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IN THE CONTERMINOUS UNITED STATES

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ON THE REGIONALIZATION OF GROUND MOTION ATTENUATION IN THE CONTERMINOUS UNITED STATES*

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ABSTRACT

Attenuation results from geometric spreading and from absorption. The former is almost independent of crustal geology or physiographic region. The latter depends strongly on crustal geology and the state of the Earth's upper mantle. Except for very high-frequency waves, absorption does not affect ground motion at distances less than 25 to 50 km. Thus, in the near-field zone, the attenuation in the eastern United States will be similar to that in the western United States. Most of the differences in ground motion can be accounted for by differences in attenuation caused by differences in absorption. The other important factor is that for some western earthquakes the fault breaks the Earth's surface, resulting in larger ground motion. No eastern earthquakes are known to have broken the Earth's surface by faulting. The stress drop of eastern earthquakes may be higher than for western earthquakes of the same seismic moment, which would affect the high-frequency spectral content. But we believe this factor is of much less significance than differences in absorption in explaining the differences in ground motion between the East and the West.

INTRODUCTION

The attenuation of ground motion originating from an earthquake is complex in nature and it consists of a combination of harmonic motions of varying frequency. The composition of frequencies and the percentage of strong motion caused by different waves (i.e., P waves, SH and SV waves, and surface waves, etc.) undergoes changes along the transmission path. In the near and intermediate field (say, distances to about 150 km from the source), the body waves carry a significant portion of strong motion in the intermediate field and diminish in amplitudes as R^{-n} (where R is the distance from the source). Surface waves carry a significant portion of strong motion in the far field and diminish as R^{-m}, $n>m$, which partially account for the change in attenuation rates between the intermediate and far-fields. The amplitude and duration of surface waves at these distances may be significant for structures with a high natural period.

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During their travel along the transmission path, the amplitudes of both body and surface waves decay with distance due to geometric spreading and internal damping within the material. Within the conterminous United States, a regional propagation difference between the eastern and western parts is apparent. The much larger felt areas for similar sized earthquake in the East necessitates consideration of the effect of distant large events such as 1811 New Madrid and 1886 Charleston. In what ways is the attenuation of ground motion from these earthquakes in the East different than typical western United States earthquakes from sources less than say, 150 km away? What is the implication of "low" attenuation observed for the propagation paths in the East upon the ground motion attenuation?

Because of the potential for serious damage to structures and injury to people, predictions of future ground motion from earthquakes in the eastern United States are of much more than academic interest. The possibility of seismic activity is especially important with relation to the selection of sites for, and the design of, nuclear power plants. More than 80% of the plants in existence, under construction, or being planned are located east of the Mississippi River, and, of that 80%, a heavy concentration is in the northeastern States.

Among the design criteria for such plants, the maximum level of ground motion from regional earthquakes that the plant must withstand is probably the most difficult to establish because of our poor understanding of (1) intraplate earthquake occurrences and (2) the difficulty in quantifying attenuation of strong ground motion in the eastern United States. The approach has been to associate the maximum intensity with a maximum ground acceleration through empirical relationships and to use reported intensities of past earthquakes in the region to infer the maximum ground acceleration that would be experienced in the future at a given location.

Inferences must be made about the attenuation of ground motion in the East by studying systematic differences or similarities between the eastern United States and other regions of the world regarding information that is indirectly related to ground motion such as intensity data, local crustal models, attenuation of surface waves, etc.

In this paper, we discuss in depth the evidence developed in a number of very diverse studies which suggest that there is a significant difference in attenuation between the eastern United States (EUS) and the western United States (WUS). We will attempt to unify these studies to infer what implications different observations might have on the attenuation of strong ground motion in the EUS. Some of the important evidence are:

- Modified Mercalli (MM) intensity attenuates slower in the EUS than in the WUS, based on historical intensity data (see Refs. 1-3).

- There are higher propagation velocities at depths in the EUS than in the WUS (see Refs. 4-7).

- There are higher seismic quality Q-values (thus lower damping) in the EUS than in the WUS (see Refs. 7-15).

- There is less pronounced low Q-zone in the upper mantle in the EUS (Refs. 11-17).

- Propagation and prominence characteristics of surface waves are different in the EUS than in the WUS (see Refs. 2, 8, 10, 18).

We observe that the regional difference in attenuation (as well as velocity) is real between the eastern and western United States. We investigate the implications of this regionalization of both the attenuative property and the propagation velocities upon ground motion attenuation at near-fields. We suggest that some of the differences in ground motion might be due to source mechanism differences. The implication is that this might mean the EUS events were higher stress drop events and that these EUS events would have higher acceleration in the near-field.

EVIDENCE OF REGIONALIZATION FROM P- AND S-WAVES

Herrin and Taggart (Ref. 4) for the first time presented clear evidence of the existence of significant regional differences in P_n velocity in the conterminous United States. These authors concluded that the station corrections due to these regional variations are positive in the western conterminous United States where P_n velocities are relatively low ($\sim$7.80 km/sec), but the teleseismic travel time delays are negative in the higher P_n velocity ($\sim$8.10 km/sec) regions of the eastern conterminous United States.

The observations of Herrin and Taggart (Ref. 4) of P_n velocity, and the average short-period P-wave amplitude residuals determined by Cleary (Ref. 19) and Booth et al (Ref. 20) were used by Marshall and Springer (Ref. 21) to illustrate that there is an empirical relationship between mean station amplitude residuals and the local P_n velocities. These observations strongly suggest that the local P_n velocity is indeed a measure of regional seismic attenuation properties of the earth's upper mantle beneath the seismic station. A high local P_n velocity implies a positive station-magnitude bias beneath the station; a low local P_n velocity, a negative station bias. Also implied here is that the regions of lower P_n velocities would show higher attenuation of P amplitudes. This empirical relationship is of fundamental importance because estimates of seismic Q in the earth's upper mantle as well as of corresponding amplitude residuals can therefore be made from measured P_n velocities in many parts of the world.

Lateral variations in the structure and physical state of the crust and upper mantle of the earth bear directly on these differences. Solomon and Toksöz (Ref. 11) have observed the attenuation of P_n and S_n waves over the conterminous United States. The regions of high attenuation (thus low Q) for both types of wave lie between the Rockies and the Sierra Nevada Mountains. The regions of low attenuation (thus high Q) are the central and eastern regions and along the Pacific Coast Province. Again, it appears that the attenuation results correlate well with apparent heat flow data and seismic wave velocities.

Molnar and Oliver (Ref. 7) used the S_n velocity as a discriminant to test continuity of lithosphere in many regions of the world including the conterminous North America. S_n transmits efficiently in the Canadian shield and the entire eastern United States. However, the western North America does not either transmit S_n at all, or shows severe attenuation over paths in the western United States, the Gulf of California and the Mexican Plateau.

Cleary and Hales (Ref. 6a) observed regional variations in P-wave travel-time residuals across North America. The arrivals were as much as one second early in the central United States and up to one second late in the Basin and Range Province in the West. Doyle and Hales (Ref. 6b), in a similar study of S-waves found a range of travel-time anomalies of about eight seconds, with negative values generally occurring in the eastern and central United States. As was observed by Hales and Herrins (Ref. 5), the travel-time anomalies in the conterminous United States associated with local station conditions are: The S-wave residuals have a range of over eight seconds, while the P-wave residuals range over three seconds. The travel times of both P- and S-waves are slow relative to the Jeffreys-Bullen curves in the West, while the times are early in the East. The scatters in the data are too great to try to correlate residuals with provinces. The correlation coefficient of the P- and S-wave anomalies is seen to be 0.75 and the slope of the regression line is 3.72 ± 1.0, implying that regions of the P-wave anomaly have a similar although much larger S-wave time anomaly.

The magnitude of the times suggests that the apparent effects of partial melting in the upper mantle may be an important factor in the delay-times, since a mechanism that changes the Poisson's ratio is needed. Constant Poisson's ratio implies that the ratio of the travel-time anomalies be inversely proportional to the ratio of the P- and S-wave velocities, implying a slope of 1.7 to 1.8. What is observed from the seismic data is that the Poisson's ratio in the East is more or less constant, while in the West this ratio is about 5 to 10% higher than that in the East. These are most easily explained by lateral variations in the velocity and attenuation structures of the of the earth's upper mantle and lower crust between the East and the West.

The relative attenuation of body wave passing through the upper mantle beneath North America has been the subject of several recent studies. Solomon and Toksöz (Ref. 11) determined a differential attenuation

$$\delta t^* = \pi \int_{\text{Path}} \delta Q_\beta^{-1}(s,f)\, \beta^{-1}(s)\, ds$$

where β is the shear wave velocity and δQ_β^{-1} is the departure of the true anelasticity, at a point along the ray path, from a radially symmetric Q_β^{-1} distribution. Positive values indicate greater than normal attenuation and vice versa. The results of Solomon and Toksöz (Ref. 11) for both P- and S-waves, indicate high attenuation between the Rocky Mountains and the Sierra Nevada-Cascade ranges and low attenuation throughout most of the central and eastern portions of the conterminous United States.

Der, Masse, and Gurski (Ref. 13a) observed consistent patterns of attenuation for short-period teleseismic P- and S-waves. Their analysis shows that greater attenuation occurs for both types of waves in the western United States. In a later study, Der et al (Ref. 13b) determined average Q values for ray paths from the SALMON nuclear explosion to various LRSM stations. Fig. 1, reproduced from Ref. 13b, shows that as the NW profile crosses over the Rocky Mountain front, the dominant period of P-waves increases to about one second, while those with paths located entirely in the eastern North America have dominant periods around 0.3 to 0.5 sec. This change in spectral content illustrates dramatically the attenuation characteristics of high frequency waves for paths traversing the upper mantle of the WUS as compared to the same frequencies for the other paths (i.e., the NE and N profiles).

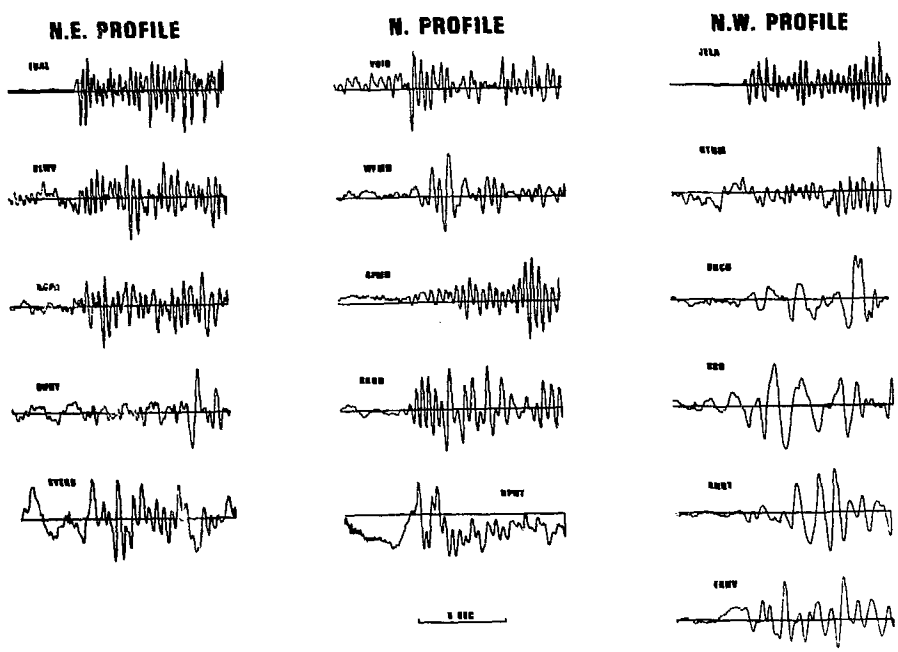

Figure 1. P-Wave Seismograms Along Three Profiles From The SALMON Nuclear Explosion (After Der and McElfresh, 1976).

EVIDENCE FOR REGIONALIZATION FROM SURFACE WAVES

From the viewpoint of obtaining regional structures, surface wave data are very valuable. In addition, the anelastic attenuation of surface waves are of practical importance to a proper description of ground motions caused by earthquakes (see Ref. 8). Love and Rayleigh wave dispersions are most sensitive to shear wave velocity distributions, which are in general difficult to measure using the usual S-waves. Also, surface waves in the period ranges detected on long period WWSSN instruments are sensitive to the details of the upper 400 km of the Earth's interior, which corresponds to body wave arrivals at less than 30 degrees distance. This range is very difficult in practice for body wave interpretations due to small amplitudes and multiple phases arriving within a short period of time on the seismograms.

Surface wave measurements over the conterminous United States were made by Pilant (Ref. 22) in the period range of 20 to 50 seconds. A sharp gradient in phase velocity of Rayleigh waves across the Rockies and much lower phase velocities in the west are prominent features at these periods. Also an evidence is seen from the records for higher phase velocities at the western coastal margins. The structure that Pilant (Ref. 22) inferred for the western region has a very low upper mantle velocity compared with the CANSD model of Brune and Dorman (Ref. 23), consistent with heating and partial melting of the upper mantle below the crust in the west. No evidence exists for the presence of low velocity zone in the east.

Biswas and Knopoff (Ref. 14) give results of two stations Rayleigh wave phase velocity measurements in the conterminous United States in the period range 20 to 250 seconds. They find three sub-regions, called the western United States, the southcentral United States, and the northcentral United States. The western United States region corresponds to the physiographic provinces west of the Rockies. Five two-station paths in the western United States were found to have phase velocities that agreed to within about 1% at all periods, but which showed differences of up to 10% compared with the eastern regions. Phase velocities for the eastern provinces are much higher, especially in the northcentral United States, where velocities become indistinguishable from the values observed in the Canadian Shield regions.

A few determinations of upper mantle Q beneath that United States have been reported, and among those are Lee and Solomon (Ref. 12) in the eastern United States and Archambeau et al (Ref. 16), Solomon and Toksöz (Ref. 11), Biswas and Knopoff (Ref. 14), Lee and Solomon (Ref. 12), and Kanamori (Ref. 24) in the western United States. The Q values are in the range of 300 to 1100 in the eastern United States and 20 to 500 in the western United States. These observations have qualitatively established that low values of upper mantle Q predominate in the eastern United States.

The only Q model pertaining to the upper mantle beneath the eastern United States, where the tectonics resemble those of shield regions, was derived by the Lee and Solomon (Ref. 12) from simultaneous inversions of surface-wave phase-velocity and attenuation data. Although the model shows much scatter, it does not require a low-Q zone in the upper mantle of the eastern United States to give an adequate fit to the data. In the western United States, the model shows clear evidence of a low-Q zone in the upper mantle.

In summary, a regional attenuation (as well as velocity) difference is apparent between eastern and western United States. Figure 2 is an interpretation of our estimated Q for paths in the 8.2 km/sec P_n velocity upper mantle as contours of Q at multiples of 250 up to Q = 1000. A more detailed contouring interval is not considered justified because of the necessity of subjectivity judging the areal extent represented by a single path determination and the estimated reliability of the determination. The Q distribution in the lower velocity materials in the western United States is given in Figure 2 with a contour interval of 100 beginning at Q = 150. What then is the implication of this regionalization of both the attenuative property and the propagation velocities upon ground motion attenuation at near-fields?

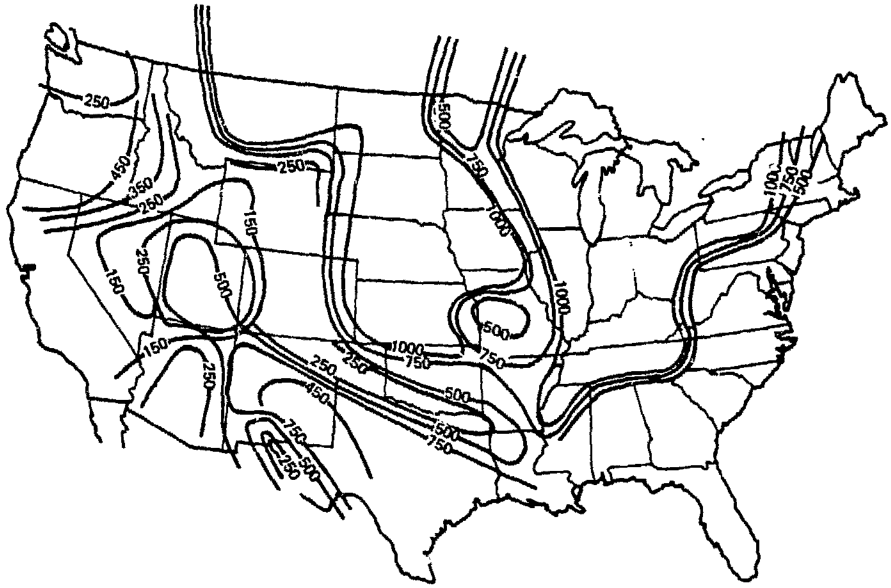

Figure 2. Contours Of Upper Mantle Q From P_n Paths In The Conterminous United States.

IMPLICATIONS FOR REGIONALIZATION OF GROUND MOTION ATTENUATION

Ground motion attenuation that we refer to herein is the decay in the amplitude of earthquake motion caused by sudden slippage on an earthquake source. Key factors influencing attenuation are the source conditions, transmission path characteristics, and local site conditions themselves. In the foregone discussion, we presented seismological envidences related principally to the transmission path characteristics. The source conditions influence the initial level of the ground motion attenuation pattern and the frequency content of the motion. Local site conditions further influence the level and frequency content of ground motion. The transmission path characteristics influence the path length and rate of amplitude decay.

For an earthquake having the Richter magnitudes greater than say, 4.0, the amplitude and duration of seismic waves at distances less than about 150 km from the source may be significant for structures with a high natural period.

In order to be able to make comparisons between ground motion from various earthquakes, we need a methodology to extrapolate -- for a given earthquake -- the observed ground motion into the near-field. The problem is too complex to directly derive the appropriate law; however, our observations from both point source and line source (salvo-type) explosions show that

$$G(r) = \frac{A_o}{r^n} \qquad (1)$$

where G = peak ground acceleration, velocity or displacement
r = hypocentral distance
n, A_o = constants dependent of the ground motion parameter, geological parameters and yield of the explosion

A single value for the constant n appears adequate to cover the range $L/r \geq 0.25$, where L/r is the ratio of the length of the salvo explosive array to the hypocentral distance, from both the near to far-field. Bernreuter (Ref. 25) gives a number of examples. Figure 3 shows typical results with data obtained from the BUGGY experiment, a salvo-type (line source) underground nuclear explosion for canal excavation performed in 1968. It is found that the exponent n is in the neighborhood of 1.7 for the peak particle velocity, and this is clearly illustrated in Figure 3. As shown by Bernreuter (Refs. 25, 26) the $n \approx 1.7$ is general for all available data obtained from the underground nuclear explosion experiments.

The attenuation of the peak ground motion with epicentral range is of particular interest for several reasons: First, the high-frequency end of the spectra is fixed by the peak acceleration. Second, estimates of the stress drop can be obtained if estimates of the peak velocity in the near-field can be made. In addition, the interpretation for the attenuation of high-frequency parts of the ground motion may be made in terms of the peak acceleration or velocity.

Figure 4 is an illustration for the attenuation of vertical ground velocity from the 1971 San Fernando earthquake event. It is seen that the velocity attenuates as the inverse r^n, where $n \approx 1.7$. In fact, it was found that $n \approx 1.8$ fits all the available earthquake data for the attenuation of peak acceleration and $n \approx 1.7$ for peak velocity. Therefore, we believe that the peak velocity attenuates with a A_o/r^n law is general.

What is implied in these findings above is that the nonlinear behavior of the ground during the passage of strong motion may not be too important compared to uncertainties introduced by other geological and seismological parameters. It is reasonable to back extrapolate far-field data into the near-field using the relation A_o/r^n. It is then possible to use far-field ground motion data to make reasonable estimates of the near-field. This becomes particularly useful in the EUS where as little near-field data are available.

Nuttli (Ref. 8) presented estimates of "peak sustained values of the vertical component of the velocity" observed at seismograph stations throughout the Midwest from the 1968 Illinois earthquake of magnitude 5.5.

These data are plotted in Figure 5 along with earthquake data from the western United States for magnitudes between 4.5 and 5.8. (It is important to note that Nuttli's data were obtained from the WWSSN seismographs. The WWSSN instruments were not designed to read velocity directly. Velocity must be estimated by determining the period and computing the (amplitude/period) ratio. Because of the inherent instrument response characteristics, the high frequency part of the ground velocity is filtered out from the seismographs. Nuttli (Ref. 8) reported the peak sustained value rather than the true peak, this must be kept in mind when interpreting the results of this study. In addition, to correct the vertical velocity to an acceleration, we make use of average values through the scatter of data; thus the final estimates of ground motion made are in essence average peak values).

Several important features are to be noted from Figure 5. First, it is evident that the form of Eq. (1) fits the data well. Second, if the WUS data are extrapolated into the far-field, we see very little difference between the vertical velocity observed in the West and in the central United States. This is surprising in view of the apparent differences in the attenuation of intensities observed between two regions. It appears then that there is an apparent paradox. On the one hand we showed that the peak ground velocity from a magnitude 5.5 earthquake in the central United States was very similar as the peak ground velocity from similar earthquakes in the West both near- and far-fields. Yet, on the other hand, the damage area and the felt area is typically much larger in the central United States than in California. This raises the important question: Why this large difference in damage and felt areas?

Part of the difference possibly can be attributed to differences in constructions. Human response to vibrations plays an important role in establishing the local Modified Mercalli (MM) magnitude scale. On would expect human reaction to be similar in both the East and West except possibly at the lower end of the MM scale. Our response to any vibrations are most sensitive to motion in the frequency bands near resonant conditions for different parts of the body. Goldman and Von Gierke (Ref. 27) have shown that the typical natural frequencies of the Thora-Abdomen system is between 3 and 4 Hz. For sitting men, the fundamental frequency of the whole body is between 4 and 16 Hz. For standing men, this frequency is between 5 and 12 Hz. Resonance of the head relative to the shoulders falls within the 20 to 30 Hz range. If, as expected, human response to vibrations is strongly influenced by the resonant conditions, one would expect a much greater human awareness in the range of 2 to 20 Hz.

It would appear that man is sensitive -- relative to the spectral content of earthquake ground acceleration -- to the higher frequencies. This fact, and a factor of two or three differences in the specific seismic attenuation factor 1/Q could well explain the differences observed in felt area between the central United States and California. Similar reasoning may follow for the differences between the EUS and WUS as well.

The ground velocity is primarily associated with the longer period waves and this is the case also in the near-field. It is generally assumed that the attenuation of seismic energy is due to geometric spreading as well as material properties which cause both attenuation and dispersion. Attenuation is expressed as (see Ref. 28).

$$a(r,f) = S(r) \exp - \frac{\pi f}{CQ} r \tag{2}$$

$a(r,f)$ = amplitude of given wave
$S(r)$ = factor to account for geometric spreading
f = frequency of wave harmonic (in Hz)
C = appropriate phase velocity
Q = seismic quality factor

It is reasonable to assume that the geometric spreading term $S(r)$ is the same in both the East and West. The major differences in amplitude of the ground motion are then related to Q and to some extent to differences in the appropriate average phase velocity. Let us assume for the sake of a simple illustration that the appropriate value of C to use in Eq. (2) is approximately the same for the East and West. The ratio of the amplitude of the various harmonics is then given by

$$\frac{a_W}{a_E} = \exp\left[\frac{f\pi}{C}\left(\frac{Q_W - Q_E}{Q_E Q_W}\right) r\right] \tag{3}$$

From the earlier sections, we take typical Q-values for the East and West as $Q_E = 660$ and $Q_W = 330$ for the purpose of our illustration. Figure 6 is a plot of Eq. (3), where Q in the East is a factor of two greater than in the West. It is seen that for periods greater than one second we observe very little difference in attenuation out to about 500 km. Because the velocity is primarily associated with longer period waves, one would indeed expect to see little difference in the far-field values of observed ground velocity between the East and West, and Figure 6 illustrates this.

The Q has a predominant effect upon the higher frequency components of the ground motion, as seen in Figure 6. Thus, in the East as the same epicentral distance, we would expect to see a higher peak acceleration for the same ground velocity than in the West.

The result of our on-going research indicates that the source size not only affects the corner frequency but also has an effect on the total seismic energy. Combining the factor related to the difference in the attenuation with the factor related to this source size results in about an order of magnitude increase of the ground motion in the East as compared to the West. We conclude then that the differences observed between the East and West may be related to an average difference in source parameters as well as the differences in absorption.

The other important factor in the apparent difference of ground motion attenuation is that for some western earthquakes the fault breaks the Earth's surface, resulting in larger ground motion. No eastern earthquakes are known to have broken the Earth's surface by faulting. The stress drop of eastern earthquakes may be higher than for western earthquakes of the same seismic moment, which would affect the high-frequency spectral content. But, we believe, most of the differences in ground motion can be accounted for by differences in attenuation caused by difference in absorption between the eastern and western United States. One of the major effects of the lower attenuation in the eastern United States will be to enrich high-frequency parts of the ground motion spectra.

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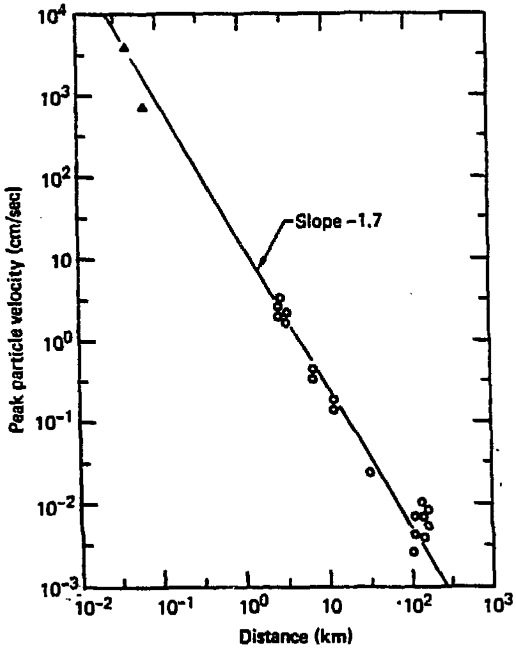

Figure 3. Peak Particle Velocity As A Function Of Distance From A Salvo-Type BUGGY Underground Nuclear Explosion Source. (O) Data Points Obtained From The Velocity Gauges And (▲) Data Points From The Spall Measurements.

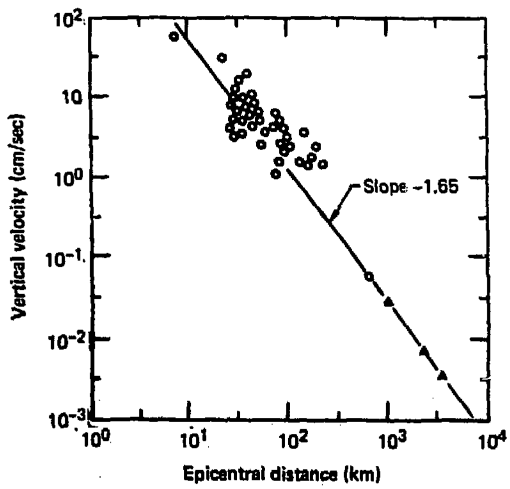

Figure 4. Attenuation Of Vertical Ground Velocity From The 1971 San Fernando Earthquake. (O) Strong Motion Data And (▲) Data Points From The Amplitude Of Ground Velocity Based On Surface Wave Magnitude.

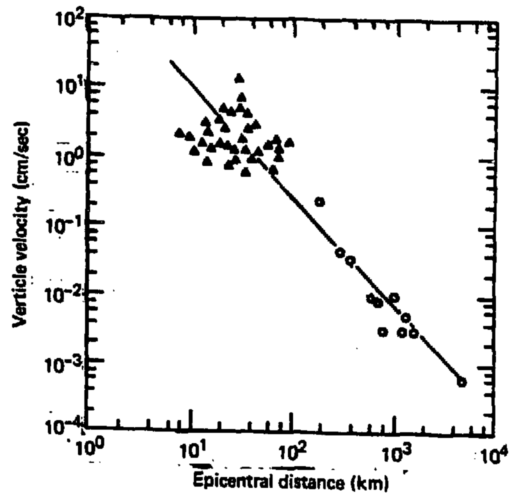

Figure 5. Comparison Of Vertical Components Of Velocity From Earthquakes In The Western United States (▲) For The Magnitude Range Of 4.5 To 5.8 And The 1968 Central Illinois Earthquake (○).

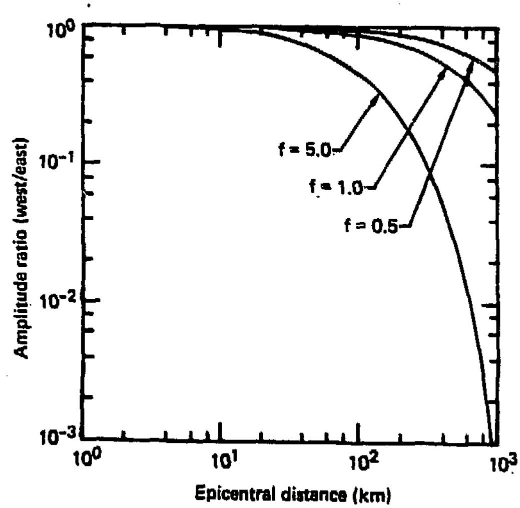

Figure 6. Effect Of Q_E vs. Q_W On Attenuation As A Function Of Epicentral Distance And Frequency. Q_E = 660 And Q_W = 330.