N. J. Arabasz

"ESTIMATION OF SEISMIC GROUND MOTION IN NORTHERN UTAH"

Contract No. 14-08-0001-19825
Robin K. McGuire
Dames & Moore
1626 Cole Blvd.
Golden, Colorado 80401
Tel. 303-232-6262

SUMMARY

Modified Mercalli (MM) intensity data in the Rocky Mountain Region provide a starting point for estimating strong ground motion in northern Utah. The available intensity data are used to calibrate attenuation equations in which site intensity near the source is constrained to equal preselected values. California strong motion data provide correlations between peak parameters (ground acceleration, ground velocity, and spectral velocity), MM intensity, and distance. These correlations, when combined with the region-specific MM intensity attenuation relation, yield quantitative estimates of ground motion which are generally consistent with other studies. The estimates are also consistent with the one available strong motion record for the northern Utah region.

FINAL REPORT

"ESTIMATION OF SEISMIC GROUND MOTION IN NORTHERN UTAH"

Name of Contractor:	DAMES & MOORE
Principal Investigator:	Robin K. McGuire
Government Technical Officer:	Steven Wegener
Short Title of Work:	Estimation of Ground Motion in Utah
Effective Date of Contract:	June 1, 1981
Contract Expiration Date:	September 30, 1983
Amount of Contract:	$44,965
Date Report is Submitted:	September 29, 1983

Sponsored by the
U.S. Geological Survey
Contract No. 14-08-0001-19825

TABLE OF CONTENTS

1. INTRODUCTION

The Salt Lake City, Utah, area is recognized to have a high seismic hazard as a result of a high population concentration near a recognized active fault zone, with little seismic resistance in existing structures and little land use planning in past years to mitigate earthquake effects. Earthquakes with magnitudes up to 6.7 have occurred in Utah, and magnitudes of 7.5 are considered possible in the Salt Lake City area (U.S. Geological Survey, 1976). One of the critical steps in quantitatively assessing the level of seismic hazard is to estimate the characteristics of ground motion which might be associated with an earthquake in northern Utah. This involves estimating ground motion at locations close to the source of energy release, and determining the attenuation of ground motion amplitudes with distance from the source.

There appear to be conflicting views on the attenuation of ground motion in northern Utah. Evernden (1975) and Howell and Schultz (1975) conclude that the attenuation of Modified Mercalli (MM) intensities in the Rocky Mountain region is lower than in California, but is higher than in the central and eastern U.S. These conclusions are based on analyses of attenuation of isoseismals throughout the contiguous U.S. King and Hays (1977), on the other hand, conclude that spectral amplitudes in northern Utah exhibit <u>higher</u> attenuation with distance than in California, at least in the frequency range 10 to 1 hz. This result is based on recorded ground motion during aftershocks of the 1975 Pocatello Valley earthquake. A third view is held by Griscom and Arabasz (1979), who conclude (based on a study of Wood-Anderson seismograph records) that ground motion attenuates approximately equally in northern Utah and California. Griscom (1980), in a subsequent study, finds that, although the available seismograph and intensity data are scattered, they tend toward the indication that ground motion in Utah attenuates faster than in California.

With these differing conclusions, it is appropriate to examine existing data and theories in order to resolve their differences and synthesize a method of estimating seismic ground motion in northern Utah. Available data include (1) MM intensities at numerous sites during several earthquakes in the northern Rocky Mountain region, 2) one, three-component accelerograph record obtained at the University of Utah during the 1962 Cache Valley earthquake, and 3) data from Wood-Anderson seismographs during numerous northern Utah earthquakes. The purpose of this study is to analyze the available data to derive a function or functions which allow predictions of ground motion characteristics of seismic shaking in northern Utah for a variety of earthquake magnitudes and source-to-site distances, throughout the frequency band of engineering interest. There will be an immediate use for these results, both in estimating ground motions and in determining the effects of shaking (identifying soil deposits which will liquefy, landslide areas, etc). A substantial uncertainty in ground motion estimates for Utah will remain until numerous strong motion data are collected for this region over a range of earthquake sizes.

2. MM INTENSITY IN THE ROCKY MOUNTAINS

One of the useful ways of developing methods of estimating strong ground motion in regions with few data is through MM intensity experienced in the region. The procedure consists of using region-specific intensity observations to develop attenuation equations estimating I_s (MM intensity at a site) as a function of earthquake size and source-to-site distance. Strong motion data are used from other regions to develop relations between quantitative strong motion parameters and MM intensity; these are mathematically combined with the intensity attenuation equations to estimate strong ground motion for the region with few data.

In Utah and the surrounding Rocky Mountain region, historical intensity data are available from the National Oceanic and Atmospheric Administration (NOAA). These data consist of a computerized list of events, intensity reports, and distances compiled from the series U.S. Earthquakes (published by NOAA and the USGS).

Table 1 shows five earthquakes in the northern Utah region, plus four additional events in the Rocky Mountain area which were selected for study. These events are well-documented with intensity data; they were selected to give a broad range of earthquake sizes. The five Utah earthquakes were analyzed separately from the entire group of nine events, as discussed below, to determine if the intensity data indicate any local characteristics of attenuation which might be different from the Rocky Mountain region as a whole.

To determine intensity attenuation, several mathematical functions are available. The most basic are of the form (e.g. Bollinger, 1977):

$$I_s = a + bI_e + c \ln \Delta + d \Delta \qquad (1)$$

$$I_s = a + bM_L + c \ln \Delta + d \Delta \qquad (2)$$

where I_e is the maximum reported intensity for the earthquake, M_L is local magnitude, Δ is epicentral distance, and a, b, c, and d are constants often fit by least-square regression analysis. A variation of this, which is appropriate if the intensity data do not appear to be appropriately related to earthquake size, is to assume the proportionality constant between I_s and I_e or M_L. For example, if it is assumed that $I_s \alpha I_e$ or $I_s \alpha 1.5 M_L$ (discussed further below), the resulting equations are:

$$I_s = a + I_e + c \ln \Delta + d \Delta \qquad (3)$$

$$I_s = a + 1.5 M_L + c \ln \Delta + d \Delta \qquad (4)$$

Procedurally, regression analyses with these forms can be accomplished by transforming the equation so that I_s and I_e (or M_L) with their assumed coefficients are on the left side, and are treated as the dependent variables.

Results of regression analyses using equations (1) through (4), for sets of the earthquakes shown in Table 1, are given in Table 2. Several conclusions can be drawn from these results.

First, the term involving Δ does not contribute significantly to explaining observations of intensity. The coefficient of Δ is very close to zero in all cases, and the residual uncertainty (σ_{Is}) is not increased by deleting Δ from the regression analyses. Further, the small values obtained for the coefficient d are generally positive, which is not realistic; if this term is to model anelastic attenuation, the value of d should be negative to be physically plausible.

The five Utah events have more disperse data, and are less-well correlated with earthquake size, than is the entire data set of nine earthquakes. The coefficient of I_e in regressions 1 and 2 (0.0376 and 0.0382, respectively) is unrealistically low. This is a result of the few earthquakes, and limited range of sizes available in Utah with numerous intensity reports. The entire data set shows about the same dependence on ln Δ as does the Utah data set, implying that attenuation in Utah is not different from the Rocky Mountain region as a whole, at least insofar as can be determined by MM intensities. We conclude that the entire set of nine earthquakes is more appropriate to examine further than the subset of Utah data alone; the entire data set provides more observations over a wider range of earthquake sizes than does the Utah data alone.

Calculated values of coefficient b_M are about 0.34 for the Utah data and 0.88 for the entire data set. This is unrealistically low: typically, a change of one magnitude unit implies a change of more than one intensity unit, so the coefficient b_M should be greater than 1. The values obtained probably result from the relatively small number of earthquakes available. We conclude that the use of an assumed dependence of I_s on I_e or M_L is more appropriate than calculating a coefficient from the data available.

To relate I_s to I_e, it is typical to assume a one-to-one dependence between the two (e.g. Bollinger, 1977; Howell and Schultz, 1975; Evernden, 1975). To determine an appropriate dependence between I_s and M_L, values of I_e and M_L for the nine events were plotted as shown in Figure 1. Also shown are two dashed lines representing least-square fits to the data:

$$M_L = 1.5 + 0.59\ I_e \text{ (Utah data)} \qquad (5)$$

$$M_L = 2.4 + 0.47\ I_e \text{ (all data)} \qquad (6)$$

Also shown for comparison is the relation developed for southern California by Gutenberg and Richter (1956):

$$M_L = 1 + 0.67\ I_e \qquad (7)$$

The dashed lines in Figure 1 are not significantly different from the general relation developed by Gutenberg and Richter based on a large number of earthquakes. Equation (6) is heavily influenced by the Hebgen Lake earthquake (I_e = X, M_L = 7.1) where I_e = X was assigned based on landslide effects near the epicenter. Such determinations are known to be highly unreliable: slopes have been known to fail during low levels of shaking as well as intense shaking. Thus the observation of a landslide caused by a particular earthquake is not, by itself, a good indicator of the size of that event. We conclude that equation (7) is an appropriate relation to use to estimate M_L from I_e (and vice versa). Involving the assumption that there is a one-to-one relationship between I_s and I_e, the proper scaling between I_s and M_L can be approximated by $I_s \alpha 1.5 M_L$ (this form has been assumed by others, for example Howell and Schultz, 1975). This scaling forms the basis for equation (4) and regression analyses 7, 8, 15, and 16.

From all of these considerations, regressions 12 and 16 appear to hold the most promise for representing MM intensities in the Rocky Mountain region. They do not have a superfluous Δ term, and their dependence on earthquake size (represented by I_e and M_L, respectively) is appropriate.

Figures 2 through 10 show MM intensity data for each earthquake, plotted versus distance. Also shown are lines representing estimates from regressions 12 and 16. The estimates are inaccurate in some cases, particularly at close distances, where it is clear that the few available intensity observations in those regions are insufficient to specify the curves accurately.

To make better estimates of ground motion in the near-source region, it is common to assume that ground motion within some distance of the epicenter does not vary in the mean. To incorporate this assumption in regression analyses on MM intensity, the curves can, effectively, be fixed to give some predetermined intensity at $\Delta = \Delta_T$. The regression analysis can then be conducted in the usual way to determine the remaining parameters of the assumed equation.

Two constraints for near-source intensities were examined:

$$I_s = I_e \qquad \text{for } \Delta \leq \Delta_T \qquad (8)$$

$$I_s = 1.5\, M_L - 1.5 \qquad \text{for } \Delta \leq \Delta_T \qquad (9)$$

The latter constraint is obtained by inverting equation (7) and assuming that $I_s = I_e$ for $\Delta \leq \Delta_T$. Procedurally, these constraints are applied by combining the original regression equation with one expressing the constraint as a function of Δ_T, to eliminate one of the original regression coefficients. A new set of dependent and independent variables is obtained, the regression analysis is conducted, and the resulting coefficients can be used to transform the equation back to the original set of dependent and independent variables. This sometimes results in a term involving the product of two independent variables.

The results of these regressions are given in Table 3, for Δ_T = 10, 15, and 5 km. In general, calculated values of coefficients b_I and b_M are more realistic with these regressions. The dependence on Δ of some of these equations (involving coefficient d and sometimes f_1 or f_3) is usually unrealistically positive, reinforcing the previous conclusion that equations without a Δ term are preferable. The residual standard deviations shown in Table 3 are calculated from the data after the coefficients have been determined, because the transformation of variables does not readily lend itself to a determination of the uncertainty in observations of I_s. For theoretical reasons we prefer the mathematical forms which do not fix b_I or b_M, but rather which only impose a constraint on I_s at $\Delta = \Delta_T$. Thus, for example, regression 22 implies fewer constraints than does regression 24, and fits the data as well.

Figures 11 through 19 show the earthquake intensity data and estimates from several of the regression equations. We select and plot estimates from regressions 18 and 22 as the best equations using I_e and M_L (respectively), for Δ_T = 10 km; also shown are estimates for the analogous equations for Δ_T = 15 km (regressions 26 and 30), and Δ_T = 5 km (regressions 34 and 38). All estimates fit the near-source data better (in a visual sense) than the unconstrained models (Figures 2 through 10). For a final choice we prefer regression 22 because it is a function of M_L (an instrumental measure of earthquake size) rather than I_e, and because Δ_T = 10 km seems consistent with the intensity data available. Also, earthquakes in the Rocky Mountain region appear to have depths on the order of 10 km, which is consistent with a leveling-off of average ground motion estimates for $\Delta_T <$ 10 km.

These results suggest that a distance parameter which accounts for depth of energy release might be more appropriate to account for near-source intensities. Accordingly, the regression analyses were repeated assuming a focal depth of 10 km and using hypocentral distance instead of epicentral distance. The computed coefficients are shown in Table 4. These generally indicate the same trends as shown in Table 3 and exhibit slightly larger residual standard deviations. Regression 46 is preferred over the alternatives among the analyses done using hypocentral distance for the reasons stated above. Estimates from regression 46 are also plotted on Figures 11 through 19.

A final set of comparisons is shown in Figures 20 and 21. In these cases we compare observations for the Hebgen Lake, MT, earthquake and the Pocatello Valley, ID earthquake, against regressions 22 and 46, and several other equations available. The USGS (1976) used the following relation to characterize intensities in Utah:

$$I_s = I_e - 4 \log_{10} [(\Delta^2 + h^2)^{1/2}/h] \quad (10)$$

where h is depth (which we select to be 10 km for the two comparisons shown). For Figures 20 and 21 we invert equation 7 to estimate I_e from M_L rather than using the observed value of I_e (this results in $I_e \simeq$ IX for Hebgen Lake and $I_e \simeq$ VII-VIII for Pocatello Valley). This procedure contributes to the underestimation of intensities by equation (10) for these two earthquakes; this underestimation is real if one wishes to characterize the size of events by M_L rather than I_e.

Also shown in Figures 20 and 21 are estimates using the following equation from Howell and Schultz (1975):

$$\ln I_s = \ln (1.5\, M_L - 1.5) + .434 - .152 \ln \Delta - .00053\, \Delta \quad (11)$$

which is their preferred equation when M_L is used to characterize earthquake size. This equation overestimates intensities for the Pocatello event because equation (11) was derived from isoseismal data, rather than intensity observations. Regressions 22 and 46 provide a better fit to the data than either of the alternatives, which is expected because the data were used to determine the coefficients of regressions 22 and 46. In making visual comparisons between equations and data, it should be noted that the purpose here is to predict a mean intensity at a given distance, not the mean distance for a given intensity. Thus the best fit does not necessarily pass through, for example, the average distance of I_s = V for any given earthquake.

3. ESTIMATES OF GROUND MOTION FROM MM INTENSITY

Quantitative relationships between ground motion parameters and MM intensity can be developed using data from areas in which strong ground motion records are abundant (e.g., McGuire, 1977). To use such relationships for ground motion estimation in Utah, the internal Dames & Moore statistical data base was augmented by recent strong motion data from California, Alaska, and Nicaragua. These data will hereby be referred to as the California data in this report.

Table 5 shows the earthquakes used in this study, their magnitudes, and the number of stations for which site intensity determinations were available. The magnitudes used are moment magnitudes published by Joyner and Boore (1981); these are generally equivalent to M_L for values less than about 6. For larger values the difference is not significant for the purposes of estimating ground motion in Utah, given all of the other uncertainties and extrapolations involved. The source-to-station distances used for the strong motion data are the distances reported by Joyner and Boore (1981); these are actually the closest distance to surface faulting. Much of this data set is comprised of near-source observations, for which a surface distance is an inappropriate characterization of the distance from the energy release. Joyner and Boore (1981), for example, explore a depth term to obtain a proper distance-dependence for the near-source data. For the purposes of this study we use the expedient of calculating an effective distance from the energy release by

$$R = (R_s{}^2 + 100)^{1/2} \qquad (12)$$

where R_s is the surface distance to faulting given by Joyner and Boore (1981). This amounts to assuming that the energy causing the observed ground motion is released at a point on the fault closest to the site, at a depth of 10 km.

For each strong motion record, the MM intensity in the locale of the station was determined from the NOAA Earthquake Intensity file (which contains all data in the annual publication, "U.S. Earthquakes"). If an intensity report was not available for the town in which an instrument was located, the assignment was based on an intensity (not an isoseismal) map for that earthquake.

The following parameters are considered important characterizations of ground motion for which estimates were desired:

Parameter	Symbol	Units
peak acceleration	a_g	(g)
peak velocity	v_g	(cm/sec)
response amplitude (5% damping, 10 hz)	$PSRV_{10}$	(in/sec)
response amplitude (5% damping, 5 hz)	$PSRV_5$	(")
response amplitude (5% damping, 2 hz)	$PSRV_2$	(")
response amplitude (5% damping, 1 hz)	$PSRV_1$	(")
response amplitude (5% damping, .5 hz)	$PSRV_{.5}$	(")

In general discussions about these parameters, they will be referred to with the symbol y.

There are several ways in which relationships between y and I_s in a region with abundant data can be combined with I_s attenuation in a region without such data, in order to estimate y in the latter region (McGuire, 1977). The most straightforward method is:

$$\ln y = a + b\, I_s \tag{13}$$

This has the advantage of simplicity but does not recognize that the relationship between y and I_s may be a function of earthquake size and distance. For example, if y is a high-frequency measure of ground motion such as peak acceleration, a magnitude 5.2 earthquake might generate intensity VII at the epicenter with relatively high accelerations. On the other hand, a magnitude 7.5 earthquake might generate intensity VII at a distance of 100 km where the high frequencies of the motion have been largely attenuated. Equation (13) does not recognize this effect.

Several alternative mathematical forms can be devised in an attempt to incorporate a magnitude and distance dependence into the relationship between y and I_s. Available forms include:

$$\ln y = a + bI_s + c \ln R \tag{14}$$

$$\ln y = a + bI_s + d M_L \tag{15}$$

$$\ln y = a + bI_s + c \ln R + d M_L \tag{16}$$

There are theoretical reasons why each of these may not be appropriate or justifiable (McGuire, 1981). Notwithstanding all of the above objections, equations (13) through (16) are candidates for making practical ground motion estimates in Utah.

Table 6 shows the results of regression analyses on the California earthquake data. In these regressions, two variables have been added to equation (13) through (16). Variable S is a binary variable equal to zero (for a soil site) or unity (for a rock site), where the designations used in this study are those of Joyner and Boore (1981). The second variable is V, which is zero for horizontal components and unity for vertical components.

Two conclusions are immediate from the results shown in Table 6. First, for the high frequency parameters (a_g, $PSRV_{10}$, and $PSRV_5$), there is not much dependence on soil condition. This is consistent with the results reported by Joyner and Boore (1981) for a_g. Second, inclusion of both the M and ln R terms in the regression makes the I_s term redundant and meaningless; it generates a coefficient close to zero.

For estimating values of y in Utah, the form involving y as a function of I_s and ln R is preferable: it gives attenuation of ground motion in Utah which compares reasonably well with available observations. The alternatives (y as a function of I_s or of I_s and M_L) do not yield such comparisons with reasonable distance attenuation.

An interesting result from Table 6 is that regressions for ln v_g indicate that peak velocity is dependent on both I_s and ln R (regressions 59 and 60). This is in contrast to earlier results (McGuire, 1977) which showed v_g as a function of I_s to be independent of ln R. Figure 22 shows values of ln v_g (horizontal data only) normalized by I_s (from regression 60) plotted vs distance; this illustrates the dependence on distance calculated by the analysis. It is caused (or at least emphasized) by recent earthquake records at near-source distances which show relatively high values of ln v_g - .382 I_s.

4. COMPARISONS OF GROUND MOTION ESTIMATES WITH OBSERVATIONS

To generate ground motion estimates for Utah, we selected regressions 52, 59, 68, 76, 83, 91, and 99 to represent variables a_g, v_g, and $PSRV_{10}$ through $PSRV_{.5}$. These are functional forms of the type shown by equation (14). They are based on the closest distance to the energy release; for maximum consistency we use intensity estimates from regression 46 (which are based on hypocentral distance). Combining the ground motion equations with the intensity estimates gives the equations shown in Table 7. These can be compared to available data and equations for Utah to determine their validity.

Figure 23 shows a comparison between acceleration estimates from Table 7 and from Campbell (1982) for Utah. The curves from this study show somewhat more rapid attenuation of acceleration with distance than do the Campbell curves, which are based on a theoretical model and consideration of anelastic attenuation in the Utah region. The curves from this study show less sensitivity to magnitude than the Campbell curves; in fact the Table 7 equation indicates $a_g \propto \exp(.371\ M_L)$ which is low by comparison to California data. Joyner and Boore (1981) and Hanks and McGuire (1981) both find (from empirical and theoretical bases, respectively) that $a_g \propto \exp(0.6\ M_L)$. Using alternative regressions (such as regression 50) to estimate a_g results in a higher dependence of a_g on M_L, but a lower attenuation, which is considered unrealistic.

Estimates of peak velocity may be compared to observations of Wood-Anderson seismograph response to determine the validity of the estimates. The local magnitude scale is defined by

$$M_L = \log_{10} A(R) - \log_{10} A_o(R) \qquad (17)$$

where A(R) is one-half the maximum peak-to-peak amplitude of the ground motion record on a Wood-Anderson seismograph at distance R. For California, the standard amplitudes $A_o(R)$ which define the scale can be approximated by (Trifunac, 1976):

$$\log_{10} A_o(R) = \begin{cases} -1.4 - R/50 & 0 < R < 75 \text{ km} \\ -2.525 - R/200 & 75 < R < 350 \text{ km} \end{cases} \qquad (18)$$

Also, Boore (1983) finds that peak velocity (in cm) is related to A(R) (in m) by:

$$v_g = .77\ A(R) \qquad (19)$$

Putting aside for the moment the question of whether the California definition of $A_o(R)$ is applicable to Utah, we can use equations (17), (18), and (19) to show that:

$$\log_{10}\ v_g \ \alpha \ \begin{cases} R/50 & 0 < R < 75 \text{ km} \\ R/200 & 75 < R < 350 \text{ km} \end{cases} \qquad (20)$$

This dependence is shown on Figure 24, along with estimates of v_g from Table 7 for $M_L = 6.5$. The Table 7 equation shows somewhat less attenuation with distance ($v_g \ \alpha \ R^{-1.15}$) than the local magnitude scale definition (which can be approximated by $v_g \ \alpha \ R^{-1.23}$). Given the uncertainty in intensity attenuations and the potential unknown biases introduced by the conversions used here, this difference is not resolvable. Griscom (1980) argues that attenuation is greater in Utah than in California, and that positive corrections to the standard local-magnitude definition should be applied to Utah stations. Figure 24 does not support this conclusion, but given the uncertainties in the steps described above, the results derived here are insufficient to refute that conclusion.

One strong motion record is available in Utah for comparison to estimates. This record was obtained during the 1962 Cache Valley earthquake (M_L = 5.7) at a distance of 29 km. The instrument was located in the basement of a building on the Utah State University campus; the site is underlain by sediments from Lake Bonneville.

Table 8 shows a comparison of observed values of a_g, v_g, and spectral velocity (as reported by Smith and Lehman, 1979) with estimates from Table 7. The agreement is excellent, given the uncertainty in observations from component to component. Figures 25, 26, and 27 show response spectra for each component and the estimated values from Table 8. Again the comparison is excellent, given the variation in spectral amplitudes from component to component and frequency to frequency. While comparison of the estimates from equations in Table 7 with one strong motion record is indicative of agreement, it should not cause blanket acceptance of those equations. There are substantial uncertainties which remain in estimating strong ground motion in Utah, which only the collection of numerous quantitative data will resolve.

5. CONCLUSIONS

Available MM intensity data in the Rocky Mountain region, and correlations of ground motion parameters with intensity from California, allow estimates of ground motion parameters for Utah. The attenuation of MM intensity in the northern Utah area does not appear to be different from that in the Rocky Mountain region as a whole. Because of the sparsity of data in the near-source region, the most useful intensity equations are those which constrain site intensity I_s to be equal to a predetermined value near the epicenter. We prefer characterization of the earthquake size with M_L rather than epicentral intensity; the most applicable constraint is one which requires I_s to equal 1.5 M_L - 1.5 (this comes from the early work of Gutenberg and Richter) at a distance of 10 km.

Several functional forms are available to estimate ground acceleration, ground velocity, and spectral velocity from site intensity, distance, and magnitude. The equation which gives the most reasonable estimates when combined with intensity attenuation is that which relates ground motion parameters to site intensity and distance.

Comparison of the derived equations with other studies and data indicate general agreement throughout the frequency band of engineering interest. The magnitude dependence of acceleration is less than what is typical in California, but the acceleration values estimated for M_L = 6.5 are in general agreement with an independent set of curves. The velocity attenuation with distance is slightly less than that of the local magnitude definition for California, although that difference is not significant given the uncertainties in using MM intensity to characterize local attenuation, and in deriving quantitative ground motion measures from MM intensity. Estimates of spectral velocity are consistent with the one available strong ground motion record (from the 1962 Cache Valley earthquake) in Utah.

Given all of the uncertainties inherent in estimating ground motion for areas where few data exist, the equations derived here should be used with caution. They are, however, useful means of predicting strong ground motion in Utah which are consistent with MM intensity observations in the Rocky Mountain area, and with available correlations between strong ground motion and intensity.

REFERENCES

Bollinger, G.A. (1977), "Reinterpretation of the Intensity Data for the 1886 Charleston, South Carolina, Earthquake", in "Studies Related to the Charleston, South Carolina, Earthquake of 1886 - A Preliminary Report", ed. D.W. Rankin, U.S. Geological Survey Prof. Paper 1028, p. 17-32.

Boore, D.M. (1983), "Stochastic Simulation of High Frequency Ground Motions based on Seismological Models of the Radiated Spectra", Bull. Seis. Soc. Am., Vol. 73, No. 6, Dec. (in press).

Campbell, K.W. (1982), "A Preliminary Methodology for the Regional Zonation of Peak Ground Acceleration", Proc., 3rd Inter. Conf. on Microzonation, p. 365-376.

Evernden, J.F. (1975), "Seismic Intensities, 'Size' of Earthquakes and Related Parameters", Bull. Seis. Soc. Am., Vol. 65, No. 5, p. 1287-1313, Oct.

Griscom, Melinda (1980), "Space-Time Seismicity Patterns in the Utah Region and an Evaluation of Local Magnitude as the Basis of a Uniform Earthquake Catalog", Unpublished M.S. Thesis, University of Utah.

Griscom, Melinda and W.J. Arabasz (1979), "Local Magnitude (M_L) in the Wasatch Front and Utah Region: Wood-Anderson Calibration, Coda-Duration Estimates of M_L, and M_L versus m_b", in Earthquake Studies in Utah, 1850 to 1978, ed. by W.J. Arabasz, R.B. Smith, and W.D. Richins, Univ. of Utah, Dept. of Geology and Geophysics, p. 433-444.

Gutenberg, Beno, and C.F. Richter (1956), "Earthquake Magnitude, Intensity, Energy, and Acceleration", Bull. Seis. Soc. Am., Vol. 46, No. 2, p. 105-145, April.

Hanks, T.C. and R.K. McGuire (1981), "The Character of High-Frequency Strong Ground Motion", Bull. Seis. Soc. Am., Vol. 71, No. 6, p. 2071-2095.

Howell, B.F. Jr., and T.R. Schultz (1975), "Attenuation of Modified Mercalli Intensity with Distance from the Epicenter", Bull. Seis. Soc. Am., Vol. 65, No. 3, p. 651-665, June.

Joyner, W.B. and D.M. Boore (1981), "Peak Horizontal Acceleration and Velocity from Strong Motion Records Including Records from the 1979 Imperial Valley, California, Earthquake", Bull. Seis. Soc. Am., Vol. 71, No. 6, p. 2011-2038.

King, K.W. and W.W. Hays (1977), "Comparison of Seismic Attenuation in Northern Utah with Attenuation in Four Other Regions of the Western United States", Bull. Seis. Soc. Am., Vol. 67, No. 3, p.781-792.

McGuire, R.K., (1977), "The Use of Intensity Data in Seismic Hazard Analysis", Proc., 6th World Conf. on Earthquake Eng., New Delhi.

McGuire, R.K. (1981), "Seismic Ground Motion Estimation in the Central and Eastern United States", in "Seismic Hazard Analysis Review Panel, Ground Motion Panel, and Feedback Results", NUREG CR/1582, Vol. 5, p. E11-E23.

Smith, R.B., and J.A. Lehman (1979), "Ground Response Spectra from the M_L 5.7 Logan (Cache Valley) Earthquake of 1962", in Earthquake Studies in Utah 1850 to 1978, ed. by W.J. Arabasz, R.B. Smith and W.B. Richins, Univ. of Utah, Dept. of Geology and Geophysics, p. 487-495.

Trifunac, M.D. (1976), "Preliminary Analysis of the Peaks of Strong Earthquake Ground Motion - Dependence of Peaks on Earthquake Magnitude, Epicentral Distance, and Recording Site Conditions", Bull. Seis. Soc. Am., Vol. 66, No. 1, p. 189-219, Feb.

U.S. Geological Survey (1976), "A Study of Earthquake Losses in the Salt Lake City, Utah Area", USGS Open-File Report 76-89.

TABLE 1

EARTHQUAKES USED FOR MM INTENSITY STUDY

Event	M_L	Maximum MMI	No. of MMI Repts
NORTHERN UTAH REGION:			
Hansel Valley, UT (3/12/34)	6.6*	VIII	105
Cache Valley, UT (8/30/62)	5.7*	VII	218
Magna, UT (9/5/62)	5.2*	VI	43
Marysvale, UT (10/4/67)	5.2*	VII	55
Pocatello Valley, ID (3/28/75)	6.0*	VIII	209
ROCKY MT. REGION:			
Helena, MT (10/19/35)	6.3**	VIII	242
Helena, MT (10/31/35)	6.0**	VIII	160
Hebgen Lake, MT (8/18/59)	7.1**	X	785
Cimarron, CO (10/11/60)	5.5**	VI	45

* Source: Griscom (1980)
** Source: NOAA Data File

TABLE 2

RESULTS OF REGRESSION ANALYSES OF FORM

$$I_s = a + b_I I_e + b_m M_L + c \ln \Delta + d \Delta$$

Data	No.	a	b_I	b_m	c	d	σ_{Is}
	1	7.86	0.0376	x	-0.783	0.0006	0.85
	2	7.68	0.0382	x	-0.726	x	0.85
	3	1.73	1*	x	-1.00	0.0005	1.02
	4	1.58	1*	x	-0.956	x	1.02
5 Utah Events	5	6.31	x	0.336	-0.810	0.0005	0.84
	6	6.15	x	0.340	-0.765	x	0.84
	7	0.104	x	1.5*	-0.935	0.0004	0.96
	8	0.0927	x	1.5*	-0.931	x	0.96
	9	5.20	4.20	x	-0.807	0.0004	0.72
	10	4.90	0.429	x	-0.738	x	0.72
	11	-1.86	1*	x	-1.06	-0.0009	0.92
	12	2.47	1*	x	-1.23	x	0.92
9 Rocky Mountain Region Events	13	3.25	x	0.869	-0.817	0.0005	0.72
	14	2.82	x	0.888	-0.735	x	0.72
	15	0.0688	x	1.5*	-0.959	-0.0002	0.78
	16	0.167	x	1.5*	-0.985	x	0.78

* Assumed Value

TABLE 3
RESULTS OF REGRESSION ANALYSES WITH NEAR-SOURCE CONSTRAINTS

$$I_s = a + b_I I_e + b_m M_L + c \ln \Delta + d \Delta + f_1 I_e \Delta + f_2 I_e \ln \Delta + f_3 M_L \Delta + f_4 M_L \ln \Delta$$

No.	Form of Reg. Eq.	Constraint	a	b_I	b_m	c	d	f_1	f_2	f_3	f_4	σ
17	Same as #9	$I_s = I_e$ @ 10 km	3.73	1.06	x	-1.89	0.0622	-0.006	x	x	x	1.13
18	Same as #10	"	0.595	1.29	x	-0.258	x	x	-0.126	x	x	.84
19	Same as #11	"	3.50	1*	x	-1.53	0.00300	x	x	x	x	1.02
20	Same as #12	"	3.11	1*	x	-1.35	x	x	x	x	x	.92
21	Same as #13	$I_s = 1.5\ M_L - 1.5$ @ 10 km	1.806	x	1.57	-1.68	0.0562	x	x	-0.00743	x	.87
22	Same as #14	"	-0.0577	x	1.71	-0.626	x	x	x	x	-0.0912	.80
23	Same as #15	"	1.84	x	1.5*	-1.47	0.00414	x	x	x	x	.91
24	Same as #16	"	1.29	x	1.5*	-1.21	x	x	x	x	x	.80
25	Same as #9	$I_s = I_e$ @ 15 km	7.24	1.14	x	-3.25	0.104	-0.009	x	x	x	1.90
26	Same as #10	"	2.16	1.26	x	-0.799	x	x	-0.0953	x	x	.96
27	Same as #11	"	6.82	1*	x	-2.60	0.0147	x	x	x	x	2.18
28	Same as #12	"	4.40	1*	x	-1.62	x	x	x	x	x	.97
29	Same as #13	$I_s = 1.5\ M_L - 1.5$ @ 15 km	4.84	x	1.69	-2.92	0.105	x	x	-0.0127	x	1.78
30	Same as #14	"	2.48	x	1.49	-1.47	x	x	x	x	0.002	.89
31	Same as #15	"	5.00	x	1.5*	-2.48	0.0144	x	x	x	x	1.90
32	Same as #16	"	2.45	x	1.5*	-1.46	x	x	x	x	x	.89
33	Same as #9	$I_s = I_e$ @ 5 km	1.58	1.02	x	-1.11	0.0400	-0.0044	x	x	x	1.25
34	Same as #10	"	-0.138	1.21	x	0.080	x	x	-0.130	x	x	.76
35	Same as #11	"	1.48	1*	x	-.912	-0.0024	x	x	x	x	.93
36	Same as #12	"	1.72	1*	x	-1.07	x	x	x	x	x	.93
37	Same as #13	$I_s = 1.5\ M_L - 1.5$ @ 5 km	-0.0838	x	1.52	-0.980	0.032	x	x	-0.0048	x	.92
38	Same as #14	"	-1.36	x	1.72	-0.086	x	x	x	x	-0.14	.77
39	Same as #15	"	-0.0796	x	1.5*	-0.878	-0.0015	x	x	x	x	.80
40	Same as #16	"	0.039	x	1.5*	-0.956	x	x	x	x	x	.78

* Assumed Value

TABLE 4

RESULTS OF REGRESSION ANALYSIS WITH NEAR-SOURCE CONSTRAINTS

$$I_s = a + b_I I_e + b_m M_L + c \ln R + d R + f_1 I_e R + f_2 I_e \ln R + f_3 M_L R + f_4 M_L \ln R$$

No.	Form of Reg. Eq.	Constraint	a	b_I	b_m	c	d	f_1	f_2	f_3	f_4	σ
41	Same as #9	$I_s = I_e$ @ R = 14.14	5.84	1.10	x	-2.63	.0796	-.0072	x	x	x	1.30
42	Same as #10	"	1.58	1.30	x	-0.596	x	x	-0.111	x	x	0.92
43	Same as #11	"	5.46	1*	x	-2.10	.0080	x	x	x	x	1.37
44	Same as #12	"	4.13	1*	x	-1.56	x	x	x	x	x	0.95
45	Same as #13	$I_s = 1.5 M_L - 1.5$ @R = 14.14	3.63	x	1.63	-2.34	.075	x	x	-.0094	x	1.08
46	Same as #14	"	1.56	x	1.60	-1.16	x	x	x	x	-.0375	0.86
47	Same as #15	"	3.71	x	1.5*	-2.01	.0087	x	x	x	x	1.25
48	Same as #16	"	2.21	x	1.5*	-1.40	x	x	x	x	x	0.85

* Assumed Value

TABLE 5

CALIFORNIA EARTHQUAKE DATA

DATE	EVENT	M	No. of Stations With I_s Determined
05/18/40	Imperial Valley, CA	7.0	1
07/21/52	Kern County, CA	7.4	4
03/22/57	Daly City, CA	5.3	1
06/28/66	Parkfield, CA	6.1	7
04/09/68	Borrego Mt., CA	6.6	6
09/12/70	Lytle Creek, CA	5.3	4
02/09/71	San Fernando, CA	6.6	20
07/30/72	Sitka, AK	7.7	1
12/23/72	Managua, Nicaragua	6.2	1
02/21/73	Point Mugu, CA	5.6	1
11/28/74	Hollister, CA	5.2	3
08/01/75	Oroville, CA	6.0	1
08/13/78	Santa Barbara, CA	5.1	3
08/06/79	Coyote Lake, CA	5.8	8
10/15/79	Imperial Valley, CA	6.5	21
01/24/80	Livermore Valley, CA	5.5	3
01/27/80	Livermore Valley, CA	5.8	4
02/25/80	Horse Canyon, CA	5.3	11

TABLE 6
RESULTS OF REGRESSION ANALYSES
ON CALIFORNIA EARTHQUAKE DATA

$$\ln y = a + bI_g + c \ln R + dM + eS + fV$$

No.	y	No. of Data	a	b	c	d	e	f	σ
49	a_g	239	-6.01	0.627	x	x	0.0083	-0.527	0.93
50	a_g	239	-6.01	0.627	x	x	x	-0.527	0.93
51	a_g	239	-0.424	0.232	-0.969	x	-0.0394	-0.530	0.58
52	a_g	239	-0.430	0.232	-0.968	x	x	-0.530	0.58
53	a_g	239	-4.51	0.632	x	-0.243	0.0151	-0.528	0.92
54	a_g	239	-4.51	0.633	x	-0.243	x	-0.528	0.92
55	a_g	239	-3.15	0.0894	-1.28	0.731	-0.0750	-0.530	0.47
56	a_g	239	-3.15	0.0887	-1.28	0.728	x	-0.529	0.47
57	v_g	245	-1.39	0.629	x	x	-0.439	-0.844	0.82
58	v_g	245	-1.45	0.624	x	x	x	-0.844	0.84
59	v_g	245	2.12	0.383	-0.612	x	-0.483	-0.846	0.69
60	v_g	245	1.99	0.382	-0.601	x	x	-0.845	0.71
61	v_g	245	-2.48	0.625	x	0.177	-0.436	-0.844	0.82
62	v_g	245	-2.57	0.620	x	0.181	x	-0.843	0.84
63	v_g	245	-1.24	0.204	-1.01	0.912	-0.495	-0.846	0.53
64	v_g	245	-1.36	0.204	-0.993	0.907	x	-0.845	0.57
65	$PSRV_{10}$	233	-3.54	0.616	x	x	0.0112	-0.272	1.07
66	$PSRV_{10}$	233	-3.54	0.616	x	x	x	-0.272	1.07
67	$PSRV_{10}$	233	2.78	0.183	-1.12	x	-0.0536	-0.275	0.70
68	$PSRV_{10}$	233	2.77	0.182	-1.12	x	x	-0.275	0.69
69	$PSRV_{10}$	233	-2.16	0.634	x	-0.233	-0.0053	-0.272	1.07
70	$PSRV_{10}$	233	-2.16	0.634	x	-0.233	x	-0.272	1.06
71	$PSRV_{10}$	233	-0.604	-0.0548	-1.55	0.978	-0.0084	-0.274	0.55
72	$PSRV_{10}$	233	-0.607	-0.0550	-1.55	0.979	x	-0.274	0.55
73	$PSRV_5$	233	-2.14	0.539	x	x	0.0784	-0.609	0.95
74	$PSRV_5$	233	-2.14	0.541	x	x	x	-0.609	0.94
75	$PSRV_5$	233	3.18	0.174	-0.948	x	0.0237	-0.612	0.65
76	$PSRV_5$	233	3.18	0.174	-0.948	x	x	-0.612	0.65
77	$PSRV_5$	233	-1.49	0.547	x	-0.111	0.0705	-0.609	0.95
78	$PSRV_5$	233	-1.47	0.549	x	-0.114	x	-0.609	0.94
79	$PSRV_5$	233	-0.120	-0.0582	-1.36	0.954	0.0678	-0.611	0.50
80	$PSRV_5$	233	-0.0960	-0.0564	-1.36	0.951	x	-0.611	0.50
81	$PSRV_2$	233	-2.12	0.639	x	x	-0.343	-0.959	0.86
82	$PSRV_2$	233	-2.15	0.631	x	x	x	-0.958	0.87
83	$PSRV_2$	233	1.80	0.371	-0.697	x	-0.383	-0.960	0.69
84	$PSRV_2$	233	1.72	0.365	-0.689	x	x	-0.960	0.71
85	$PSRV_2$	233	-2.30	0.637	x	0.0320	-0.341	-0.958	0.86
86	$PSRV_2$	233	-2.42	0.628	x	0.0469	x	-0.958	0.87
87	$PSRV_2$	233	-1.22	0.158	-1.08	0.874	-0.343	-0.960	0.58
88	$PSRV_2$	233	-1.34	0.150	-1.08	0.888	x	-0.959	0.60
89	$PSRV_1$	233	-2.67	0.740	x	x	-0.544	-0.817	0.85
90	$PSRV_1$	233	-2.72	0.728	x	x	x	-0.816	0.88

TABLE 6 (Concluded)

No.	y	No. of Data	a	b	c	d	e	f	σ
91	$PSRV_1$	233	0.172	0.545	-0.506	x	-0.573	-0.818	0.77
92	$PSRV_1$	233	0.0546	0.537	-0.494	x	x	-0.817	0.81
93	$PSRV_1$	233	-4.35	0.719	x	0.285	-0.524	-0.816	0.84
94	$PSRV_1$	233	-4.53	0.705	x	0.308	x	-0.815	0.87
95	$PSRV_1$	233	-3.39	0.295	-0.954	1.03	-0.526	-0.817	0.63
96	$PSRV_1$	233	-3.57	0.282	-0.953	1.05	x	-0.816	0.67
97	$PSRV_{.5}$	233	-2.47	0.695	x	x	-0.749	-0.747	0.95
98	$PSRV_{.5}$	233	-2.54	0.678	x	x	x	-0.746	1.00
99	$PSRV_{.5}$	233	0.152	0.515	-0.466	x	-0.776	-0.749	0.89
100	$PSRV_{.5}$	233	-0.0069	0.503	-0.451	x	x	-0.747	0.95
101	$PSRV_{.5}$	233	-5.42	0.657	x	0.501	-0.713	-0.746	0.91
102	$PSRV_{.5}$	233	-5.67	0.638	x	0.533	x	-0.745	0.96
103	$PSRV_{.5}$	233	-4.38	0.196	-1.04	1.31	-0.715	-0.747	0.68
104	$PSRV_{.5}$	233	-4.63	0.178	-1.04	1.34	x	-0.746	0.74

TABLE 7

GROUND MOTION ESTIMATES FOR UTAH

	Regressions Used
$\ln a_g$ (g) $= -.0681 + .371\, M_L - 1.24 \ln R - .0087\, M_L \ln R - .530\, V$	46, 52
$\ln v_g$ (in/sec) $= 2.72 + .613\, M_L - 1.06 \ln R - .0144\, M_L \ln R - .846\, V - .483\, S$	46, 59
$\ln PSRV_{10}$ (in/sec) $= 3.05 + .291\, M_L - 1.33 \ln R - .0068\, M_L \ln R - .275\, V$	46, 68
$\ln PSRV_5$ (in/sec) $= 3.45 + .278\, M_L - 1.15 \ln R - .0065\, M_L \ln R - .612\, V$	46, 76
$\ln PSRV_2$ (in/sec) $= 2.38 + .594\, M_L - 1.13 \ln R - .0139\, M_L \ln R - .960\, V - .383\, S$	46, 83
$\ln PSRV_1$ (in/sec) $= 1.02 + .872\, M_L - 1.14 \ln R - .020\, M_L \ln R - .818\, V - .573\, S$	46, 91
$\ln PSRV_{.5}$ (in/sec) $= .955 + .824\, M_L - 1.06 \ln R - .0193\, M_L \ln R - .749\, V - .776\, S$	46, 99

TABLE 8
ESTIMATES AND OBSERVATIONS OF STRONG
GROUND MOTION AT UTAH STATE UNIVERSITY
FOR THE 1962 CACHE VALLEY EARTHQUAKE
(M_L = 5.7, _ = 29 km)

Parameter	Table 7 Estimate (depth = 10 km) Horiz./Vert.	Observations: N-S	E-W	VERT
a_g (g)	0.094/0.055	0.092	0.125	0.049
v_g (cm/sec)	10.0/4.3	6.2	10.7	4.2
$PSRV_{10}$ (in/sec)	1.0/0.78	0.82	0.98	0.58
$PSRV_5$ (in/sec)	2.6/1.4	2.0	2.8	1.2
$PSRV_2$ (in/sec)	5.1/2.0	4.9	9.0	3.1
$PSRV_1$ (in/sec)	5.5/2.4	4.8	10.1	3.1
$PSRV_{.5}$ (in/sec)	5.2/2.5	2.5	4.5	2.1

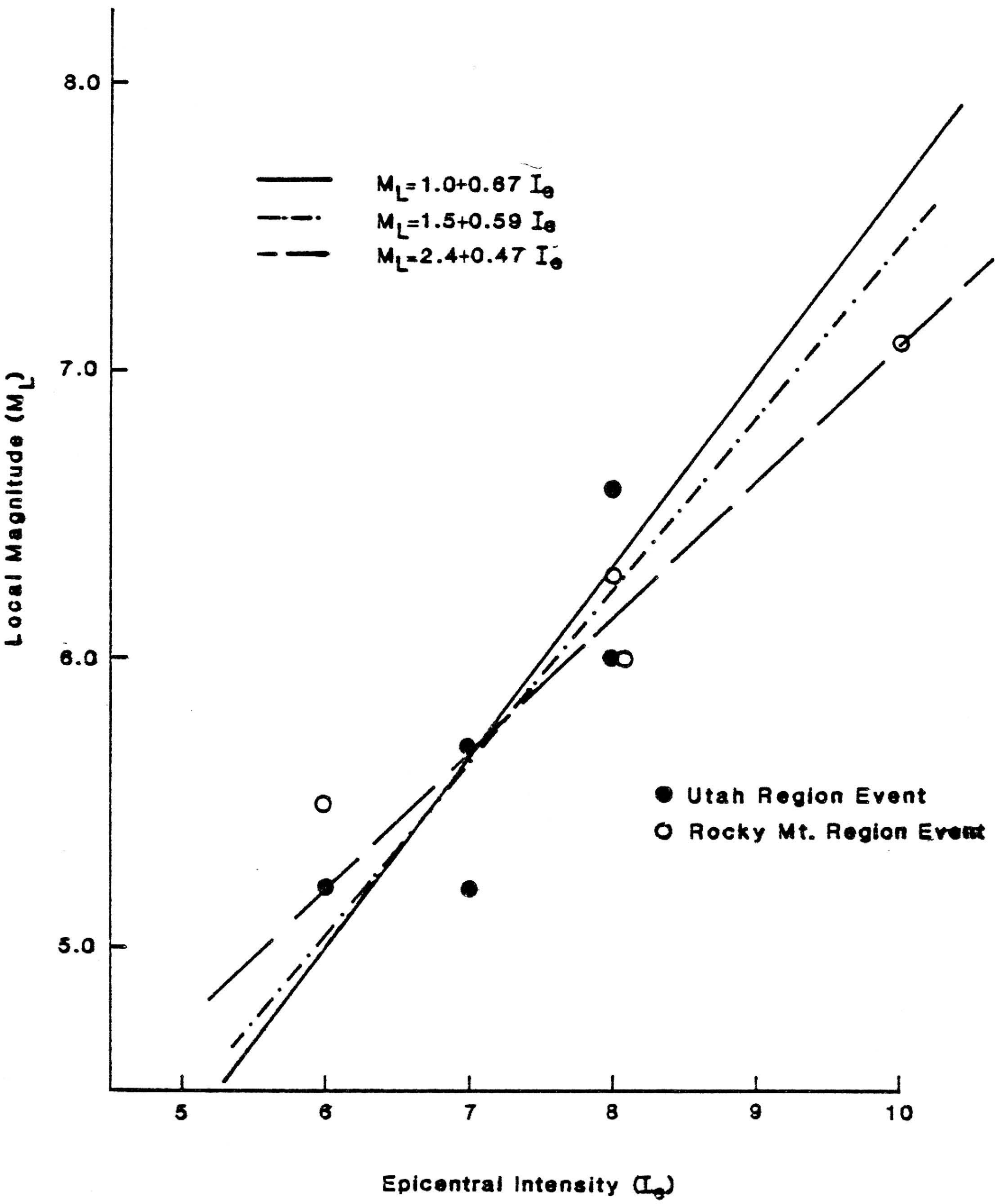

Figure 1.

Local Magnitude versus Maximum Intensity for for Rocky Mountain Earthquakes

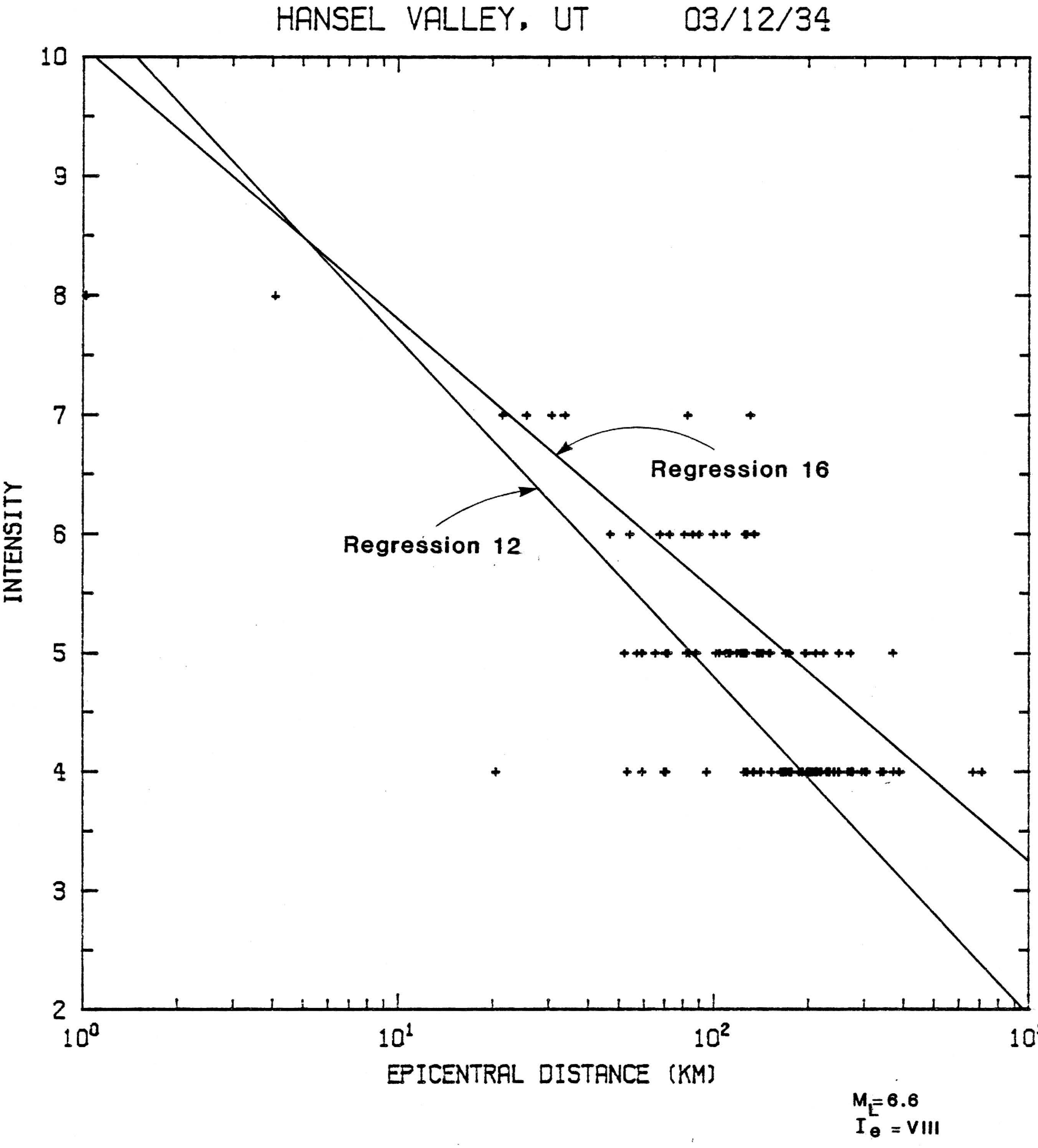

Figure 2.

Hansel Valley, Utah Earthquake Intensities vs. Distance with Lines for Regressions 12 and 16

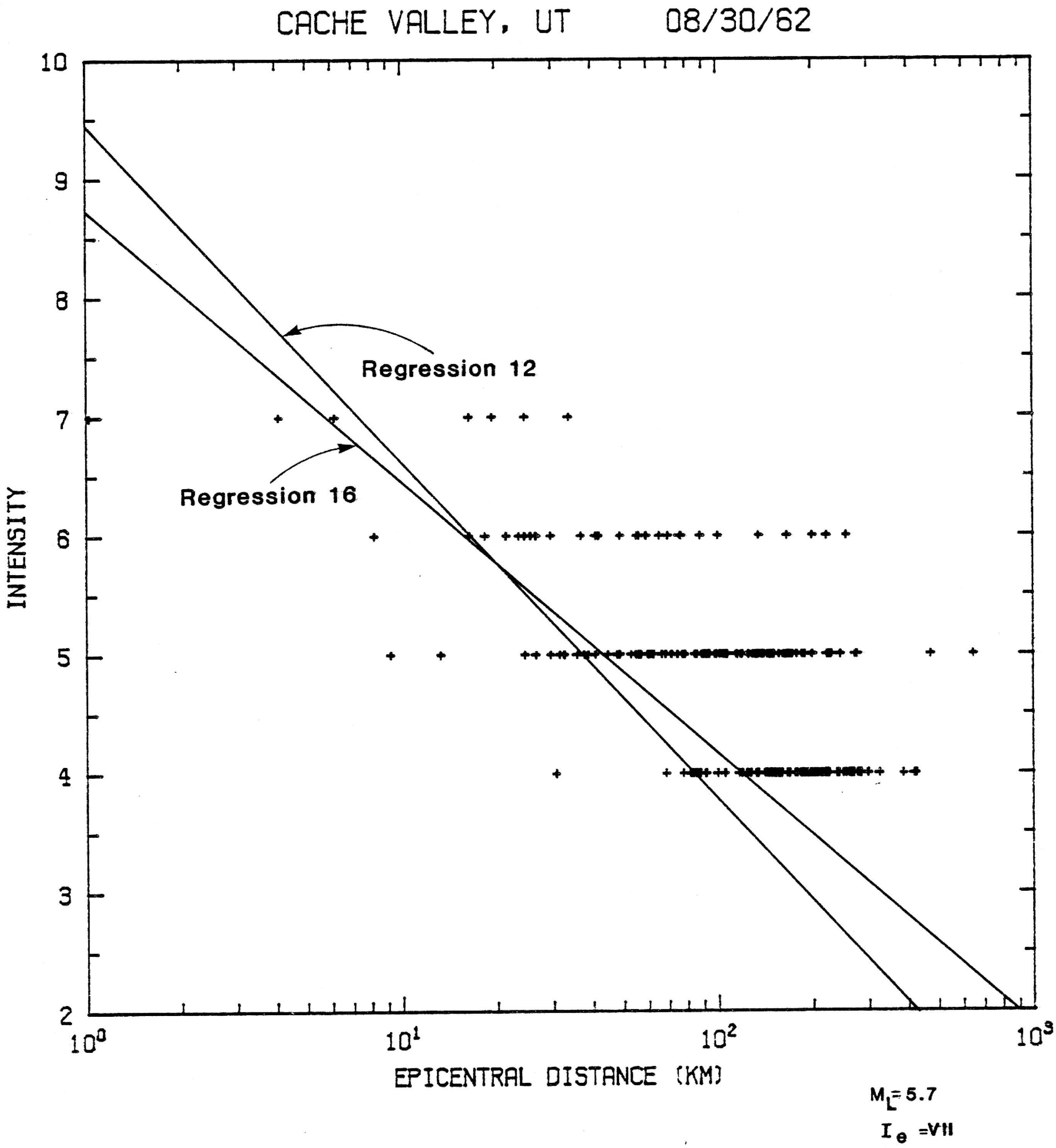

Figure 3.

Cache Valley, Utah Earthquake Intensities vs. Distance with Lines for Regressions 12 and 16

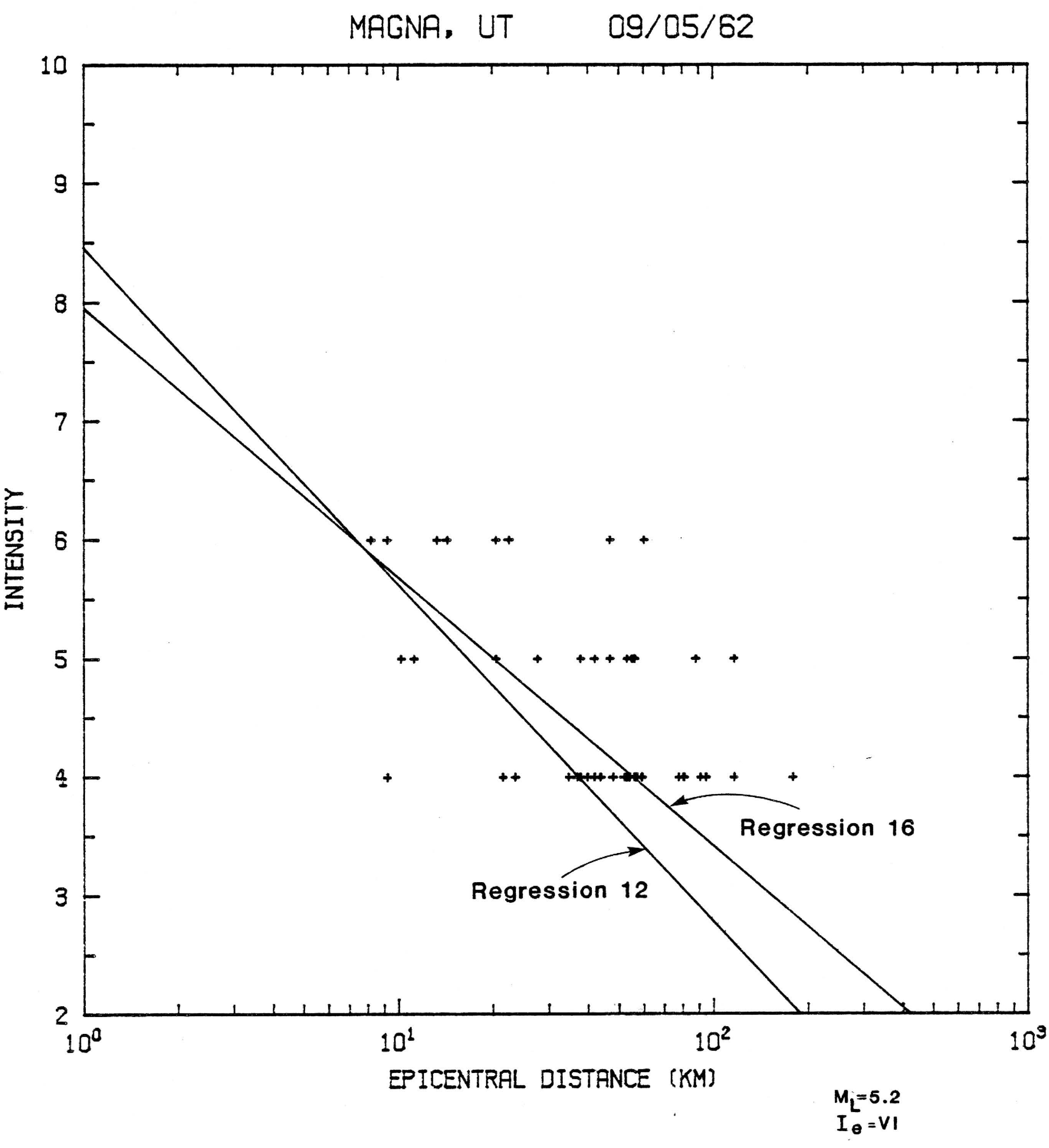

Figure 4.

Magna, Utah Earthquake Intensities vs. Distance with Lines for Regressions 12 and 16

Dames & Moore

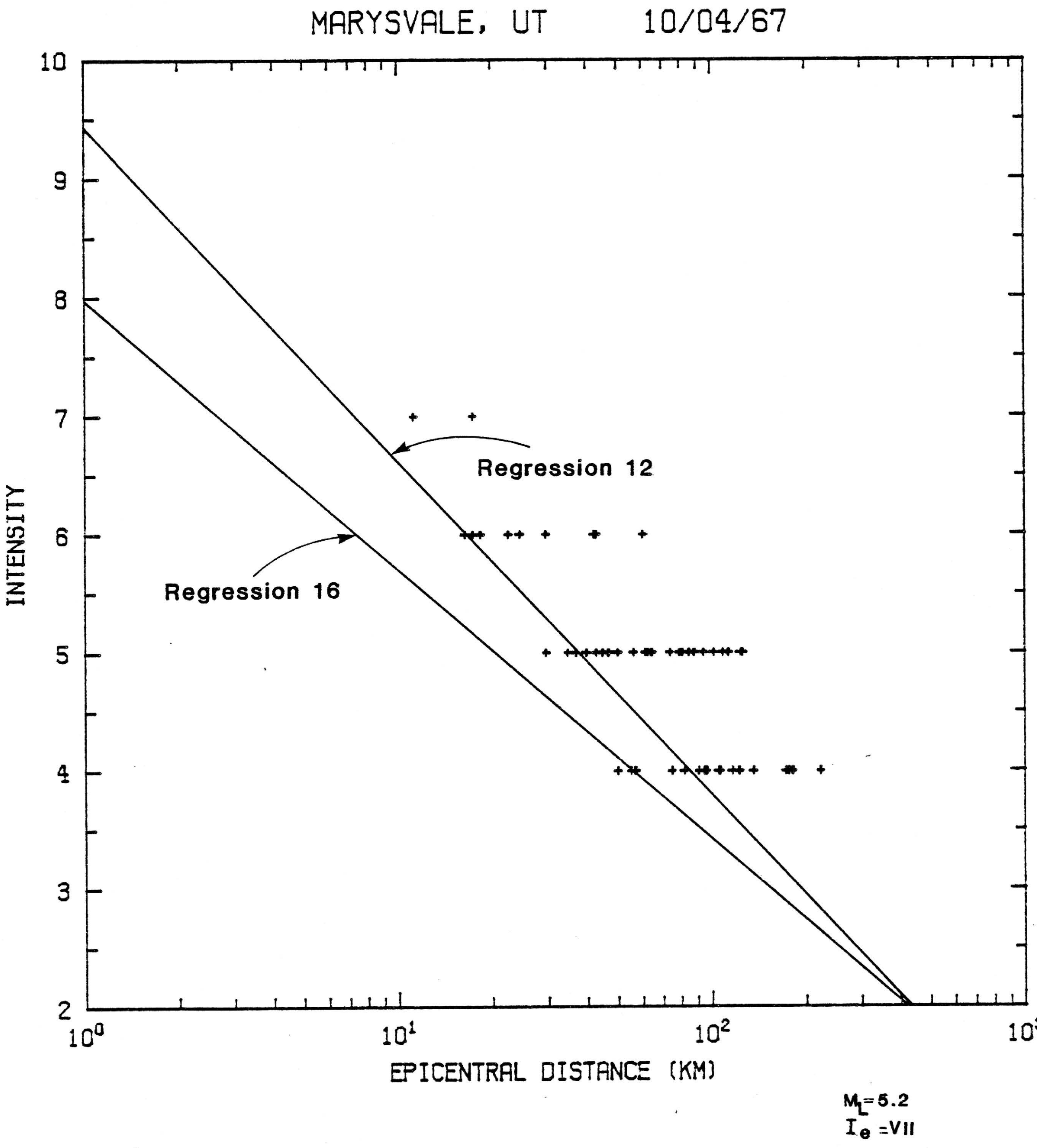

Figure 5.

Marysvale, Utah Earthquake Intensities vs. Distance with Lines for Regressions 12 and 16

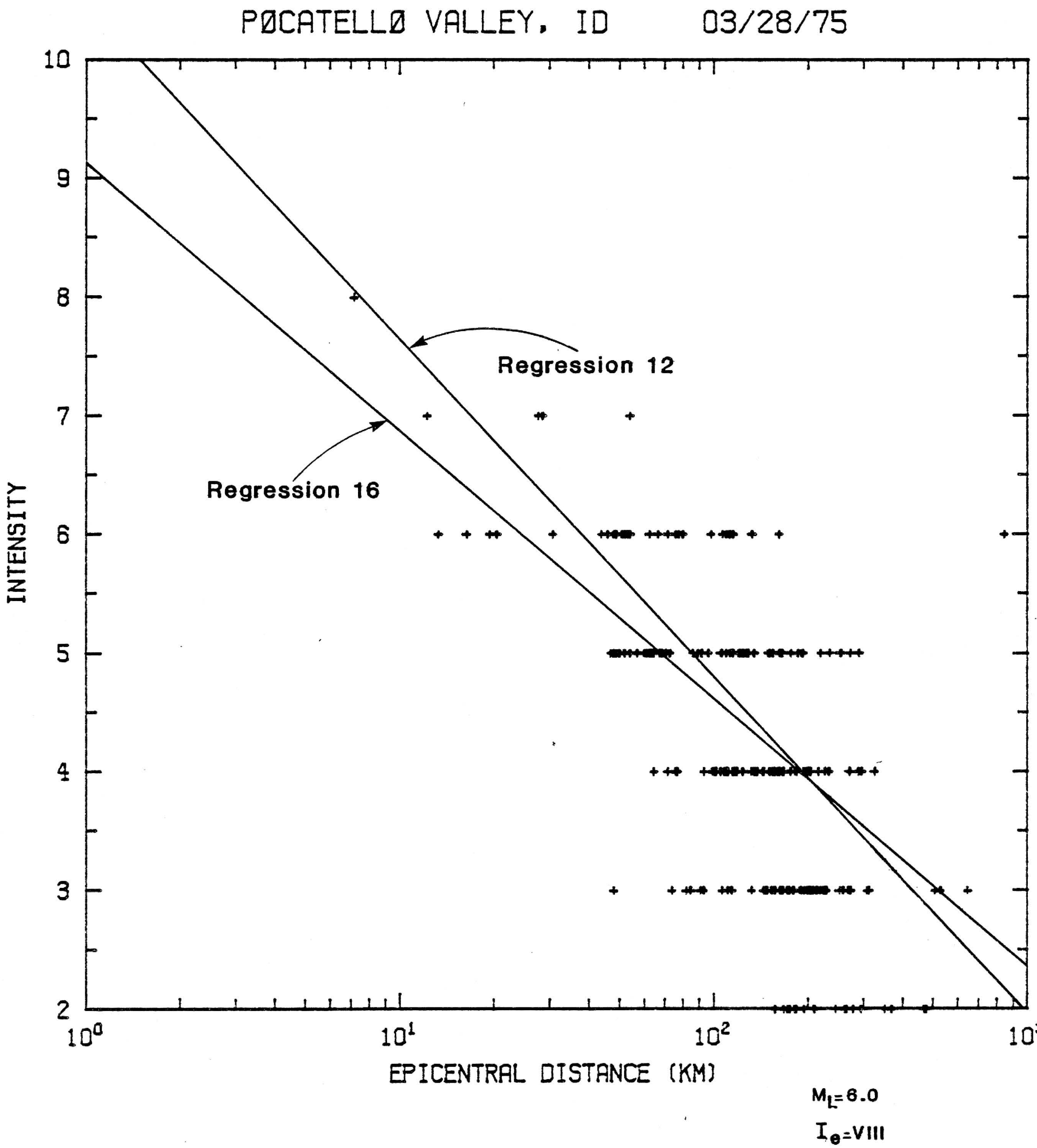

Figure 6.

Pocatello Valley, Idaho Earhquake Intensities vs. Distance with Lines for Regressions 12 and 16

Dames & Moore

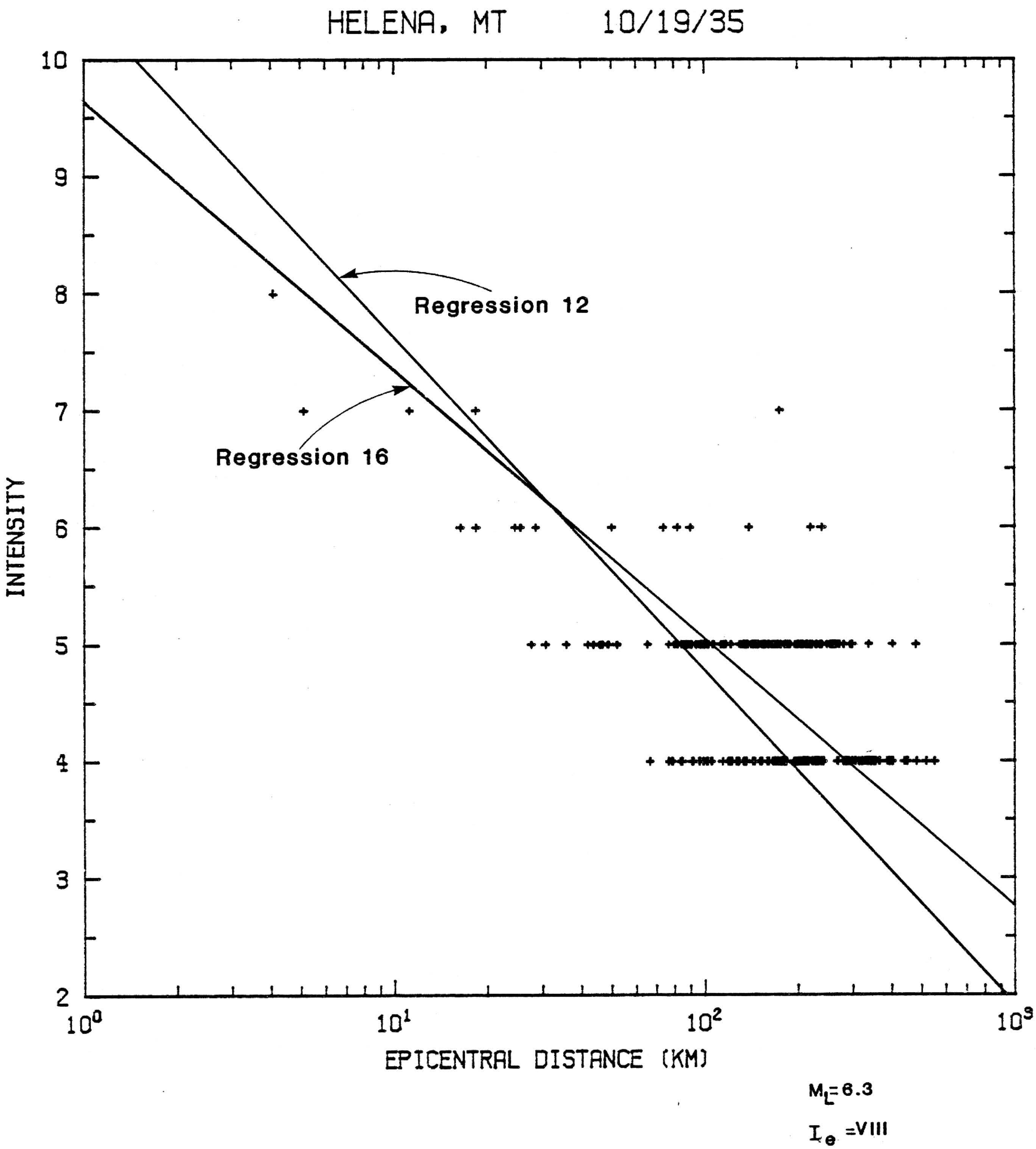

Figure 7.

Helena, Montana Earthquake Intensities vs. Distance with Lines for Regressions 12 and 16

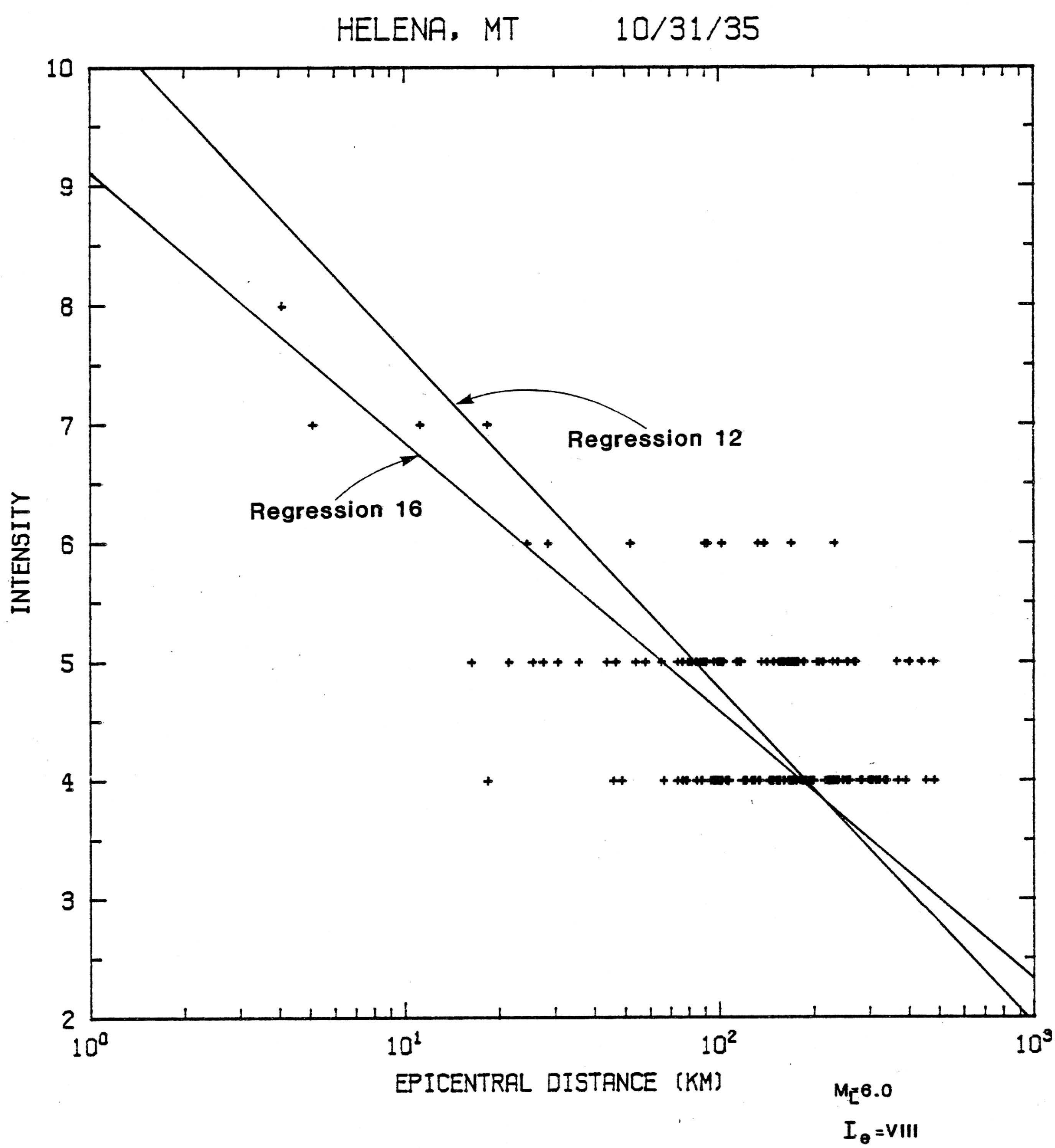

Figure 8.

Helena, Montana Earthquake Intensities vs. Distance with Lines for Regressions 12 and 16

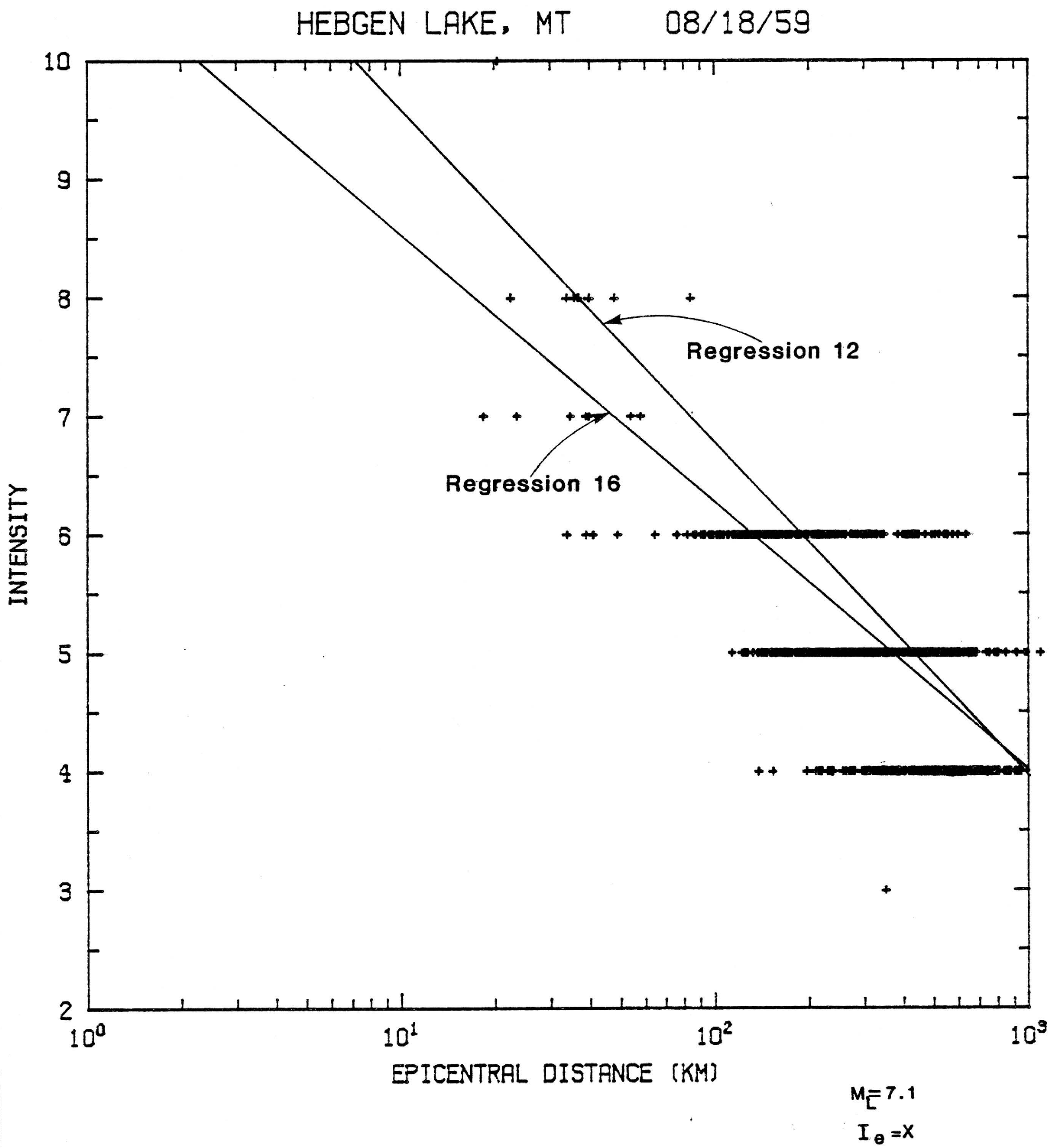

Figure 9.

Hebgen Lake, Montana Earthquake Intensities vs. Distance with Lines for Regressions 12 and 16

Dames & Moore

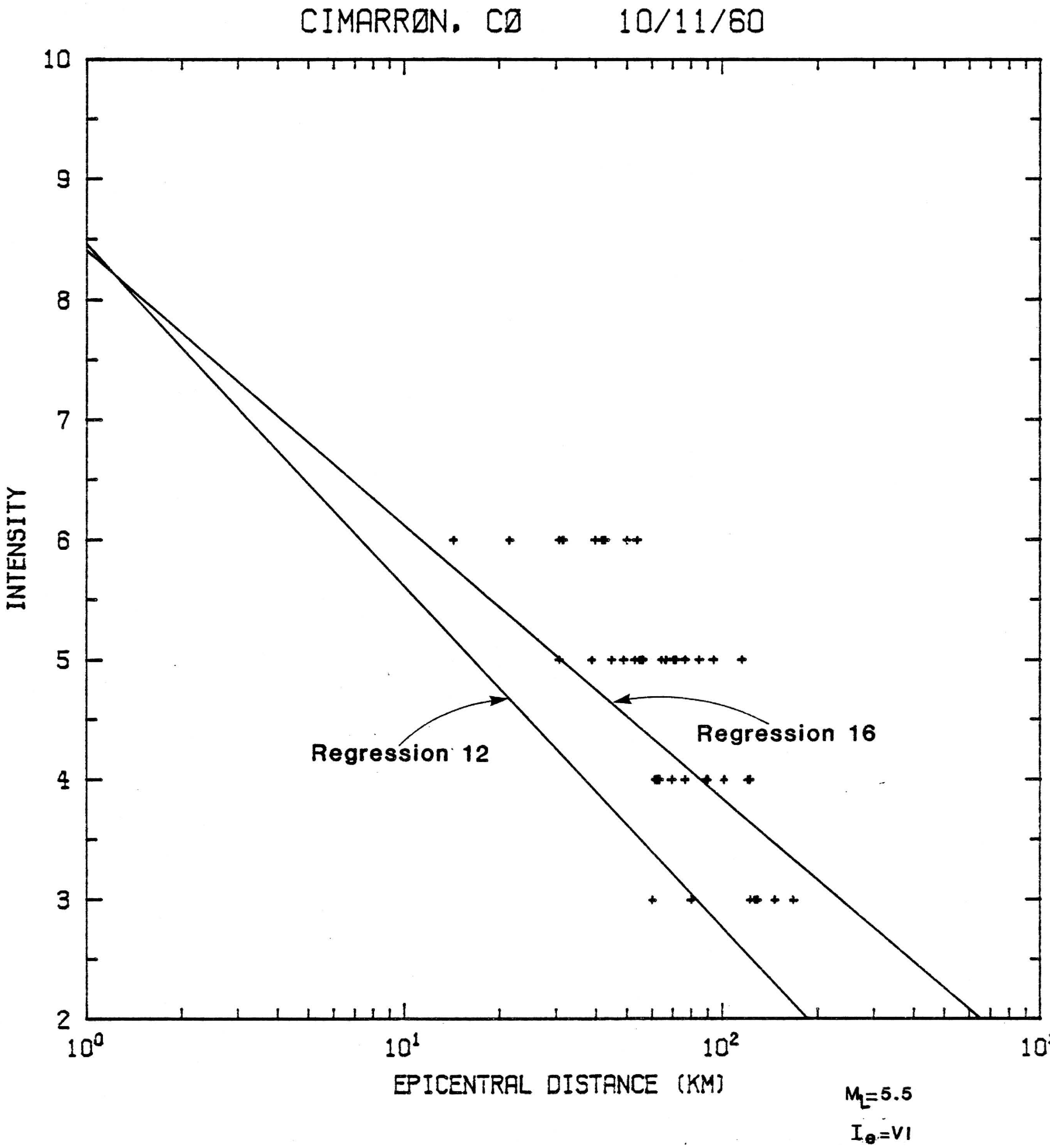

Figure 10.

Cimarron, Colorado Earthquake Intensities vs. Distance with Lines for Regressions 12 and 16

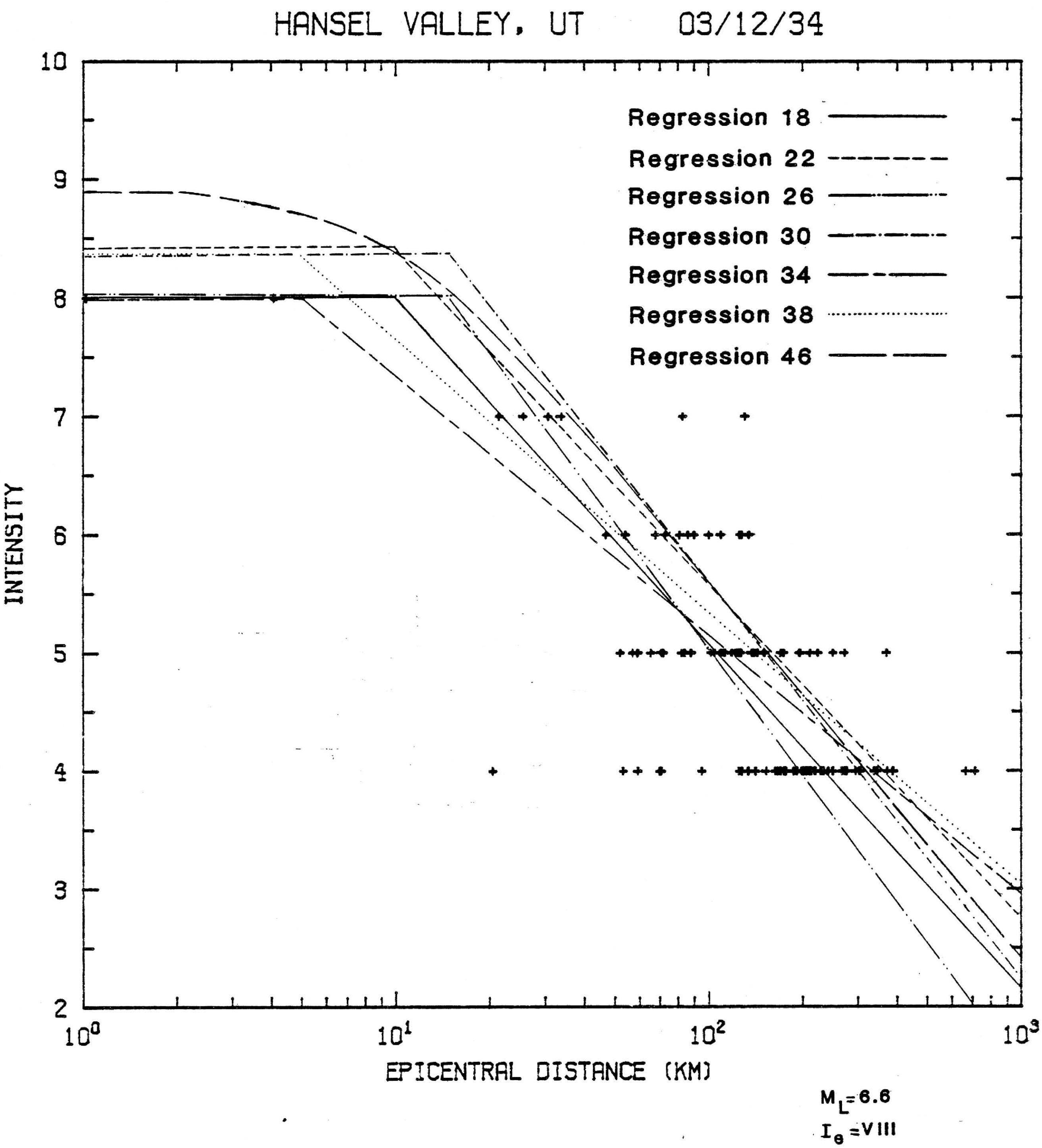

Figure 11.

Hansel Valley, Utah Earthquake Intensities vs. Distance with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

Dames & Moore

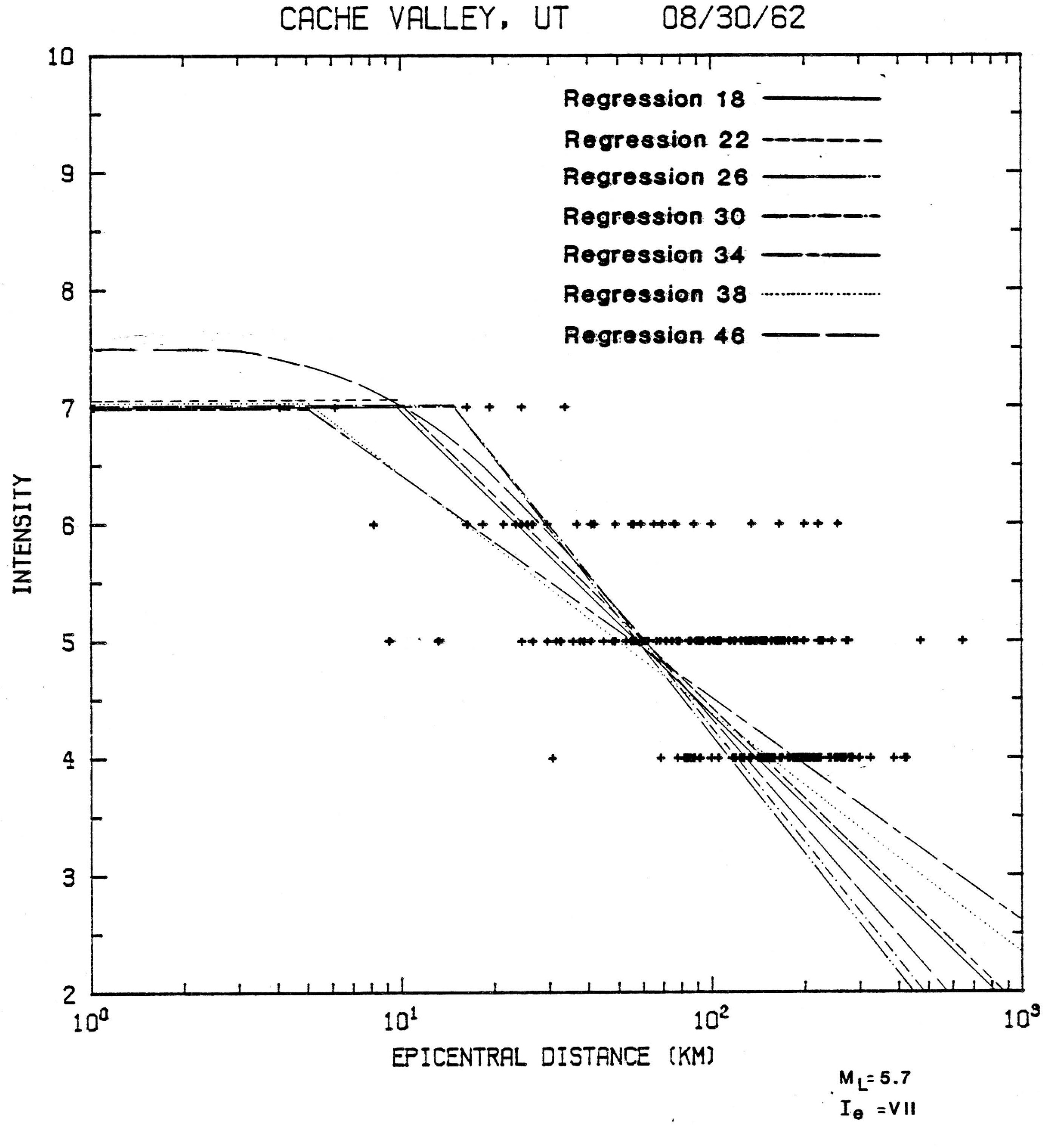

Figure 12.

Cache Valley, Utah Earthquake Intensities vs. Distance with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

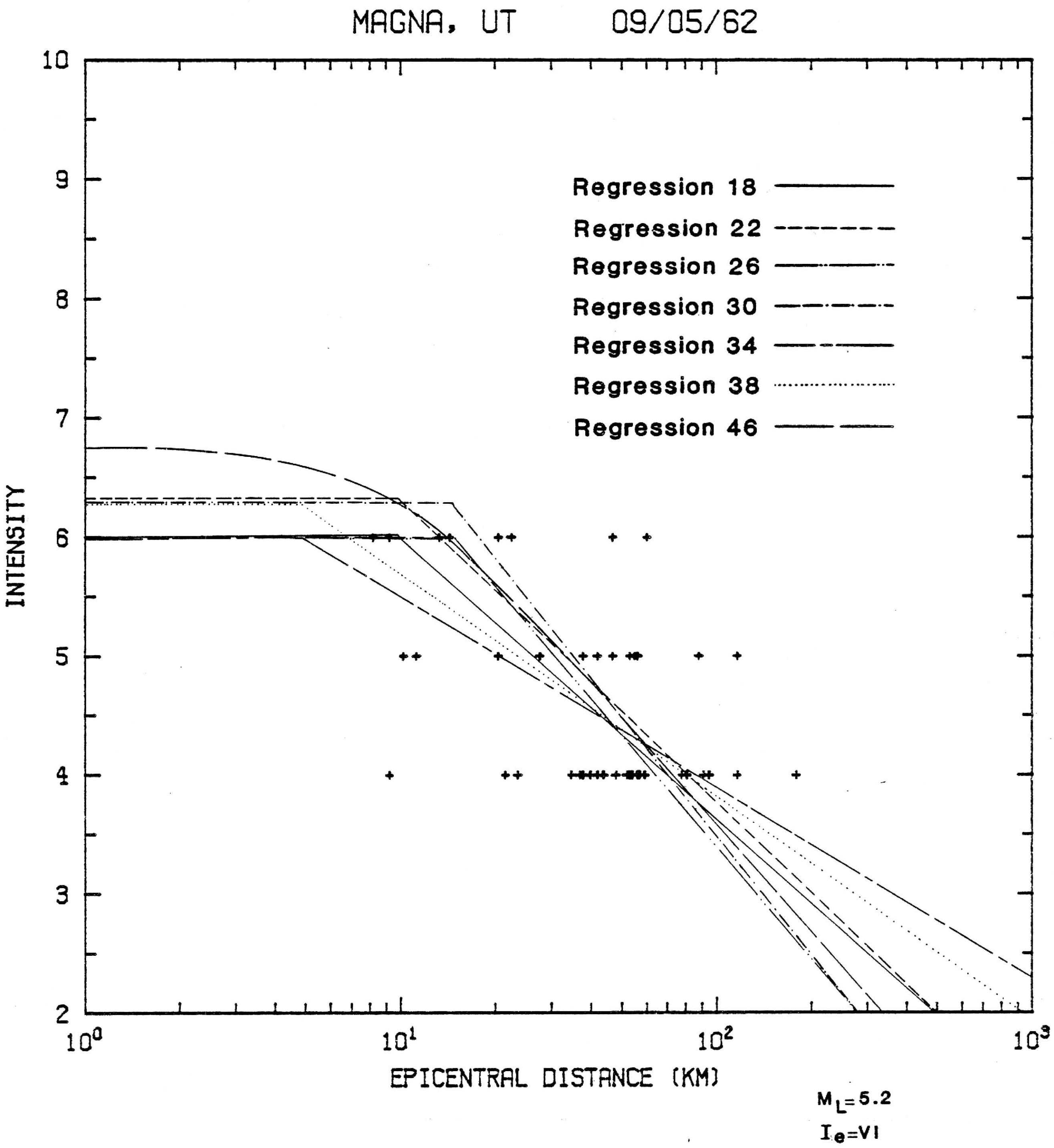

Figure 13.

Magna, Utah Earthquake Intensities vs. Distance with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

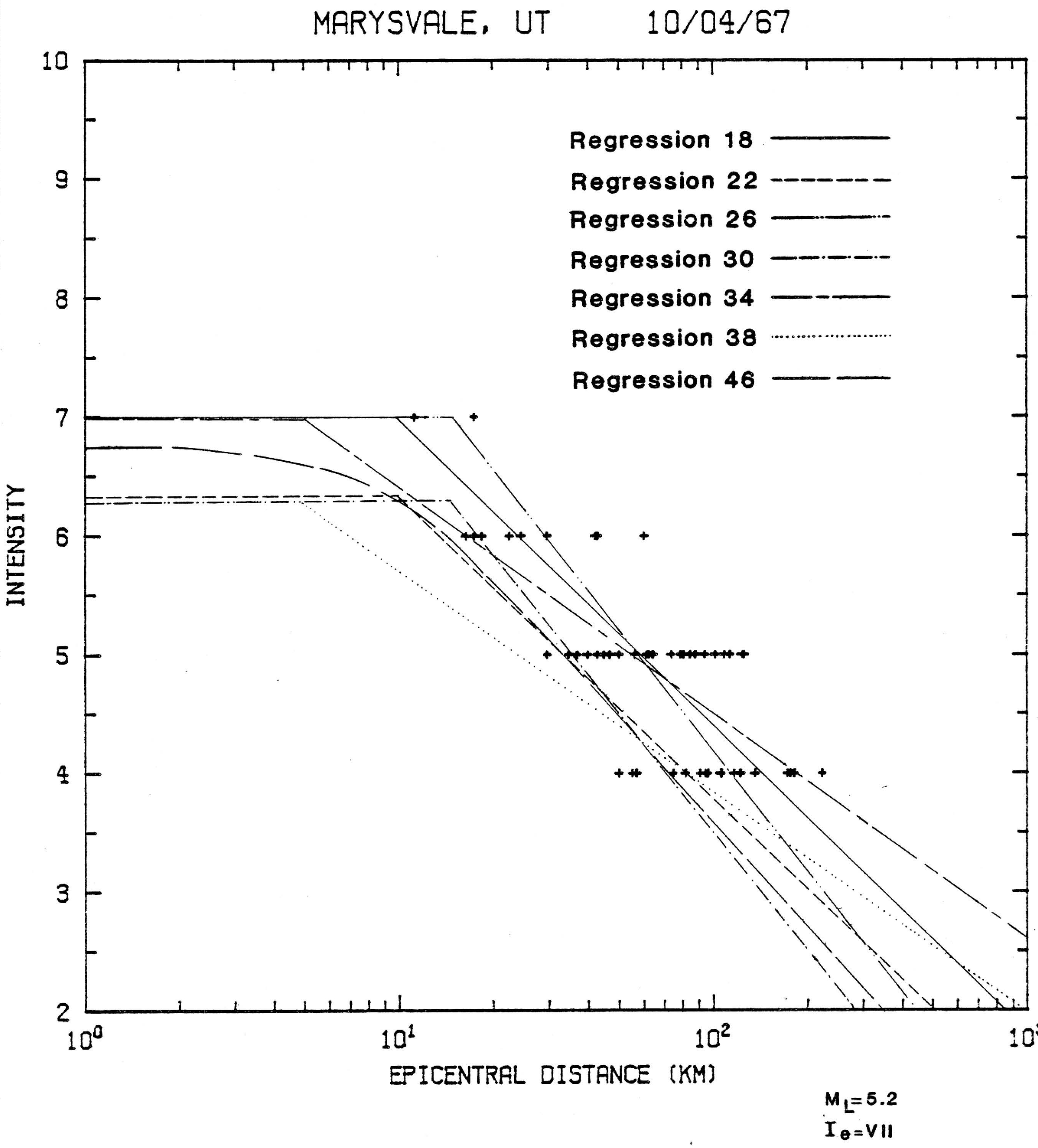

Figure 14.

Marysvale, Utah Earthquake Intensities vs. Distance with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

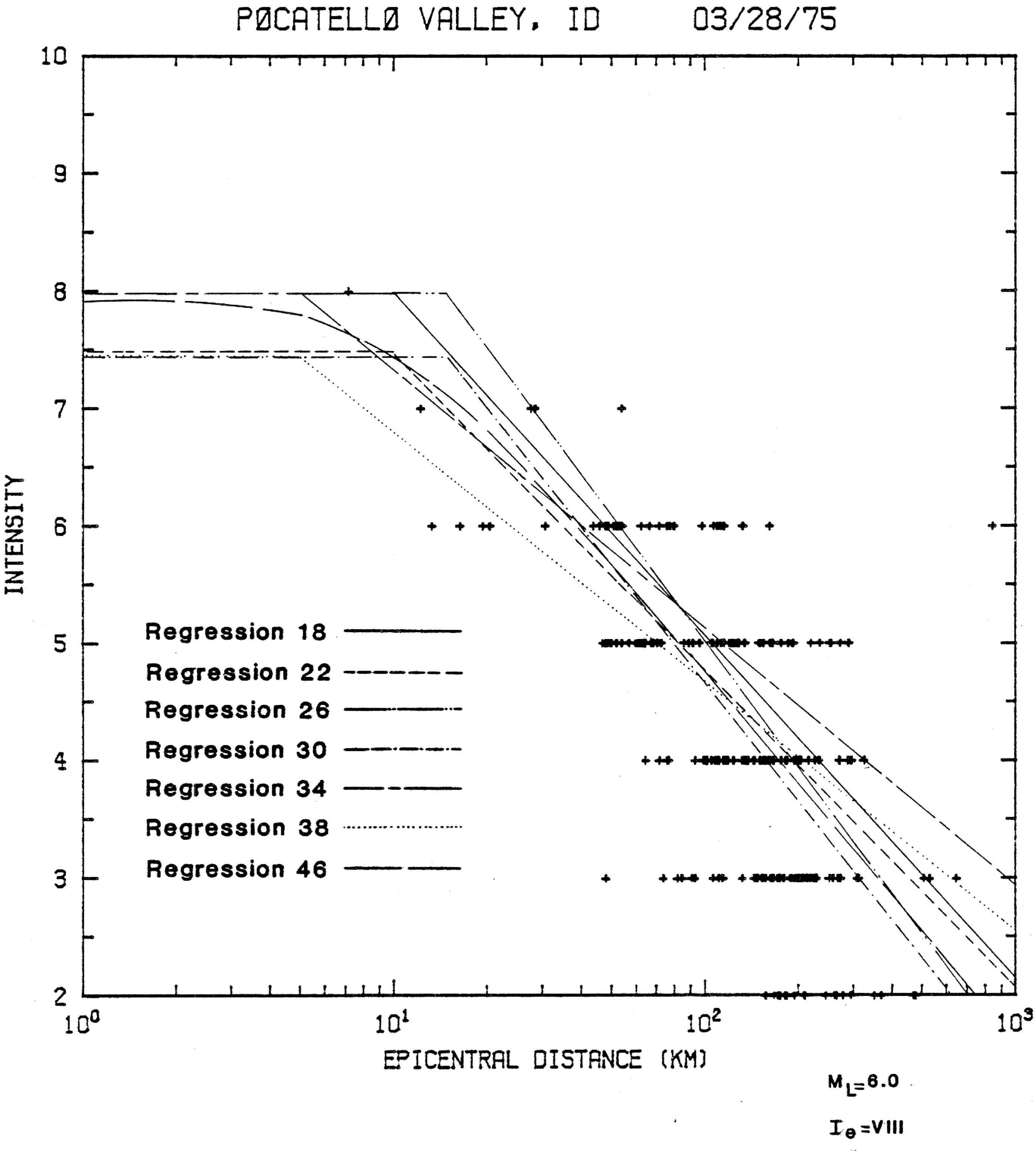

Figure 15.

Pocatello Valley, Idaho Earthquake Intensities vs. Distance with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

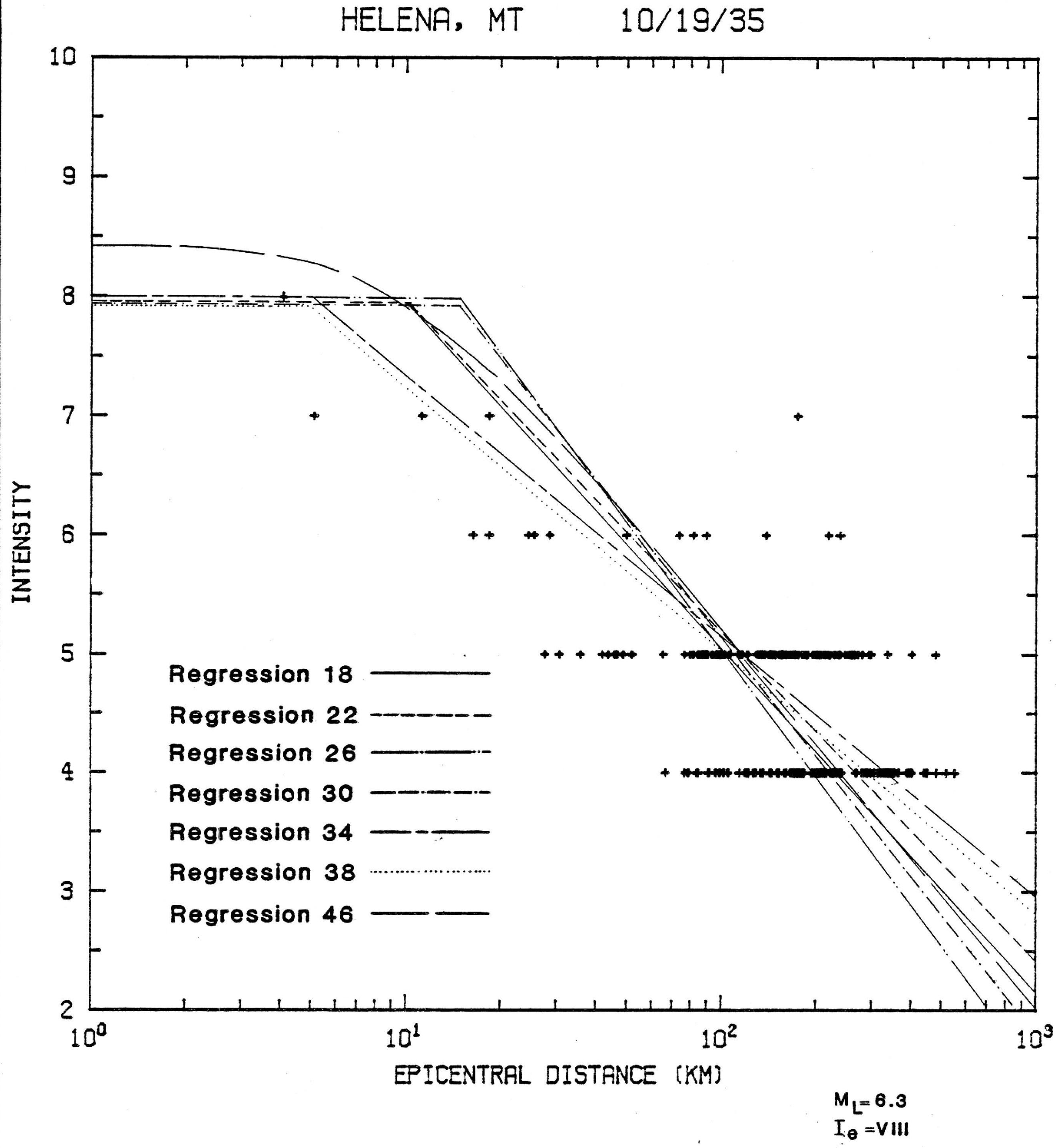

Figure 16.

Helena, Montana Earthquake Intensities vs. Distance
with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

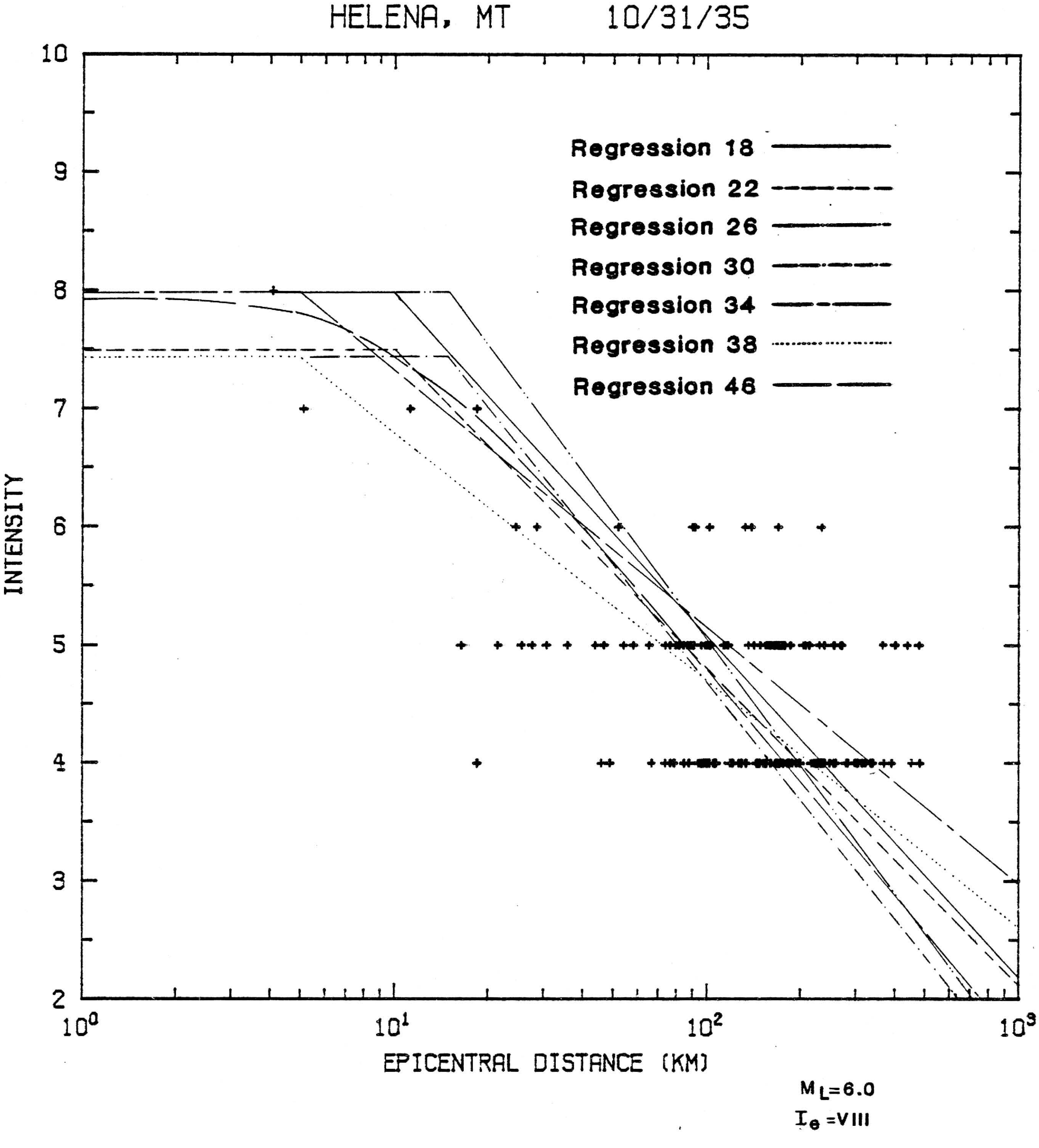

Figure 17.

Helena, Montana Earthquake Intensities vs. Distance with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

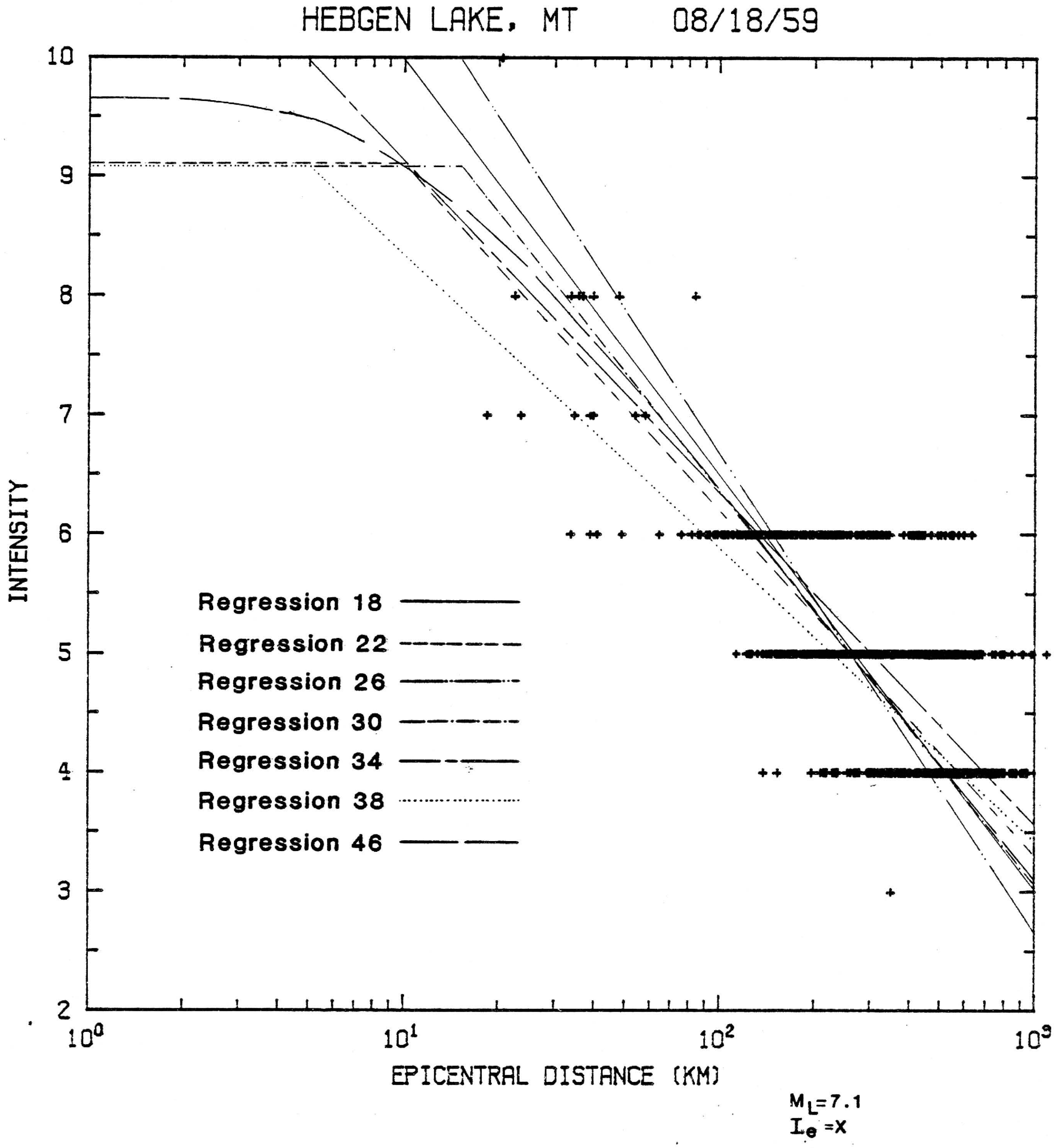

Figure 18.

Hebgen Lake, Montana Earthquake Intensities vs. Distance with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

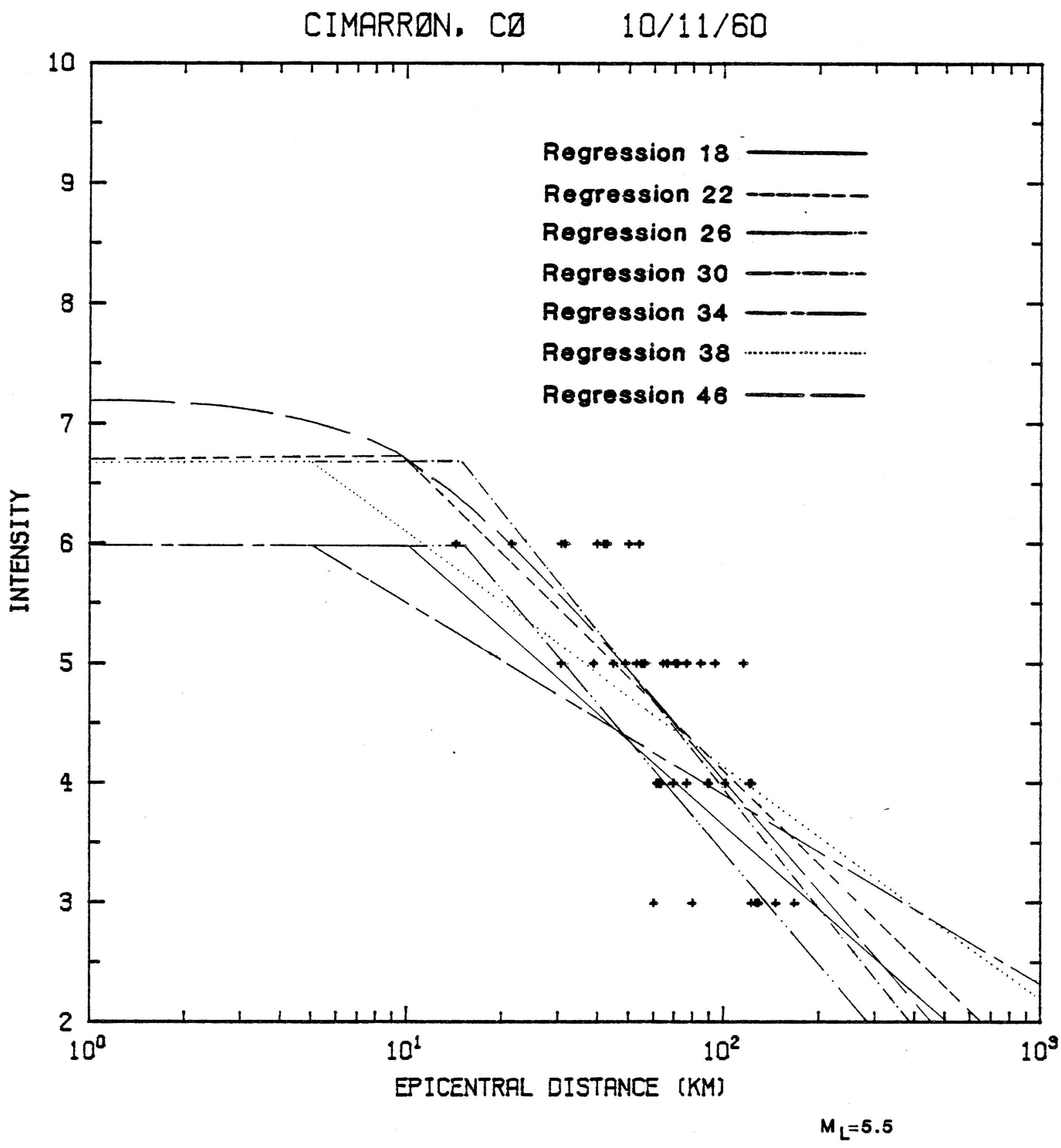

Figure 19.

Cimarron, Colorado Earthquake Intensities vs. Distance with Lines for Regressions 18, 22, 26, 30, 34, 38, and 46

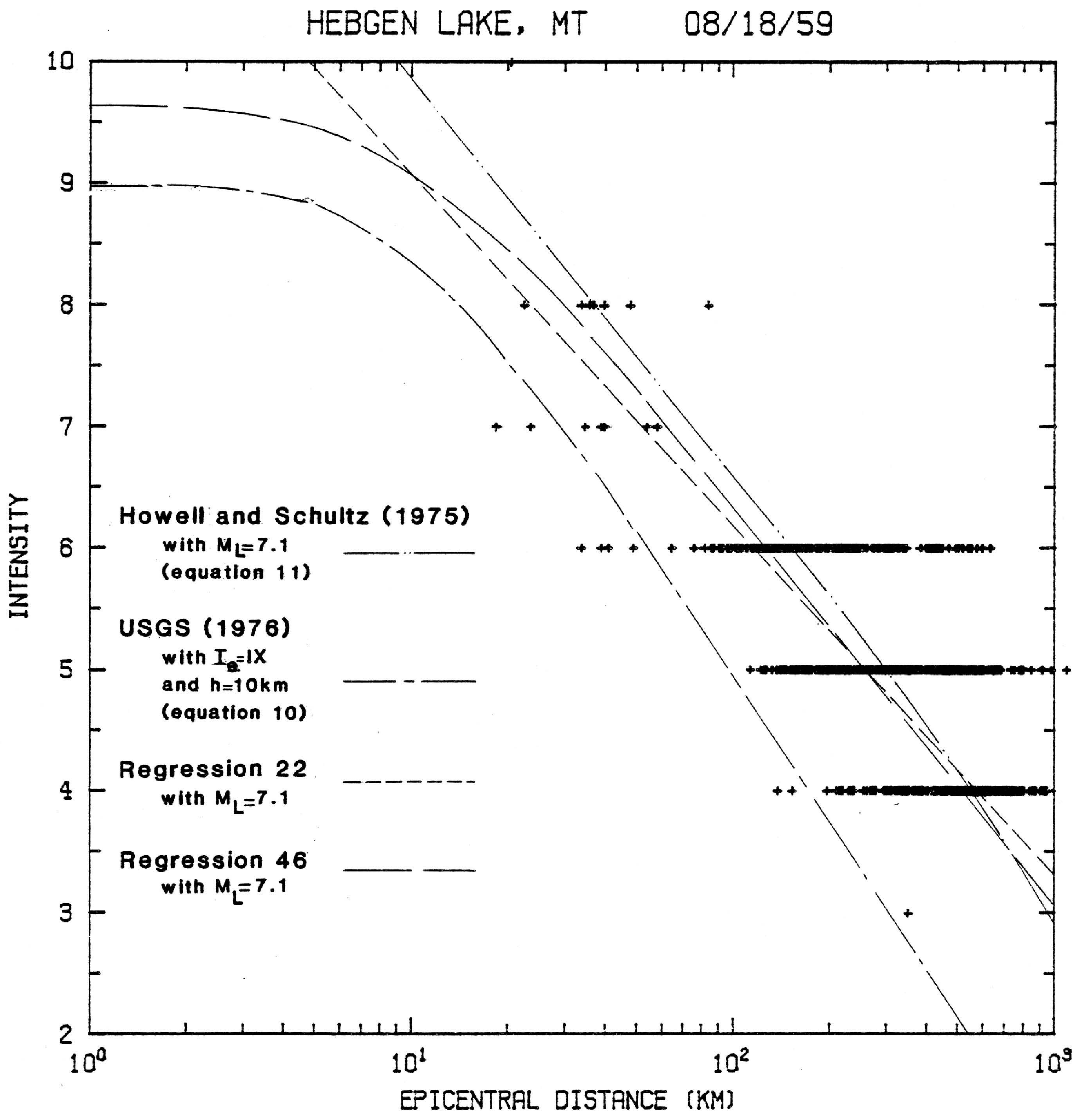

Figure 20.

Comparison of Intensity Curves for Hebgen Lake, Montana

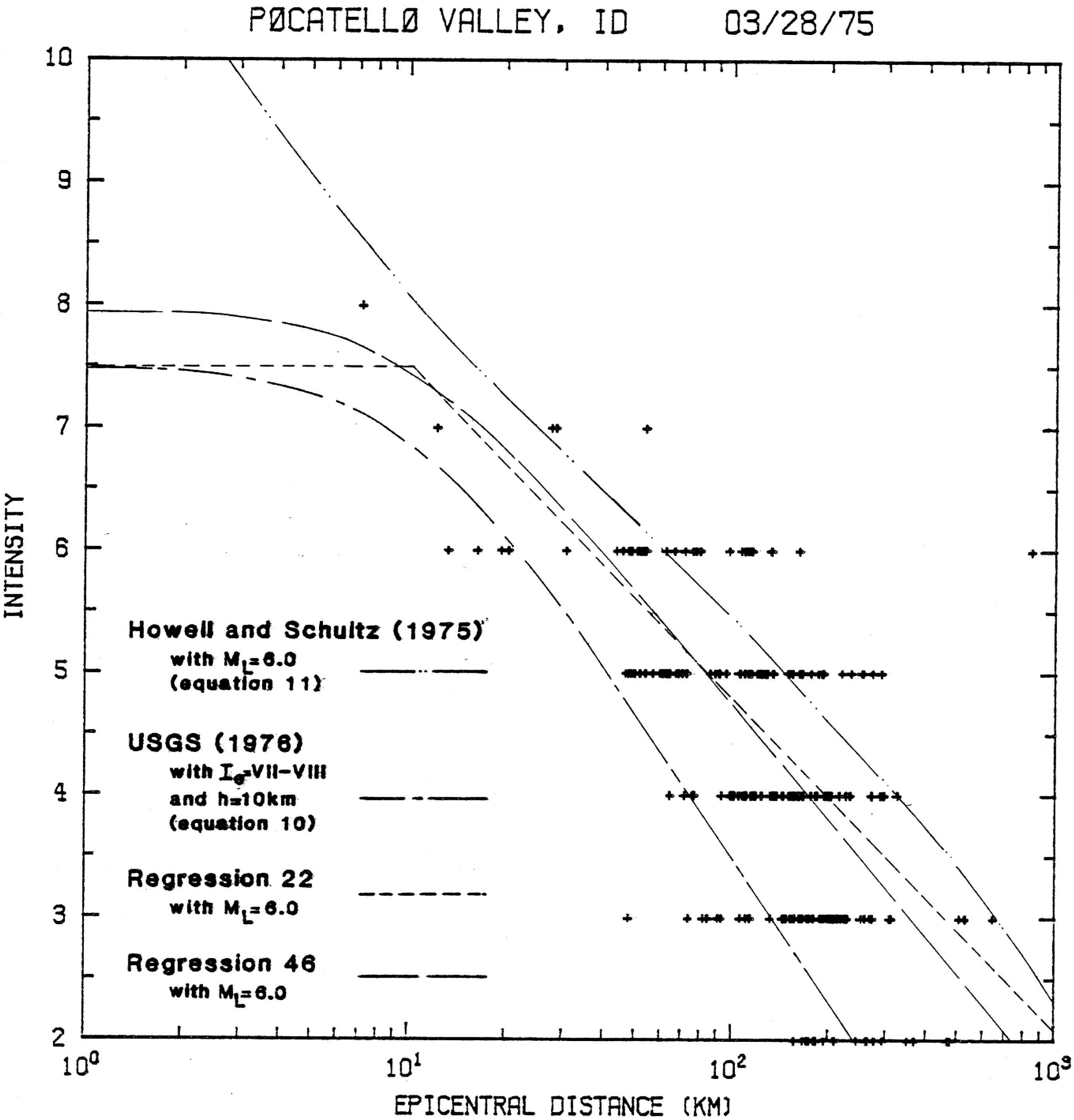

Figure 21.

Comparison of Intensity Curves for Pocatello Valley, Idaho

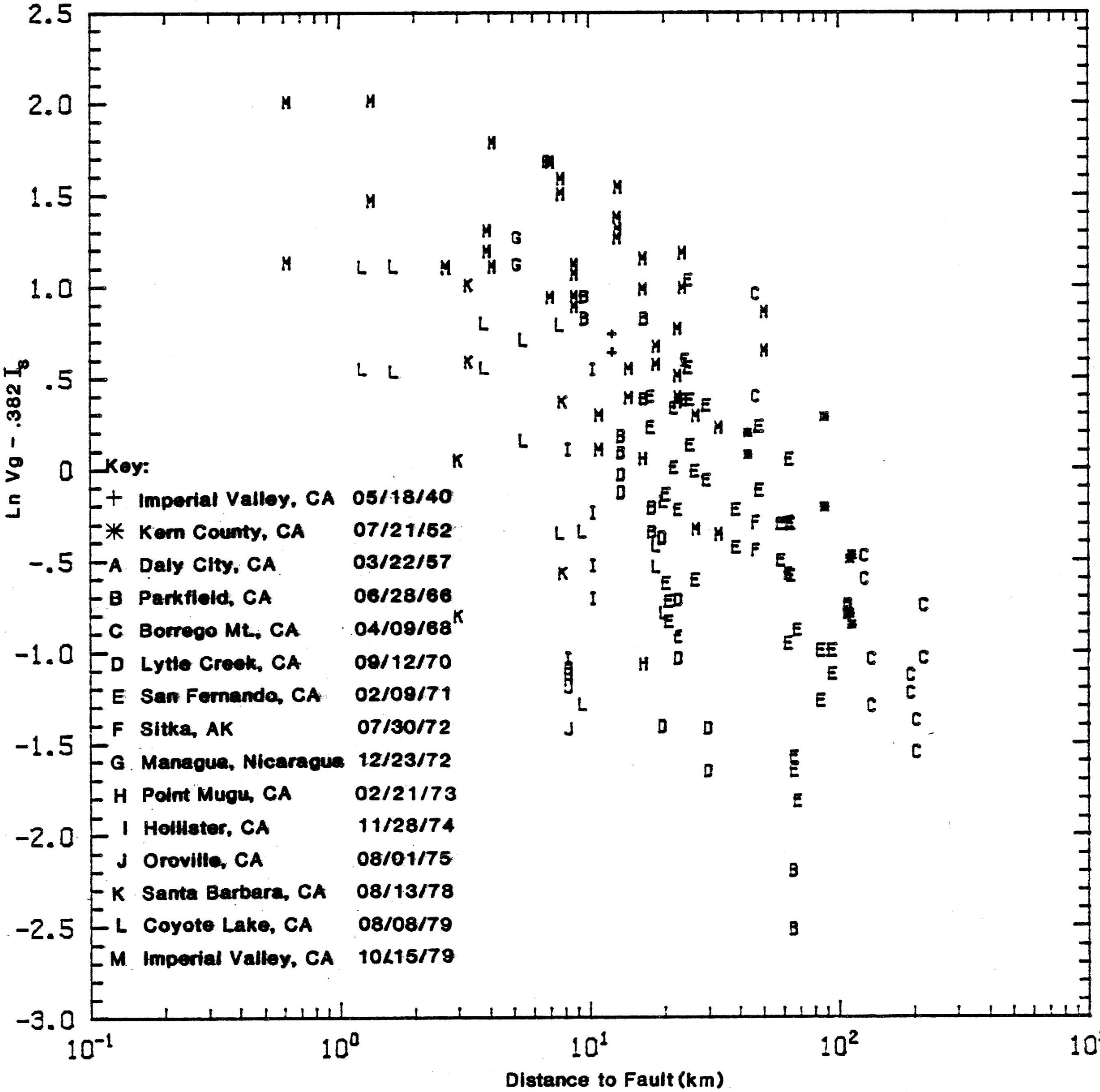

Figure 22.

Normalized Velocity vs. Distance (Horizontal Data Only)

Dames & Moore

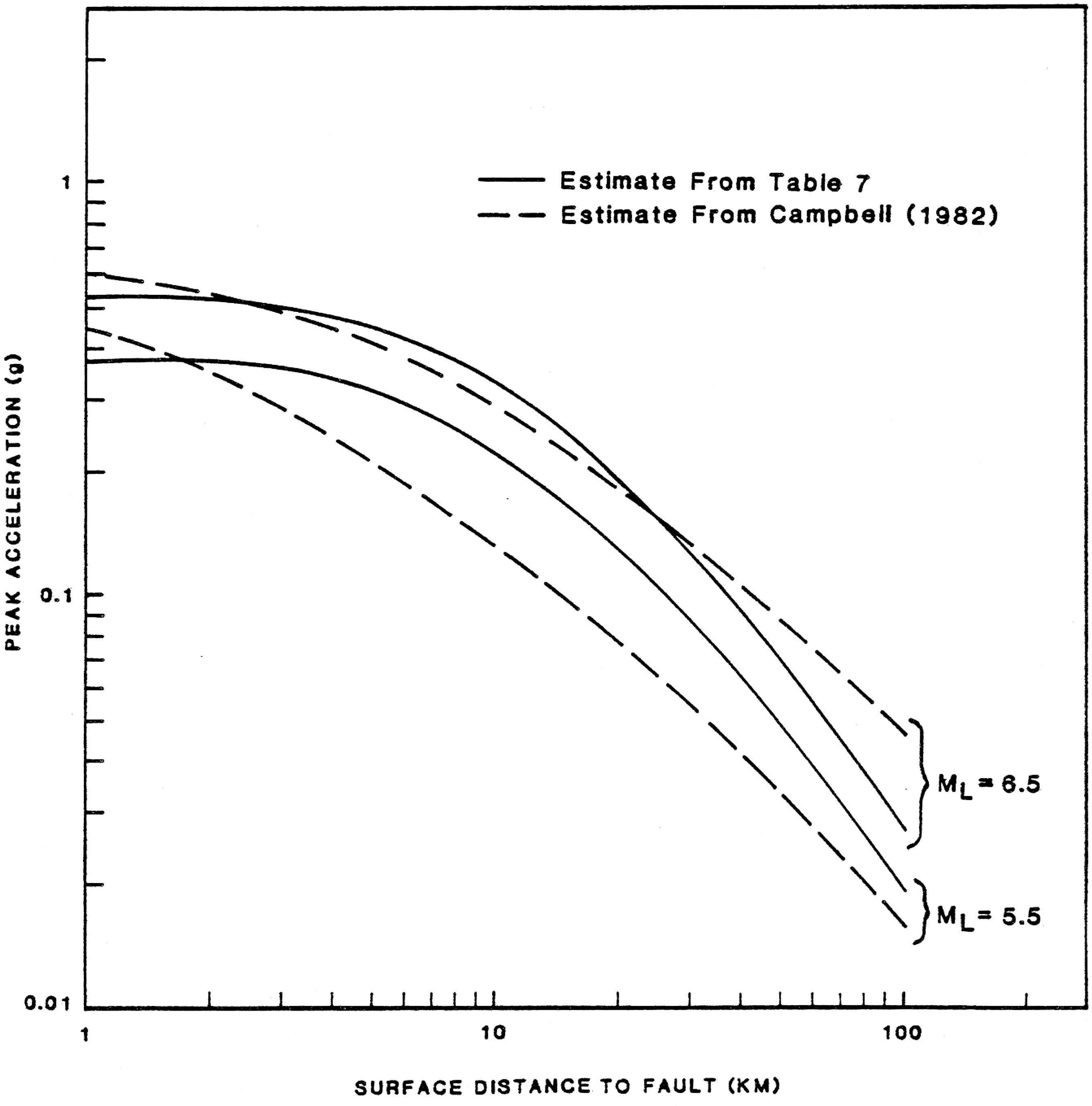

Figure 23.
Comparison of Peak Acceleration Estimates for Utah

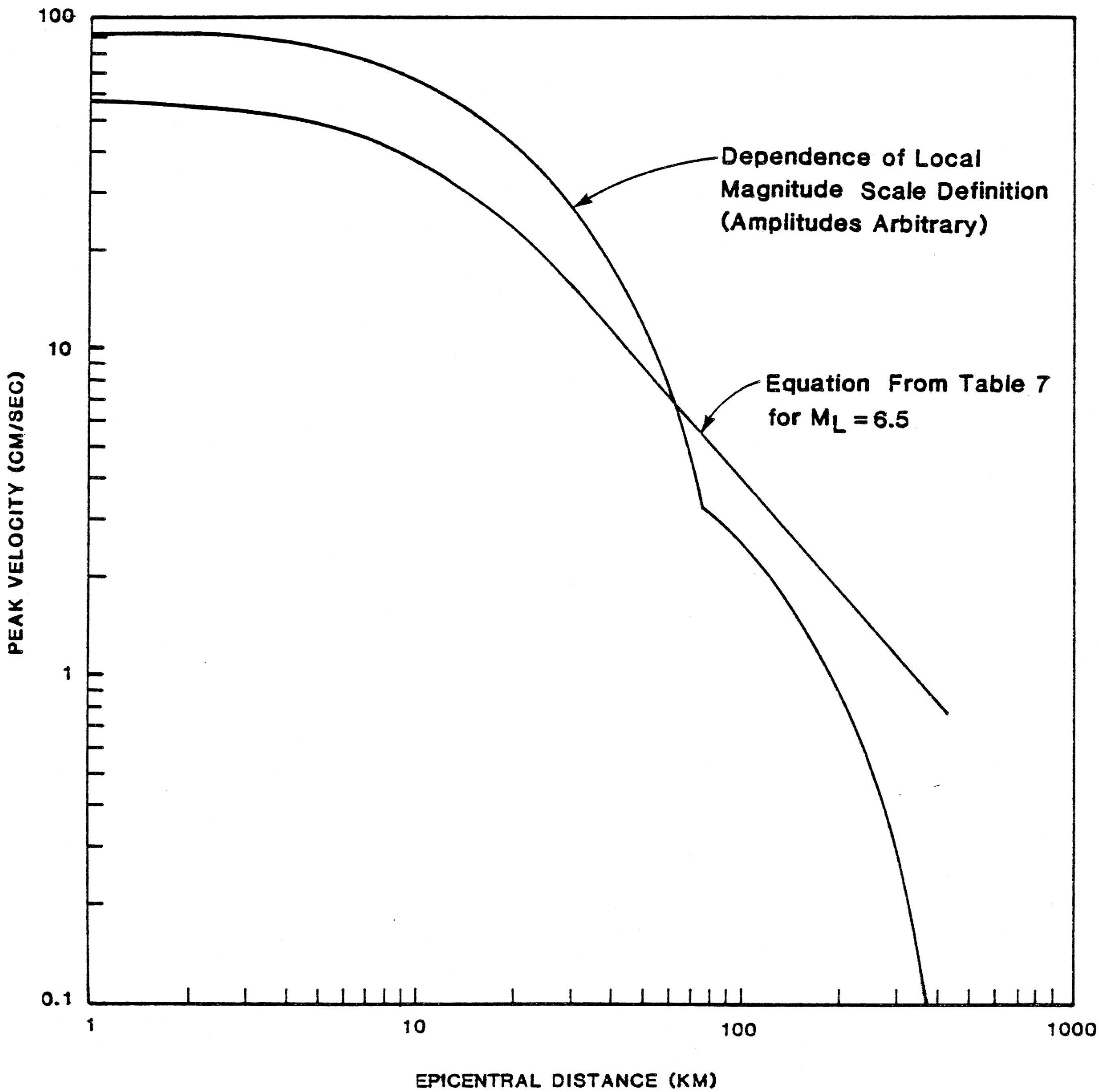

Figure 24.

Comparison of Peak Velocity Estimates and California Local Magnitude Scale Definition

Dames & Moore

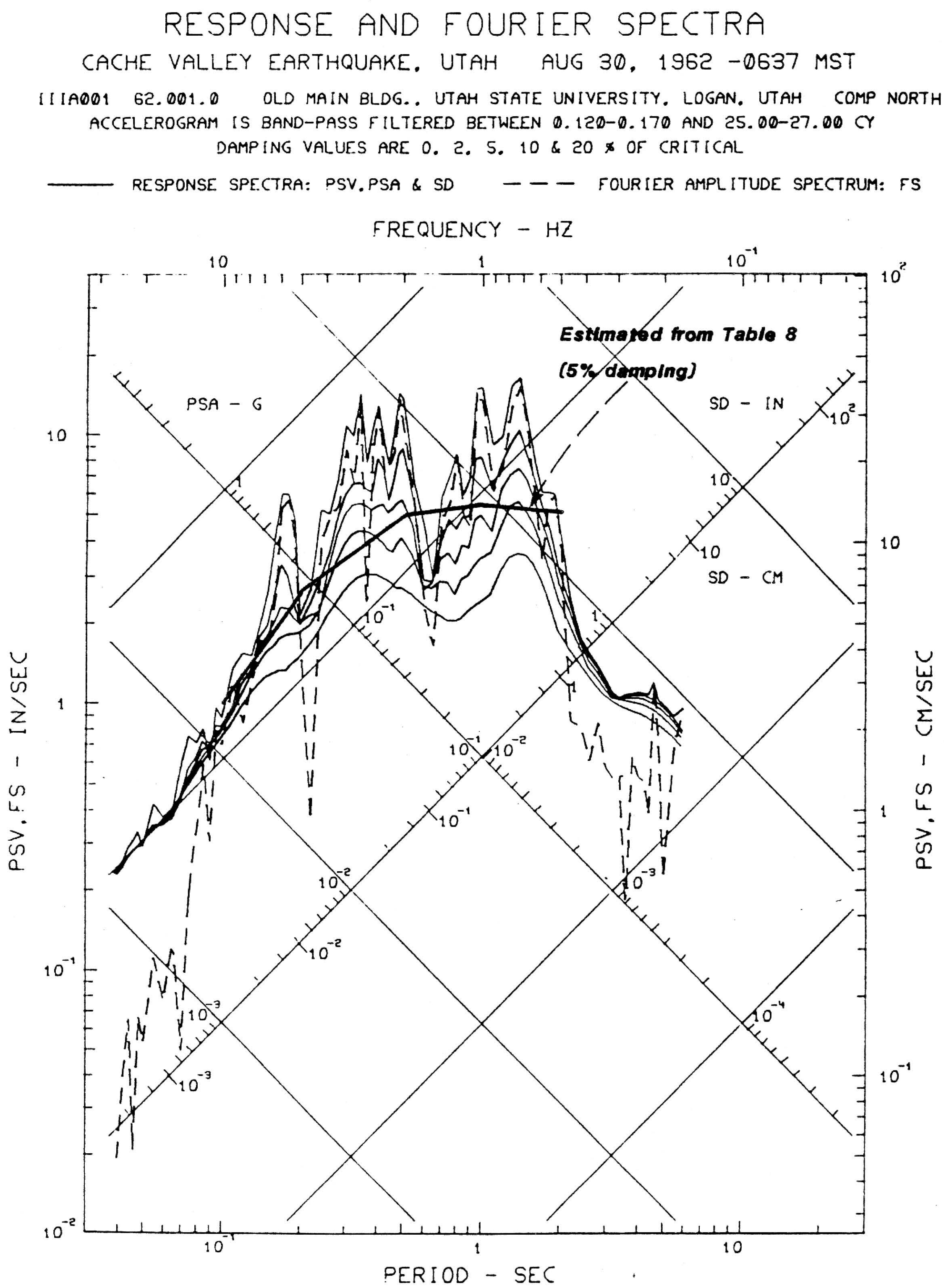

Figure 25.

Cache Valley Earthquake Record (N-S Component)

(After Smith and Lehman, 1979)

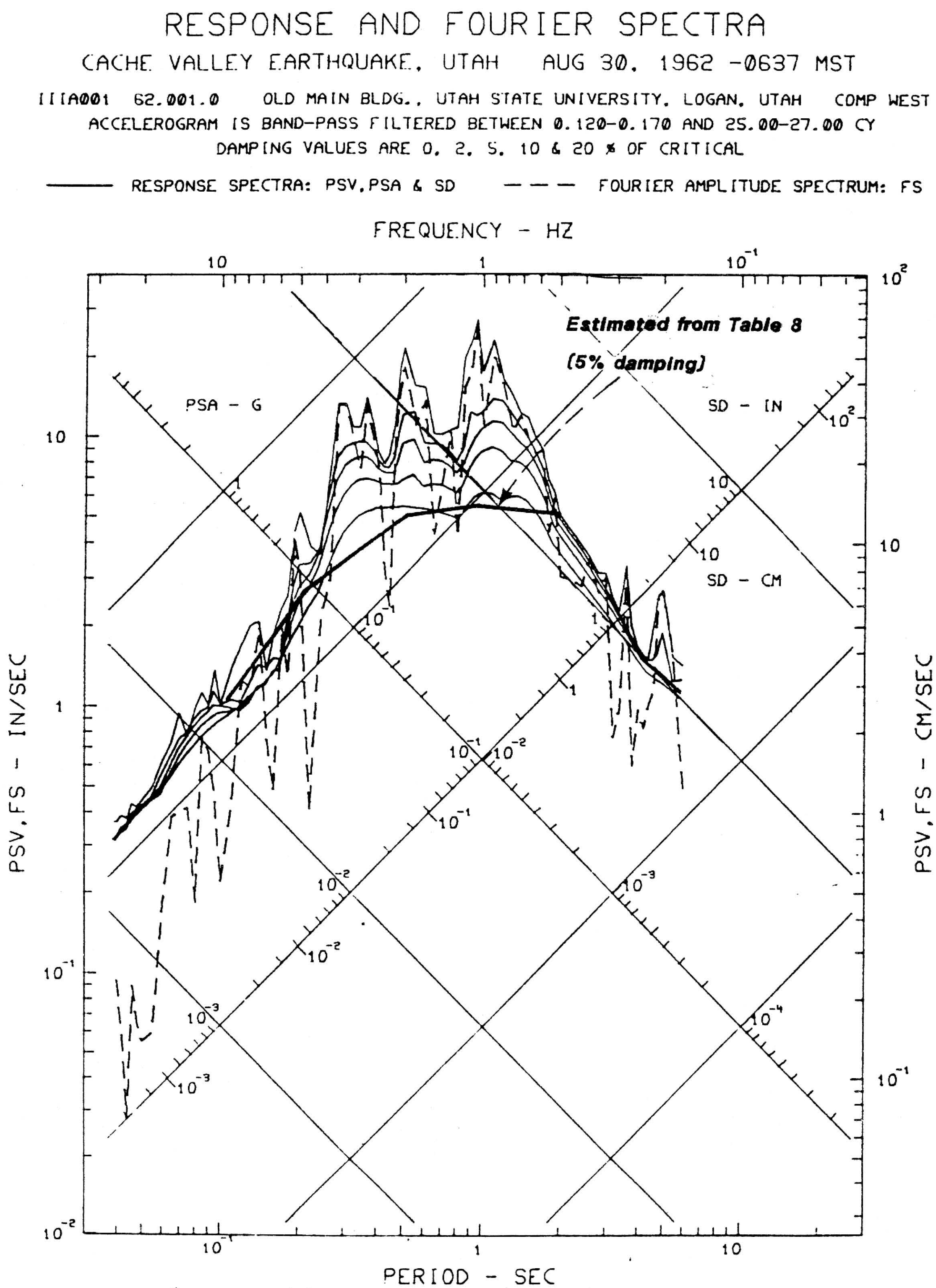

Figure 26.

Cache Valley Earthquake Record (E-W Component)

(After Smith and Lehman, 1979)

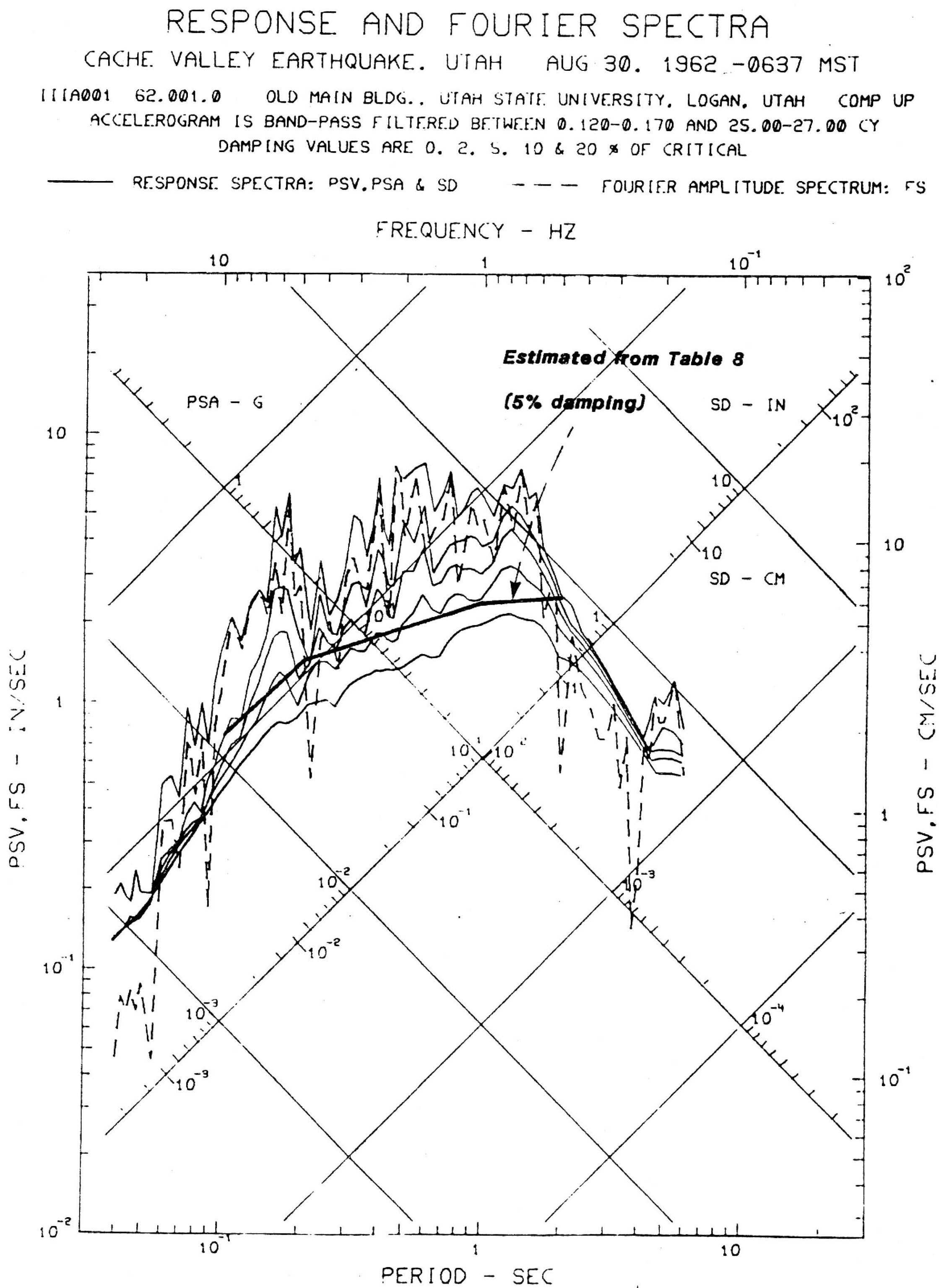

Figure 27.

Cache Valley Earthquake Record (Vertical Component)

(After Smith and Lehman, 1979)