research is to determine by observation whether a change with depth in the high-frequency spectral slope can be observed and, if observable, whether it can be used to predict the maximum depth of faulting (and hence maximum magnitude) of an induced earthquake.

During the first half-year of work, the major portion of the research time was expended in finding and evaluating data sources, in developing analysis techniques and programs, and in pre-examining acquired data. The remaining project period was used to process as much data as time or resources would allow.
II. The Search for Data

The high-frequency slope of the seismic source spectra is not the usual objective for seismic data acquisition. Hence, the availability of data from which these spectral data can be obtained is limited. Ideally, seismic data used for studying the high-frequency spectral content of earthquakes should be recorded on instruments having a wide dynamic range and a broad-band frequency response. Such instruments have been used recently, but the lack of standardized methods for data transfer make access and acquisition difficult.

Telemetry systems which are now used routinely for studies of small earthquakes have a typical dynamic range of about 40 dB and a reasonably flat response to particle velocity for frequencies between 1 and 25 or 30 Hz. Because most systems are set at maximum usable gain levels, most large events at close range will saturate the records. Such non-linear response renders the spectral estimates invalid. At greater distances, absorptive attenuation renders the high-frequency data uninterpretable. The lower-magnitude events have corner frequencies which may be too high to fall within the frequency band-width of the system.

Aside from recording characteristics, other effects can also degrade the seismic data and earthquake spectra. Perhaps
most influences on the high-frequency slope occur as a result of the propagation path taken. Surface wave contamination, wave scattering, lateral inhomogeneities, and most importantly inelastic attenuation are path effects which can bias the recorded data. Also, near-source effects, multiple arrivals, and the selection of the time window can affect the frequencies observed.

In an exercise to identify the range of valid earthquake data, Long and Johnston (1979) combined the characteristics of the recording system, along with expected amplitudes and corner frequencies for certain magnitude events, with the effects of attenuation. Frequencies recorded to about 4 times the corner frequency are needed to derive the spectral slope beyond the corner frequency. In addition, frequencies which are greater than or equal to 20 percent of the peak in instrument response or original unattenuated amplitude are desired to preserve the source effects and keep the signal-to-noise ratio high. Utilizing the mentioned factors, the size and range of events that can be successfully analyzed for high-frequency slope were determined for typical telemetry systems. Average values for corner frequency versus magnitude and amplitude versus distance, from mb(Lg) equation, were used to set limits. The results indicate that the WWSSN data are almost unusable. The results for telemetry systems (Figure 1) show a small but usable range of events. A wider dynamic range as available in
digital systems or from recording at lower gains allows events of larger magnitude or at closer range to be used. The minimum magnitude event that can be utilized is determined by the high-frequency instrument response. The lower limit of $mb=1.5$ in figure (1) is for a frequency response up to 60 Hz. Systems with usable response at higher frequencies (e.g., 200 Hz) will allow use of events smaller than $mb=-1$. The search for data was undertaken with the above guidelines as conditions for usable data.
Figure 1: Range of magnitude versus distance that can be used for determination of the spectral slope past the corner frequency for a typical telemetry system using $m_b(Lg)$. Wide dynamic range or low-gain systems would reduce the possibility of trace saturation at larger magnitudes.
III. Areas Considered

Through a search of the literature and conversations with many individuals, the following potential sources of data were investigated. Many others were not considered because of obvious shortcomings in the available data.

1. **New Brunswick, Canada aftershock data.** This data is available as multiple-station digital event recorder data. However, it was not released to open file in time for our analysis. In addition, the stations are spaced too far apart to resolve shallow depths.

2. **Oregen, California.** There is extensive tape recorded analog data but, according to one reference, much of it is clipped. The larger events are recorded on strong motion instruments, but the high-frequency response may be uncertain. Digital event-recorder data also exist but their general availability was not determined. Some analysis of this data has been published, and it was not known whether sufficient additional data existed to allow establishing a depth-versus-spectra relation.

3. **Arkansas.** This was a very active epicentral zone at which a short monitoring trip in the field could have yielded usable data. The area was tested during a single trip which
covered two days. The data gathered are mostly on smoked paper. The single component data were unable to resolve depth of focus.

4. *Dalashitiao*, China. No data were found.

5. *Jennette Lenses*, California. This is not an area of induced seismicity but instead a concentrated area of high-level activity. The cause may be related to thermal stresses induced by a hypothesized magma body. Many events have been recorded on multiple digital event recorders. A tape of this data was received in 1982 from the U.S.G.S. The data were not used in this study but may be utilized in future analysis as part of thesis research.

6. *Monticello*, South Carolina. The Monticello Reservoir has induced many small events (M<3.0). These have been recorded on analog tape from a seismic telemetry net and, in limited surveys, on digital event recorders. The data recorded on the telemetry net are unlikely to satisfy the criteria for measuring the high-frequency slope. The digital data are available from the U.S.G.S. in Menlo Park and satisfy most of the requirements for spectral analysis.
IV. Data Chosen

We decided to use the data from the Monticello Reservoir area since digital event recorders were used and the stations were close enough together (2.0 to 3.0 Km) to resolve the depth to these shallow events.

We selected 51 microearthquake events located in the vicinity of the reservoir for this study. The 51 events are among the several hundred events recorded in 1979 that are on data tapes provided by the U.S.G.S. One criteria in the selection process is that the time between the P and S phases be less than 0.2 seconds. This corresponds to a distance of less than about 1.0 Km. For these events the probability of a near-vertical ray path was greatest and we expected to obtain more-reliable depth estimates. In addition, the data need to be within the dynamic range of the instrumentation. To insure this, only events that have been plotted and checked for evidence of clipping are used in this study. In the Monticello area, the digital event recorders were placed in an array with a spacing of 2.0 to 3.0 Km. Since the events during the recording period were concentrated in two general areas, only two stations (DUC and SNK) were close enough to the epicenters to record data arriving at near-vertical angles of incidence. However, only a few events were recorded by both stations; therefore, comparison of results between stations was limited.
V. Analysis of Data

The three-component digitized data were sampled at 200 cps and recorded sequentially over the sampling period. Therefore, in our analysis all the data were interpolated to simulate simultaneous sampling. The seismic data were replotted on a wider time scale so that accurate arrival times of the initial P-pulse and S-phase could be obtained.

In order to study the variability in particle motion with time, the covariance matrices, the principle axes, and the direction cosines (Table 1) are computed for successive 10-point (0.08 seconds) windows of data. For this and other windowing schemes in the analysis, the first window ends at the end of the first cycle of the P-wave arrival. Observations of the directions of the principle axes of the particle motion indicate that the arrivals following the first cycle of P-wave motion are complex. The simplest explanation is that the later arrivals are predominately scattered phases. An example of this can be seen in Figure (2) which is a plot of the temporal variation of the apparent angle of incidence of one of the 51 events. This angle is taken as the arc-cosine of the vertical direction cosine (V) of the greatest principle axis (S) of the data within the 10-point window. The plot shows that, after the first motion, the apparent angle of incidence is variable and somewhat periodic as it fluctuates between 30 degrees of
Table 1. Equations

Covariance Matrix $C_{jk}$ $j=1,3$ and $k=1,3$

$C_{11} = \Sigma_i (x_i - \bar{x})^2$; $C_{12} = \Sigma_i (x_i - \bar{x})(y_i - \bar{y})$; $C_{13} = \Sigma_i (x_i - \bar{x})(z_i - \bar{z})$

$C_{22} = \Sigma_i (y_i - \bar{y})^2$; $C_{23} = \Sigma_i (y_i - \bar{y})(z_i - \bar{z})$; $C_{33} = \Sigma_i (z_i - \bar{z})^2$

$x_i =$ east-west component; $y_i =$ north-south component; $z_i =$ vertical component

$i = 1,n$; $n =$ number of sample points for which the covariance matrix is being computed

$\bar{x}$, $\bar{y}$, $\bar{z}$ are the mean values of $x_i$, $y_i$, and $z_i$

The principle axes $S_1$, $S_2$, $S_3$ are computed from a third order polynomial:

$$S^3 - I_1 S^2 + I_2 S - I_3 = 0$$

$I_1 = C_{11} + C_{22} + C_{33}$

$I_2 = C_{11}C_{22} + C_{22}C_{33} + C_{33}C_{11} - C_{12}^2 - C_{23}^2 - C_{31}^2$

$I_3 = C_{11}C_{22}C_{33} + 2C_{12}C_{23}C_{31} - C_{11}C_{23}^2 - C_{22}C_{13}^2 - C_{33}C_{12}^2$

$S_1 = I_1/3 + 2(-a/3)^{1/2} \cdot \cos(\phi/3)$; $a = I_2 - I_1^2/3$

$S_2 = I_1/3 + 2(-a/3)^{1/2} \cdot \cos((\phi+2\pi)/3)$; $\phi = \cos^{-1} \left( (-b/2 \cdot (-a^3/27)^{1/2}) \right)$

$S_3 = I_1/3 + 2(-a/3)^{1/2} \cdot \cos((\phi+4\pi)/3)$; $b = -1/27 \cdot (2I_1^3 - 9I_1I_2 + 27I_3)$

Three directional cosines for each principle axes:

EW = east-west component; NS = north-south component; V = vertical component

$E_W = [1/(1 + G_j^2 + F_j^2)]^{1/2}$; $N_S = G_j \cdot E_W$; $V_j = F_j \cdot E_W$; $j = 1,3$

$G_j = [C_{11}C_{23} - C_{23}S_j - C_{12}C_{13}] / [C_{23}C_{13} - C_{13}S_j - C_{12}C_{23}]$

$F_j = [C_{12}C_{23} - C_{13}(C_{22}-S_j)] / [(C_{33}-S_j)(C_{22}-S_j) - C_{23}^2]$

$\theta_C = 90^\circ - \arccos \left( (1.5 \cdot (1 - \sin(90^\circ - \theta_I)))^{1/2} \right)$

$\theta_C$ is the confined angle of incidence and $\theta_I$ is the apparent angle of incidence.

Pseudo depth (PD) = $1.37 \cdot (6 \text{ Km/sec}) \cdot (S-P) \cdot \cos(\theta_C)$; where $(S-P)$ is the time between initial S and P-phases and assumed velocity is 6 Km/sec.
Figure (2): Vertical scale is apparent angle of incidence. $0^\circ$ is vertical and $90^\circ$ is horizontal. The horizontal axis is time in seconds.
vertical and almost totally horizontal. This temporal variability in the angle of incidence was observed in the other 50 events as well.

The first cycle is used to determine the apparent direction of propagation at the surface. The apparent direction of propagation at the surface is then used to compute the particle motion in the direction of propagation by projecting the three-component data onto this assumed propagation direction. The directional cosines \((E_W, N_S, V)\) of the greatest principle axis of the first motion were used to convert the three directional components \((X_i, Y_i, Z_i)\) to a single trace \(P_T_i\):

\[
P_T_i = X_i*E_W + Y_i*N_S + Z_i*V
\]

Spectral estimates are determined from this trace.

Computations of the displacement spectra of this trace are made using window lengths of 16, 32, 64, and 128 data points. Starting at the initial P-pulse, data within the window is transformed to the frequency domain using a fast Fourier transform algorithm to get the velocity spectrum. The displacement spectrum is obtained by dividing by angular frequency in the frequency domain and \(\log_{10}(\text{displacement}/\text{max displacement})\) versus \(\log_{10}(\text{frequency})\) is plotted. The window in the time domain is then shifted half its length and the procedure repeated until 1 second (200 points) of data is
covered. The 16-point windows gave inconsistent results as their length is comparable to a single cycle of particle motion. The 64 and 128 point length traces tend to include the entire P-wave or both the P- and S-wave arrivals. Therefore, the 32-point window is used for the remaining analysis. The windows are not tapered in the time domain.

The corner frequency and the high-frequency slope are determined from spectral data using a curve fitting technique. Figure (3) illustrates the technique used. Only frequencies less than the corner of the anti-aliasing filter (50 or 70 Hz) are used in the slope determination. The low-frequency asymptote and the high-frequency slope are estimated by fitting the data points to one of a set of theoretical curves for source displacement functions. The corner frequency is taken as the intersection of the low-frequency asymptote and the high-frequency slope.

The window position is a factor in high-frequency slope and corner frequency determination for the S-phase. Since the initial window is set to end at the end of the first cycle of the P-wave arrival, the window position is not a factor in the variation of the spectral estimates in this part of the spectral trace. However, with variations of 20 to 40 sample points (0.1 to 0.2 seconds) in S-P time, the initial S-phase may be either at the leading edge or in the middle of the
Figure (3): Vertical scale is Log10(displacement/max displacement) and horizontal scale is Log10(frequency). Top: Displacement spectrum of initial P-pulse cycle from one station. Corner frequency is 32 Hz and spectral slope is -5. Bottom: Sample high-frequency slopes.
32-point window. The corner frequency and spectral slope increase by an average of 0.7 Hz and by a power of 2 respectively by moving the window from the edge of the initial S-phase to a position where the S-phase is near the window center. When there is a choice of window positioning, the window with its leading edge near the end of the initial S-phase is chosen for the spectral estimate study. Doing this avoids contamination by surface waves.

In studies of the seismic coda, Duckworth (1983) finds that attenuation due to scattering is greater than the attenuation due to inelastic Q. In addition, it is unlikely that Q will be a factor in our analysis of the earlier sections of the seismic trace. In the Piedmont Province, where Monticello is located, the Qp is approximately 750 and the Qs approximately 400 (Long and Johnston, 1979). At these high values, inelastic attenuation does not strongly affect traces recorded at close range.

A rough estimate of the source depth, or pseudo depth (PD), is calculated using angle of incidence for direction and S-P time for distance. In order to take into account the free surface effect, a confined angle of incidence is computed from the apparent angle of incidence. Equations for pseudo depth and confined angle of incidence are given in Table 1. For purposes of comparing the data only, we have assumed a constant
velocity medium in which the confined angle of incidence and S-P time can be used to calculate depth directly. However, a velocity gradient increasing with depth would permit shallower depths for the less steep angle of incidence, but the relative position of the depths should change only slightly.

The spectral estimates of the initial P-wave and S-wave arrivals are used in examining the relationships among the high-frequency slope, pseudo depth, and corner frequency for each of the 51 events.

Figure (4) shows plots of corner frequency versus high-frequency slope for the initial P and S phases. Changes in the corner frequency and slope appear to be positively correlated in that they tend to increase or decrease together. This is more frequently true for the P-phase than for the S-phase when the data are segregated by station. Since corner frequency is inversely proportional to fault radius, this may be an indication that earthquakes with a smaller fault plane area have fewer asperities on the fault plane. Because our data are so restricted in depth, the corner frequency versus spectral slope could be related to the area available on existing fault planes for free or unrestrained slip at the depth of these events (estimated at 1.0 Km). Our data imply a corner frequency of 15 to 20 Hz for an \( w \)-square decay. The inferred radius of faulting is about 0.1 Km and the magnitude
Figure (4): Vertical scale is spectral slope $n$ and horizontal scale is corner frequency in Hz.
is equivalent to 1.5. The relation between corner frequency and spectral slope needs further study to assure that instrument response has been completely removed. This relation also needs to be evaluated in other areas to see if it may be used to predict the maximum earthquake.

The spectral slope of the first motion or P-pulse is usually greater than $w^{-6}$ with values ranging from $w^{-6}$ to $w^{-9}$. Cubic decay would indicate a susceptibility to induced seismicity, and $w^{-4}$ to $w^{-6}$ decay or less would indicate tectonic processes or induced seismicity of deeper focus earthquakes (Long, 1982).

Figures (5 and 5) show the relation between pseudo depth and high-frequency slope. There appears to be no depth variation in the slope, although there may be an increase in slope with increased depth for corner frequencies above the median corner frequency. This relation was not expected and may relate to ambient stress conditions. However, the slight relation between depth and slope can not be given much weight since the result could be a consequence of the limited depth range in the data and other restrictions on choice of appropriate data. No relationship is found between the pseudo depth and corner frequency (see figure 7).
Figure (5): P-phase Vertical scale is pseudo depth in Km and horizontal scale is high-frequency slope for: A. both stations, B. station SNK, C. station DUC, D. frequencies above the corner frequency median (27.5 Hz), E. frequencies below this median.
Figure (6): S-phase Vertical scale is pseudo depth in Km and horizontal scale is high-frequency slope n for: A. both stations, B. station SNK, C. station DUC, D. frequencies above the corner frequency median (26.5 Hz), E. frequencies below this median.
Figure (7): Vertical scale is pseudo depth in Km and horizontal scale is corner frequency in Hz.
Source depth estimates and high-frequency slope data from Monticello are superimposed onto similar data from Oroville, California, whose earthquakes are also believed to be reservoir induced (Long and Johnston, 1979). Source depths at Oroville are greater than the apparent depths determined here for Monticello. Figure (8) shows the superposition for the initial P- and S-phases. The relatively limited range in data from Monticello is clearly seen. The relationship that was sought between slope and depth is not obvious in the P-phase data; however, the trend for spectral slope to decrease with increasing source depth is observed in the S-phase data.
Figure (8): Superposition of Monticello data (●) with Oroville data (★). Vertical scale is depth in Km. Horizontal scale is spectral slope.
VI. Summary

A wide station spacing and the resulting need to limit data to those with a short S-P time has restricted the data set. This restriction was necessary to obtain a near-vertical ray path and use single station location techniques to locate the events.

Evaluations of particle motion indicate that the arrivals following the first cycle of P-wave motion are complex. This may be an indication of later arrivals being predominately scattered phases. The first cycle is used to determine the direction of propagation which in turn is used to compute the particle motion along the ray path. Spectral estimates are determined from the trace obtained by projecting the three-component seismic data onto the computed propagation direction.

Window lengths used in the transformation of the trace to the frequency domain must be selected carefully. A 32-point window is selected after eliminating a 16-point window for being comparable in length to a single cycle of particle motion and eliminating 64- and 128-point windows for containing the entire P-wave or both the P and S arrivals. Also, the window position is crucial. Corner frequencies and spectral slopes vary as the window is moved along the trace.
Comparisons among the 51 selected events indicate that the changes in corner frequency and spectral slope are positively correlated in that both tend to increase or decrease together. This may indicate that earthquakes with a smaller fault plane area have fewer asperities on the fault plane.

For the P-pulse, the spectral slopes are generally greater than w-cube. In addition, there is no depth variation in the high-frequency slope for either the P- or S-phases. Both of these results may be consequences of the narrow depth range of the data. The depth range is increased when the Monticello data are superimposed onto data from Brawley, California, whose earthquakes are up to 3 times as deep as those chosen for this study. The clustering of Monticello data over a narrow range of depths is clearly seen. The trend for high-frequency slopes to decrease with increasing depth is observed for the superimposed S-phase data, but it is not so obvious for the P-phase data.


Bibliography


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