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ATTENUATION OF INTENSITIES IN IRAN

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ABSTRACT

The isoseismal maps for twelve earthquakes in different parts of Iran were analyzed to study the attenuation of intensities with distance. The attenuation is quite sensitive to the selection of epicentral intensities, I_0, for the earthquakes considered. A graphical method presented in this paper provides a consistent basis for the estimation of epicentral intensities, for different earthquakes, from the intensity-distance plot. The attenuation relations were derived by using an iterative least squares fit procedure, wherein an initial approximate estimate of epicentral intensity for each earthquake is successively improved. The isoseismal maps for a number of earthquakes are elongated in the direction of local structural trend/causative faults. Therefore, three different attenuation relations were derived.

$$I(R) = I_0 + 6.453 - 0.00121\,R - 4.960\log(R + 20) \qquad R < 120 \text{ km}$$
average attenuation

$$I(R) = I_0 + 4.824 - 0.00548\,R - 3.708\log(R + 20) \qquad R < 160 \text{ km}$$
parallel to the isoseismals

$$I(R) = I_0 + 8.729 + 0.01158\,R - 6.709\log(R + 20) \qquad R < 110 \text{ km}$$
transverse to the isoseismals

where $I(R)$ is the intensity at a distance R (km) from the epicenter.

The average attenuation of intensities in Iran is slightly more rapid than the San Andreas province attenuation.

INTRODUCTION

In deterministic seismic risk analysis, once a causative fault or a tectonic province has been identified and a corresponding design earthquake has been associated with it, the problem reduces to determining the ground motion at a site (usually located at some distance from the fault or tectonic province) resulting from the occurrence of such an earthquake. This involves an understanding of the attenuation properties of the transmitting medium from the earthquake focus to the site under consideration.

The probabilistic risk analysis procedures of Algermissen and Perkins (1976) and McGuire (1976) make use of relations developed for the attenuation of intensities with distance.

The paucity of strong motion instrumentation in most seismic regions of the world makes it essential to understand the attenuation of intensities for a meaningful seismic risk analysis. Even though assignment of intensities and preparation of isoseismal maps is a subjective process, the intensities are usually the only seismic data which may be related to structural damage.

Attenuation characteristics and falloff of intensity with distance varies from region to region. Various authors, such as Brazee (1972), Cornell and Merz (1974), Howell and Schultz (1975), Gupta and Nuttli (1976), and Gupta (1976), have developed

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attenuation relations for specific regions and have emphasized the need for developing independent relations for other parts of the world.

Data

To determine the attenuation relation, isoseismal maps for 12 earthquakes occurring in different parts of Iran were analyzed. The locations of these earthquakes are shown in Figure 1. The isoseismal maps are presented in Figures 2 to 4. The pertinent information on these earthquakes and the isoseismal references are

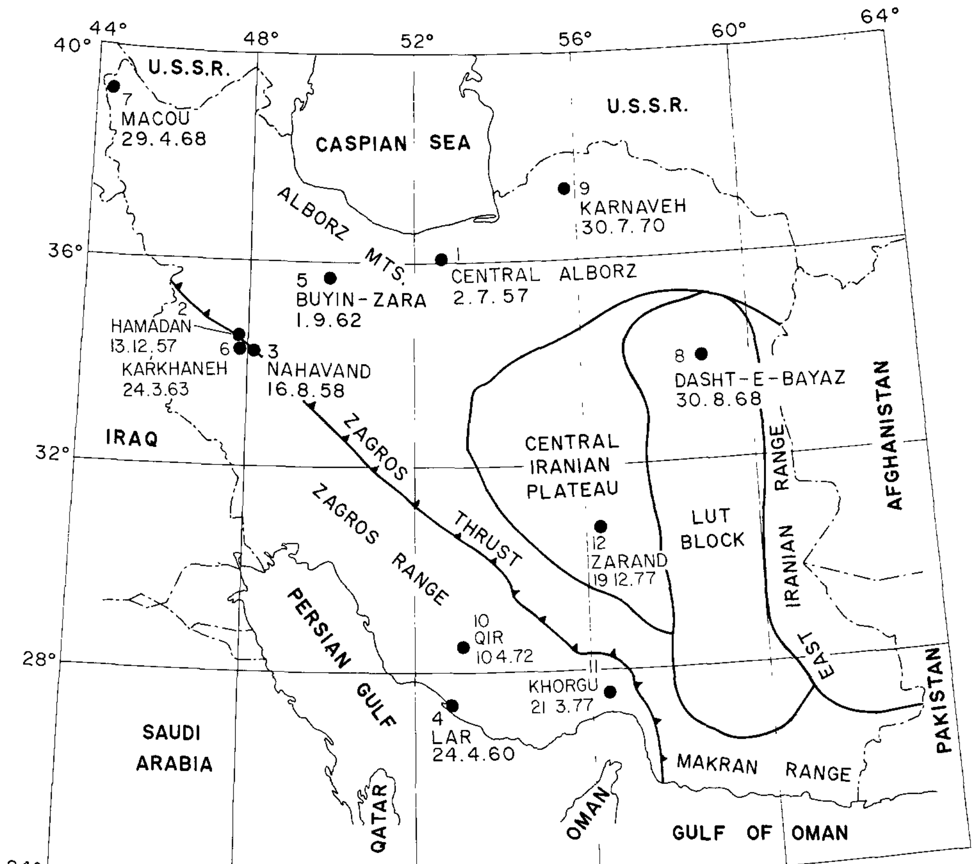

FIG. 1. An index map showing location of earthquakes (solid circles) selected for attenuation study. For each earthquake, an identification number, name of the nearest locality, and date of occurrence is given. Large scale regional tectonic features are also displayed on the map.

summarized in Table 1. The left hand column contains an identification number for each earthquake, which is shown on Figure 1.

The areas covered by each isoseismal line were accurately measured by using a planimeter. From the measured areas, radii for an equivalent circular region were calculated.

Three isoseismal maps were available for the Hamadan earthquake of December 13, 1957. The map prepared by Ambraseys *et al.* (1973) was used in this study. This map, prepared in coauthorship with F. Peronaci, was considered to be an improvement over the map published earlier by Peronaci (1958). The map published by Hagiwara and Naito (1959) is rather speculative.

For event 5, the Buyin-Zara earthquake of September 1, 1962, isoseismal maps from four different sources were available. The intensity-distance curve derived from the isoseismal map published by the Institute of Geophysics (1963) shows a very sharp falloff of intensity with distance. The isoseismal map prepared by

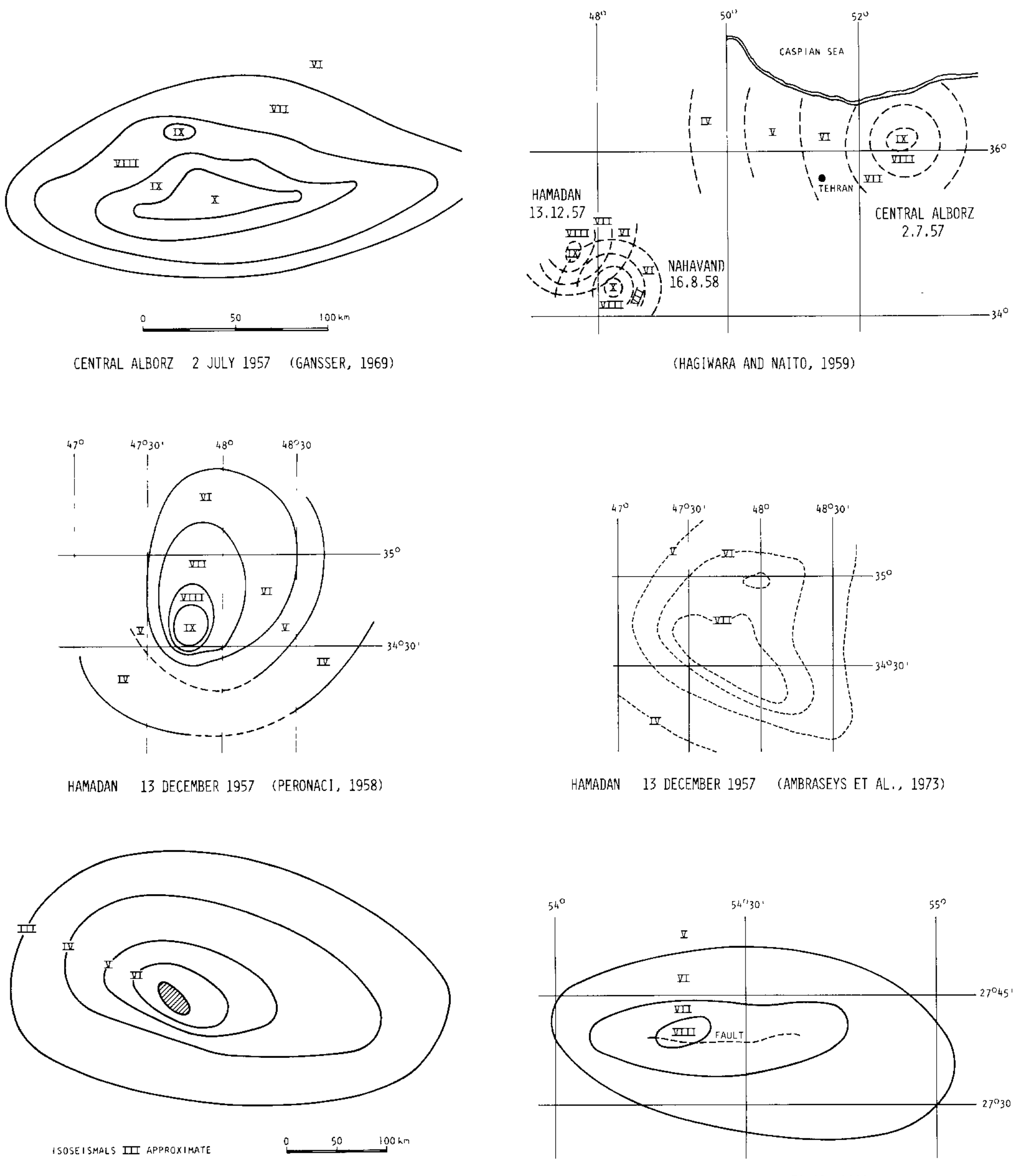

FIG. 2. Isoseismal maps of earthquakes in Iran.

Mohajer and Pierce (1963) shows an unusually regular pattern of isoseismal lines. The intensity maps prepared by Ambraseys (1963) and Omote *et al.* (1965) appear more realistic. In this study, intensity-distance data derived from Ambraseys' (1963) map were used.

For the Dasht-e Bayaz earthquake of August 31, 1968 (event 8) the average of the equivalent radii for the corresponding intensities obtained from the isoseismal maps presented by Bayer *et al.* (1969) and the Institute of Geophysics (1969), were used. The intensity-distance curves derived from the two isoseismal maps are quite close to each other.

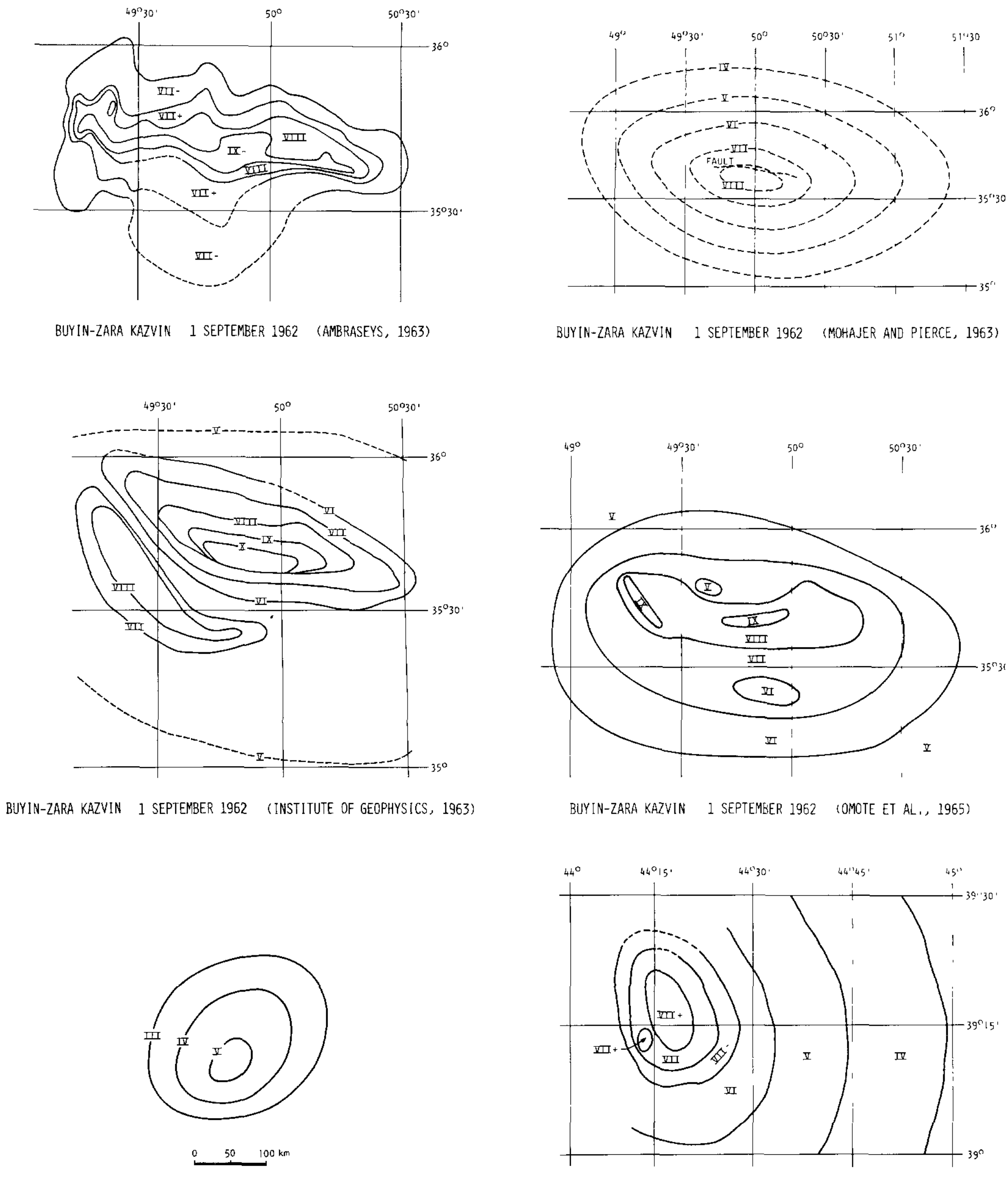

FIG. 3. Isoseismal maps of earthquakes in Iran.

A plot of MM intensity, I, versus distance, R (or equivalent circle radius of the isoseismal line concerned), is presented in Figure 5. It is observed that even though the magnitude of event 1 is only $7\frac{1}{4}$ to $7\frac{1}{2}$ (PAS), its epicentral intensity, by extrapolation, would be estimated as more than 11.0. This earthquake caused considerable damage to mountain villages and resulted in landslides. The isoseismals were constructed on the basis of visible destruction, partly on the basis of data collected by the staff of Alexander Gibbs Company (Gansser, 1969). It appears that the intensities based mainly on geological effects, such as landslides, rockfalls, ground rupture, etc., may be overrated on the Modified Mercalli scale. For event 8, the intensity-distance curve below intensity VIII is generally parallel to the curves for other events. The curve is very steep above intensity VIII. The earthquake occurred in the northern part of the Lut block and was accompanied by a fresh fault

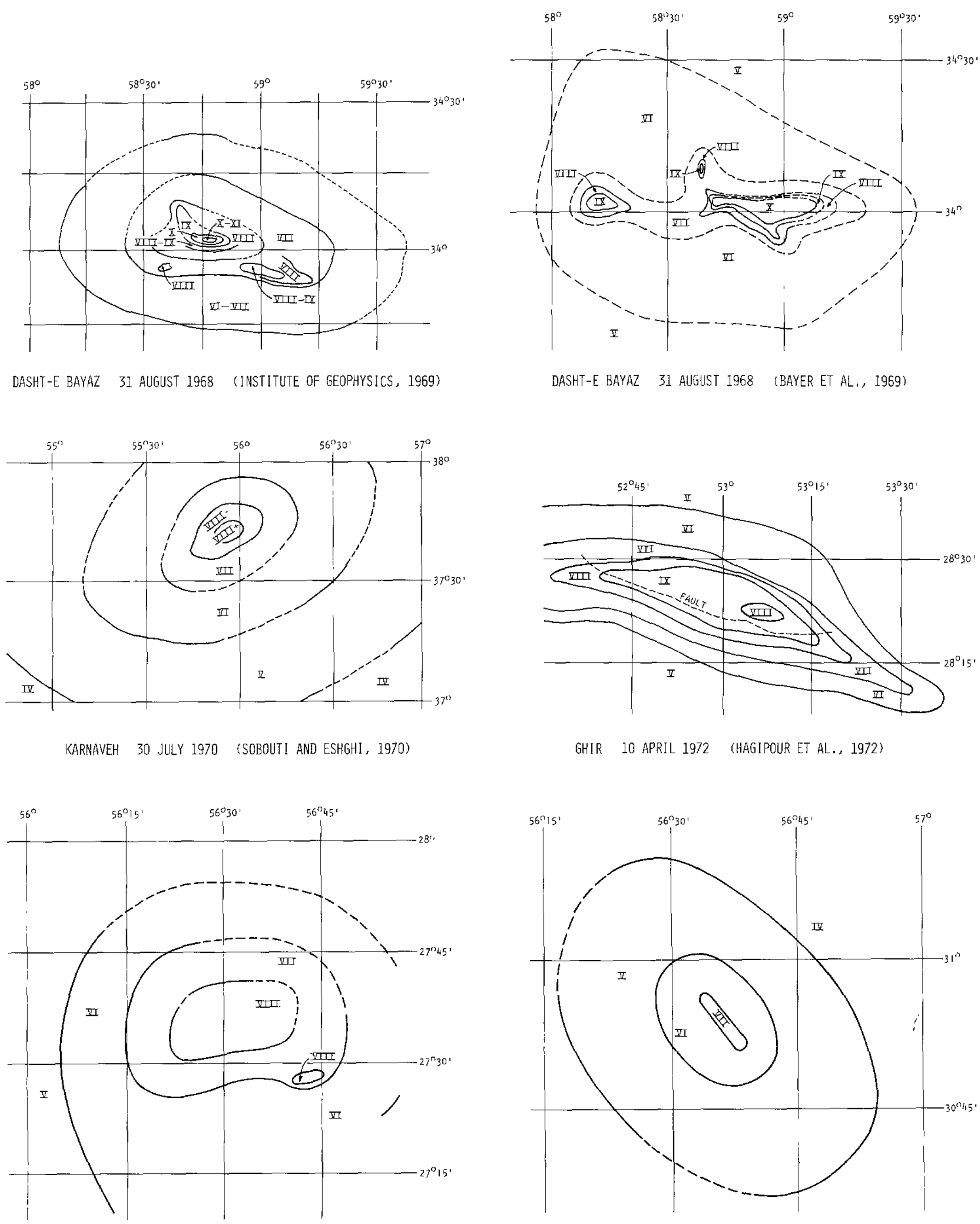

Fig. 4. Isoseismal maps of earthquakes in Iran.

TABLE 1

EARTHQUAKES SELECTED FOR ATTENUATION STUDY

No.	Date	Origin Time h	m	s	Latitude °N	Longitude °E	Location	Magnitude[1]	G - R Computed I_o	Graphical I_o	Calculated[2] I_o	Isoseismal Reference
1	2 July 1957	00	42	22.0	36.1	52.7	Central Alborz	7¼-7½ PAS	9.57			Gansser (1969)
2	13 Dec. 1957	01	45	05.0	34.6	47.8	Hamadan	7¼ PAS	9.38	8.46	8.63	Peronaci (1958), Hagiwara and Naito (1959), Ambraseys et al. (1973)
3	16 Aug. 1958	19	13	44.0	34.3	48.2	Nahavand	6-3/4 PAS	8.63	8.03	8.16	Hagiwara and Naito (1959), Ambraseys and Moinfar (1974a)
4	24 Apr. 1960	12	14	26.0	27.7	54.4	Lar	6.0 PAS	7.50	8.19	8.37	Afshar (1960)
5	1 Sept. 1962	19	20	38.5	35.6	49.8	Buyin-Zara Kazvin	7¼ PAS	9.38	9.21	9.39	Inst. Geophysics (1963), Ambraseys (1963), Mohajer and Pierce (1963), Omote et al. (1965)
6	24 Mar. 1963	12	44	00.5	34.3	47.8	Karkhaneh	5.3	6.45	7.17	7.28	Ambraseys and Moinfar (1974b)
7	29 Apr. 1968	17	01	57.6	39.3	44.3	Macou	5.3	6.45	7.48	7.67	Nabavi (1970)
8	31 Aug. 1968	10	47	37.4	34.0	59.0	Dasht-e Bayaz	7.3 MS	9.45	8.74	8.93	Bayer et al. (1969), Inst. Geophys. (1969)
9	30 July 1970	00	52	19.5	37.7	55.9	Karnaveh	6-3/4 BRK	8.63	8.66	8.83	Sobouti and Eshghi (1970)
10	10 Apr. 1972	02	06	53.2	28.4	52.8	Qir	7.1 BRK	9.15	8.92		Hagipour et al. (1972)
11	21 Mar. 1977	21	18	54.2	27.6	56.4	Khorgu	7.0 PAS	9.00	8.57	8.76	Nowroozi et al. (1977)
12	19 Dec. 1977	23	34	34.2	30.8	56.3	Zarand	5.5	6.75	6.90	7.09	Mohajer-Ashjai et al. (1978)

Notes:

[1] In the magnitude column, the references are abbreviated as, PAS - California Institute of Technology, Pasadena; BRK - Seismographic Station, University of California, Berkeley; MS - surface wave magnitude determined by U.S. Geological Survey using IASPEI formula; when no reference is given, body wave magnitude determined by U.S. Geological Survey is assumed.

[2] Obtained as a byproduct in the derivation of equation (3).

rupture with a 4.6 *m* left lateral displacement (Niazi, 1968). It seems that intensities IX and above were overrated because of surface faulting effects and, therefore, were not considered in the following analysis.

Event 10 shows an anomalously high attenuation in contrast to the attenuation indicated by events in other parts of Iran and by events 4 and 11 which occurred in the same geological and tectonic environment as event 10. It is possible that this anomalously high attenuation for event 10 is due to the subjectivity involved in assigning intensity values at different observation points and in drawing an isoseismal map, rather than due to a difference in attenuation properties of the ground in the vicinity of the epicenter.

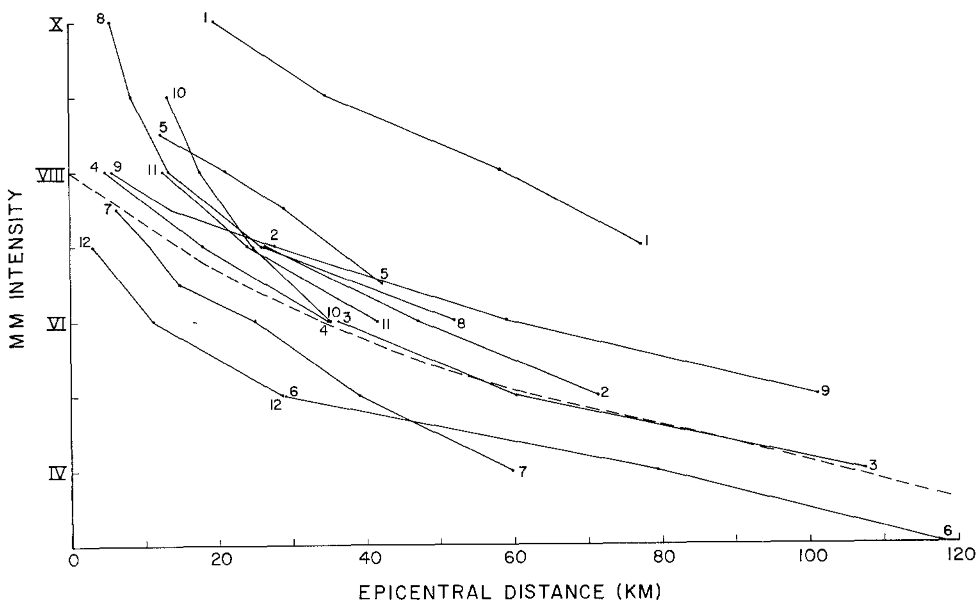

FIG. 5. Intensity-distance plot for different earthquakes. The earthquake identification numbers correspond to those given in Table 1. The dashed curve is drawn by following the general parallelism of the intensity-distance curves for different earthquakes. The level of this curve is arbitrarily set at intensity VIII, at distance zero.

ANALYSIS

For describing the attenuation of intensity with distance, a number of authors (Howell, 1974; Howell and Schultz, 1975; Gupta and Nuttli, 1976; Gupta, 1976; Bollinger, 1977) have recently used relations of the form

$$I(R) = I_0 + a + bR + c \log R \qquad (1)$$

where $I(R)$ is the site intensity at a distance R from the epicenter of an earthquake of epicentral intensity I_0, and a, b, and c are constants appropriate to the region under consideration.

To derive an attenuation relation from the intensity-distance data for a number of earthquakes, the parameters a, b, and c in equation (1) were determined by a least-squares fit between $I - I_0$ and R. Using these values of a, b, and c in equation (1), new average epicentral intensities for each earthquake were determined by calculating I_0 from various isoseismal intensities and corresponding R. Using these values of I_0 for different earthquakes, which were considered to be an improvement over the previous I_0 set, revised values of a, b, and c were determined by a least-squares fit between $I - I_0$ and R. The procedure was repeated until no significant improvement in the standard error of $I - I_0$ in the least-squares analysis could be made.

However, it may be noted that because of the constant term a in equation (1), the level, though not necessarily the shape, of the attenuation curve is quite sensitive to the initial I_0 set selected for the analysis. The empirical relation will show a high rate of attenuation if the initial I_0 set contains systematically high values of I_0 for different earthquakes, and a low rate of attenuation if the initial I_0 set contains systematically low values of I_0. Two methods were used to estimate the initial I_0 set.

Graphical. Following the general parallelism of the intensity lines, a smooth curve, shown by the dashed line in Figure 5, passing through intensity VIII at zero distance (ordinate line) was drawn. Using this curve as a reference, for a particular earthquake, I_0 at each isoseismal intensity point was estimated and then averaged.

Computational. The epicentral intensity, I_0, for each earthquake was estimated by using Gutenberg and Richter's (1956) relation

$$M = 1 + (\tfrac{2}{3})I_0 \tag{2}$$

where M is the earthquake magnitude.

Both, graphically estimated and G-R computed [using equation (2)], I_0 sets as initial approximations are given in Table 1. Although intensities are defined in integral units as Roman numerals, for computational purpose, in Tables 1 to 3 and throughout the text, these are given in decimal units. This should not be interpreted to imply an accuracy of 0.01 units in intensity, but is an obvious necessity when it is realized that a change in intensity from VII to VIII corresponds in the mean to a change in acceleration from 0.13 to 0.26 g (Trifunac and Brady, 1975).

With each of the two I_0 sets, attenuation relations for the following combinations of earthquake data were considered: (1) all of Iran, (2) all of Iran excluding event 10 in recognition of the anomalously high attenuation for this event, and (3) region northeast of the Zagros Ranges, i.e., excluding events 4, 10, and 11. Data for event 1 were not considered in the derivation of attenuation relations because, as discussed earlier, the quality of the isoseismal map of this earthquake is suspect.

The empirical attenuation relations derived from different combinations of earthquake data, using graphically estimated as well as G-R computed I_0 sets as initial approximation, are summarized in Table 2. Numerically, all of these relations give almost identical results. At a distance of 25 km, for example, the spread in $I - I_0$ values computed from these equations is 0.06 intensity unit. Thus, the graphical and G-R computed I_0 sets as initial approximate intensities yield almost identical results. Also, based on the available data, it seems unnecessary to consider the region northeast of the Zagros ranges as a separate attenuation province. In a broad framework, all of Iran can be treated as one unit.

The relations presented in Table 2 apply only in the far-field region. The mathematical form of the ground motion in the near-source region is quite complicated because it must take into account a finite area of the fault and the effect of rupture propagation. The formulation of these effects in a simple form, such that an empirical attenuation relation may be derived, is not possible. Under the circumstance, we make the assumption that a relation of the form given by equation (1) applies to the near-source region as well, but with R replaced by $R + D$, where D is a suitably chosen constant. This change avoids the singularity at $R = 0$ in equation

(1), in recognition of the fact that the earthquake focus is always at some depth below the surface. The value $D = 20$ km was considered reasonable for this study.

With the above change, the constraint $I = I_0$ at $R = 0$ was applied. The following attenuation relation, using the graphically estimated initial I_0 set and excluding data for events 1 and 10, was derived for Iran in general

$$I - I_0 = 6.453 - 0.00121\, R - 4.960 \log(R + 20) \qquad R < 120 \text{ km.} \qquad (3)$$

The study yielded improved estimates of epicentral intensity, presented in Table 1, for each of the earthquakes. The attenuation curve given by the equation (3) and the corresponding data points are shown in Figure 6.

TABLE 2

EVALUATION OF REGIONAL EFFECTS AND INITIAL I_0 SETS ON ATTENUATION RELATIONSHIPS

	Standard Deviation of $I - I_0$	Attenuation Relation
A. Attenuation relations of the form		$I(R) = I_0 + a + b\,R + c \log R$
Graphical initial I_0 set		
All of Iran	0.32	$I(R) = I_0 + 1.723 - 0.01250\,R - 2.204 \log R$
All of Iran, event 10 excluded	0.24	$I(R) = I_0 + 1.475 - 0.01365\,R - 2.005 \log R$
Region northeast of Zagros Ranges, events 4, 10, and 11 excluded	0.25	$I(R) = I_0 + 1.465 - 0.01322\,R - 2.009 \log R$
G-R computed initial I_0 set		
All of Iran	0.32	$I(R) = I_0 + 1.761 - 0.01254\,R - 2.204 \log R$
All of Iran, event 10 excluded	0.24	$I(R) = I_0 + 1.544 - 0.01370\,R - 2.006 \log R$
Region northeast of Zagros Ranges, events 4, 10, and 11 excluded	0.25	$I(R) = I_0 + 1.523 - 0.01325\,R - 2.011 \log R$
B. Attenuation relations of the form		$I(R) = I_0 + a + b\,(R + D) + c \log\,(R + D)$
All of Iran		
Equivalent circle, event 10 excluded	0.23	$I(R) = I_0 + 6.453 - 0.00121\,R - 4.960 \log\,(R + 20)$
Parallel to the isoseismals, events 5 and 10 excluded	0.27	$I(R) = I_0 + 4.824 - 0.00548\,R - 3.708 \log\,(R + 20)$
Transverse to the isoseismals, events 5 and 10 excluded	0.19	$I(R) = I_0 + 8.729 + 0.01158\,R - 6.709 \log\,(R + 20)$

Note: Event 1 was not used in deriving any of the attenuation relations.

It should be noted that the epicentral intensities derived in this study are based on an analysis of all intensity reports at different observation points, and therefore do not depend on a subjective judgment of assigning intensity at one observation point. The study also paves the way for obtaining a reasonable estimate of epicentral intensity for those historical earthquakes also for which very scant information is available. For example, we can now make an estimate of epicentral intensity from such descriptions as, "a violent earthquake occurred near Kashan and was felt as far as Isfahan with an MM intensity of IV."

For want of precise depth determinations, we have not considered the effect of differences in hypocentral depths on the spatial distribution of intensity. Therefore, the attenuation relation derived in this study assumes a more or less similar focal depth for all earthquakes considered.

Effect of Elongation or Ellipticity of Isoseismals on Attenuation

Most of the earthquakes, with the exception of events 2, 3, and 6, show a general elongation of isoseismals in the direction of local structural trend. This observation

TABLE 3

Differential Attenuation, from 30 km to 100 km, for Various Regions

Region	$I(30) - I(100)$
Iran	1.97
Central United States	1.16
Eastern Province	1.40
Cordilleran Province	1.39
San Andreas Province	1.81

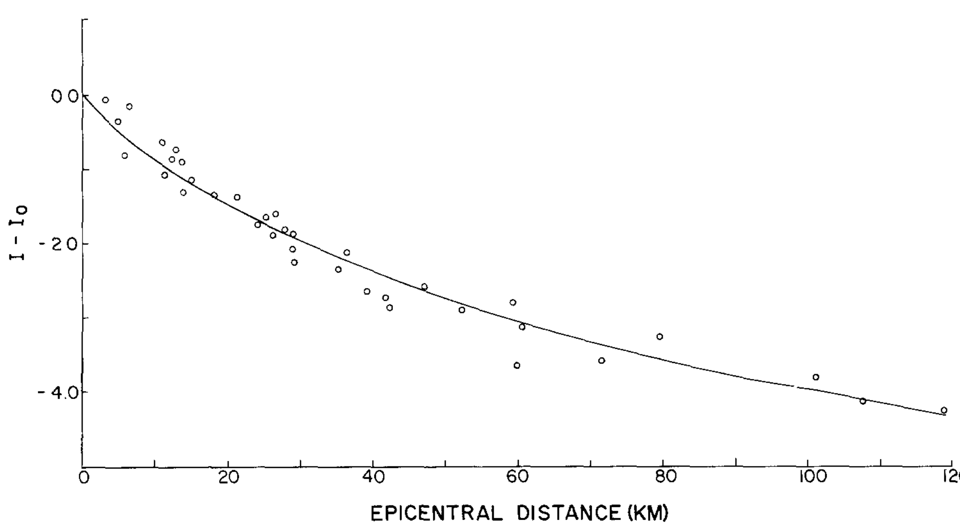

Fig. 6. Attenuation curve for Iran obtained by a least-squares fit between $I - I_0$ and epicentral distance. I_0 set obtained in the last (fifth) iteration is used in this plot.

is apparently related to the orientation of the causative fault, source dimension and the effect of rupture propagation.

Attenuation of intensities parallel and transverse to the isoseismals was investigated by making distance measurements, for each isoseismal line for the various events, in these directions. Event 5, the Buyin-Zara earthquake of September 1, 1962, shows a very strong elongation effect compared to other earthquakes and, therefore, it was not considered in this analysis of average regional attenuation characteristics. Events 1 and 10 were not considered for the reasons stated in the previous section. The following attenuation relations were derived by using graphically estimated initial I_0 set given in Table 1.

$$I - I_0 = 4.824 - 0.00548\, R - 3.708 \log (R + 20) \qquad R < 160 \text{ km} \tag{4}$$

parallel to the isoseismals

$$I - I_0 = 8.729 + 0.01158\,R - 6.709\log(R + 20) \qquad R < 110 \text{ km} \tag{5}$$
transverse to the isoseismals.

The attenuation curves parallel and transverse to the isoseismals, along with the one derived by using equivalent circle radius, equation (3), are shown in Figure 7.

When the design earthquake can be identified with a particular fault or tectonic structure, the attenuation relation parallel to the isoseismals should be used when the site is located along or close to the strike of the fault, and the attenuation relation transverse to the isoseismals should be used when the site is located within the lines drawn perpendicular to the strike of the fault at its two extremities. When neither the relation parallel to the isoseismals nor the relation transverse to the isoseismals is suitable, or when the design earthquake cannot be identified with a particular fault or tectonic structure and the concept of tectonic province must be evoked, the attenuation relation obtained from equivalent circle radius should be used.

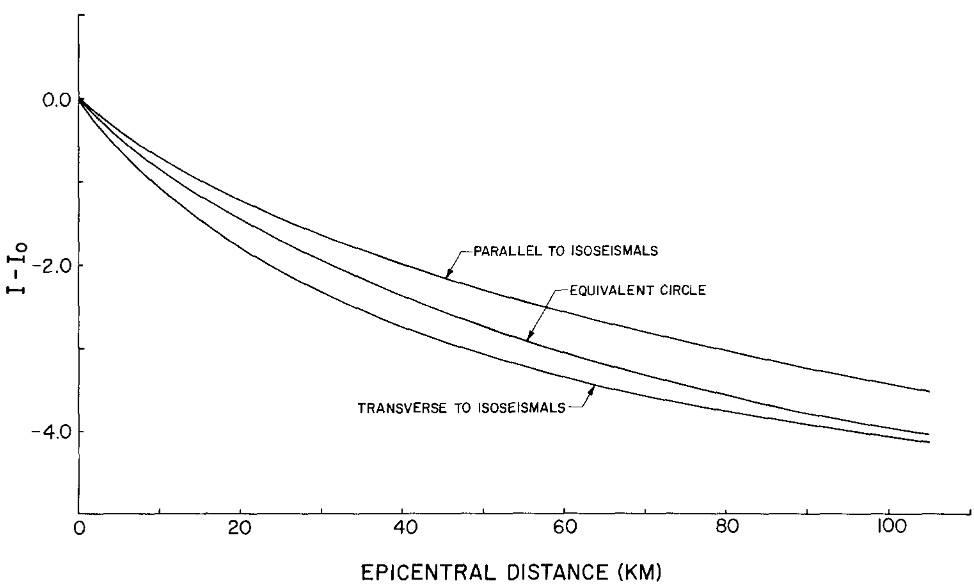

FIG. 7. Attenuation of intensities parallel and transverse to the isoseismals. Average attenuation obtained by using equivalent circle radii for epicentral distances is also shown.

COMPARISON OF THE ATTENUATION OF INTENSITIES IN IRAN WITH DIFFERENT ATTENUATION PROVINCES OF THE UNITED STATES AND SOUTHERN CANADA

It is of some interest to compare the attenuation of MM intensities in Iran with some other regions of the world. Howell and Schultz (1975) studied the attenuation of intensities for United States and southern Canada. They divided the entire region into three attenuation provinces, the San Andreas Province, Cordilleran Province, and Eastern Province. Among the various forms of attenuation relations, they considered equation (1). Their results are as follows

$$I(R) = I_0 + 0.874 - 0.0186\,R - 0.422\ln R \tag{6}$$
for San Andreas Province

$$I(R) = I_0 + 1.802 - 0.0090\,R - 0.628\ln R \tag{7}$$
for Cordilleran Province

$$I(R) = I_0 + 3.278 - 0.0029\,R - 0.989\ln R \tag{8}$$
for Eastern Province.

The attenuation curves corresponding to equations (6) to (8) are plotted in Figure 8, along with the attenuation curve for Iran. The attenuation of central United States was also studied by Gupta (1976). He gave the following attenuation relation

$$I(R) = I_0 + 2.35 - 0.00316\,R - 1.79\log R \tag{9}$$

which is also plotted in Figure 8.

At first sight, it would appear that the attenuation of MM intensities in Iran is much more severe than that for any of the attenuation provinces of United States and southern Canada described by Howell and Schultz (1975). It is realized, however, that in the studies of Howell and Schultz (1975) and Gupta (1976), the epicentral

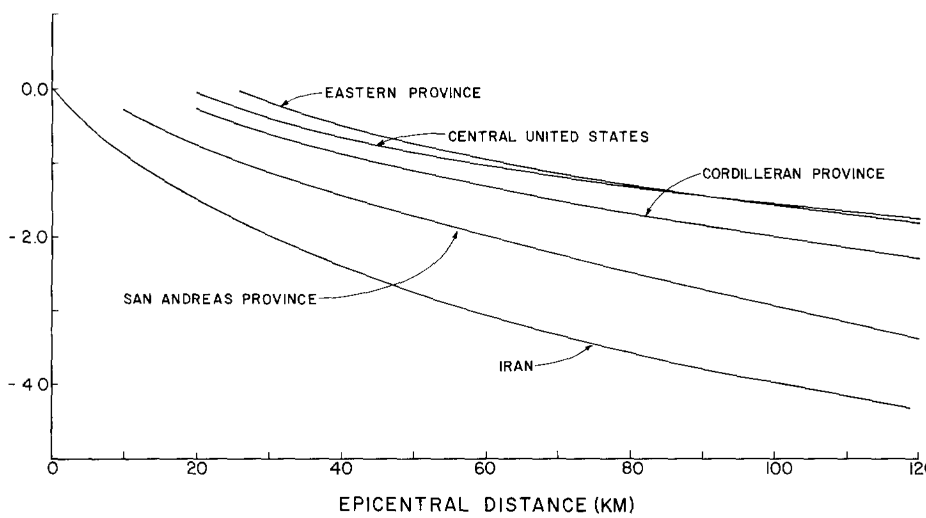

FIG. 8. Comparison of the attenuation of intensities in Iran with different attenuation provinces of the United States and southern Canada. The latter cases show low attenuation for the reasons stated in the text. Differential attenuation, $I(R_1) - I(R_2)$, between two given distances, R_1 and R_2, provides a better means of comparison of attenuation for different regions.

intensities, I_0, for different events were taken to be the maximum isoseismal intensities. Thus, systematically low values of I_0 were selected. The attenuation relations derived by these authors, therefore, yield low values for the falloff of intensity, $I_0 - I(R)$, at any given distance. Instead of comparing absolute values of $I_0 - I(R)$, a better method of comparing the attenuation for different regions would be to look at the differential attenuation, $I(R_1) - I(R_2)$, for two given distances R_1 and R_2. The differential attenuations from a distance of 30 km to 100 km, for different regions, are summarized in Table 3.

It is observed that the attenuation of intensities in Iran is slightly more rapid than the San Andreas province attenuation.

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