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# Attenuation of Intensity with Epicentral Distance in the Iberian Peninsula

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Abstract We have classified the attenuation of the Medveded, Sponheuer, and Karnik (MSK) intensity into five types (each as a function of the epicentral intensity $I_o$) based on the mean radii of 254 isoseismal maps, mainly historical earthquakes in the Iberian Peninsula. Geographically representing each earthquake with its corresponding attenuation tendency, it can be seen that those with low attenuation lie west of the Peninsula and those with high attenuation in the south and east. This regionalization seems to be due as much to the seismotectonic characteristics (different crustal types and size of the earthquakes) as to the different construction types in each region. These attenuation values are similar to those of southern Europe, but much higher than those found in the United States. From the point of view of seismic-hazard evaluation, these laws represent an improvement with regard to those used so far. We have extended previous attenuation studies to the whole of the Iberian Peninsula, and, in some points, differences of attenuation assignment of almost two degrees of intensity have been corrected.

## Introduction

In countries with moderate seismicity, such as the Iberian Peninsula, the number of accelerograms is not only small but also refers to low-magnitude earthquakes located in only some of its seismic areas (Carreño *et al.*, 1999). Therefore, the attenuation of the seismic energy must be carried out with the seismic intensity MSK because more data are available, a greater area is covered, and the earthquakes are larger. Thus, the low-quantitative nature of the intensity versus that of the acceleration nonetheless has the advantage of increasing the number, size ($I_o$), and regional extent of the data. Although in most earthquakes with isoseismal maps the hypocentral and even the epicentral distance are ignored, they nevertheless provide information on the intensity attenuation of the largest and most destructive earthquakes.

Attenuation laws are usually based on considering that the intensity is proportional either to (1) the logarithm of the energy density or (2) to a power of it. In both cases, coefficients representing the source characteristics, the geometric spreading, and the exponential absorption can be obtained. In Europe (Karnik, 1969; Ambraseys, 1985), and particularly in Spain (Munoz, 1974; Martin, 1984), the first hypothesis has often been chosen. In most cases, the term related to the exponential absorption is rejected, and the epicentral distance is considered rather than the hypocentral one.

Mathematical methods used in the estimation of the parameters of these laws range from simple linear regression to sophisticated models of nonlinear regression. The variations in the exponential absorption and geometric-spreading coefficients are used for regionalization (Howell and Schultz, 1975).

In the Iberian Peninsula, Martin (1984) obtained intensity attenuation laws for the SW San Vicente Cape and several regions of Spain (general, south, southeast, south–south-east, and Azores) using the catalog of isoseismal maps for the Iberian Peninsula (Mézcua, 1982). Most of these maps refer to historical earthquakes and their size, assigned by the Instituto Geográfico Nacional de España (IGN), is given by epicentral intensity ($I_o$). Martin (1984) used a method of nonlinear regression, considering both large ($I_o \geq$ VIII) and small ($I_o <$ VIII) earthquakes for the fit. Similar methods have been used by other authors (Muñoz, 1974; Lopez Casado *et al.*, 1992), in some cases covering all of Spain and in other only some regions of it.

The treatment of uncertainties is currently an important step in the evaluation of seismic hazard. They can either be included with the data in computer programs through the Monte Carlo process or analyzed with the results, using the logic-tree method (EPRI, 1985; Araya and Der Kiureghian, 1988; Giner, 1996; Van Eck and Stoyanov, 1996; Schenk *et al.*, 1997). Attenuation laws contribute to the uncertainties through the value of their standard deviation and of their greater or lesser regionalization inside the study area (Perulla *et al.*, 1996). The standard deviation (random variable representing the uncertainty in the prediction of ground motion) is a function of the data quality and the fitting model. The regionalization is a function of the variation in geometric spreading, anelastic attenuation, and the source characteristics of each study area (Howell and Schultz, 1975; Ambraseys, 1985), and for its evaluation, good quality data from each region is required.

Therefore, regional-attenuation laws that are adjusted to real data represent an improvement in seismic-hazard evaluations, reducing statistical and modeling uncertainties. They also provide a simpler treatment of uncertainties and sensitivity analysis.

This work is a continuation of previous attenuation research using a larger theoretical base, introducing a wider regionalization of the law, a greater number of parameters in the fitting, and also an update of the isoseismal maps catalog, including 132 new maps of Spain and Portugal.

## Theory

The laws relating the intensity of an earthquake to distance (Karnik, 1969; Milne and Davenport, 1969; Stepp, 1971; Brazee, 1972; Muñoz, 1974; Martin, 1984; Campbell, 1985) can be obtained considering one of the following two hypotheses.

The first hypotheses states that the intensity, $I$, is proportional to the logarithm of the energy density, $E$, according to the expression:

$$I = c_1 + c_2 \cdot \ln E, \tag{1}$$

where $c_1$ and $c_2$ are empirical constants and the energy decays with distance by means of the relationship:

$$E = (E_0/4\pi)\Delta^{-b} e^{-c\Delta}, \tag{2}$$

where $E_o$ is the total energy released, $\Delta$ the hypocentral distance, $b$ is a constant representing the geometric spreading, and $c$ is a constant representing the rate of absorption. Parameter $b$ is 5/6 for the airy phase, 2 for body waves, and 1 for surface waves. Parameter $c$ is always positive and smaller than unity.

Substituting (2) in (1), we obtain the relationship:

$$I = c_1 + c_2 \ln (E_0/4\pi) - c_2 b \ln \Delta - c_2 c\Delta. \tag{3}$$

If we assume that at the epicenter of an earthquake, $\Delta = h$ (focal depth) and $I = I_o$ (epicentral intensity), then

$$I_0 = c_1 + c_2 \ln (E_0/4\pi) - c_2 b \ln h - c_2 ch \tag{4}$$

Now, we define the parameters $a_1$, $a_2$, and $a_3$, as:

$$a_1 = c_2 b \ln h + c_2 ch \tag{5}$$

$$a_2 = c_2 b \tag{6}$$

$$a_3 = c_2 c \tag{7}$$

and, eliminating $E_o$ between (4) and (3), we obtain the relationship:

$$I = I_0 + a_1 - a_2 \ln \Delta - a_3\Delta \tag{8}$$

where $a_2$ and $a_3$ represent the terms related to geometric spreading and the rate of absorption, respectively. When $a_1$, $a_2$, and $a_3$ are obtained in the regression, then the rate of absorption can be estimated, eliminating $c_2$ between (6) and (7), and assuming a fixed $b$-value by means of the relationship:

$$c = (a_3/a_2)b \tag{9}$$

The attenuation law from equation (8) has been used by Gupta and Nutli (1976), Chandra *et al.* (1979), VanMarcke and Shi-Sheng (1980), Martin (1984), Ambraseys (1985), and Tilford *et al.* (1985), among others.

The second hypothesis states that the intensity is proportional to a positive power ($d < 1.0$) of the seismic-energy density according to the relationship:

$$I = c_3 E^d \tag{10}$$

which, keeping in mind once again equation (2) of the energy decay with distance and after taking the natural logarithm, gives us the equation (Howell and Schultz, 1975):

$$\ln I = \ln I_0 + b_1 - b_2 \ln \Delta - b_3\Delta, \tag{11}$$

where:

$$b_1 = bd \ln h + cdh, \tag{12}$$

$$b_2 = bd, \tag{13}$$

$$b_3 = cd. \tag{14}$$

In equation (11), $b_2$ and $b_3$ represent the terms related to geometric spreading and the rate of absorption, respectively.

According to equations (5) and (12), the source parameters $a_1$ and $b_1$ depend on the focal depth ($h$). Due to the lack of depth assignments for most of our isoseismal maps, we have not studied this dependence. However, we do analyze their relationship with the size ($I_o$) of the earthquake.

The difference between (8) and (11) lies in the curvature of the variation in intensity with distance. Equation (11) is steeper near the epicenter and flatter at greater distances than equation (8).

An inspection of the isoseismal maps suggests that an increase in the earthquake size translates into an increase in the mean radius of the epicentral isoseismal, in which attenuation does not take place (for large earthquakes the focus cannot be considered as a point). Therefore, the fitting of the aforementioned parameters was carried out in two steps (Joyner and Boore, 1981) to separate the dependence of the data on the distance and on the epicentral intensity. In the first step, using the method of Marquardt (1963), according to Draper and Smith (1981), the parameters of equation (8) or (11) are obtained for the different degrees of intensity ($I_o > \mathrm{III}$). In the second step, the parameters $a_1$ or $b_1$ are used to establish the epicentral intensity-dependence according to

$$a_1 \text{ or } b_1 = e_1 + e_2 I_0 + e_3 I_0^2$$

The quadratic nature of this function was chosen for the best fit to the data. A similar circumstances occurs with the relationship between the epicentral intensity, $I_o$, and the magnitude, $m_b$, in the Iberian Peninsula. According to Giner (1996), the best relationship between the epicentral intensity and the magnitude is a quadratic relation on the epicentral intensity. Thus, the final mathematical models to be obtained in this work are:

$$I = f(I_0) - a_2 \ln(R^2 + R_0^2)^{1/2} - a_3(R^2 + R_0^2) \quad (15)$$

$$\ln I = g(I_0) - b_2 \ln(R^2 + R_0^2)^{1/2} - b_3(R^2 + R_0^2) \quad (16)$$

We have substituted the hypocentral distance $\Delta$ by $(R^2 + R_o^2)^{1/2}$, where $R$ is the epicentral distance, $R_o$ is a value that improves the adjustment and that it is used due to the scarcity of data for the focal depth, and $f(I_o)$ and $g(I_o)$ are quadratic functions on the epicentral intensity that consider its dependence on the size of the earthquake.

## The Data

The isoseismal maps catalog of earthquakes for the Iberian Peninsula (Mézcua, 1982) has been used to obtain the radii of the circles of equal area of mapped isoseismal lines for 257 earthquakes, of which 132 are new for this work (Table 1) and the remaining 125 correspond to those used by Martin (1984), although 18 of them have been slightly modified (we have obtained the radii of more isoseismal lines). The use of these mean radii supposes that the laws that will be obtained consider isotropic attenuation. This working hypothesis has been made due to the consideration that we cannot obtain any other type of information due to the quality of the isoseismal maps in our catalog. For the same reason, we have taken the epicentral distance instead of the hypocentral one. The regression was not done on the intensity observations themselves, as we are interested in average attenuation values regardless of the direction. Moreover, for some of the maps these data are not available, and furthermore we are aware of the evident lack of accuracy in the localization of the epicenter of historical earthquakes, giving rise to intensities with erroneous distances to the epicenter. Table 2 presents a classification of the 254 earthquakes with known epicentral intensity (MSK) according to the number of calculated radii, their focal depth, and their body-wave magnitude (Mézcua and Martínez Solares, 1983; updated to 1997).

In off-shore earthquakes such as those of the Azores-Gibraltar fault, the Alboran Sea and the Mediterranean Sea, where the isoseismal line of the epicentral area does not exist, the epicentral intensity has been assigned as a function of the body-wave magnitude ($m_b$), using the relation obtained by Giner (1996):

$$m_b = 2.86 + 0.035 I_0^2$$

In three earthquakes with no assigned magnitude, the assignment was carried out as a function of their distance to the coast. To do so, we used the attenuation laws obtained by Martin (1984) to simulate earthquakes with a fixed epicentral intensity. We compared the simulated isoseismal map with the drawn one and choose the epicentral intensity that best simulated the drawn isoseismals.

For the very energetic off-shore earthquakes (Lisbon 1755, Atlantic Ocean 1941, and southwest S. Vicente Cape 1969) we have not assigned an epicentral intensity *a priori*. In these cases, the epicentral intensity is a parameter to be obtained in the regression. The intensity obtained was compared with that given by the aforementioned relationship ($I_o$–$m_b$). These three earthquakes are not included in Table 2.

A particular characteristic of the seismicity in the Iberian Peninsula is the presence of intermediate deep earthquakes ($20 < h < 180$ km). However, not enough data are available to include them when fitting, although we will attempt to consider their influence on the laws obtained. These areas, with intermediate focal depth, are the Gulf of Cádiz, the West Alborán Sea, southwest Málaga province, and the Pyrenees.

## Methodology

The methodology in this work was designed based on the use of the new isoseismal maps, to obtain (1) the different types of attenuation in the Iberian Peninsula, (2) a better fitting of old and new data, and (3) a regionalization of the above tendencies. It consists of the following steps:

(1) The pairs of values ($I - I_o$, $R$) were represented graphically for all the earthquakes. The observed dispersion is estimated to be a result of (i) a mixture of earthquakes of different size, (ii) isoseismal maps with errors, and (iii) the existence of a geographic regionalization of the attenuation in the Iberian Peninsula. To minimize (i), we used only earthquakes with $I_o \leq \mathrm{VIII}$ (Fig. 1), since we assume that the effects of the size of the earthquakes on the distance between isoseismal lines are less for small earthquakes than for large earthquakes. We neglect (ii) as we assume that these errors are random. Then, we can assume that the dispersion of our data is mainly due to (iii). Therefore, in equa-

# Table 1

## Mean Radii of the Isoseismal Obtained in this Work*

| Information | | | | | Mean Radii of the Isoseismal in Kilometers | | | | | | | | | Information | | | | | Mean Radii of the Isoseismal in Kilometers | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Year | M/D | Lon. | Lat. | Io | X | IX | VIII | VII | VI | V | IV | III | II | Year | M/D | Lon. | Lat. | Io | X | IX | VIII | VII | VI | V | IV | III | II |
| 1428 | 0202 | −2.20 | 42.40 | IX | | 8.29 | 25 | 60 | 120 | | | | | 1958 | 0116 | 0.60 | 38.10 | VI | | | | | 7 | 12 | 20 | 29 | |
| 1522 | 0922 | 2.50 | 36.92 | IX | | 11.65 | 20.9 | 42.05 | 62.3 | 90 | | | | 1958 | 0121 | 7.10 | 41.60 | V | | | | | | 3.61 | 7.81 | 10 | |
| 1531 | 0126 | 9.00 | 38.95 | IX | | 39 | 64 | | | | | | | 1958 | 0607 | 2.93 | 36.70 | V | | | | | | 9.35 | 21.36 | | |
| 1680 | 0109 | 4.66 | 36.68 | IX | | | 55 | 86 | 140 | 201 | 460 | 590 | | 1958 | 0618 | 1.53 | 38.90 | VI | | | | | 1.96 | 4.46 | 8.39 | | |
| 1722 | 1227 | 7.58 | 37.17 | VIII | | | 15.5 | 31.92 | 51.08 | 75.69 | | | | 1958 | 1125 | −0.10 | 42.87 | VII | | | | 12 | 30 | 50 | 65 | 80 | |
| 1755 | 1101 | 10.00 | 36.00 | **XII** | 150 | 222 | 312 | 600 | 792 | 1020 | | | | 1959 | 0414 | 0.50 | 38.00 | V | | | | | 3.27 | 11.41 | 24.05 | | |
| 1804 | 0113 | 2.83 | 36.83 | VIII | | | 4.5 | 14.25 | 25 | 50 | 80 | 100 | | 1960 | 0229 | 9.62 | 30.45 | X | 1.7 | 3.87 | 5 | 13.08 | 32.64 | | | | |
| 1804 | 0825 | 2.80 | 36.80 | IX | | 5.7 | 15 | 28 | 48 | 70 | | | | 1960 | 0503 | 8.40 | 41.50 | V | | | | | | 5.75 | 15.31 | | |
| 1817 | 0318 | 2.08 | 67.25 | VIII | | | 7.18 | 21.9 | 41.8 | 74.74 | 160.57 | 270 | | 1960 | 0507 | 9.75 | 39.63 | **VI** | | | | | | 8.29 | 18.95 | 31.51 | |
| 1845 | 1007 | −0.70 | 41.00 | VI | | | | | 5.27 | 17.38 | 30.38 | | | 1960 | 1105 | 8.80 | 41.70 | V | | | | | | 2.02 | 16.12 | 25.73 | |
| 1858 | 1111 | 9.00 | 38.20 | IX | | 44 | 70 | 115 | 180 | 270 | 370 | 500 | | 1962 | 0211 | 8.62 | 37.17 | IV | | | | | | | 11.06 | 19.28 | |
| 1863 | 1006 | 1.90 | 37.40 | VI | | | | | 9 | 16 | 35 | | | 1962 | 0503 | 7.02 | 43.88 | **VII** | | | | | 50 | 65 | 80 | | |
| 1901 | 0424 | 7.66 | 36.83 | V | | | | | | 60 | 160 | 230 | | 1962 | 0609 | −0.58 | 41.97 | V | | | | | | 3.45 | 7.06 | 10.62 | |
| 1901 | 0525 | 3.50 | 36.70 | VII | | | | 10 | 40 | 80 | | | | 1962 | 0831 | 9.25 | 39.47 | V | | | | | | 8.92 | 16.93 | 30.9 | 40 |
| 1901 | 1000 | −3.00 | 41.75 | IV | | | | | | | 20 | 50 | 70 | 1962 | 0904 | 8.76 | 36.67 | **VII** | | | | | | | 60.71 | 75 | |
| 1903 | 0420 | −3.28 | 42.30 | VI | | | | | 13.23 | 35.9 | 65 | | | 1962 | 1226 | 10.65 | 39.35 | **VIII** | | | | | | 130 | 216 | 293 | |
| 1903 | 0809 | 9.00 | 38.30 | VII | | | | 63 | 140 | 330 | | | | 1962 | 1231 | −1.03 | 41.98 | V | | | | | | 1.91 | 8.24 | 21.2 | |
| 1904 | 0808 | 9.38 | 38.78 | V | | | | | | 30 | 100 | 140 | 180 | 1964 | 0315 | 7.75 | 36.13 | **IX** | | | | 108 | 162 | 232 | 297 | 400 | 556 |
| 1909 | 0423 | 8.82 | 38.95 | IX | | 18.5 | 36.16 | 74 | 150 | 296 | | | | 1964 | 0403 | 1.10 | 38.20 | V | | | | | | 4.28 | 11.46 | 21.11 | |
| 1909 | 0701 | 0.67 | 38.00 | VII | | | | 6.91 | 15.36 | 21.48 | 39.04 | | | 1964 | 0411 | 1.33 | 38.10 | V | | | | | | 7.46 | 16.69 | 21.58 | |
| 1910 | 0616 | 3.37 | 36.67 | VIII | | | 12.67 | 56.13 | 124.4 | 191.49 | 275.01 | | | 1964 | 0509 | 1.10 | 38.20 | V | | | | | | 6.91 | 16.92 | 23.29 | |
| 1910 | 1124 | 8.25 | 43.53 | VII | | | | 11.28 | 70 | 120 | | | | 1964 | 0516 | 8.53 | 41.17 | V | | | | | | 3 | 10 | 19 | 28 |
| 1912 | 1018 | 8.38 | 41.38 | VI | | | | | 12.62 | 31.41 | 43.24 | | | 1964 | 0829 | 0.06 | 43.12 | VI | | | | | 50 | 60 | 68 | | |
| 1912 | 1116 | 8.45 | 41.37 | VI | | | | | 11.97 | 33.01 | 45.74 | | | 1965 | 0708 | −1.82 | 42.52 | IV | | | | | | | 14.4 | 23.8 | 34 |
| 1913 | 0504 | 8.50 | 36.50 | **VII** | | | | | 55 | 95 | 160 | 220 | | 1966 | 0114 | −2.10 | 42.30 | V | | | | | | 3.45 | 6.96 | 11.11 | |
| 1913 | 0609 | 9.25 | 39.33 | VI | | | | | 25 | 50 | 100 | | | 1966 | 0826 | 8.70 | 38.05 | VI | | | | | 8.6 | 16.69 | 97.24 | 150 | |
| 1913 | 0807 | 9.45 | 38.62 | VI | | | | | 30 | 50 | 100 | | | 1966 | 1218 | 3.30 | 43.25 | VI | | | | | 5.43 | 10.36 | 18.38 | | |
| 1916 | 0427 | −2.55 | 41.57 | V | | | | | | 0.64 | 2.76 | 4.52 | | 1967 | 0813 | 0.68 | 43.30 | VIII | | | | | 5.17 | 11 | 32.9 | 47.87 | |
| 1916 | 0615 | −2.58 | 41.58 | V | | | | | | 1.69 | 2.82 | 6.91 | | 1968 | 0225 | −1.20 | 40.93 | **V** | | | | | | | | 38 | 61 |
| 1917 | 0928 | −3.27 | 42.48 | VI | | | | | 10 | 20 | 40 | 60 | | 1969 | 0505 | 10.40 | 36.00 | **VIII** | | | | | | 135 | 252 | 375 | |
| 1917 | 1023 | −2.28 | 41.20 | VI | | | | | 2.82 | 12 | 20 | 27 | | 1969 | 0716 | −0.46 | 38.28 | V | | | | | | | 12 | 15.31 | |
| 1918 | 0401 | 8.27 | 37.08 | V | | | | | | 15 | 50 | | | 1969 | 0906 | 12.30 | 36.90 | **IX** | | | | | | | 300 | 480 | |
| 1920 | 1126 | 8.60 | 42.40 | VII | | | | 45 | 90 | 150 | | | | 1969 | 0918 | 8.60 | 39.97 | IV | | | | | | | 39.03 | 149 | |
| 1923 | 0909 | −2.30 | 42.37 | V | | | | | | 2.82 | 19.13 | | | 1969 | 1021 | 8.76 | 39.97 | V | | | | | | 19.59 | 57.15 | 123.4 | |
| 1923 | 1107 | 3.00 | 41.00 | **VII** | | | | | | | 85 | 107 | 130 | 1970 | 0105 | 0.65 | 38.07 | V | | | | | | 4.5 | 7.5 | 10 | |
| 1923 | 1119 | −0.83 | 42.68 | VIII | | | 25.23 | 60.13 | 97.72 | 143.62 | 220 | | | 1970 | 0316 | −2.13 | 42.30 | V | | | | | | 15.96 | 41.03 | 80 | |
| 1924 | 0227 | −0.78 | 42.68 | VI | | | | | 6.58 | 25.23 | 59.71 | | | 1971 | 0126 | 0.49 | 38.10 | IV | | | | | | | 17.62 | 26.95 | |
| 1926 | 0612 | 2.37 | 36.77 | VI | | | | | 25 | 40 | 60 | | | 1972 | 0614 | 8.56 | 36.65 | **VII** | | | | | | | 55 | 78 | 102 |
| 1926 | 1218 | 9.17 | 38.78 | VI | | | | | 10.5 | 50 | 105 | | | 1973 | 0124 | 0.80 | 38.03 | IV | | | | | | | 6.03 | 10 | |
| 1927 | 0312 | −2.47 | 41.70 | VII | | | | 2.82 | 6.91 | 13.53 | | | | 1973 | 0523 | 0.63 | 38.13 | IV | | | | | | | 5.59 | 17.65 | |

(*continued*)

Table 1
Continued

| Information | | | | | Mean Radii of the Isoseismal in Kilometers | | | | | | | | | Information | | | | | Mean Radii of the Isoseismal in Kilometers | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Year | M/D | Lon. | Lat. | Io | X | IX | VIII | VII | VI | V | IV | III | II | Year | M/D | Lon. | Lat. | Io | X | IX | VIII | VII | VI | V | IV | III | II |
| 1928 | 1124 | −2.37 | 41.50 | IV | | | | | | | 4.22 | 10.43 | | 1973 | 1026 | −3.00 | 41.50 | IV | | | | | | | 11.97 | 22.75 | |
| 1928 | 1128 | −2.33 | 41.52 | VI | | | | | 1.6 | 5.97 | 14.63 | | | 1974 | 0311 | 7.88 | 40.40 | IV | | | | | | | 42 | 60 | 76 |
| 1928 | 1213 | −2.33 | 41.52 | V | | | | | | 1.6 | 5.97 | 14.63 | | 1975 | 0806 | 8.20 | 41.13 | IV | | | | | | | 38.9 | 57.4 | |
| 1929 | 0203 | −2.43 | 41.53 | V | | | | | | 5.98 | 15.71 | 25 | | 1976 | 0926 | 0.59 | 38.88 | VI | | | | | 5.75 | 17.73 | 28.47 | 38.94 | |
| 1930 | 0511 | −2.85 | 41.72 | IV | | | | | | | 13.52 | 23.6 | | 1979 | 0220 | 7.87 | 42.32 | IV | | | | | | | 8.88 | | |
| 1934 | 0907 | −1.72 | 36.23 | IX | | 8.44 | 16.45 | 24.59 | 50 | 70 | 90 | 120 | | 1979 | 0320 | 3.80 | 37.17 | VI | | | | | 3.8 | 8.69 | 13.58 | 17.54 | |
| 1941 | 0404 | 0.58 | 38.00 | V | | | | | | 7.72 | 19.6 | 26.31 | | 1979 | 0429 | 3.60 | 37.12 | IV | | | | | | | 1.77 | 4.33 | |
| 1941 | 1125 | 19.02 | 37.42 | **XII** | | | | 512 | 675 | 850 | 1150 | | | 1979 | 0514 | 2.46 | 37.60 | V | | | | | | 19.85 | 27.1 | 37.1 | |
| 1942 | 0410 | 0.60 | 38.10 | IV | | | | | | | 10.05 | 18.71 | | 1979 | 0525 | −2.63 | 41.95 | V | | | | | | 2.3 | 4.4 | 6.3 | |
| 1943 | 0326 | −2.80 | 41.67 | V | | | | | | 9.77 | 48.42 | 124.9 | | 1979 | 0819 | 3.70 | 37.20 | V | | | | | | 1.65 | 3.56 | 5.8 | |
| 1945 | 1116 | 0.73 | 38.15 | V | | | | | | 3.97 | 15.2 | 23.03 | | 1979 | 1025 | 0.77 | 38.02 | V | | | | | | 22.79 | 46.16 | | |
| 1946 | 0703 | 0.27 | 38.45 | V | | | | | | 7.12 | 22.53 | 41.98 | | 1979 | 1125 | 3.77 | 36.87 | VI | | | | | 9.87 | 17.12 | 24.19 | 32.16 | |
| 1946 | 0930 | 0.63 | 37.93 | **V** | | | | | | | 13 | 21.5 | 33.5 | 1980 | 0523 | 7.47 | 37.20 | V | | | | | | 6.08 | 24.48 | 48.86 | |
| 1946 | 1129 | 2.65 | 36.97 | V | | | | | | 7.24 | 16.61 | 30.64 | | 1980 | 0927 | 3.10 | 36.78 | IV | | | | | | | 6.43 | 12.13 | |
| 1947 | 0609 | 0.40 | 38.25 | **V** | | | | | | | 8.8 | 13.2 | 18 | 1980 | 1006 | 8.67 | 42.33 | V | | | | | | 9.77 | 30.59 | 60 | |
| 1950 | 0131 | −0.20 | 43.12 | VII | | | | 11.28 | 28 | 50.46 | | | | 1980 | 1010 | −1.45 | 36.15 | IX | | 15 | 30 | 45 | 88 | 110 | 168 | 204 | 232 |
| 1951 | 0211 | −2.82 | 41.62 | V | | | | | | 5.64 | 12.3 | 21.3 | | 1980 | 1111 | 5.22 | 37.83 | VI | | | | | 3.26 | 12.68 | 21.31 | 33.5 | |
| 1954 | 0521 | 0.20 | 38.52 | V | | | | | | 2.52 | 6.5 | 19.19 | | 1980 | 1203 | 5.67 | 36.92 | V | | | | | | 28.21 | 47.04 | 62.19 | |
| 1955 | 1120 | −1.70 | 40.90 | **VI** | | | | | | | 40 | 80 | | 1980 | 1208 | 2.12 | 35.98 | **VI** | | | | | | | 80 | 92 | |
| 1956 | 0731 | −2.33 | 41.55 | V | | | | | | 5 | 15 | 22 | | 1981 | 0122 | 2.65 | 37.05 | V | | | | | | 3.61 | 17.56 | 24.27 | |
| 1956 | 0814 | 2.09 | 37.08 | IV | | | | | | | 7.5 | 18 | | 1981 | 0305 | −0.22 | 38.50 | V | | | | | | 20 | 70 | 115 | |
| 1956 | 1228 | 2.71 | 36.75 | V | | | | | | 1.91 | 7.54 | 20.85 | | 1984 | 0624 | 3.74 | 36.83 | V | | | | | | 15.48 | 66.01 | 108.5 | |
| 1957 | 0629 | 1.34 | 36.32 | **VII** | | | | | | | 85 | 110 | 135 | 1989 | 1220 | 7.39 | 37.23 | VI | | | | | 9 | 40 | 110 | 190 | |
| 1957 | 0715 | 7.49 | 36.35 | **V** | | | | | | | 95 | 145 | | 1993 | 1223 | 2.94 | 36.78 | VII | | | | 3.95 | 15.98 | 49.36 | 90.24 | | |
| 1958 | 0105 | 0.68 | 38.88 | V | | | | | | 2.07 | 6.68 | 14.08 | | 1994 | 0104 | 2.82 | 36.57 | VII | | | | 7.73 | 23.87 | 44.11 | 79.29 | | |
| 1958 | 0107 | 8.18 | 38.32 | V | | | | | | 2.26 | 23.92 | 34.81 | | 1994 | 0526 | 4.00 | 35.27 | VI | | | | | 18 | 40 | 99 | 155 | 195 |

*M/D, month and day; Lon., Longitude; Lat., Latitude. $I_0$, epicentral intensity (MSK scale) (Mézcua and Martínez Solares, 1981; updated to 1997); Bold type, off-shore earthquake (epicentral intensity assigned in this work).

tions (8) and (11), $a_1$ and $b_1$ can be considered approximately constant. Then, the attenuation is divided into five tendencies, two extreme attenuations, very high and very low, which take into account earthquakes whose attenuation is very different from the rest of the data, and three tendencies, high, medium, and low, which take into account those earthquakes with the three characteristic types of attenuation in this region. To do so, we use the third and fourth mean radii values for earthquakes with three or more isoseismal lines and the second mean radius for earthquakes with only two isoseismal lines. This represents isoseismals VI and V for earthquakes with $I_o$ = VII, V, and VI for $I_o$ = VII, and VI and III for $I_o$ = VI (we use damages and preceptibility areas to obtain different attenuation tendencies). The results are shown in Table 3. These radii are used instead of the first one because the surfaces they represent are less affected by ground amplification (site effect) and or source effects.

(2) In the next step, following Howell and Schultz (1975), an attenuation law using equation (11) is obtained for all the earthquakes selected in step (1). In the fitting, we neglect the terms related to the source ($b_1$) and to the geometric spreading ($b_2$) and substitute the hypocentral distance ($\Delta$) by the epicentral distance ($R$). That is to say, we fit the data to the equation: $\log I = \log I_o - b_3 R$. This fitting is done to obtain an approximate estimate of the absorption coefficient ($b_3$). Finally, the different absorption coefficients obtained are averaged with respect to the type of attenuation. Table 3 shows that these values follow a tendency similar to that proposed for the attenuation. That is, the mean absorption coefficient diminishes as the attenuation decreases. This result obtained for each earthquake justifies the above division of the attenuations and our graphical criteria.

(3) Subsequently, the two models of attenuation law that give the best overall fit to our data are determined. To do so, all the earthquakes selected in step (1) are fitted, in accordance with each attenuation tendency, to equations (8) and (11), substituting $\Delta = (R + R_o)^{1/2}$, by means of a nonlinear regression algorithm (Marquardt, 1963). The coefficients obtained were statistically significant to a confidence level of 90% or higher. As equation (8) provided the best results (smaller rms), it was chosen for our work (Table 4). To estimate the parameters of the fitting, the physical restrictions of energy attenuation were kept in mind: both the coefficient related to the geometric spreading ($a_2$ and $b_2$) and that of the exponential absorption ($a_3$ and $b_3$) must be positive.

Before continuing, we should note that step (1) has been done to classify the data due to the great dispersion; step (2) has been performed to justify that classification; and step (3) has been carried out to choose a physical attenuation model. As we have seen, equation (8) best represents small to moderate earthquakes ($I_o \leq$ VIII), but in the Theory section we have said that to represent the attenuation of earthquakes of any size, equation (8) must be replaced by equation (15). To calculate the coefficients of the fitting of each attenuation tendency, we have used a two-step-regression analysis (as per Joyner and Boore, 1981). The following steps explain the methodology to obtain the parameters of equation (15).

(4) In the first step of the regression, the values of parameters $a_1$, $a_2$, and $a_3$ were obtained in equation (8) for earthquakes with $I_o \leq$ VIII. This was done assuming that coefficient $a_1$ is approximately constant for small to moderate earthquakes ($I_o \leq$ VIII) and taking $I - I_o$ as the dependent variable. Due to the dispersion of the data, we will suppose that the group of data ($R$) corresponding to each ($I - I_o$) value follows a normal distribution of more likely value, the mean value. Once this value has been obtained, we exclude those data outside the standard deviation and repeat the process until a stable mean value is obtained. This final value is noted as $\bar{R}$. Then, we fit all the pairs of values ($I - I_o$, $\bar{R}$) by means of a nonlinear regression algorithm (Marquardt, 1963; according to Draper and Smith, 1981). The $a_1$, $a_2$, and $a_3$ values obtained were statistically significant to a confidence level of 90% or higher.

(5) In the second step of the regression, we fix the values of $a_2$ and $a_3$, obtained in step (4) in equation (15) and fit the pairs of values ($I$, $\bar{R}$) for each degree of epicentral intensity ($I_o >$ V) to obtain a f($I_o$) value for each degree of epicentral intensity. In this step we use only earthquakes with $I_o \geq$ VI since they have more drawn isoseismals, possibly fewer errors and are the most important in the evaluation of seismic hazard. In addition, we have fitted the earthquakes with $I_o >$ VIII that were not studied in previous sections. To include earthquakes with known $I_o >$ VIII and the very energetic earthquakes within an attenuation tendency, we plotted them to determine their fit in one of the five selected tendencies. The lowest rms in the fitting of each earthquake was the criteria to choose their attenuation tendency and their f($I_o$) value.

(6) Then, the pair of (f($I_o$), $I_o$) values were fitted to a quadratic function on $I_o$ (Fig. 2a). In the case of the very energetic off-shore earthquakes, included in the very low attenuation tendency as a result of section (4), and whose epicentral intensity is not assigned, we were giving different values to their epicentral intensity and making the (f($I_o$), $I_o$) regression. The coefficients of the f($I_o$) function and the $I_o$ values of the mentioned earthquakes ($I_o$ = XII for the 1755 Lisbon earthquake, $I_o$ = XII for the 1941 Atlantic Ocean earthquake and $I_o$ = XI for the 1969 S. Vicente Cape earthquake) were chosen based on the lowest rms.

(7) The standard deviation of the residuals and the coefficient of determination were obtained comparing the predicted values $\hat{I}$ obtained from equation (15) with real values ($I$). In this step we have used all the earthquakes with $I_o >$ V and not the pair of fitted values ($I$, $\bar{R}$) to show the real scattering of the data sample used.

(8) Finally, all the earthquakes with their attenuation tendencies were represented on the map of the Iberian Peninsula. We then proceeded to a possible geographic regionalization of the attenuation laws as a function of their different seismotectonic characteristics (change in crust and earthquake size behavior in the Iberian Peninsula).

Table 2
Data of the Isoseismal Maps

| | No. of Mean Radius Calculated | | | | | | | | | Focal Depth (km) | | | | | Magnitude $m_b$ | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $I_0$ | 8 | 7 | 6 | 5 | 4 | 3 | 2 | 1 | Off* | 1–20 | 21–40 | 41–60 | 61–80 | >80 | 6.9–6.0 | 5.9–5.0 | 4.9–4.0 | 3.9–3.0 | 2.9–2.0 |
| X | | | | 2 | 1 | | | | | | | | | | 1 | | | | |
| IX | 1 | 2 | 1 | 4 | 1 | 1 | 2 | | 2 | 1 | 1 | | | | 2 | 2 | | | |
| VIII | | | 12 | 4 | 3 | 3 | | | 4 | 4 | 1 | | | | | 6 | 7 | | |
| VII | | | | 13 | 5 | 12 | 1 | | 9 | 7 | 1 | | | | | 2 | 11 | 2 | |
| VI | | | | 1 | 29 | 19 | 9 | | 8 | 16 | 1 | 1 | | | | 3 | 21 | 6 | |
| V | | | | | 4 | 76 | 19 | 1 | 8 | 16 | 1 | | | | | 2 | 24 | 30 | 2 |
| IV | | | | | | 3 | 24 | 1 | 4 | 10 | | | | 1 | | | 5 | 14 | |
| Tot. | 1 | 2 | 13 | 24 | 43 | 114 | 55 | 2 | 35 | 54 | 5 | 1 | | 1 | 3 | 15 | 68 | 52 | 2 |

*Number of off-shore earthquakes.

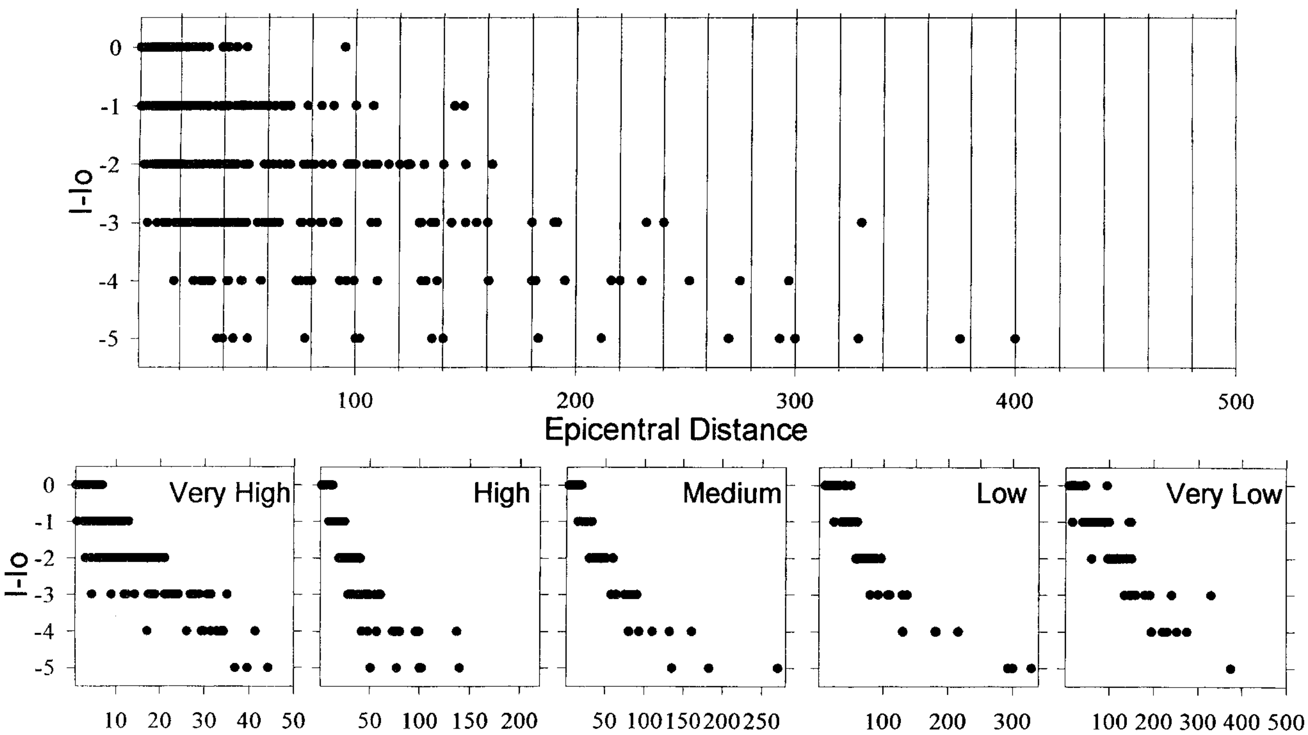


Figure 1. $I - I_o$ versus epicentral distance for earthquakes with $I_o \leq$ VIII in the Iberian Peninsula.

## Results

As already indicated, for the Iberian Peninsula and adjacent areas, we have defined five attenuation laws whose equations appear in Table 5 (three decimals should be considered in the $a_1$ coefficient to maintain the best fit to the data). The goodness of fit to the data is shown in Figure 2: (b) very high; (c) high; (d) medium; (e) low; and (f) very low attenuation. The percentage of earthquakes included in each of the above tendencies is: 32%, very high; 30%, high; 13%, medium; 11%, low; and 14%, very low, in agreement with the number and the geographical distribution of the data (more on the east than on the west)

We have compared our different attenuations (Fig. 3a) and represented our laws graphically together with those obtained by other authors (Figs. 3b–f). It can be observed that our law for very high attenuation is similar but faster to those obtained by Martin (1984) (Fig. 3b). The high attenuation is similar to those of Ambraseys (1985) for northern Europe, with the value of $h$ between 5 and 10 km, and with those of Grandori *et al.* (1987) for Italy (Fig. 3c). The medium attenuation compares well with those of Chandra *et al.* (1979) for Iran, and Vanckmarcke and Shi-Sheng (1980) for the Philippines (Fig. 3d). The low and very low attenuations are correlated with those of Howell and Schultz (1975) for the San Andreas, Cordilleran, and the eastern United States, and

Table 3
Regionalization of the Isoseismal Maps

| Attenuation | Second R* | Third R† | Fourth R† | $b_3 \cdot 10^4$‡ |
|---|---|---|---|---|
| Very High | ≤10 | ≤20 | ≤30 | 400 ± 200 |
| High | 10 < R ≤ 20 | 20 < R ≤ 40 | 30 < R ≤ 60 | 160 ± 40 |
| Medium | 20 < R ≤ 40 | 40 < R ≤ 60 | 60 < R ≤ 90 | 80 ± 20 |
| Low | 40 < R ≤ 60 | 60 < R ≤ 80 | 90 < R ≤ 130 | 50 ± 10 |
| Very Low | >60 | >80 | >130 | 33 ± 8 |

*Earthquakes with two mean radii only.
†R, mean radius in kilometers.
‡$b_3$, mean absorption coefficient ($\ln I = \ln I_0 - b_3$ R; Howell and Schultz, 1975).

Table 4
Comparison between Attenuation Models Proposed in Equations (3) and (8)*

| Attenuation | | A | B | C | $R_0$ | r-sq. | $\sigma$ |
|---|---|---|---|---|---|---|---|
| Very High | (3) | 0.836 | 0.700 | 0.06584 | 2 | 0.80 | 0.53 |
| | (8) | 0.202 | 0.177 | 0.01111 | 2 | 0.73 | 0.57 |
| High | (3) | 2.884 | 1.440 | 0.00562 | 4 | 0.88 | 0.43 |
| | (8) | 0.475 | 0.260 | 0.00131 | 2 | 0.80 | 0.50 |
| Medium | (3) | 4.232 | 1.578 | 0.00394 | 10 | 0.93 | 0.34 |
| | (8) | 0.658 | 0.261 | 0.00121 | 7 | 0.83 | 0.45 |
| Low | (3) | 5.844 | 1.716 | 0.00370 | 21 | 0.92 | 0.38 |
| | (8) | 0.866 | 0.269 | 0.00099 | 14 | 0.83 | 0.47 |
| Very Low | (3) | 7.033 | 1.765 | 0.00324 | 47 | 0.88 | 0.48 |
| | (8) | 1.322 | 0.329 | 0.00085 | 47 | 0.84 | 0.50 |

*A, $a_1$ coefficient in equation (3) and $b_1$ coefficient in equation (8); B, $a_2$ coefficient in equation (3) and $b_2$ coefficient in equation (8); C, $a_3$ coefficient in equation (3) and $b_3$ cofficient in equation (8); $R_0$, assumed value which gives best fit; r-sq, coefficient of determination; $\sigma$, standard deviation in intensity.

with those of Ambraseys (1985) with $h$ between 13 and 30 km (Figs. 3e and 3f). Very high attenuation laws, like these, are also seen in southern Italy (Perulla *et al.*, 1996). These comparisons show that the coefficient of absorption for the Iberian Peninsula falls within the range of values obtained in other countries (Table 6).

The regionalization of the attenuation laws is shown in Figure 4, according to which (1) earthquakes with low or very low attenuation are concentrated in Portugal and the Azores-Gibraltar fault (57% very low and 18% low versus 11% medium, 11% high, and 3% very high); (2) very high attenuation in the Granada basin (90% very high versus 10% medium) to high attenuation in southeast Spain, the coast of Algeria, the Levante, the Balearic Islands, and NE Spain (36% very high and 46% high versus 7% medium, 7% low, and 3% very low); (3) low attenuation in the area between Huelva and Málaga (17% very low, 66% low, and 17% medium), that is, the behavior is intermediate between that of Portugal (1) and that of Spain (2); (4) in the Alboran Sea, low (37%) and very low (37%) attenuations; (5) high attenuation in the Iberian System (62%); (6) very high attenuation in the southern part of the Iberian Meseta; (7) high to medium attenuation in the northern part of the Iberian Meseta and the Cantabrian range; and (8) low and very low attenuation in the central Pyrenees (25% very low, 25% low, 33% medium, 8% high, and 8% very high). Not fitting in the aforementioned model is the south coast of Portugal with high attenuation and the north coast of Portugal with very high (12.5%), high (25%), medium (25%), low (25%), and very low (12.5%) attenuation.

As can be observed in Figure 4, for specific areas (for instance, seismic sources used in the seismic hazard evaluation) within the above regions, the percentages are even greater. This comprises not only an improvement in the seismic-hazard evaluation for these areas but also allows a better treatment of uncertainties. Using the Monte Carlo method, the procedure for expressing the input-model parameters as weighted distributions, rather than as single valeus, is now easier (Mallard and Woo, 1993), which is of interest to obtain confidence levels for the results of seismic-hazard evaluations (Kulkarni *et al.*, 1984). Using logic-tree methodology, we can obtain more real probabilities for the branches of the logic tree corresponding to the attenuation laws.

A similar situation of regionalization is given in the United States, with low attenuation in the west and high in the east (Hank and Johnston, 1992; Bollinger *et al.*, 1993; Rizzo *et al.*, 1995). However, except in the case of our very low law, our attenuations are faster.

Apart from the seismotectonic differences of the two areas (west and east of the Iberian Peninsula), the different types of construction, that is to say, the distinct vulnerability of buildings in one area or another, and the geographical distribution of the population could also account for this variability in the behavior. Overestimation of the epicentral intensity and the remaining intensities, due to the poor quality of buildings or to their antiquity, together with the proximity of populations, can give rise to a sharp decrease in the damage area, therefore showing a faster attenuation of the intensity.

This regionalization is clearly correlated with the seismotectonic characteristics of each area. Thus, very low and low attenuations can be associated with the Hercynian domain in the west and center of the Peninsula; high attenuation with the Alpine domain in the south, east, and northeast; and very high attenuation with certain Neogene basins. A similar result was obtained by Lana *et al.* (1999), who found high values of attenuation and of the coefficient $Q_\beta^{-1}$ for fundamentally Alpine or Neogene areas and low values of attenuation and of the coefficient $Q_\beta^{-1}$ for mainly Hercynian domains.

A detailed comparison of the laws we have found inside each region with those corresponding to Martín (1984) reveals clear differences. Taking, for example, our very high attenuation in the Granada basin and comparing it with Martin's Southern attenuation (1984), which corresponds to the same data, according to their geographical location, we find large differences between the two laws and between the

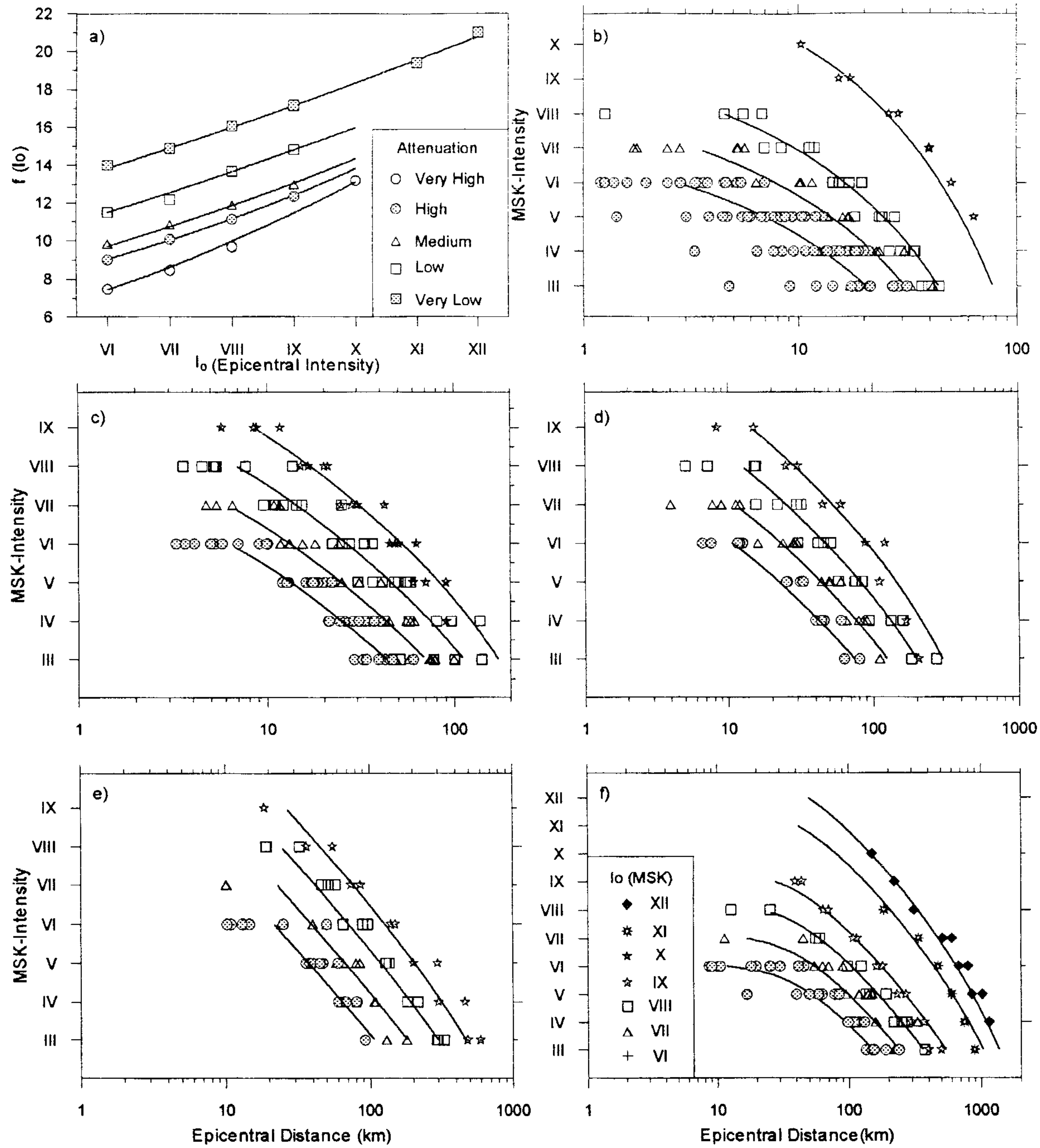


Figure 2. (a) Relation between the f($I_o$) coefficient and the epicentral intensity $I_o$. Goodness of fit to our data: (b) very high attenuation; (c) high attenuation; d) medium attenuation; d) low attenuation, and (e) very attenuation.

Table 5
Best Solutions of Equation: $I = \mathrm{f}(I_0) - a_2 \ln \Delta - a_3 \Delta$

| Attenuation | f($I_0$) | $a_2$ | $a_3$ | $R_0$ | $\sigma$ | r-sq. |
|---|---|---|---|---|---|---|
| Very High | $3.606 + 0.171 \cdot I_0 + 0.078 \cdot I_0^2$ | 0.920 | 0.07615 | 2 | 0.49 | 0.86 |
| High | $6.016 + 0.090 \cdot I_0 + 0.069 \cdot I_0^2$ | 1.477 | 0.01035 | 4 | 0.46 | 0.91 |
| Medium | $4.927 + 0.571 \cdot I_0 + 0.037 \cdot I_0^2$ | 1.445 | 0.00609 | 6 | 0.39 | 0.94 |
| Low | $5.557 + 0.902 \cdot I_0 + 0.014 \cdot I_0^2$ | 1.762 | 0.00207 | 2 | 0.59 | 0.81 |
| Very Low | $7.900 + 0.902 \cdot I_0 + 0.014 \cdot I_0^2$ | 2.075 | 0.00201 | 40 | 0.46 | 0.91 |

$R_0$, assumed which gives best fit in kilometers, $\Delta = (R^2 + R_0^2)^{1/2}$; r-sq, coefficiente of determination; $\sigma$, standard deviation in $I$.

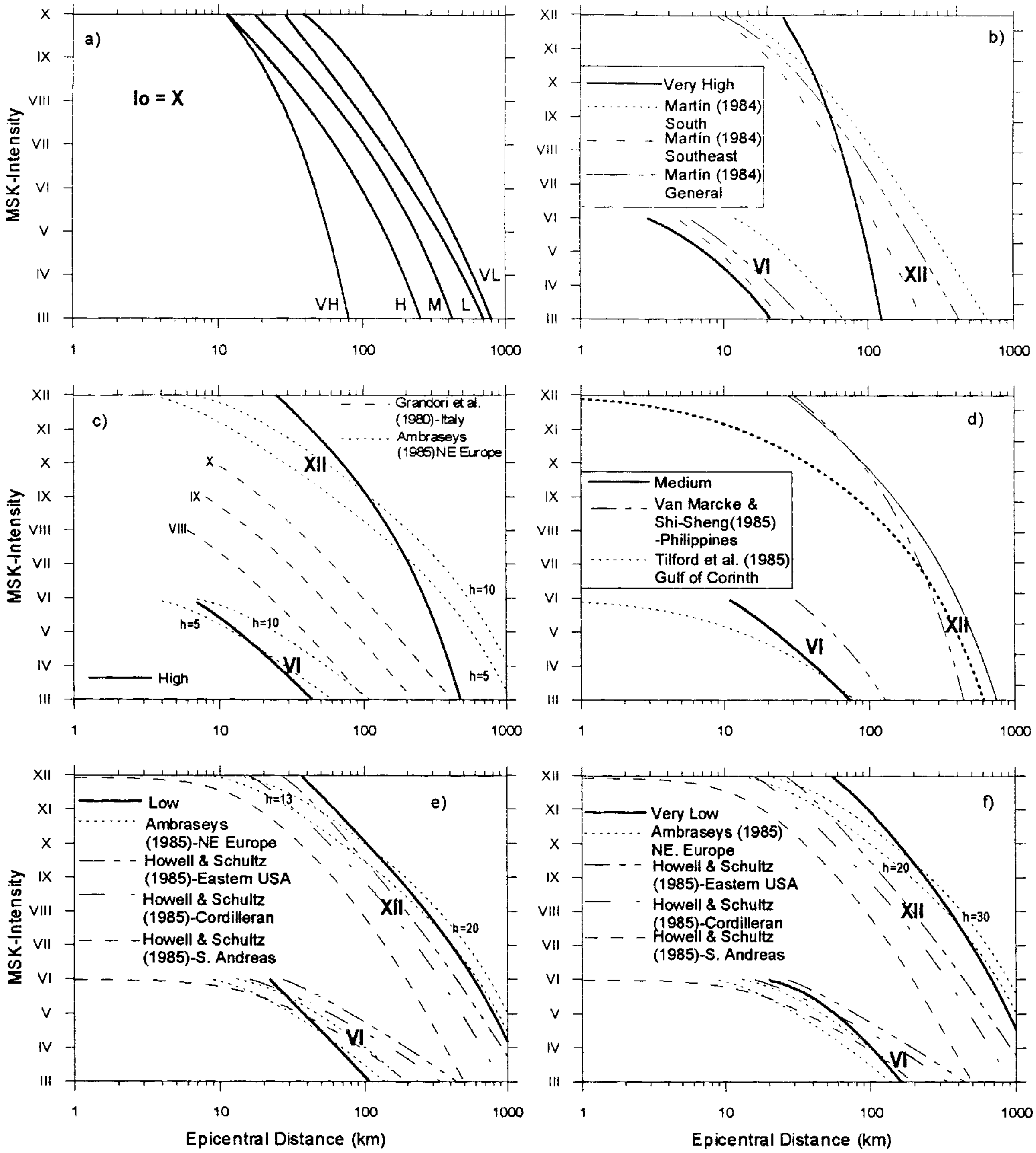


Figure 3. (a) Behavior of attenuation laws obtained in this work ($I_o$ = X). Comparison of our laws with those obtained by different investigators: (b) very high, Martín (1984); (c) high ($h$ = 5, 10 km) (Ambraseys, 1985; (1980); Grandori *et al.* 1987); (d) medium, Tilford *et al.* (1985) and Van Marcke and Shi-Sheng (1980); (e) low, Howell and Schultz (1975) and Ambraseys (1985) ($h$ = 13, 20 km); (f) very low, Howell and Schultz (1975) and Ambraseys (1985) ($h$ = 20, 30 km).

types of fit to real data, with our fit providing a clear improvement (see Fig. 5). It is obvious, therefore, that our regionalization is more accurate than previous ones. The new maps used and the improvement implied in the new physical model plainly reduce both statistical and modeling uncertainties (McGuire, 1993; Toro. *et al.,* 1997).

Of the four earthquakes with depths between 20 and 40 km that appear in Table 2, two have very low attenuation ($h$ = 23 and 30 km), another one low ($h$ = 27 km), and the last one high attenuation ($h$ = 28 km). The latter is located in the anomalous area of the Cape of San Vicente (high attenuation with shallow and intermediate earthquakes), so

Table 6
Attenuation Coefficient, $c$ (Different Evaluation Sources)

| Attenuation | $c$ (km$^{-1}$) | Method | Source |
|---|---|---|---|
| Very High | 0.0828 | $a_3/a_2$ | Table 5 |
| High | 0.0070 | $a_3/a_2$ | Table 5 |
| Medium | 0.0042 | $a_3/a_2$ | Table 5 |
| Low | 0.0012 | $a_3/a_2$ | Table 5 |
| Very Low | 0.0009 | $a_3/a_2$ | Table 5 |
| Eastern | 0.0031 | $a_3/a_2$ | Howell and Schultz (1975) |
| Cordilleran | 0.0063 | $a_3/a_2$ | Howell and Schultz (1975) |
| S. Andreas | 0.0150 | $a_3/a_2$ | Howell and Schultz (1975) |
| NW Europe | 0.0070 | Fit ($I_0 - I$, $\Delta$) | Ambraseys (1985) |

there is most probably an error in the calculated depth or in the isoseismal map (only two incomplete radii are given). Therefore, either the earthquakes are shallow or there is an error in the calculation of the mean radii. The remaining two deep earthquakes have low ($h = 60$ km) and very low ($h = 100$ km) attenuation.

Table 7 presents the distribution of the earthquakes used according to their size and tendency. Although it is not very clear, it seems that the large earthquakes associate better with low attenuation and the small ones with high attenuation. This also correlates very well with the regionalization since, given the seismotectonic characteristics of the area, the more energetic earthquakes are in the west and the less energetic ones in the east.

## Conclusions

From the analysis of the results obtained according to the methodology we have developed in function of the data used and of the proposed objectives, the following conclusions can be drawn.

First, an attenuation law with distance, based on the proportion between the intensity and the logarithm of the energy density and not between the intensity and a power of the energy density (lognormal distribution of the intensity), is best adapted to the intensity-attenuation data of the Iberian Peninsula.

Second, different degrees of attenuation of intensity with distance occur within the Iberian Peninsula. The greatest difference is found in the west, the Gulf of Cádiz, Portugal, and Galicia, with low and very low attenuation, and in the south and east, with high and very high attenuation. The Alborán Sea and the Central Pyrenees, with low and very low attenuation, are two areas that deviate from their

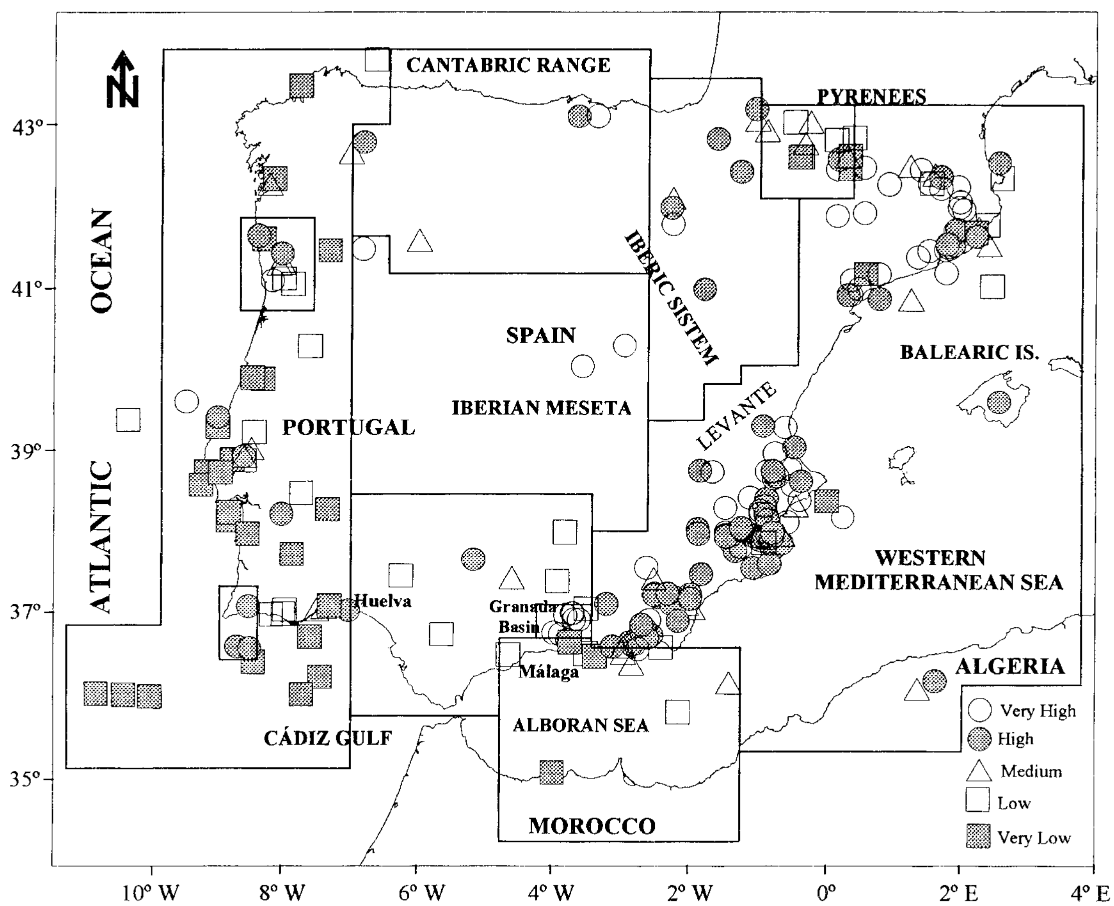


Figure 4. Regionalization of the isoseismal maps.

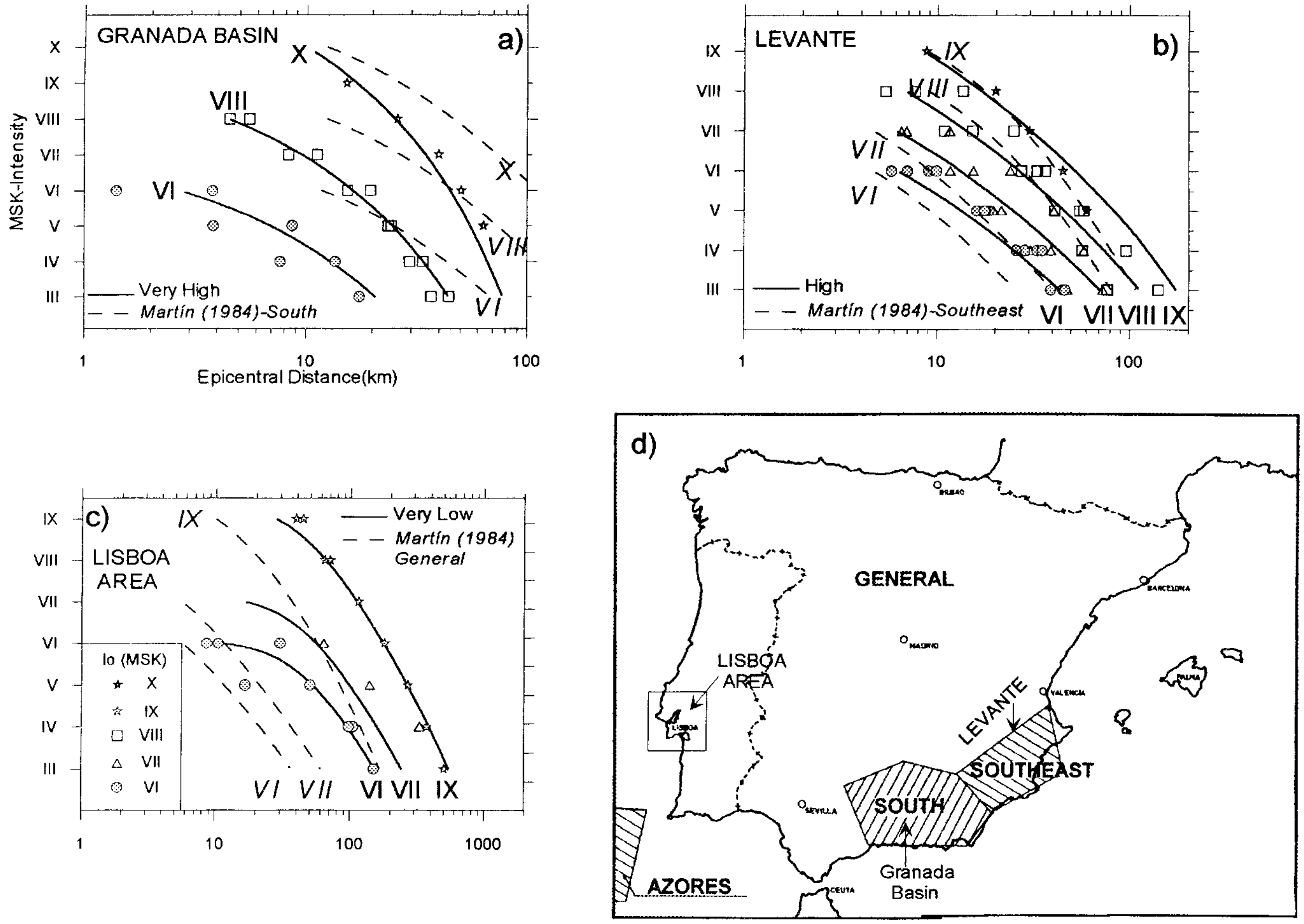


Figure 5. (a) Goodness of fit to Granada basin data and comparison with the Southern law obtained by Martin (1984); (b) goodness of the fit to Levante data and comparison with the southeastern law obtained by Martín (1984); (c) goodness of fit to the Lisbon-area data and comparison with the general law obtained by Martín (1984).

regional models. Note that these two areas have intermediate earthquakes.

Third, the attenuation laws obtained here, as well as the absorption coefficient, fall within the range of values given for other countries. Only the very high attenuation, characteristic of some Neogene areas in the south and east of the Iberian Peninsula, falls outside this margin, with an absorption coefficient 5 to 10 times higher than the other ones. This anomalous coefficient, together with their seismotectonic origin (high $Q_\beta^{-1}$, small magnitude, and shallow depth of the earthquakes) could be due to the greater vulnerability of buildings in these areas, giving rise to an overestimation of the epicentral and following intensities over very short distances.

From the tectonic point of view, the difference in the attenuation of the earthquakes of the Iberian Peninsula seems to be a function of the existence or not of a well-consolidated basement. In the western part of the Peninsula, in Portugal, the low attenuation seems to be due to the relatively cratonized Hercynian materials in the region. However, the eastern part of the Peninsula corresponds to materials affected by the Alpine orogeny—less cratonized and much more discontinuous in its blooming—due to which the attenuation is unavoidably much higher. The fact that earthquakes of larger magnitude are common in the west and not in the east may be an additional cause for these different types of attenuation.

Fourth, for the Iberian Peninsula, our attenuation laws represent an improvement with regard to previous works. We have obtained not only a greater regionalization and extension of the attenuation laws, but also a better fitting to the data. Thus, of the five regions: (general, south, southeast, south–southeast, and the Azores) given by Martin (1984) that affected only Spain and the southwest area of the San Vicente Cape, we have categorized nine regions affecting the whole Iberian Peninsula and the area from the Azores to the Gulf of Cádiz. Moreover, in the regression, including the coefficient due to the size of the earthquake in the fitting, we have divided the earthquakes according to their size, in-

Table 7
Percentages of Attenuation Tendencies for each Epicentral Intensity Degree

| $I_0$ | Very High | High | Medium | Low | Very Low |
|---|---|---|---|---|---|
| IV | 16% | 44% | 24% | 12% | 4% |
| V | 42% | 35% | 7% | 7% | 9% |
| VI | 37% | 19% | 11% | 14% | 19% |
| VII | 23% | 30% | 17% | 13% | 17% |
| VIII | 18% | 32% | 18% | 18% | 14% |
| IX | | 36% | 10% | 27% | 27% |
| X | 100% | | | | |
| XI | | | | 100% | |
| XII | | | | | 100% |

stead of separating them into only two groups (Martín, 1984): large ($I_o \geq$ VIII) and small ($I_o <$ VIII).

Fifth, the improvement in the fit of the data, obtained from a correct selection of the attenuation equation (reducing the modeling uncertainties), and for more real regionalization (reducing statistical uncertainties), indicate that these new laws guarantee a decrease in the uncertainties in seismic-hazard evaluations. Furthermore, the determination from the regionalization map of the percentages of the tendency type given in each area provides us with a quantitative real base of the certainty of each of the laws given in the corresponding area. This also allows a rigorous treatment of the uncertainties when using the Monte Carlo or logic-tree methods.

Sixth, the low correlations obtained have not allowed the introduction either of the depth or the magnitude in the attenuation laws. However, the low and very low attenuation in the Central Pyrenees and Alboran Sea could be associated to a greater depth ($20 < h < 180$ km) of the earthquakes in those areas.

Finally, when enough accelerogram data are available to perform accelerogram attenuations in the fitting of the data law, we must keep in mind the results of the intensity-attenuation law obtained in this research.

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