Weston Geophysical
CORPORATION

August 3, 1978

Mr. R. G. Domer
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Attention: Mr. R. O. Barnett

Gentlemen:

In accordance with our contract TV-43410A, a study was conducted on the "Prediction of Strong Motions for Eastern North America on the Basis of Magnitude".

Preliminary data have been previously submitted; this is a formal presentation of our findings.

Sincerely,

Richard J. Holt

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RJH:djc

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Supplemental Report:

PREDICTION OF STRONG MOTIONS FOR EASTERN NORTH AMERICA ON THE BASIS OF MAGNITUDE

to the

TENNESSEE VALLEY AUTHORITY
REPORT

JUSTIFICATION OF THE SEISMIC DESIGN CRITERIA
USED FOR THE SEQUOYAH, WATTS BAR, AND
BELLEFONTE NUCLEAR POWER PLANTS

PHASE II

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The fundamental research of this report was sponsored by the Boston Edison Company and the Tennessee Valley Authority. The present text and especially its appendices represent a substantial revision of the original report submitted under the Pilgrim Docket for the Boston Edison Company. The basis of the revisions is in response to questions and comments formulated by the Nuclear Regulatory Commission and their consultants concerning the original report.

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PREDICTION OF STRONG MOTIONS FOR EASTERN NORTH AMERICA ON THE BASIS OF MAGNITUDE

INTRODUCTION

The present study develops an alternate method for predicting ground motion spectra at specific sites in eastern North America. In the design of critical structures, maximum accelerations resulting from local and/or regional earthquakes constitute an essential parameter. Presently, the technique used depends heavily on the Modified Mercalli (MM) intensity-acceleration relationships derived from observations obtained mostly in western United States. Both the theoretical and experimental weaknesses of this technique are witnessed by the large scatter of the intensity-acceleration data. This has led to a very conservative interpretation and use of the data. The new approach eliminates many shortcomings of the intensity-acceleration technique by having recourse to the magnitude concept in the prediction of ground motion level for bedrock sites at specified epicentral distances.

The magnitude concept uses instrumental measurements and empirically established distance corrections to scale the relative sizes of earthquakes. Historically, the magnitude concept was devised by Richter (1935) and Gutenberg and Richter (1936) as a more objective estimation of the energy released at the source; its superiority over the epicentral intensity concept is derived from the fact that many external biases of soil conditions, particularities of construction, as well as the subjectivity of observers, are excluded in favor of calibrated measurements of ground motions. By using the m_{bLg} magnitude scale (Nuttli, 1973a) that was derived and confirmed for eastern North America, the present method has the advantage of taking into consideration the appropriate attenuation. The differences between the eastern and western parts of the continent are well known for wave attenuation, type of mechanism, and other source parameters.

Briefly outlined, the proposed method starts from a specified 1-second period m_{bLg} magnitude (and its associated sustained ground motion amplitude) and a description of the short-period, higher-mode displacement spectrum within the period range of interest. The spectral amplitude is then scaled to the desired epicentral distance and converted to the acceleration power spectral density function applicable to the top of bedrock at the site. At this point, $S_v(2\pi, 0.1)$, the (response) spectral velocity at 10% damping and natural period of T_n = 1-second ($\omega=2\pi$) is predicted by random vibration analysis. Finally, peak ground acceleration a_g is estimated from $S_v(2\pi, 0.1)$ and standard unit response spectra are adopted. At present, this latter step is necessarily based on strong-motion data from outside the eastern United States because such data are not yet available for eastern United States earthquakes.

The proposed method utilizes a spectral source model derived from eastern North American earthquakes (Street and Turcotte, 1977). The data recorded at seismographic stations located on bedrock makes the method strictly applicable to bedrock sites within eastern North America; however, it could readily be extrapolated to other regions by use of the appropriate parameters. It could also be used for structures on soil foundations, provided the modifying influence of the soil column is taken into consideration.

NEEDS FOR AN ALTERNATE METHOD

Before describing the details of the proposed method, some problems related to the intensity-acceleration approach will be reviewed. These implicitly point out the advantages of this alternate technique.

First, the peak acceleration-intensity relationship carries all the problems and difficulties related to the determination of a maximum epicentral intensity: 1) varying amounts of effort made to compile and evaluate intensity reports (field observations, questionnaires, telephone calls, newspaper reports, etc.); 2) varying degrees of observational subjectivity; 3) biasing influence of population distribution; 4) local construction practices and styles; and 5) soil amplifications, etc.

Secondly, the migration of intensity data (and their associated peak accelerations) to other locations is also questionable since it generalizes what is particular by definition. Clearly, the maximum intensity of an individual earthquake, even if it were correctly determined, remains a function of the source parameters, the transmitting paths, and the site conditions. Because these particular factors influence the resulting observed accelerations, the migration of this observed acceleration is valid only to the extent that the potential earthquake can be truly associated with the same particulars. The process of prediction becomes more questionable when it is used with a site intensity (I_S) value obtained by attenuating an epicentral intensity (I_O) through the distance separating the site and the potential (or historical) earthquake; this suggests that the site intensity (I_S) is equal to an epicentral intensity (I_O). For example, the acceleration spectrum resulting from a nearby epicentral Intensity VII(I_O) at a given site (I_S) is theoretically different from an acceleration spectrum of an Intensity VII(I_S) taken at a site, but resulting from an event with Intensity IX(I_O) located 100 km away and attenuated to the site. Yet, the currently used technique does not account or correct for the gross assumption that near-field and far-field attenuated intensity-acceleration values are the same.

Attenuation laws are functions of distance, frequency, and regional geology. Unless a method predicting ground motion accounts for these parameters, it should be used with great caution, and whenever possible, replaced by other methods which consider these particulars. A recognized fact (based on experimental and calibrated magnitude information) is that events of similar magnitude have different felt areas in the west and east in the United States and Canada. This fact shows the importance of regional attenuation; it stresses the uncertainty of transposing from one region to another the intensity-fall-off data as well as the corresponding intensity-acceleration relationship.

The authors of the currently used intensity-acceleration relationship (Trifunac and Brady, 1975) have strongly stressed "that the physical basis for correlating an earthquake intensity scale with the recorded levels of strong motion is dubious indeed". They "emphasized the weaknesses in carrying out such correlations, as well as the wide scatter of the measured peak values"; they "do not recommend the use of these average trends for routine engineering design".

Recognizing the weakness of the peak acceleration versus intensity correlations, other parameters have been suggested for use in conjunction with peak values to explain the observed intensities. For example, Bolt (1973), Kobayashi (1973), Cloud (1973), and Bolt et al (1975) have pointed out the importance of duration of strong shaking. Studies by Seed and Idriss (1971) and Youd et al (1976) have demonstrated the particular importance of this parameter with respect to liquefaction. Ploessel and Slossen (1974), by suggesting that repeatable high ground accelerations are of more engineering significance than the maximum peak acceleration, likewise imply the importance of the duration of the ground motion.

Other authors, such as Puchkov (1963), Kanai (1967), Necioglu and Nuttli (1971), and McGuire (1977a), by arguing that the velocity of the particle ground motion correlates as well as, or better than, peak acceleration with intensity data are, in effect, taking the frequency content of the ground motion into consideration. For example, when discussing liquefaction, Puchkov emphasizes the importance of the period of the ground motion as well as its associated amplitude, and he concluded this was best achieved by particle velocity measurements.

It is realized from the standpoint of producing damage to buildings and other effects upon which MM intensity ratings are based that all of these parameters (peak ground motion, frequency content, and duration of ground motion) are of varying significance depending upon such variables as the type of structure, foundation material, etc. Theory as well as observations demonstrate the inadequacies of a single parameter, such as measured peak acceleration values, in explaining observed intensities. Therefore, MM intensity alone cannot be expected to predict accurately the peak ground acceleration.

ALTERNATE METHOD OF PREDICTING GROUND MOTION

Up to this point, the more troublesome problems of the intensity-acceleration approach have been pointed out, particularly for eastern North America for which so few strong motion data exist. The present study develops an alternate approach. Its major advantage is a fundamental dependence on instrumentally-observed data.

The prediction of ground accelerations is achieved through the following six steps:

1. for a specified m_{bLg} value (that of the selected design earthquake), calculate the corresponding 1-second ground displacement amplitude from the source spectrum $S(\omega)$;
2. for a specified epicentral distance, obtain the site displacement spectral level for the 1-second period $\Omega_r(2\pi)$;
3. obtain the acceleration spectral density function ordinate at $\omega=2\pi$, $G_r(2\pi)$;
4. apply random vibration analysis to predict $S_V(2\pi, 0.1)$, the ordinate of the response spectrum at the natural frequency $\omega=2\pi$ (i.e., natural period, T_n = 1 second) and 10% damping;
5. utilize western strong motion data to predict peak ground acceleration a_g from $S_V(2\pi, 0.1)$;
6. scale a set of standard response spectra (such as those in Regulatory Guide 1.60).

Step 1 - Calculations of the $S(2\pi)$ Displacement Spectral Level

The present method assumes that the design earthquake can be defined in terms of Nuttli's magnitude (m_{bLg}). It is not the object of this study to describe the various techniques from which intensity data can be used to establish a corresponding magnitude estimate; studies by Nuttli (1973b), Nuttli and Zollweg (1974), Street and Turcotte (1977), using intensity fall-off, total felt area and intensity $\geq$IV felt area, respectively, have achieved a certain degree of coherency in this matter. Bollinger (1977) has combined techniques to assign magnitude estimates for some important historical events, including those of Charleston, South Carolina and Giles County, Virginia. In the present application of the method, the Giles County, Virginia 1897 earthquake, chosen for establishing the design earthquake for the nuclear sites of the Tennessee Valley Authority (TVA), is best characterized by an equivalent m_{bLg}=5.8, as suggested by and discussed with the Nuclear Regulatory Commission (NRC). Additional research by Weston Geophysical (Document 1) on the peripheral limits of the felt area was conducted in order to establish beyond doubt the value of the total felt area found by Hopper and Bollinger (1971) and used by Bollinger (Personal Communication, 1978) in his magnitude estimate given to NRC.

In eastern North America, the most appropriate magnitude scale is the 1-second period m_{bLg} scale of Nuttli (1973a).

The scale was derived from data of central United States earthquakes. A check on the applicability of the scale for earthquakes in northeastern North America (Street, 1976) showed its suitability, provided the epicentral distances are restricted to 20 degrees. Bollinger (1973) has also confirmed the applicability of Nuttli's scale to southeastern United States.

Magnitudes, by definition, are based on an instrumental measure of the ground motion of a given wave at a given period. In the case of Nuttli's scale, the measurements are made on the higher-mode surface wave Lg phase, which Herrman and Nuttli (1975) have found to carry, in most cases, the greatest energy levels. For design engineering purposes, periods other than the 1-second period are also of interest; it is thus necessary to relate the 1-second period ground motion to that of the other adjacent periods, the shorter ones in particular. This can be done if the approximate spectral shape of the Lg phase is known.

Street et al (1975), studying the spectral behavior of the Lg phase of 78 earthquakes in central United States, had found that the generalized spectral shape is characterized by ω^{o} and ω^{-2} trends, and occasionally by ω^{o}, ω^{-1} and ω^{-2}.

A later study by Street and Turcotte (1977), of 32 earthquakes in northeastern North America, containing much of the significant larger historical events since the instrumental era, e.g. La Malbaie, Grand Banks, Ossipee, Timiskaming, Cornwall, etc., has confirmed the validity of the ω^{o} and ω^{-2} model and found that the Lg spectra are remarkably well behaved in shape and level, as a function of magnitude. These conclusions are based on a comparative study of average observed spectra.

In their study, all the source spectra are scaled to a reference epicentral distance, r_o = 100 km, in order to observe the behavior of the 1.0-second period amplitude level and the corner frequency as a function of magnitude. For scaling the amplitudes, the following relationships were used:

$$S(\omega) = \begin{cases} 4\pi\rho\hat{\beta}^3 r_o (r/r_o)\,\Omega_r(\omega)\exp\ -\gamma(D-0.9) & \text{for } r \leq r_o \quad (1) \\ \\ 4\pi\rho\hat{\beta}^3 r_o (r/r_o)^{\frac{1}{2}}\,\Omega_r(\omega)\exp\ -\gamma(D-0.9) & \text{for } r \geq r_o \quad (2) \end{cases}$$

where ρ = 2.5 gm-cm^{-3}, $\hat{\beta}$ = 3.5 km-sec^{-1}, r_o = 100 km, r is the epicentral distance in km, D is the epicentral distance in degrees, and γ, the anelastic attenuation coefficient, is 0.11 deg^{-1} (Street, 1976). $\Omega_r(\omega)$ is the observed spectrum, and $S(\omega)$ is the source spectrum at r_o distance.

The reasons for selecting these relationships are given in the earlier study by Street et al (1975) on the Lg spectra of some central United States earthquakes, as well as the basis between $S(\omega=0)=M_o$.

In the present proposed approach to predict ground motion, it is only necessary to establish the spectral level at the 1-second period $(\omega=2\pi)$, and confirm that the spectral shape about the 1-second period is well behaved. It is not necessary to know the corner period specifically, but only where it lies with respect to the 1-second period. From Street and Turcotte (1977) data, it can be observed that for events with m_{bLg} ≈ 4.5 and greater, the corner period T_{02} (intersection of ω^o and ω^{-2}) is always greater than 1 second. Thus, if we limit the practical use of this approach to events with $m_{bLg} \geq 5.0$, i.e., those having any design significance, it will be possible to calculate the $S(2\pi)$ and assume safely that adjacent frequencies, particularly the higher ones, are located on ω^{-2} slope, and thus relatively well behaved.

From Table 1 of Street and Turcotte (1977), where observed m_{bLg} and corresponding $S(2\pi)$ are given for the data set, the following relation

$$S(2\pi) = 10^{(17.5 + m_{bLg})} \qquad (3)$$

can be derived. Such experimental relationships can be applied to the design earthquake magnitude.

Step 2 - Calculation of $\Omega_r(2\pi)$ Site Displacement

The site specific displacement spectral level $\Omega(2\pi)$ for a specified distance r can be readily obtained in terms of $S(2\pi)$:

$$\Omega_r(2\pi) = S(2\pi)\ g(r)^{-1} \tag{4}$$

where g(r) is

$$g(r) = 4\pi\rho\hat{\beta}^3 r_o(r/r_o)^c = 1.35 \times 10^{25}(r/r_o)^c \tag{5}$$

$$\begin{aligned} \text{and} \qquad & c = 1 && \text{for } r \leq r_o \\ & c = \tfrac{1}{2} && \text{for } r \geq r_o \\ & r_o = 100 \text{ km} \end{aligned}$$

in accordance with the constants given in Formulas (1) and (2).

The term exp $\left[\gamma(D-0.9)\right]$ of formulas (1) and (2) does not appear in Formula (5) because at distances within the range of interest to this study, it approaches one.

Step 3 - Prediction of Spectral Density Level $G_r(2\pi)$

Because acceleration is the second derivative of displacement, the site displacement spectral amplitude $\Omega_r(2\pi)$ can be simply converted to the acceleration amplitude spectrum $A_r(2\pi)$ by

$$A_r(2\pi) = \Omega_r(2\pi)\ (2\pi)^2 \tag{6}$$

At some stage in the analysis, it is necessary to convert from the vertical motion used in the m_{bLg} to horizontal strong motions. The conversion can be done at this point. Based on a set of 70 strong motion accelerograms (discussed in Appendix A), the mean of the ratio of the horizontal* $A(2\pi)$ to the vertical $A(2\pi)$ was 2.4. This value is comparable to typical reported average values of the ratio peak horizontal acceleration to the peak vertical acceleration. The 2.4 factor is also compatible with the findings of Street and Turcotte (1977, p. 609), based on a study of horizontal and vertical components of eastern United States seismograph records. In subsequent pages, $A(2\pi)$ refers to the horizontal component.

Next, the acceleration Fourier amplitude spectrum $A_r(\omega)$ can be related to the spectral density function of the ground motion. Specifically, the one-sided spectral density function $G_r(\omega)$ is proportional to the squared Fourier amplitude spectrum (Bendat, 1958)**:

$$G_r(\omega) = \frac{1}{2\pi} \frac{2}{s} \left| A_r(\omega) \right|^2 \tag{7}$$

*For each station/site case, there is one vertical recording and two horizontal recordings. One of the two horizontal records was randomly chosen to calculate the ratio of horizontal $A(2\pi)$ to vertical $A(2\pi)$.

**This relationship implies that G_r (ω) can be estimated by smoothing (or averaging) the squared Fourier acceleration amplitude spectrum.

in which s denotes strong-motion duration. $G_r(\omega)$ is the spectral density of the power (or energy per unit time) in the ground motion. If the total duration of the motion is used in Formula (7), it would imply that the "energy" in the motion is distributed uniformly over the entire duration. It is appropriate for what follows to select, as the value for s, a reduced time interval of strong motion. $G_r(\omega)$ then becomes the power spectral density averaged over this reduced time. Appendix B examines the influence of the choice of the duration s, and proposes a definition for the duration of strong motion which will be denoted by s_o (Vanmarcke and Lai, 1977).

Figure 1 illustrates the steps leading from the source spectrum $S(\omega)$ to the site spectral density function $G_r(\omega)$. The dotted lines in the plots labeled (c) and (d) in Figure 1 suggest different possible patterns of spectral amplitude decay caused by attenuation of high frequency components of ground motion.

Step 4 - Response Spectra and $S_V(2\pi, 0.1)$ Prediction

Response spectra are plots of the maximum response of a linear one-degree-of-freedom system as a function of the natural frequency ω_n for different values of damping ζ. Of particular interest are the response spectra for relative displacement (S_D), pseudo-velocity ($S_V = \omega_n S_D$), and pseudo-acceleration ($S_A = \omega_n^2 S_D$).

Our specific objective is to predict the response spectrum ordinate at the period $T_n = 1$ second ($\omega_n = 2\pi$) for a specified damping value, when the ground motion is characterized by a spectral density function $G_r(\omega)$ and the strong-motion duration $s = s_o$. For our purpose, it is sufficient to know $G_r(\omega)$ (and hence $A_r(\omega)$) in only a relatively narrow frequency band neighboring ω_n; here $\omega_n = 2\pi$.

The response spectra prediction can be accomplished by a nonstationary random vibration analysis (Vanmarcke, 1976) which is outlined in Appendix C. Briefly, the earthquake ground motion is represented as a segment of duration s_o of a stationary Gaussian random function with smooth spectral density function $G_r(\omega)$. The analysis assumes that the one-degree linear oscillator is at rest at the start of the ground motion, and predicts the buildup of the variance of the (pseudo-velocity) response from zero (at the start of the earthquake) to a maximum value σ_V^2 at time s_o.

The general solution expresses the median* response spectra $S_V(\omega_n, \beta)$ as a multiple of the response standard deviation, σ_V, as illustrated in Figure 2:

$$S_V(\omega_n,\beta) = \alpha \cdot \sigma_V \tag{8}$$

*The median response spectra corresponds to 50% probability of being exceeded. Other fractiles of the probability distribution of response spectra can also be obtained (see Appendix C).

where α is a dimensionless peak factor. Both α and σ_v depend on the one-degree system parameters ω_n and β and on the ground motion parameters. Only the specific results appropriate for moderate natural frequencies (including $\omega_n = 2\pi$) and for relatively large damping values (such as $\beta = 0.1$) are needed here:

$$\sigma_v = \left[\frac{\pi}{4\beta\omega_n} G_r(\omega_n) \right]^{\frac{1}{2}} \tag{9}$$

$$\alpha = \left[2\ln\left(\frac{2.8\,\omega_n s_o}{2\pi}\right)\right]^{\frac{1}{2}} \tag{10}$$

The choice of damping is not critical. Other damping values could be used; the results are available for arbitrary damping values. For very light damping, however, the peak response (in particular σ_v) becomes more sensitive to duration (see Appendix C and Appendix A, Part III).

Inserting expressions (9) and (10) into Formula (8), taking $\omega_n = 2\pi$ and $\beta = 0.1$, yields:

$$S_v(2\pi, 0.1) = \left[2.5\ln(2.8 s_o)\, G_r(2\pi)\right]^{\frac{1}{2}} \tag{11}$$

In terms of the one-second ($\omega = 2\pi$) acceleration amplitude, the result is:

$$S_v(2\pi, 0.1) = \left[\frac{2.5}{\pi} \frac{\ln(2.8 s_0)}{s_0} \right]^{\frac{1}{2}} A_r(2\pi) \qquad (12)$$

Finally, collecting all of the above results in terms of m_{bLg} and r, the median horizontal (pseudo-velocity) response spectrum (in units cm/sec) corresponding to $\omega_n = 2\pi$ and 10% damping is:

$$S_v(2\pi, 0.1) = \left[\frac{2.5}{\pi} \frac{\ln(2.8 s_0)}{s_0} \right]^{\frac{1}{2}} \frac{2.4 (2\pi)^2 10^{17.5}}{1.35 \times 10^{25}} 10^{m_{bLg}} \left(\frac{r}{r_0}\right)^{-c} \qquad (13)$$

in which c = 1 for $r \leq r_0$ = 100 km, and c = ½ for $r \geq r_0$.

Step 5 - Estimation of a_g From $S_V(2\pi, 0.1)$

Previous sections have shown how to predict, given magnitude m_{bLg} and distance r, the (median) peak response of a simple oscillator with natural frequency 2π and 10% critical damping, Equation (13). The nature of eastern United States instrumental data and magnitude scales has promoted our focus upon spectral ordinates at $\omega=2\pi$, but structural design demands response spectra at other frequencies and damping values as well. Nuclear power plant design requires relatively high frequency results; therefore, we use peak ground acceleration a_g as the parameter to scale standardized response spectra. A final step then will be to predict the peak ground acceleration from $S_V(2\pi, 0.1)$. The proposed procedure used is an empirical one, directly parallel to the current practice of estimating peak ground acceleration from the MM intensity (Trifunac and Brady, (1975). In both cases, because of a lack of appropriate eastern records, the prediction is based on a set of western United States strong ground motion accelerograms. Implicitly, both procedures assume that the relative frequency content of strong motions is the same in the east as in the west. The use of $S_V(2\pi, 0.1)$ is to be preferred to MM intensity, however, because in contrast to the latter, the former is an unambiguous, instrumental value. Also, $S_V(2\pi, 0.1)$ will be demonstrated numerically to be the preferred predictor of a_g (see Appendix A, Part III).

Using a set of strong motion data (described in more detail in Appendix A), the ratio $a_g/S_V(2\pi, 0.1)$ was studied. Letting

$$a_g = k\, S_V(2\pi, 0.1) \tag{14}$$

the best estimate of k was 10.46 with S_V in units cm/sec and a_g in units cm/sec^2.

Combining Equations (13) and (14) yields the following equation for the horizontal peak acceleration on rock, for r < 100 km:

$$a_g = 2.07 \times 10^{-5} \left[\frac{\ln(2.8 s_o)}{s_o} \right]^{\frac{1}{2}} 10^{m_{bLg}} \left(\frac{100}{r}\right) \tag{15}$$

Step 6 - Scaling of Standard Response Spectra

In parallel with current practice, this value of a_g can now be used with standard response spectra such as those in the NRC Regulatory Guide 1.60, or any others that may become available and possibly be more appropriate.

SPECIFIC APPLICATIONS

1. The proposed approach will now be applied to the selected design earthquake, i.e., a hypothetical repetition of the Giles County event. The event is given a magnitude m_{bLg} = 5.8 and assumed to occur near the sites (15 km). This magnitude and distance have been discussed with and suggested by NRC.

Two strong motion durations S_o have been used in Equation (15) to calculate the horizontal peak accelerations presented in Table 1. First, a more conservative value S_o = 2.51 sec is the mean duration of three Friuli aftershocks with an average Magnitude 6, recorded at the rock site of S. Rocco. The second value S_o = 4.85 sec is the mean duration of 22 strong motion recordings at 11 United States rock sites.

Table 1 shows the predicted maximal horizontal accelerations for the selected values of magnitude and distance and also for smaller and larger ones.

TABLE 1

PREDICTED ACCELERATIONS (g)

Distance	10 Km		15 Km		20 Km	
Strong Motion Duration (Sec)	2.51	4.85	2.51	4.85	2.51	4.85
m_{bLg}						
5.6	.07	.06	.05	.04	.04	.03
5.8	.12	.10	.08	.06	.06	.05
6.0	.18	.15	.12	.10	.09	.08

2. As a supporting example, let us consider the horizontal components of motion recorded on hard rock at the S. Rocco station during three aftershocks of the 1976 Friuli, Italy earthquake sequence. Relevant information about these records is summarized in Table 2.

TABLE 2

FRIULI AFTERSHOCKS

DATE M D Y	COMP.	MAG.	DIST. (Km)	DURATION S_o (Sec)	ACCELERATION (g) OBSERVED	PREDICTED
9 11 76 16h 21GMT	NS EW	5.9	15.7	1.30 2.68	0.09 0.085	0.105 0.091
9 15 76 3h 15GMT	NS EW	6.1	12.7	5.96 2.10	0.061 0.119	0.141 0.188
9 15 76 9h 21GMT	NS EW	6.0	23.2	2.43 0.59	0.137 0.242	0.08 0.082
Mean		6.0	17.2	2.51	0.122	0.115
Std. Dev.					0.06	0.04

Table 2 includes a tabulation of strong-motion durations estimated on the basis of the procedure described in Appendix B. Assuming that the Magnitude M assigned to the aftershocks by the Italian seismologists is just as characteristic of the relative size of the events as the m_{bLg} scale, and using in Equation (15), the tabulated values of M, r and S_o, we arrive at a very interesting comparison of the predicted with the observed accelerations, as shown in the last two columns of Table 2.

The means of the observed and predicted maximum accelerations are 0.122g and 0.115g, respectively, and the corresponding standard deviations are 0.06g and 0.04g.

An alternative approach is to calculate the mean magnitude (M=6), the mean distance (17.2 km) and the mean strong-motion duration (S_o = 2.51 sec) for the six records. Inserting these values into Equation (15) gives the prediction a_g = 0.120g, which is practically identical to the mean of the observed maximum accelerations (ag = 0.122g).

CONCLUSION

On the basis of the present method of predicting the peak acceleration for a design earthquake, a value a_g = 0.08g was calculated for a magnitude m_{bLg} = 5.8 at 15 km. If this acceleration value is used to anchor the NRC Regulatory Guide 1.60, all the design spectra for Sequoyah, Watts Bar and Bellefonte are well above their required design. In view of the excellent agreement between the accelerations predicted by the method applied to some Friuli aftershocks and the observed accelerations from the same events, it is concluded that the selected design bases for Sequoyah, Watts Bar and Bellefonte are conservative.

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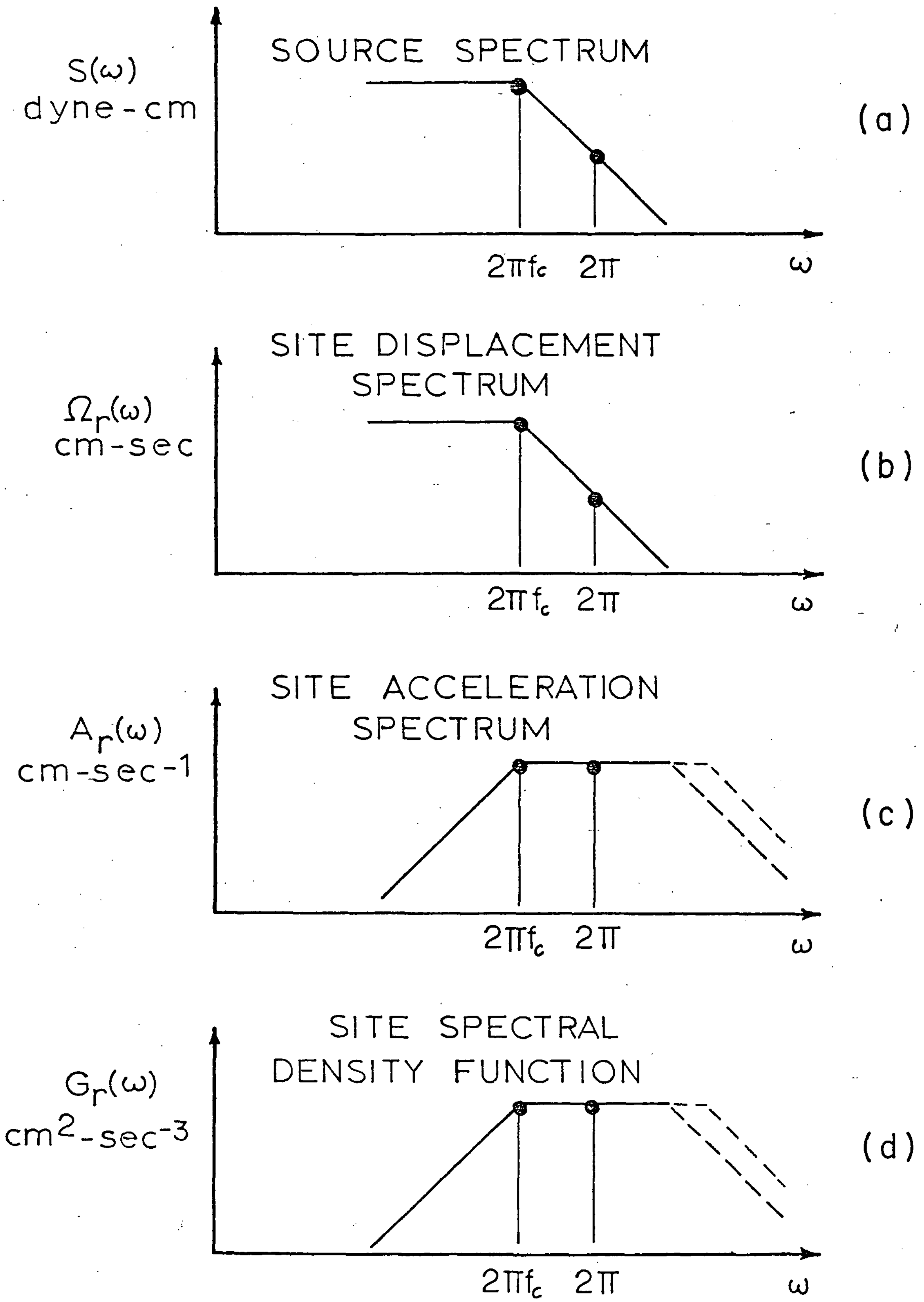
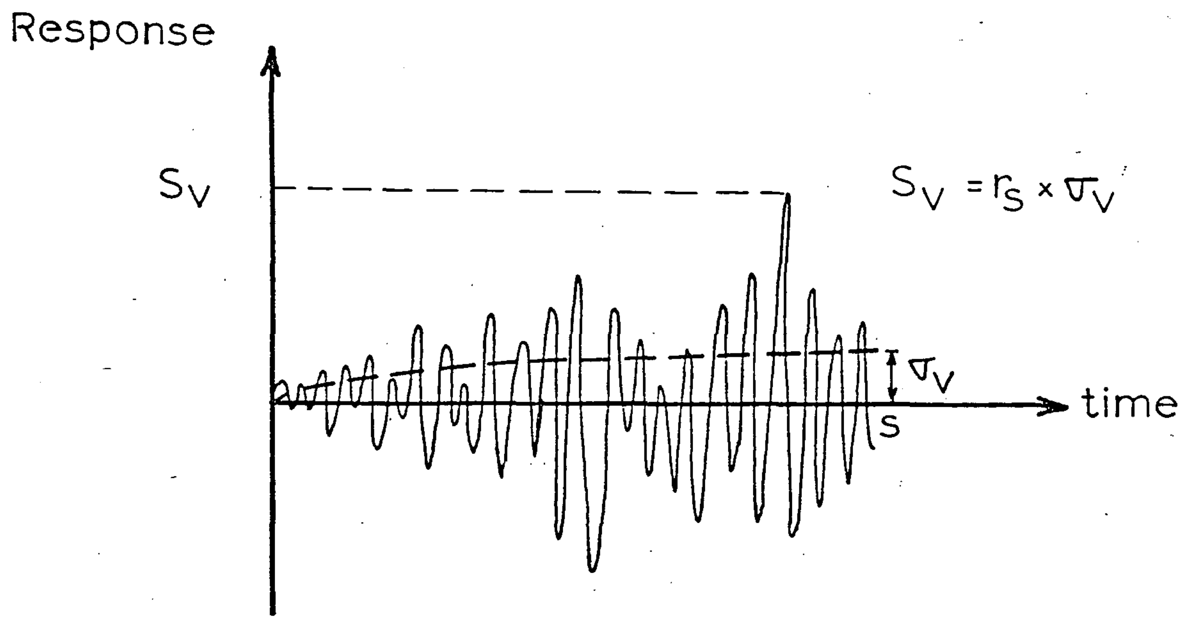

SOURCE SPECTRUM
S(ω)
dyne-cm
2πfc
2π
ω
(a)
SITE DISPLACEMENT
SPECTRUM
Ωr(ω)
cm-sec
2πfc
2π
ω
(b)
SITE ACCELERATION
SPECTRUM
Ar(ω)
cm-sec-1
2πfc
2π
ω
(c)
SITE SPECTRAL
DENSITY FUNCTION
Gr(ω)
cm2-sec-3
2πfc
2π
ω
(d)

Figure 1

ONE-DEGREE SYSTEM RESPONSE TO

EARTHQUAKE GROUND MOTION



S_V = absolute maximum response
σ_V = standard deviation of the response
r_s = peak factor

Figure 2

APPENDIX A

APPENDIX A

ANALYSIS OF DISPERSION

In this appendix, we study the dispersion associated with the several steps in the proposed analysis. In addition, we compare this dispersion with that implicit in the parallel steps in current ground motion prediction procedures.

Specifically, we compare the proposed method with the current procedure in use in the eastern United States, in which (1) an MM intensity is predicted at the site (given the epicentral MM intensity and the epicentral distance, r); and (2) the value of a_g is predicted from MM intensity at the site. Note that, in the proposed procedure, $S_V(2\pi, 0.1)$ plays a role analogous to that of the MM intensity at the site. Therefore, we want to compare how well these two variables can be predicted from a given event size and distance, and how well each variable predicts a_g. These comparisons are, of course, only part of the story; the proposed method has all the advantages of avoiding the use of MM intensity (both epicentral intensity as a measure of the event size and site intensity as a measure of motion intensity) because it replaces intensity by instrumentally-defined variables.

Note that the proposed procedure utilizes the same philosophy with respect to explicit incorporation of dispersion as is currently done. Specifically, in both procedures, median prediction values are utilized at all steps (except in the common last step; i.e., utilization of the NRC Regulatory Guide 1.60 standard spectra). We wish to show that the proposed procedure provides predictions of comparable dispersion and therefore can be used in the same design context with the same confidence.

PART I: Dispersion in A(2π) Given m_{bLg} and r

The nature of the problem makes the direct assessment of the dispersion in A(2π) difficult; the low seismicity of the eastern United States implies that we have very few records of events of engineering interest ($m_{bLg} > 4.5$ to 5), especially at closer distances. It is anticipated, however, that there should be less dispersion in estimating the A(2π) amplitude of the Lg wave that would be measured by a typical short-period seismometer at the distances cited in this proposal, than the dispersion normally involved in estimating strong-motion peak accelerations in the west. This conclusion follows because measurements are made in the far field and because seismograph foundation conditions are more uniform. The lower dispersion is confirmed by the comparatively low station-to-station variability in the data from a single event, when all stations are used to estimate event magnitude by a typical magnitude versus maximum sustained amplitude and distance formula. It is further confirmed by the comparatively low dispersion in the empirically fit constants of those formulas; the fitting being done over several events. Seismologists commonly report magnitude estimate bounds of $\pm\frac{1}{4}$ magnitude units; inverted to give the bounds in amplitude from a given magnitude (at a given distance), this implies a factor of 10 to a power $\pm 1/4$, i.e. multiply or divide by a factor 1.8. Assuming these bounds represent about two standard deviations, we might estimate, therefore, a standard deviation of about ± 0.12 log base 10 units or about 0.3 log base e units in the Fourier amplitude $A(2\pi)$. This implies a coefficient of variation in $A(2\pi)$ of about 30 percent (see below).

Another approach to estimate the approximate level of the dispersion in $A(2\pi)$ has also been used. A set of 32 eastern events of m_{bLg} magnitude greater than 3.0 (up to 4.5) was available, together with one or more short-period seismometer traces for each. Seismic moment estimates (Mo) and corner period estimates (T_{o2}) were available from Fourier amplitude displacement spectra of these traces. The level of the ω^{-2} fall-off portion of each spectrum, which, as discussed in Step 1, is the portion of greater interest in the case of larger events of engineering interest (where T_{o2} is about 1-second or more), is proportional to the moment divided by the square of the corner period. These observed "ω^{-2} fall-off levels" were compared to predicted values. The predicted values were obtained by using the composite estimate of M_o for each event and a predicted corner period for each event (the prediction is that provided by Figure 6 of Street et al (1975), which plots M_o versus T_{o2} for eastern United States events). There were a total of 95 cases in which M_o and the corner period were observable; the standard deviation of the ratio of observed-to-predicted values (of the level of the ω^{-2} fall-off) was 0.58 (the mean was 1.43, implying a coefficient of variation of 41%).

A final estimate of the dispersion associated with predicting $A(2\pi)$, given magnitude and distance, was obtained by using western strong-motion accelerometer data. A data set of 70 event/site pairs (140 horizontal, 70 vertical traces) was utilized; it is the same set selected by McGuire (1977a) and McGuire and Barnhard (1977) to avoid bias due to unusually large numbers of data points from one event (e.g., San Fernando, 1971). Lacking a full set of m_{bLg} values, M_L values were used instead (there is some justification for saying they both "sample" roughly comparable frequency ranges in ground motion traces for magnitudes of roughly six and less). The value of $A(2\pi)$ used was the numerical average of the calculated $A(\omega)$ values over the range $\omega = 0.8\pi$ to $\omega = 4\pi$ radians (0.4 to 2 cps), as the latter were reported in Volume IV of the California Institute of Technology processed records (see California Institute of Technology, 1972).

With this data, a regression of the same form commonly used for attenuation of instrumental ground motion intensity (such as peak ground acceleration) was used:

$$\ell n\, A(2\pi) = C_1 + C_2 M_L - C_3 \ell n(r+25) \tag{A1}$$

in which r is focal distance in kilometers. The dispersion in ln A(2π) was found to be 0.67, implying a coefficient of variation* of A(2π) of about 0.75. The results presented here are for horizontal A(2π).

We conclude then by proposing that the results above suggest that the coefficient of variation in predicting A(2π) (vertical or horizontal**) given m_{bLg} and r, is probably about 0.7 or less. This conclusion assumes that there is no bias in the equation, i.e., that it makes an unbiased (mean or median value) prediction of A(2π) over the distance and magnitude range of interest.

In passing, we observe that, by using regression analysis on the same data, we can predict a_g with a standard deviation of ln a_g of 0.57 or a coefficient of variation of only 0.62.

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*This value is based on coefficient of variation = $\sqrt{\exp\ \sigma^2\ \ -1}$, in which σ is the standard deviation (dispersion) of the natural log. This equation assumes an approximately normal distribution of residuals of ln A(2π); this form is indeed typically observed.

**The coefficient of variation (c.o.v.) of the ratio of horizontal A(2π) to vertical A(2π), mentioned in Step 3, was found to be 0.37. This is a relatively small portion of the total dispersion in the horizontal A(2π) because these dispersion measures (c.o.v.'s) combine approximately using an SRSS rule. So if, for example, the coefficient of variation in predicting vertical A(2π) is 0.65 and the coefficient of variation of the ratio is 0.37, the coefficient of variation of the horizontal A(2π) is (assuming independence) approximately 0.75; that is, both vertical and horizontal A(2π) have about the same dispersion, 0.7, to one significant figure, which is as close as this estimate is valid. Therefore, in subsequent discussion, we shall not distinguish between the dispersions in vertical and horizontal A(2π) values.

This latter value is typical of the values of 0.6 to 0.7 reported by various investigators who have made similar statistical studies of attenuation of a_g. Unfortunately, such data is lacking for the eastern United States, prompting this study. We shall return to such comparisons at the end of this appendix.

PART II: Confirmation of and Dispersion in the Prediction of $S_V(2\pi, 0.1)$ from $A(2\pi)$

In Step 4 of this report, the median value of $S_V(2\pi, 0.1)$ is predicted to be (Equation (12))

$$S_V(2\pi, 0.1) = \left[\frac{2.5}{\pi} \; \frac{\ell n(2.8 s_o)}{s_o} \right]^{\frac{1}{2}} A(2\pi) \qquad \text{(A2)}$$

This result is quite insensitive to strong motion duration s_o. For s_o = 20, 10, and 5 seconds, the median value of $S_V(2\pi, 0.1)$ is predicted to be 0.4, 0.52, and 0.65 times $A(2\pi)$, respectively. The result is equally valid for either vertical or horizontal motions (provided both S_V and A are in the same direction).

The 140 horizontal strong motion records cited previously were used to study the statistics of the ratio $S_V(2\pi, 0.1)/A(2\pi)$. The mean* of the ratio was found to be 0.50; its coefficient of variation was only 0.30. This mean value is consistant with a separate study which found that the average "equivalent duration" of records in the data set was 9.3 seconds (Vanmarcke and Lai, 1977). The theory used in Step 4 apparently works very well to predict $S_V(2\pi, 0.1)$ from $A(2\pi)$.

In addition, the dispersion (c.o.v.) in the prediction is very low (0.30) compared to other values we encounter (e.g., perhaps as high as 0.7 for prediction of $A(2\pi)$). One

*For this low (30%), a coefficient of variation (c.o.v.), the mean and median differ by only a few percent, the mean being $\left[1 + (\text{c.o.v.})^2 \right]^{\frac{1}{2}}$ times the median, assuming a log-normal distribution.

important implication is that we can effectively treat $S_V(2\pi, 0.1)$ and $A(2\pi)$ as deterministically interchangable variables. Therefore, any conclusions reached about $A(2\pi)$ will apply to $S_V(2\pi, 0.1)$ and vice versa. For example, the western data cited above led to a c.o.v. of 0.75 in predicting $A(2\pi)$ from M_L and r; based on the findings here, we should expect virtually the same c.o.v. in predicting $S_V(2\pi, 0.1)$ from M_L and r, and indeed even approximately the same dependence on M_L and r. This conclusion has been confirmed by direct regression of ln $S_V(2\pi, 0.1)$ on M_L and r, where the c.o.v. is 0.79.

PART III: <u>Dispersion in Predicting</u> a_g <u>from</u> $S_V(2\pi, 0.1)$

The basic set of 140 strong motion horizontal records yielded a coefficient of variation of the ratio $k = a_g/S_V(2\pi, 0.1)$ of 0.72. This value is to be compared directly with the dispersion implied in currently used empirical predictions of a_g from MM Intensity (I). A direct comparison is available using the same data set discussed above*. Using the typical regression form

$$\ell n\, a_g = b_o + b_1 I \tag{A3}$$

the standard deviation of ln a_g was found to be 0.76, implying a coefficient of variation of a_g of 0.88. This value is consistent with values found by other investigators (e.g., Trifunac and Brady, 1975a) and is substantially larger than the value of 0.72 associated with using $S_V(2\pi, 0.1)$ to predict a_g. Although not as yet commonly used, it has been observed by several investigators recently (e.g., McGuire, 1977a, and Computer Science Corporation, 1977) that the relationship between a_g and MM intensity may be significantly dependent on focal (or epicentral) distance. The physical argument supporting the observation is based on the

*As discussed by McGuire and Barnhard (1977), the MM intensity values used here were observed values near the instruments, not values read from MM intensity contour maps, which may be biased by the contouring process.

relatively faster attenuation of higher frequency waves (as represented by a_g) than moderate frequency waves (as represented roughly by I). A similar phenomenon is observed in the data in the relationship between a_g and $S_v(2\pi, 0.1)$. In this case, $S_v(2\pi, 0.1)$ is, without doubt, an (instrumental) measure of the intensity of moderate frequency waves. The regression

$$\ln a_g = C_0 + C_1 \ln(r + 25) + C_2 S_v(2\pi, 0.1) \tag{A4}$$

yields $C_o = 6.205$, $C_1 = -0.776$, and $C_2 = 0.624$ with a standard deviation of ln a_g of only 0.43 (or a c.o.v. of 0.45). Accounting for the r-dependence of the relationship substantially reduces the dispersion in the prediction. For levels of values of a_g of about 200 cm/sec^2, this regression will give <u>lower</u> values of a_g for a given $S_v(2\pi, 0.1)$ value at distances in excess of about 10 km than will the simple ratio $a_g/S_v(2\pi, 0.1) = 10.46$; therefore, it is conservative to use the ratio in the range of interest here.

It has also been conjectured that the relationship between higher frequency intensity measures and moderate frequency intensity measures (e.g., a_g and I, or a_g and $S_v(2\pi, 0.1)$) may depend on the soil conditions at the site. In the 70 cases, there were 11 which McGuire (1977b) has classified as "rock" sites. The average ratio of $a_g/S_v(2\pi, 0.1)$ was 15.61 for these 11 cases (with a coefficient of variation of 0.70). This is somewhat higher than the ratios for all sites (10.46) and for the 59 soil sites only (9.50), but, due to the small rock site sample size, the standard error of the difference between the rock and soil mean ratios is high (3.4), implying that the difference 15.61-9.50 = 6.1 is not statistically significant at the 5% significance level.

As mentioned in Step 4, the use of 10% damping is relatively arbitrary; other values could be used. Relatively light dampings are to be avoided, however, as they are less sensitive to the peak ground acceleration. For example, the ratio of a_g to $S_v(2\pi, 0)$ has a c.o.v. of 100%.

PART IV: Total or Coupled Uncertainty in a_g.

In this section, we attempt to compare the total uncertainty faced in predicting a_g when the size and distance of the design event have been specified. Recall that for western data a direct regression of ln a_g on M_L and r is possible (and is typically used in western U.S. applications). The standard deviation of ln a_g was found (see Part I) to be 0.57 for this data set.

In the east we lack such data. The current practice is to estimate the MM intensity at the site from the epicentral intensity I_o and the distance r, using attenuation laws developed from observed isoseismals in the east. Here, we continue to use our data set of 70 western cases in order to insure internal consistency in the dispersion calculations. An attenuation law of the typical form

$$I = C_0 + C_1 I_0 - C_2 \ln(r+25) \tag{A5}$$

yielded a standard deviation on the site MM Intensity (I) of 0.66. This dispersion would be even higher if C_1 were constrained to be unity, as many investigations assume. As discussed above, the regression of ln a_g on I yielded a standard deviation on ln a_g of 0.76, with the regression

$$\ln a_g = 0.251 + 0.631 \tag{A6}$$

Substituting one concludes that the total (net) dispersion in ln a_g (predicted in this way) is*

$$\sqrt{(0.63)^2(0.66)^2+(0.76)^2} = \sqrt{0.17+0.58} = \sqrt{0.75} = 0.87 \qquad \text{(A7)}$$

or a coefficient of variation of 1.06. Note that the major contribution to the uncertainty comes from predicting a_g given I. Therefore, even if r is zero, i.e., even if I_o is "migrated to the site", the standard deviation of ln a_g will be 0.76 (and its coefficient of variation 0.89).

Under the proposed method, we first predict vertical $A(2\pi)$ from m_{bLg} and r, then horizontal $A(2\pi)$ from vertical, then $S_V(2\pi, 0.1)$ from $A(2\pi)$, and finally a_g from $S_V(2\pi, 0.1)$. As discussed above, we cannot provide a strong data base to support the estimate of the dispersion in vertical or horizontal $A(2\pi)$, but we estimate it to be 0.7 or less. The dispersion (c.o.v.) in the ratio of $S_V(2\pi, 0.1)$ to $A(2\pi)$ was 0.30. The c.o.v. of the ratio of a_g to $S_V(2\pi, 0.1)$ was

*There should, in general, be a correlation term under the radical equal to $\rho(0.63)(0.66)(0.76)$, in which ρ is the correlation between the residuals. This correlation was found to be negative (-0.24) leading to a term -0.15 under the radical, reducing the standard deviation of ln a_g to 0.77 and the c.o.v. to 0.91. Keeping this term will not alter the conclusions. All such correlations were not available in later calculations.

found to be 0.72. Neglecting correlations, these c.o.v.'s combine approximately* by the SRSS rule:

$$\sqrt{(0.7)^2+(0.30)^2+(0.72)^2} =$$
$$\sqrt{0.49+0.09+0.52} = \sqrt{1.10} = 1.05 \qquad \text{(A8)}$$

Note that, if the first c.o.v. is indeed as high as 0.7, two terms contribute approximately equally to the total dispersion in a_g, namely predicting $A(2\pi)$ from M and r, and predicting a_g from $S_v(2\pi, 0.1)$.

We conclude that the coefficients of variation in predicting a_g via I_o, r, and I and in predicting a_g by the proposed method are, at worst, about equal, i.e., about 100%. As discussed, there is reason to believe our

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*Assuming independence, there are, strictly speaking, additional terms of the form

$$V_1^2\ V_2^2 + V_2^2\ V_3^2 + V_3^2\ V_1^2 + V_1^2\ V_2^2\ V_3^2$$

in which the V's are the 3 c.o.v.'s used; i.e., 0.7, 0.3, and 0.72. Inclusion of these terms increases the c.o.v. to 1.20, but based on similar calculations, it is strongly believed that negative correlations exist among these residuals, implying (as seen in the previous footnote) a <u>reduction</u> of the term within the radical and in the final c.o.v. It is presumed that these two effects approximately cancel one another out. Therefore, we use 1.05 as our estimate of the net uncertainty.

use of 0.7 for the c.o.v. of $A(2\pi)$ given M and r is conservative. There are, of course, non-quantifiable advantages to using instrumental quantities such as m_{bLg}, $A(2\pi)$, and $S_v(2\pi, 0.1)$ in place of MM intensity (epicentral I_o, and site I) to define design motions.

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APPENDIX B

APPENDIX B

STRONG-MOTION DURATION

A key ground motion parameter in the relationship between the spectral density function $G(\omega)$ and the response spectrum is the strong-motion duration s_o. No single quantitative measure of strong-motion duration is in common usage in earthquake engineering. Two crude but simple measures of duration have been mentioned in the literature. The first defines duration as the time interval between the first and last peaks equal to or greater than a given level, usually 0.05g, on the accelerogram (Page et al, 1975). The second definition is based on cumulative energy obtained by integrating over time the squared accelerations in an accelerogram; duration is the time interval required to accumulate a prescribed fraction of the total integral (i.e., the so-called Arias Intensity), for example, 95 percent (Husid et al, 1969) or 90 percent (Trifunac and Brady, 1975b) of the total energy.

This appendix presents briefly the definition and the estimation procedure for the strong-motion duration of earthquakes proposed by Vanmarcke and Lai (1977). This definition is most useful in the context of the methodology proposed here. It has been applied to the horizontal components of each of 70 strong-motion records (see Appendix A).

The records in this sample represent a broad range of seismic events, magnitudes, source-to-site distances, and site conditions ("soil" versus "rock").

Relationship Between $G(\omega)$ and $A(\omega)$

The spectral density function (s.d.f.) of a recorded ground motion a(t), can be estimated from the Fourier amplitude spectrum $A(\omega)$, as follows (see Equation 7):

$$G(\omega) = \frac{1}{\pi}\frac{1}{s_o}A^2(\omega) \qquad \text{(B1)}$$

in which s_o is the (yet to be defined) strong-motion duration. $A(\omega)$ is the absolute value (modulus) of the Fourier transform of a(t) (see Hudson, 1962):

$$A(\omega) = \left|\int_{-\infty}^{\infty} a(t)e^{-i\omega t}dt\right| = \left|\int_{o}^{t_o} a(t)e^{-i\omega t}dt\right| \qquad \text{(B2)}$$

in which ω = frequency in radians/sec, $i = \sqrt{-1}$, and t_o = duration of the digitized accelerogram. The squared modulus of the Fourier spectrum $A^2(\omega)$ indicates how the total energy in the earthquake motion is distributed over the frequency axis ($-\infty < \omega < \infty$). The integral of $A^2(\omega)$ over all frequencies is directly related to E_o, the Arias Intensity (Arias, 1969), as follows:

$$E_o = \int_o^{t_o} a^2(t)dt = \int_o^{\infty} a^2(t)dt = \frac{1}{\pi}\int_o^{\infty} A^2(\omega)d\omega \qquad \text{(B3)}$$

The equality on the right side of Equation (B3), is Parseval's theorem. The precise location of sharp peaks and valleys of the Fourier amplitude spectrum is sensitive to such factors as the choice of digitization intervals $\Delta\omega$ and Δt used in record processing. Consequently, little information of engineering significance is lost by averaging out the erratic fluctuations of the functions $A(\omega)$ or $A^2(\omega)$. By the same token, while the estimated s.d.f. may fluctuate considerably, the true s.d.f. is a well-behaved statistical quantity which describes the expected distribution of the earthquake power frequency. Equation (B1) implies an idealization of the earthquake as a segment of limited duration (s_o) of a stationary random process with spectral density function $G(\omega)$.

Recall now that the mean square acceleration σ_o^2 may be obtained by integrating $G(\omega)$ over all frequencies. Inserting Equations (B1) and (B2) into this relationship yields

$$E_o = s_o \sigma_o^2 \tag{B4}$$

In words, for such an idealized motion, the total Intensity E_o is distributed uniformly, at constant average power σ_o^2, over the strong-motion interval s_o.

The question at hand is how to determine, quantitatively, the strong-motion duration s_o and the corresponding strong-motion r.m.s. acceleration σ_o. Equation (B4) suggests that large values of s_o imply small r.m.s. values σ_o. For example, when s_o equals t_o, we obtain $\sigma_o = (I_o/t_o)^{\frac{1}{2}}$. For very small values of s_o, the r.m.s. acceleration can become as large as the peak acceleration a_g, which happens when $s_o = E_o/a_g^2$. These extreme choices for s_o are undesirable because they imply values of σ_o which are not related in a consistent manner to the peak acceleration a_g. For a typical accelerogram, the calculated peak factor a_g/σ_o may be as high as 8 or as low as 1, depending on which value of s_o is chosen. This observation is the key to the method proposed by Vanmarcke and Lai (1977) for evaluating strong-motion durations of earthquake records. The idea is that an approximate relationship must exist between σ_o and a_g. The relationship is, of course, of a probabilistic nature. The theory of stationary Gaussian random functions provides a prediction of the most probable value of the peak factor a_g/σ_o during a known time interval s_o of steady strong shaking. Specifically, the value of a_g/σ_o, which is exceeded <u>once on the average</u> during the interval s_o (or which has a probability e^{-1} of not being exceeded during

s_o), is approximately:†

$$(a_g/\sigma_o) = \sqrt{2\ell n(2s_o/T_o)} \qquad s_o \geq T_o \tag{B5}$$

in which $T_o = 2\pi/\omega_o$ = "central" period of the earthquake motion, and s_o/T_o = number of "zero crossings" with positive slope during the time interval s_o. The lower limit on s_o ensures that the peak factor will not be too low;† the peak factor predicted by Equation (B5) is relatively insensitive to the choice of T_o.

Assuming T_o is known, Equations (B4) and (B5) can be viewed as a system of the two equations and two unknowns, s_o and σ_o. The solution for s_o is implicit in the following equation:

$$s_o = \left[2\ell n(2s_o/T_o)\right](E_o/a_g^2) \qquad s_o \geq T_o \tag{B6}$$

The ratio (E_o/a_g^2) is available for all strong-motion earthquake records in the set described above.

†Equation (B5) is derived on the basis on the assumption that the crossings of a specified, relatively high threshold occur as a Poisson arrival process. The formula is inappropriate for very low thresholds. The condition $a_g/\sigma_o \leq 1.2$ inplies $s_o \geq T_o$, which is the bound set in Equation (B5).

The solution to Equation (B6) is plotted in Figure B-1. Note that the duration s_o is nearly a linear function of E_o/a_g^2 for a given value of T_o. For example, for the north-south component of the 1940 El Centro record, the ratio E_o/a_g^2 is 0.975, and T_o is about 0.3 sec. From Figure B-1, it follows that s_o is about 7.5 sec. From Equation (B5), the corresponding peak factor is $r=2.54$, and the r.m.s. acceleration is $\sigma_o=a_g/r = 0.32g/2.54 = 0.13g$.

The main features of this definition of the strong-motion duration s_o are that (1) the total motion energy E_o is preserved, and (2) a consistent relationship between a_g and the strong-motion r.m.s. acceleration σ_o is guaranteed. The proposed definition also provides a link between the earthquake ground motion measures E_o, a_g and σ_o. Finally, the strong-motion duration s_o plays an all-important role in the relationship among the functions $A(\omega)$, $G(\omega)$, and the response spectrum (see Appendix C).

Using the procedures just outlined, strong-motion durations were obtained for the horizontal components of three aftershocks of the 1976 Friuli, Italy earthquakes recorded at the S. Rocco station, a hard rock site (Muzzi and Pugliese, 1977). The E-W component of the motion recorded on September 11, 1976, is shown in Figure B-2 together with the calculated strong-motion duration S_o = 2.7 sec.

The Arias intensity is $E_o = 2734\ cm^2/sec^3$, the r.m.s. strong-motion acceleration = $\sigma_o = 0.034g$, and the peak factor $r = 2.53$. The recorded maximum accelerations, magnitudes, distances, and strong-motion durations of six horizontal components of motion at S. Rocco are presented in Table 2.

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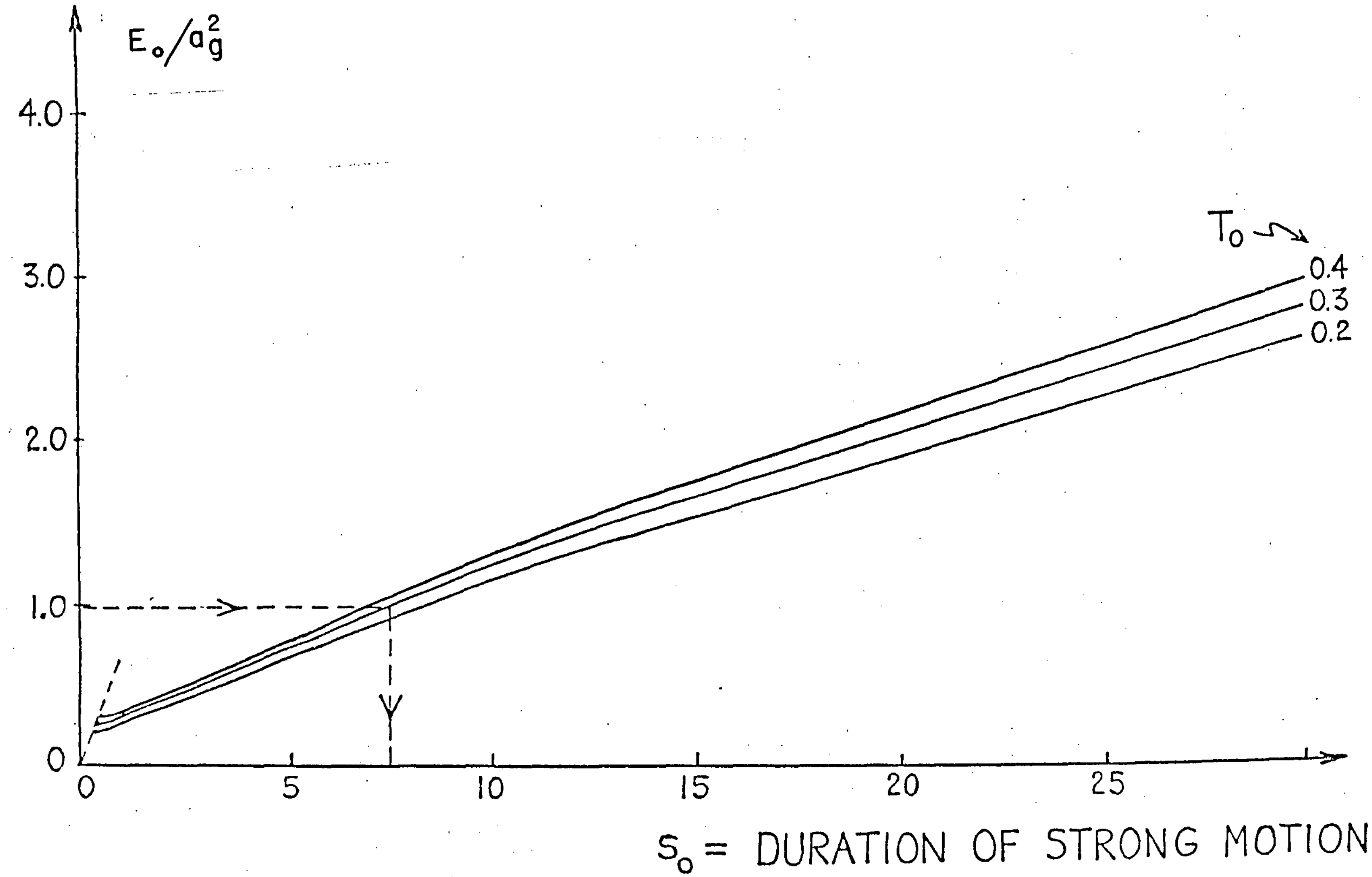
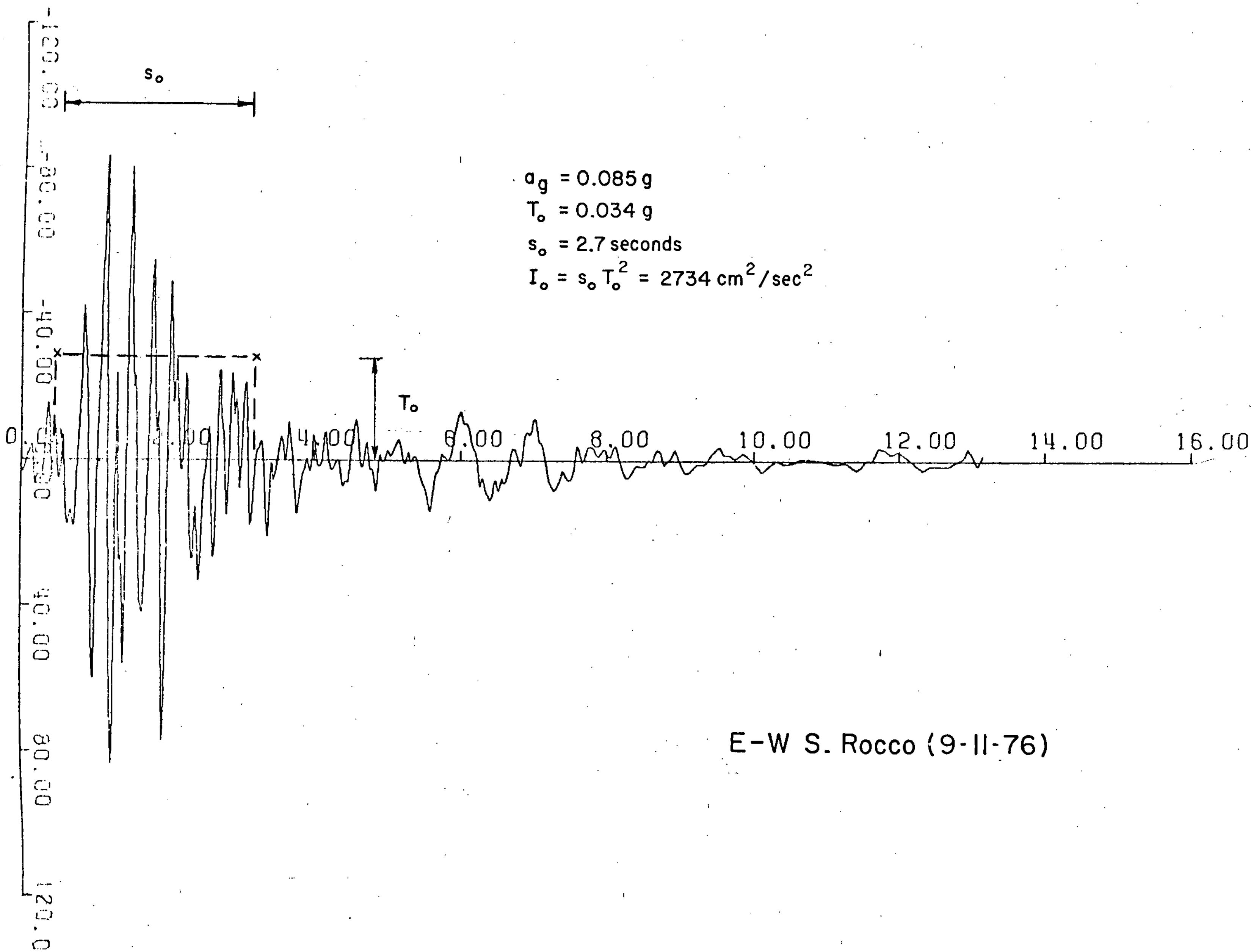

E_o/a_g^2
4.0
3.0
2.0
1.0
0
T_0
0.4
0.3
0.2
0
5
10
15
20
25
S_o = DURATION OF STRONG MOTION

Fig. B-1



S₀
$a_g = 0.085\,g$
$T_o = 0.034\,g$
$s_o = 2.7$ seconds
$I_o = s_o T_o^2 = 2734\ cm^2/sec^2$
T_o
E-W S. Rocco (9-11-76)
0
4.00
6.00
8.00
10.00
12.00
14.00
16.00
-120.00
-80.00
-40.00
40.00
80.00
120.0

Fig. B-2

APPENDIX C

APPENDIX C

RELATIONSHIP BETWEEN SPECTRAL DENSITY FUNCTION AND RESPONSE SPECTRA

This section summarizes the results of a random vibration analysis of linear one-degree systems with different frequencies ω_n (from 0 to ∞) and damping ratios β (including $\beta = 0$). The emphasis is on the major concepts and practical results. A more detailed treatment may be found elsewhere (Vanmarcke, 1976). The results given below permit the derivation of response spectra from a specified s.d.f. $G(\omega)$ and vice versa.

The earthquake is modeled as a suddenly applied "stationary" acceleration characterized by the s.d.f. $G(\omega)$ and the strong motion duration s_o. The standard deviation and the "central" frequency of the input acceleration are σ_o and ω_o, respectively. The aim is to predict the pseudo-acceleration† response spectra $S_A(\omega_n, \beta)$ corresponding to a given probability of no exceedance, p. The solution takes the following general form:

$$S_A(\omega_n, \beta) = \alpha_p \times \sigma_A(s_o) \tag{C1}$$

†The pseudo-acceleration response is, by definition, ω^2 times the relative displacement response and ω_n times the pseudo-velocity response.

in which $\sigma_A(s_o)$ = time-dependent standard deviation of the pseudo-acceleration response, evaluated at $t = s_o$; and α_p = dimensionless peak factor.

The standard deviation $\sigma_A(t)$ builds up from zero when the earthquake strikes to its maximum value $\sigma_A(s_o)$. This maximum value need not necessarily be close to the steady-state value which would be reached if the excitation were to continue and provided the system had some damping. The variance σ_A^2 (t) is obtained by integrating the time-dependent spectral density function of the response, $G_A(\omega, t)$, over all frequencies with the following approximate result:

$$\sigma_A^2(t) = \int_0^\infty G_A(\omega,t)d\omega \simeq \int_0^\infty G(\omega)\,\omega_n^4\left[(\omega^2-\omega_n^2)+4\beta_t^2\omega_n^2\omega^2\right]^{-1}d\omega$$

$$\simeq G(\omega_n)\,\omega_n\left[\frac{\pi}{4\beta_t}-1\right]+\int_0^{\omega_n}G(\omega)d\omega \qquad \text{(C2)}$$

Note that the dependence on time enters solely through the quantity β_t which may be interpreted as a "fictitious" damping parameter. It measures the bandwidth of the one-degree response after t seconds of exposure to the earthquake, and it depends on the true system damping β and on the product $\omega_n t$, as follows:

$$\beta_t = \beta(1-e^{-2\beta\omega_n t})^{-1} \qquad \text{(C3)}$$

Note that β_t approaches β when the product $\beta\omega_n t$ is large. This is when the stationary response condition develops. When β exceeds 0.05, it is usually appropriate to replace β_t by β for systems with low or moderate natural periods. For undamped systems, β_t becomes $(2\omega_n t)^{-1}$, and the stationary condition is never reached. Also, when the damping is not too large, the factor $\left(\frac{\pi}{4\beta_t} - 1\right)$ may be replaced by $\pi/4\beta_t$ in Equation (C3).

It is convenient to express the result for the response standard deviation $\sigma_A(s)$ in terms of the ground motion standard deviation σ_o, as a ratio:

$$\frac{\sigma_A(s_o)}{\sigma_o} = \left\{ G^*(\omega_n)\, \omega_n \left[\frac{\pi}{4\beta_{s_o}} - 1 \right] + F^*(\omega_n) \right\}^{\frac{1}{2}} \tag{C4}$$

in which $G^*(\omega) = \sigma_o^{-2}\, G(\omega)$, i.e., a normalized (unit-area) s.d.f., and $F^*(\omega_n) = \int_o^{\omega_n} G^*(\omega)\, d\omega$, a normalized cumulative earthquake spectrum. Furthermore, to facilitate the evaluation of the peak factor α_p in Equation (C1), it is useful to define the fraction q as the ratio of the contribution $\sigma_o^2\, F^*(\omega_n)$ (the second term in Equation (C2)) to the response variance $\sigma_A^2(s_o)$:

$$q = \frac{\sigma_o^2\, F^*(\omega_n)}{\sigma_A^2\,(s_o)} \tag{C5}$$

$q \simeq 0$ for low and moderate natural frequencies (including $\omega_n = 2\pi$) and $q \simeq 1$ at very high values of ω_n.

The peak factor α_p, although less important than the standard deviation, also plays a significant role in response spectra prediction. It depends on the damping as well as on the natural frequency of the system. While exact solutions for α_p do not exist, an approximate solution is available which has been extensively checked against simulation results and peak factors inferred from the response spectra of real earthquakes (Vanmarcke, 1975; 1976). The solution takes the form of an expression for the peak factor α_p as a function of $n = (s_o/T_A)\ (-\ell n\ p)^{-1}$ for different values of a bandwidth factor δ_A. Recall that p denotes the probability of no exceedance. If the probability $p = 0.5$ (to obtain median responses), the factor $(-\ell n\ p)^{-1}$ becomes 1.4 (while for more conservatively chosen probabilities $p =$ 0.9 and 0.99, this factor becomes 10 and 100, respectively). The approximate expression for α_p is

$$\alpha_p = \left[2\ell n \left\{2n\left[1-\exp(\delta_A^{1.2}\sqrt{\pi\ell n(2n)})\right]\right\}\right]^{\frac{1}{2}} \qquad \text{(C6)}$$

The period and bandwidth parameters T_A and δ_A depend on the value of the fraction q as follows:

$$\begin{aligned} T_A &= T_o q + (2\pi/\omega_n)(1-q) \\ \delta_A &= q + \left[(4/\pi)\beta_{s_o}\right]^{\frac{1}{2}}(1-q) \end{aligned} \qquad \text{(C7)}$$

in which T_o = "central" period of the ground motion, and β_{s_o} = damping parameter β_t evaluated at $t = s_o$ (see Equation (C3)). At intermediate periods, e.g., $\omega_n=2\pi$, the fraction $q \simeq 0$, and the median peak factor becomes approximately $\alpha = \alpha\ 0.5 = \sqrt{2\ \ell n(1.4\ \omega_n\ s_o/\pi)}$ (see Equation 10).

REFERENCES

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DOCUMENT 1

SUMMARY OF INVESTIGATIONS OF THE FELT AREA

EARTHQUAKE OF MAY 31, 1897

GILES COUNTY, VIRGINIA

Weston Geophysical has attempted to document the total perceptible felt area of the Giles County, Virginia, earthquake of May 31, 1897. Based upon an earlier study by Hopper and Bollinger (1971), the total felt area had been calculated at 280,000 square miles. Included in their summary of this event was an isoseismal map of the area affected by the earthquake. This area included portions of fifteen states.

The outer limit of the felt area, as determined by Hopper and Bollinger (1971), seems to have been established on the basis of dispatches to the New York Times, the Washington Post, and the major Roanoke, Virginia, newspapers. There is no direct indication of a research effort conducted in these peripheral regions.

Weston Geophysical has attempted to complement the information already provided by Hopper and Bollinger (1971) for the peripheral region by contacting directly local sources of information either by telephone calls, letters, or in some cases, by personal visits to local repositories.

Using the Hopper and Bollinger isoseismal map as a basis for research, the peripheral region in need of further investigation comprised areas of the following states:

Alabama, Delaware, Georgia, Indiana, Kentucky, New Jersey, North Carolina, Ohio, Pennsylvania, South Carolina, and Tennessee.

The initial phase of research involved establishing which towns in these states were on the periphery of the felt area and would have had a local newspaper in 1897. A telephone canvass was conducted of newspaper offices, historical societies and libraries in these towns to verify available newspaper coverage. In addition, state libraries, historical societies and universities were contacted, as they sometimes retain extensive local newspaper collections. A listing of all repositories contacted regardless of the result obtained is included as Table 1.

As a result of this study, Weston Geophysical has been able to document further the original felt area provided by Hopper and Bollinger (1971) and at the same time, to define its limits more clearly. The limit of the felt area has been more thoroughly documented in the northeast area of the isoseismal map, particularly in Pennsylvania, Ohio, and New Jersey. Local newspaper accounts have provided confirmation of effects of the earthquake for a number of localities not previously mentioned. Negative findings have been provided from towns whose local newspaper reports did not mention the earthquake or did not indicate that the event was felt locally. These findings are included as Table 2.

These additional findings have been added to the original Hopper and Bollinger (1971) isoseismal map. They seem to indicate a slightly different orientation to the elliptical isoseismal pattern than Hopper and Bollinger had originally provided. At the same time, this information reconfirms the original felt area of approximately 280,000 square miles. The revised boundary of the felt area has been included on Figure 1, where both Weston Geophysical data and the original results of Hopper and Bollinger (1971) are presented.

REFERENCES

Hopper, Margaret G. and G. A. Bollinger, The Earthquake History of Virginia: 1774 to 1900, Virginia Polytechnic Institute and State University, Blacksburg, Virginia, pp. 54-66, 1971.

TABLE 1

TOWNS AND LOCAL REPOSITORIES CONTACTED DURING FELT AREA INVESTIGATION

<u>ALABAMA</u>	
Anniston	Public Library of Anniston and Calhoun County Calhoun County Courthouse The Anniston Star Office The John H. Forney Historical Society
Cullman	Cullman County Public Library Cullman College Library The Cullman Times Office The Cullman Tribune Office Cullman County Courthouse
Decatur	Wheeler Basin Regional Library John C. Calhoun State Community College Library Northern Alabama Historical Society Morgan County Courthouse
Florence	Muscle Shoals Regional Library
Fort Payne	De Kalb County Library The Fort Payne Times-Journal Office
Gadsden	Gadsden Public Library
Guntersville	Guntersville Public Library The Advertiser-Gleam Office
Heflin	Cleburne County Public Library Cleburne County Courthouse
Huntsville	Northern Alabama Cooperative Library System
Montgomery	Alabama State Library
Moulton	Lawrence County Public Library
Oneonta	Oneonta Public Library The Southern Democrat Office Blount County Historical Society
Pell City	St. Clair County Library The St. Clair News Aegis
Phil Campbell	Northwest Alabama State Junior College Library
Roanoke	Roanoke City Public Library

ALABAMA (Continued)	
Scottsboro	Scottsboro Public Library
University	Geological Survey of Alabama
DELAWARE	
Dover	Delaware Division of Libraries Dover Public Library State Department of Archives
Georgetown	Georgetown Public Library Delaware Technical and Community College
Laurel	Laurel Public Library
Milford	Milford Public Library
Seaford	Seaford District Library
Wilmington	Historical Society of Delaware Wilmington Institute Library
GEORGIA	
Americus	Lake Blackshear Regional Library
Athens	University of Georgia at Athens Library
Brunswick	Brunswick-Glynn County Regional Library Brunswick Junior College
Columbus	Chattahoochee Valley Regional Library
Douglas	Satilla Regional Library
Fitzgerald	Carnegie Library
Fort Valley	Thomas Public Library
La Grange	Troup-Harris-Coweta Regional Library
Macon	Middle Georgia Regional Library
Waycross	Okefenokee Regional Library The Waycross Journal-Herald

INDIANA

Crawfordsville	Crawfordsville District Public Library
Evansville	Evansville Public Library and Vanderburgh County Public Library Indiana State University, Evansville Library Evansville Courier and Evansville Press Libraries The Evansville Courier Office The Willard Library of Evansville
Frankfort	Frankfort Community Public Library
Huntington	Huntington Public Library
Indianapolis	Indiana State Library
Kokomo	Kokomo Public Library
Oakland City	Oakland City - Columbia Township Public Library Oakland City College Library The Oakland City Journal
Peru	Peru Public Library
Petersburg	Barrett Memorial Library
Princeton	Princeton Public Library
Rockville	Rockville Public Library
Sullivan	Sullivan County Public Library
Vincennes	Vincennes University Junior College Library
Wabash	Wabash Carnegie Public Library Wabash Plain Dealer Office The Wabash Museum

KENTUCKY

Bowling Green	Bowling Green Public Library Western Kentucky University
Brandenburg	Meade County Public Library The Meade County Messenger Office
Frankfort	Kentucky Department of Library and Archives
Glasgow	Mary Wood Weldon Memorial Library
Leitchfield	Leitchfield Public Library
Lexington	University of Kentucky Lexington Public Library
Munfordville	Hart County Public Library
Scottsville	Allen County Public Library

MARYLAND & D.C.

Annapolis	Maryland State Library Public Library of Annapolis and Anne Arundel County Historic Annapolis, Inc. Maryland Hall of Records
Baltimore	Enoch Pratt Free Library Maryland Historical Society Library Johns Hopkins University Library
Bel Air	Harford County Library Harford Community College
Cambridge	Dorchester County Public Library
Chestertown	Kent County Public Library
Easton	The Star-Democrat Talbot County Free Library
Elkton	Cecil County Public Library

MARYLAND & D.C. (Continued)

Princess Anne	Somerset County Library System University of Maryland - Eastern Shore Lib
Salisbury	Wicomico County Free Library
Snow Hill	Worcester County Library Carroll County Public Library
Washington, D.C.	The Library of Congress

NEW JERSEY

Cape May	Cape May City Library
Cape May Court House	Cape May County Library Cape May County Clerk
Salem	Salem Historical Society The Salem Sunbeam Salem County Clerk's Office
Trenton	New Jersey State Library

NORTH CAROLINA

Durham	Duke University
Raleigh	North Carolina State Library
Wilmington	Wilmington Public Library

OHIO

Ashtabula	Ashtabula County District Library
Cleveland	Western Reserve Historical Society
Columbus	Ohio Historical Society Library State Library of Ohio
Dayton	Dayton and Montgomery County Public Library
Findley	The Courier Office

OHIO (Continued)	
Toledo	Toledo-Lucas County Public Library
Youngstown	Reuben McMillan Free Library Association
PENNSYLVANIA	
Clarion	Clarion Free Library
Franklin	Franklin Public Library
Harrisburg	State Library of Pennsylvania
Meadville	Meadville Library Art and Historical Assn.
Oil City	Oil City Library
Philadelphia	Free Library of Philadelphia
SOUTH CAROLINA	
Beaufort	Beaufort County Library University of South Carolina
Charleston	Charleston County Library
Columbia	Columbia Newspapers Inc.
Georgetown	Georgetown County Memorial Library
Marion	Marion County Library
TENNESSEE	
Knoxville	Knoxville-Knox County Public Library
Memphis	University of Tennessee Library
Nashville	Public Library of Nashville and Davidson County Tennessee State Library and Archives Tennessee State Library University of Tennessee at Nashville

MISCELLANEOUS

Boston, Massachusetts	Boston Public Library

TABLE 2

REPORTS OBTAINED FROM INVESTIGATION OF THE FELT AREA OF THE EARTHQUAKE OF MAY 31, 1897

ALABAMA

Birmingham	Not Reported
Florence	Not Reported
Gadsden	Not Reported
Huntsville	Felt
Mobile	Not Reported
Montgomery	Not Reported

DELAWARE

Wilmington	Not Reported

GEORGIA

Americus	Not Reported
Athens	Felt
Atlanta	Felt
Augusta	Felt
Columbus	Not Reported
Eatonton	Felt
Flovilla	Felt
Jackson	Felt
LaGrange	Not Reported
Milledgeville	Felt
Sandersville	Felt
Savannah	Felt
Thomson	Felt

INDIANA

Auburn	Not Reported
Bloomfield	Not Reported
Bloomington	Not Reported
Bluffton	Not Reported
Crawfordsville	Not Reported
Denver	Not Reported
Evansville	Not Reported
Fort Wayne	Not Reported
Frankfort	Not Reported
Hartford City	Not Reported
Huntington	Not Reported

INDIANA (Continued)

Indianapolis	Felt
Kokomo	Not Reported
Madison	Felt
Peru	Not Reported
Petersburg	Not Reported
Princeton	Not Reported
Shoals	Not Reported
South Bend	Not Reported
Tell City	Not Reported
Terre Haute	Not Reported
Tipton	Not Reported
Wabash	Not Reported

KENTUCKY

Ashland	Felt
Barboursville	Felt
Beattyville	Felt
Butler	Felt
Catlettsburg	Felt
Covington	Felt
Frankfort	Not Reported
Grayson	Felt
Greenup	Felt
Hardinsburg	Not Reported
Hartford	Not Reported
Lexington	Felt
Louisville	Felt
Lynchburg	Felt
Middleboro	Felt
Maysville	Felt
Newport	Felt
Paintsville	Felt
Pikeville	Felt
Pineville	Felt
Prestonburg	Felt
Russellville	Not Reported
Salyerville	Felt
Williamsburg	Felt

MARYLAND

Baltimore	Felt
Cambridge	Felt
Ellicott City	Felt
Hagerstown	Felt
Manchester	Felt
Millersville	Felt
Roland Park	Felt
Towson	Felt
Westminster	Felt

MISSOURI

St. Louis	Not Reported

NEW JERSEY

Cape May Courthouse	Not Reported

NORTH CAROLINA

Asheville	Felt
Charlotte	Felt
Councils	Felt
Goldsboro	Felt
Greensboro	Felt
Greenville	Felt
Kenansville	Felt
Kinston	Felt
Pates	Felt
Raleigh	Felt
Wilmington	Felt
Wilson	Felt
Winston	Felt

OHIO	
Ashtabula	Not Reported
Bellefontaine	Not Reported
Chardon	Not Reported
Cincinnati	Felt
Cleveland	Felt
Columbus	Felt
Dayton	Felt
East Liverpool	Felt
Greenville	Not Reported
Lima	Not Reported
Painesville	Felt
Sidney	Not Reported
Tiffin	Not Reported
Warren	Not Felt
Wellsville	Felt
Youngstown	Felt
PENNSYLVANIA	
Altoona	Not Reported
Beaver	Not Reported
Bedford	Felt
Brownsville	Felt
Charleroi	Felt
Everett	Not Reported
Greensburg	Not Reported
Harrisburg	Not Reported
Johnstown	Not Reported
Kittanning	Not Reported
Lewistown	Not Reported
McConnellsburg	Not Reported
Mifflintown	Not Reported
Mt. Holly Springs	Felt
New Bloomfield	Not Reported
Newcastle	Not Reported
Oil City	Not Reported
Philadelphia	Not Reported
Pittsburgh	Felt
Waynesboro	Felt
Wellsboro	Not Reported
York	Not Reported

<u>SOUTH CAROLINA</u>	
Beaufort	Not Reported
Charleston	Felt
Columbia	Felt
Florence	Felt
Kingstree	Felt
Marion	Felt
Spartansburg	Felt
Wilmington	Felt
<u>TENNESSEE</u>	
Bristol	Felt
Chattanooga	Felt
Knoxville	Felt
Memphis	Not Reported

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