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Attenuation Relationship of Macroseismic Intensities in Central Europe

by Dietrich Stromeyer and Gottfried Grünthal

Abstract Two intensity-attenuation models for central Europe are presented in terms of mean relations and probability distributions of their variability. They are based on 14,714 intensity data points (IDP) of 31 earthquakes located in Germany, France, the Netherlands, and the Czech Republic. The attenuation relations are derived from crustal events with moment magnitudes M_w from 2.4 to 5.7, source depths h from 1 to 20 km, and epicentral distances R up to 400 km. Two different regression techniques are applied to derive the appropriate parameters a and b of the attenuation model $I_s - I = a \log(\sqrt{(R^2 + h^2)/h^2}) + b(\sqrt{R^2 + h^2} - h)$. A combined weighted least-squares regression utilizing intensities and distances of the IDP in a direct way provides $a = 2.80$ and $b = 0.0013$. A chi-square regression method for mean intensity-class distances estimated by intensity-level binning gives $a = 2.95$ and $b = 0.0025$. The variabilities in the derived models can be described by normal distributions centered at the mean relation with a standard deviation of about 0.7 intensity units. This specification of attenuation variability is one of the advantages compared with former relations, making both attenuation models especially suitable to be applied in seismic hazard assessments. The two independent models can be regarded as an epistemic uncertainty in logic tree based hazard calculations.

Introduction

Empirical attenuation relations of earthquake ground motion with distance are one of the basic ingredients in seismic hazard analyses. From an engineering point of view peak ground acceleration and, nowadays more and more, spectral accelerations are the appropriate physical parameters used in such attenuation models. While in the early times of hazard studies the intensity-based analyses were dominating, there also exist today special cases where seismic hazard in terms of macroseismic intensity is demanded, for example, when compatibility with previous results is required (Grünthal *et al.*, 1998) or when intensity-based risk assessments are intended (Grünthal *et al.*, 2006). The first and, for a long time, most important use of intensity-attenuation models refers to the macroseismic estimation of earthquake source parameters like location, depth, and magnitude of historical and macroseismically recorded modern events. In this case attenuation parameters are used either in a direct way (Blake, 1941; Sponheuer, 1960; Kárník, 1969; Ambraseys, 1985) or more implicitly in a common regression process to obtain the source parameters (Levret *et al.*, 1994; Bakun and Wentworth, 1997; Bakun and Scotti, 2006).

For central Europe north of the Alps only a few macroseismic attenuation investigations exist (Sponheuer, 1960; Ambraseys, 1985; Hinzen and Oemisch, 2001; Bakun and Scotti, 2006). The relation by Sponheuer (1960) in particular has frequently been used for seismic hazard assessments and macroseismic parameterizations. The parameters in this relationship were essentially moti-

vated by physical considerations of a point source with constant radiation and homogeneous wave spreading rather than by high quality macroseismic data. They were calibrated only by isoseismal radii of very few events. Recently, Bakun and Scotti (2006) developed an attenuation model for the stable continental region (SCR) of France. They are using median hypocentral distances for the integral level of intensity (intensity-level binning). We will compare their model and the attenuation relation of Ambraseys (1985) with our results.

The present study is based on a data set of carefully assigned intensity data points (IDP) of 31 earthquakes in central Europe, with most events located in Germany and some in the Czech Republic, the Netherlands, and eastern France. We use two different regression techniques to determine the model parameters. The first is a weighted least-squares method and the second a chi-square regression technique adapted based on the work of Stromeyer *et al.* (2004). Our procedures decouple the source term of the model, which describes the strength of an event given by its epicentral intensity I_0 or function containing magnitude and depth, from the intrinsic attenuation parameters. We use an intensity-calibrated event strength in our study because the historical earthquakes lacking original instrumental magnitudes form the larger part of our data base. This regression does not make use of isoseismals as they were and they are subjectively drawn with changing personal judgment from author to author.

Two different attenuation parameter sets are derived, depending on the way the IDP are incorporated into the regression process. These two independent attenuation

models can be regarded as epistemic model uncertainty, which can be handled in follow-up analyses, for example, in logic tree based probabilistic seismic hazard assessments. The two attenuation relations are presented as probability density functions that, contrary to most of the previous publications on this subject, consider all uncertainties of the regression process. Therefore, they can be directly incorporated into hazard analyses that nowadays involve comprehensive uncertainty quantifications.

Intensity Attenuation Model

Assigned intensity values are not only a function of earthquake magnitude and rupture distance. Azimuthal variations of the radiated energy due to different source mechanisms, anisotropic wave propagation, and local site effects can substantially influence the intensities. Therefore, our data were tested with respect to regional peculiarities and typical azimuthal dependencies checking computer generated contour plots of the corresponding IDP distributions and intensity distance plots, respectively. No such features could be revealed from the data and could be neglected in the models. This does not exclude that substantially larger data sets may exhibit such regional characteristics, as they are shown, for example, by the macroseismic fields of intermediate-depth earthquakes in the Vrancea region, Romania (M. Sørensen, D. Stromeyer, and G. Grünthal, unpublished manuscript, 2008). Hence, our model is isotropic (distance and depth dependent only). Furthermore, the earthquake source dimensions of the used events in the study area are small (moment magnitude $M_w \leq 5.7$) in comparison with the epicentral distances to the vast majority of the IDP sites. A point source can therefore be accepted.

Assuming a logarithmic proportionality between intensity and ground acceleration as well as a constant rate of absorption, a simple isotropic physically based intensity-attenuation model results. This was widely used, for example, in the equation based on work by Sponheuer (1960):

$$I' - I = km \log \frac{r}{r'} + ka \log(e)(r - r'), \quad (1)$$

where I and I' are two intensity assignments and r and r' are the associated hypocentral distances. m describes the geometric spreading, and a describes the absorption of energy; k is the proportionality factor between intensity and ground acceleration. The original form of equation (1) goes back to Kövesligethy (1906) and Jánosi (1907) on the basis of an empirical relationship (with $k = 3$) between intensity and peak ground acceleration established by Cancani (1904).

Besides Sponheuer (1960), a quoted references in

connection with equation (1) is Kárník (1969). The same structure of equation (1) results if one assumes a logarithmic proportionality between intensity and seismic energy density (Howell and Schulz, 1975).

The parameter k varies as a function of the supposed empirical connection between intensity and acceleration and ranges from 2 to 5 (Murphy and O'Brien, 1977). The theoretical value of the spreading coefficient for body waves is $m = 1$. Sponheuer (1960) adopted $km = 3$ for central Europe and Levret *et al.* (1994) obtained $km = 3.1$ - 3.4 from data fitting. The attenuation parameter a should be determined regionally and represents characteristic properties of the Earth's crust. It is typically in the range of $0.001 < a < 0.01$ (Sponheuer, 1960; Ahorner, 1983). Grünthal *et al.* (1998) derived $a = 0.002$ for their intensity-based seismic hazard assessments for Germany, Austria, and Switzerland.

The more than hundred year old relation type has been successfully applied in many studies, and we do not see any necessity to introduce a new structure. Physically based models of empirical relations, as in this case, should generally be preferred to purely mathematically constrained types.

Because macroseismic intensity is basically a descriptive value and not a measurement of a physical variable, relations between intensity and real physical parameters remain empirical and any interpretation is difficult. As a consequence, we have to determine the terms km and ka by a suitable regression technique from the IDP to establish an empirical attenuation relation according to the structure of equation (1). Obviously, only the products km and ka can be estimated independently. Therefore, it is common to define attenuation parameters $a = km$ and $b = ka \log(e)$ without a pure physical background.

The starting point for our regression procedure to estimate a and b is

$$I(a, b, R, h, I_*) = I_* - a \log \sqrt{\frac{R^2 + h^2}{h^2}} - b \left(\sqrt{R^2 + h^2} - h \right), \quad (2)$$

which is derived from equation (1) by replacing hypocentral distance r with epicentral distance R and depth h and by choosing a reference intensity I_* for the epicenter ($R = 0$). It should be explicitly stressed that I_* does not coincide with or mean the epicentral intensity I_0 . I_* is a calibration parameter specific for a single event and has to be determined parallel to a and b in the regression process. It can be understood as an estimator for I_0 , but the reverse is not the case. Recently, Pasolini *et al.* (2008) discussed the risk of biased estimates of I_0 and their influence on the intensity-attenuation relations used in Italy. The determination of the epicentral intensity strength I_* jointly with the attenuation parameters as mentioned pre-

viously is one of their recommendations. However, for Italy they prefer a sophisticated averaging of observed intensities in the vicinity of the epicenter.

I_* is the link between the attenuation parameters a and b and the strength of an earthquake. It can be replaced by any magnitude and depth dependent model $I_* = I_*(M, h)$, but in a first approximation the evaluation can be decoupled from the determination of the distance and depth dependent attenuation, similar to the standard methodology for the regression analysis of strong-motion data (Joyner and Boore, 1993). We do not resolve the structure of the magnitude based relation because the majority of the used events do not have an instrumental magnitude.

The parameters a and b describe the attenuation model. The estimation of I_* is only relevant for the calibration events in the regression process. When applying equation (2) I_* is given the appropriate epicentral intensity value for each earthquake.

Several procedures exist to estimate the epicentral intensity I_0 (Musson and Cenić, 2002). The accuracy of I_0 is usually comparable with that of a single well-determined IDP. Therefore, I_0 can be incorporated in the regression process as an additional data point like any other IDP. I_0 should not be used as a fixed calibration value for a specific event in any regression.

The relevance of the parameters a and b in equation (2) is changing with distance. While a controls the near-field behavior, b describes the far-field characteristics. This is depicted in Figures 1 and 2 for a typical parameter constellation in the study area. Figure 1 shows the joint

Figure 1. Joint confidence region of equation (2) (dark area) for the assumed parameter set $a^* = 3$, $b^* = 0.003$ (white cross), and $h = 15$ km. This area describes all sets (a, b) of the associated attenuation relation $I(a, b, R, h, I_*)$, which do not deviate (in the mean) more than 0.1 intensity units from $I(a^*, b^*, R, h, I_*)$. The white stars indicate two parameter sets ($a = 2.7$, $b = 0.001$ and $a = 3.3$, $b = 0.005$) inside the confidence region.

Figure 2. Intensity I versus epicentral distance R ($h = 15$ km, $I_* = 7$) for three parameter sets from the confidence region of $a^* = 3$ and $b^* = 0.003$. The most significant differences (≈ 0.2 intensity units) appear only at large distances (small intensities).

confidence region for a parameter set $a^* = 3$, $b^* = 0.003$, and a source depth $h = 15$ km defined by all possible intensity-attenuation models $I(a, b, R, h, I_*)$ with a mean deviation less than 0.1 intensity units from $I(a^*, b^*, R, h, I_*)$ in the distance range $0 \leq R \leq 300$ km. In Figure 2 the distance dependence of attenuation is plotted for a^* and b^* in comparison with two additional parameter sets from the 0.1 confidence region assuming an epicentral calibration value of $I^* = 7$. The trade-off between a and b is evident. The intensity differences are small compared with the intensity assignment in discrete classes. Only observations from larger events at greater distances can remove this ambiguity.

Calibration Earthquakes

The calculation of the intensity attenuation with distance is based on 31 seismic events in Germany, France, the Netherlands, and the Czech Republic. These events are located in the following source areas (Table 1): (a) the Lower Rhine embayment and the Middle Rhine area; (b) the uplift areas along the shoulders of the Upper Rhine graben consisting of the Black Forest, the Saulgau, and the Vosges in eastern France; (c) the seismic province of Saxothuringia with the swarm earthquake region in the Vogtland in the border region of Germany and the Czech Republic; (d) a mining induced event of the Werra potash mining district was also included in the set of thoroughly studied events with small focal depths.

As will be shown subsequently by comparing the regression errors and the differences $I_0 - I^*$, respectively, there are no significantly different characteristics in the

Figure 3. Distribution of the used IDP (circles) of the events (stars) listed in Table 1. The sizes of stars reflect their M_w magnitudes (2.4–5.7).

behavior of the data from the four different sets. The criteria for the event selection were the availability of a sufficient number of high quality macroseismic IDP and well-determined focal depths. We also tried to establish our analysis on a balanced areal distribution of IDP for each of the earthquakes within the three seismic provinces of the study area. Only IDP with an intensity assignment $I \geq 3$ are taken into account. The total set of 14,714 IDP are distributed over eastern parts of the UK, France, Belgium, the Netherlands, the northern parts of Switzerland, Germany, the western parts of Austria, the Czech Republic, and Poland (Fig. 3). The earthquakes with the largest number of IDP are the 1992 Roermond (Haak *et al.*, 1994), the 2003 Rambervillers (Cara *et al.*, 2005), and the 2004 Waldkirch (Brüstle *et al.*, unpublished manuscript, 2008) earthquakes. The data of the historical events in the

Lower and Middle Rhine area (number 1–4 in Table 1) were taken from Meidow (1995), the 1839 Unterriexingen data were taken from Fischer *et al.* (2001), and the 1872 Posterstein data were taken from Grünthal (1992) and Grünthal and Schwarz (2001). The data for all the other events are obtained from the macroseismic data archive of the former Central Institute of Physics of the Earth (ZIPE), Potsdam, which are currently being prepared for publication (G. Grünthal *et al.*, unpublished manuscript, 2008). The intensities for the events up to 1992 inclusive were assigned in terms of the Medvedev-Sponheuer-Kárník (MSK) scale (Medvedev *et al.*, 1964; Sponheuer, 1965). For the events from 1982 to 1989 the MSK scale in the version of the Ad-hoc Panel (1981) was used. For the 1992 Roermond earthquake the MSK scale and the European macroseismic scale 1992 (EMS-92), that is, the EMS-98 test version (Grünthal, 1993), were used in parallel for the German part of the shaken area. The two events in 2003 and 2004 were analyzed on the basis of the EMS-98 (Grünthal, 1998). No differences in the intensity assignments could be identified in these parallel applications. That personal judgment has a generally greater impact than the use of different modern 12-degrees scales on intensity assignments has been confirmed in a thorough study by Musson (2005).

The intensity database used contains full and half intensity values. Half intensity values are not defined in any macroseismic scale. They were introduced as uncertain values between two integer ones according to the guidelines to the EMS-98 (Grünthal, 1998). The overall number of half intensities is much lower than those of full intensities. We have visually checked their distribution in relation to integer intensity data (see, e.g., Fig. 4) and found that they can be considered in the analysis as they are.

Another issue that needs consideration is the incompleteness of IDP in the shaken area. It is obvious that data do not exist from all localities. According to our experience of collecting and interpreting macroseismic data, we conclude that the existing data are randomly distributed at least up to the intensity level of 4. Special problems can be connected with the macroseismic far field. Intensity 2 is totally underrepresented in the data set and was not considered. There is also no comprehensive coverage for not felt data. With respect to intensity 3, we observe two different characteristics. There are events where intensity 3 seems to be relatively incomplete with a tendency to underestimate the mean distance of these values from the epicenter. There are other events where we rather observe an opposite effect, that is, a certain number of observations at relatively large distances. This can have different explanations, such as a lack of observations to be assigned at intensity 2 and 3, a lack of not felt data, and for certain events a possible overestimation of intensities (as $I = 3$) due to a lack of a sufficient number of smaller macroseismic effects. Because

Figure 4. Typical example of IDP as a function of distance R for the 1935 Saulgau earthquake.

Table 1
Calibration Seismic Events and Parameter Values from Grünthal *et al.* (2008)

Date	Epicentral Area	Location		h (km)	Strength Parameters		Number of IDP*	
		Longitude (°N)	Latitude (°E)		I_0	M_w		
(a) Lower and Middle Rhine								
1	1755 December 27	Düren	6.25	50.80	18	7.0	5.4	55
2	1756 February 18	Düren	6.35	50.75	14	8.0	5.7	126
3	1846 July 29	St. Goar	7.68	50.15	10	7.0	5.1	129
4	1878 August 26	Tollhausen	6.55	50.93	9	8.0	5.5	254
5	1992 April 13	Roermond	5.95	51.17	18	7.0	5.3	2707
(b) Uplift areas along the Upper Rhine graben								
6	1839 February 7	Unterriexingen	9.02	48.90	8	5.0	3.6	24
7	1913 July 20	Ebingen	9.01	48.23	11	7.0	5.0	225
8	1935 June 27	Saulgau	9.47	48.04	9	7.5	5.4	357
9	1943 May 28	Ebingen	8.98	48.27	9	8.0	5.3	159
10	2003 February 22	Rambervillers	6.66	48.34	10	6.5	4.8	4491
11	2004 December 5	Waldkirch	8.00	48.10	9	6.0	4.6	2032
(c) Saxothuringian seismic province								
12	1847 April 7	S-Thuringia	11.14	50.46	17	6.0	4.2	46
13	1872 March 6	Posterstein	12.28	50.86	12	7.0	5.0	408
14	1897 October 29	Kraslice	12.48	50.35	8	6.0	4.1	67
15	1903 March 5	Kraslice	12.42	50.37	10	6.5	4.5	246
16	1903 March 6	Kraslice	12.47	50.34	14	6.0	4.2	210
17	1908 November 3	Kraslice	12.47	50.34	10	6.5	4.5	138
18	1908 November 4	Kraslice	12.47	50.34	9	6.5	4.5	97
19	1908 November 6	Kraslice	12.47	50.34	14	6.5	4.6	216
20	1908 December 19	Rochlitz	12.93	51.11	14	5.5	3.8	51
21	1914 June 27	Leipzig	12.43	51.36	8	6.0	4.1	144
22	1926 January 28	Stadtröda	11.76	50.88	4	6.0	3.9	135
23	1979 September 25	Schneeberg	12.69	50.81	5	5.0	2.7	23
24	1982 February 20	Leipzig	12.44	51.35	8	5.0	3.3	98
25	1983 November 5	Limbach-O.	12.68	50.81	3	4.5	2.4	20
26	1985 December 16	Vogtland	12.45	50.24	9	5.5	3.3	88
27	1985 December 20	Vogtland	12.44	50.23	9	6.0	3.2	102
28	1985 December 21	Vogtland	12.45	50.24	10	7.0	4.3	776
29	1985 December 23	Vogtland	12.45	50.24	9	6.0	3.4	432
30	1986 January 20	Vogtland	12.45	50.24	9	6.5	3.9	463
(d) Mining induced event								
31	1989 March 13	Völkershausen	10.05	50.80	1	8.0	5.4	395

*The number of IDP for $I \geq 3$.

these two data characteristics exist and according to our intention that the data be represented as they are, no modifications with respect to the lowest used intensity class were applied.

A certain number of the calibration earthquakes are either pre-instrumental or from the early instrumental period without sufficient depth control. Still, what has encouraged us to use these data is the general observation of good agreement of instrumentally and macroseismically determined focal depths in central Europe. Most of the depths estimated from macroseismic intensities were obtained by the same technique (Meidow, 1995), using average iso-seismal radii and the attenuation relation of Sponheuer (1960) which, as will be discussed later, is rather similar in its parameters to our derived models. A few examples may illustrate the good agreement of instrumentally and macroseismically derived focal depth in our study area. For the 1992 Roermond earthquake Camelbeeck *et al.* (1994) and Pelzig (1994) obtained instrumentally $h = 17$ km and $h = 19$ km, respectively; Haak *et al.* (1994) and Meidow (1995) estimated macroseismically $h = 18$ km and $h = 17$ km, respectively. Bock *et al.* (1996) determined by source

parameter modeling with synthetic seismograms of their depth phase part that $h = 7$ -10 km for the largest shocks of the Vogtland/western Bohemia 1985-1986 earthquake swarm ($h = 9$ km for the main shock 21 December 1985), which coincides very well with the macroseismic data. The same technique was applied to northern-central European earthquakes (Grünthal *et al.*, 2008) where the very good agreement between both depth determinations could also be confirmed. This is even true for the mining induced event in our study, which occurred in 0.9 ± 0.2 km depth below surface (Ahorne, 1998), whose macroseismically determined depth is 1.41 ± 0.45 km (Leydecker *et al.*, 1998).

Regression Techniques

Two different regression techniques were used to estimate the attenuation parameters a and b of model (2). The first one is a weighted least-squares method (WLQ) utilizing intensities and epicentral distances of the IDP in a direct way. This procedure determines not only the parameters a

and b of the attenuation model but simultaneously the calibration values I_* of all events. The basis of the second technique is a chi-square regression method (CHIQ) only for the parameters a and b adapted in the manner of Stromeier *et al.* (2004) using mean intensity-class radii estimated by intensity-level binning.

Weighted Least-Squares Regression

Fitting a , b , and I_* of equation (2) by a least-squares regression is equivalent to a maximum likelihood estimation using input data with normally distributed errors. This minimization problem can be written in standard form as

$$\min_x \|\mathbf{I} - \mathbf{A}\mathbf{x}\|^2. \quad (3)$$

For a single event, $\mathbf{x}$ is the parameter vector to be estimated $\mathbf{x}^T = (I_*, a, b)^T$, $\mathbf{I}^T = (I_1, \dots, I_n)^T$ is the vector of all n observed intensities, and $\mathbf{A}$ is the appropriate $(n, 3)$ design matrix with columns

$$A_{i,1} = 1, \quad A_{i,2} = -\log \sqrt{\frac{R_i^2 + h^2}{h^2}}, \quad (4)$$

$$A_{i,3} = -\sqrt{R_i^2 + h^2} + h.$$

R_i denotes the epicentral distance of I_i .

Usually intensity data sets are characterized by a great variety in the number of observations of the different intensity levels. In most cases the highest intensities are undersampled as they are restricted to a relatively small area compared to the lower intensities at larger epicentral distances. Additionally, the data can be incomplete relative to the lowest used intensity level, intensity 3. To overcome such a disproportion, a weighting scheme has been applied where each intensity class has been assigned the same weight in the regression, regardless of the number of observations within the class. Therefore, equation (3) has been expanded to a linear weighted least-squares problem:

$$\min_x \|\mathbf{W}^{-1}(\mathbf{I} - \mathbf{A}\mathbf{x})\|^2. \quad (5)$$

The weighting matrix $\mathbf{W}$ has only diagonal entries. The values are chosen in such a way that (1) they are equal for all intensities in one intensity class and (2) the sum of squared inverse weights is equal for all intensity classes (classes are identically weighted). This procedure defines the weights up to an arbitrary constant scaling factor that does not influence the regression solution $\mathbf{x}$ but is important for calculating uncertainties both for the estimated parameters and for a new intensity predicted by the model. Therefore, a rescaling of $\mathbf{W}$ has been performed in

such a way that the mean weighted and unweighted residuals are equal,

$$\mathbf{W}_{\text{new}} = \mathbf{W}_{\text{old}} \frac{\|\mathbf{W}_{\text{old}}^{-1}(\mathbf{I} - \mathbf{A}\mathbf{x})\|}{\|\mathbf{C}\|}. \quad (6)$$

For a combined regression of K earthquakes with a total number of N IDP $\mathbf{I}$, $\mathbf{W}$ and the last two columns of $\mathbf{A}$ are the stacked versions of all individual contributions. Additionally, $\mathbf{A}$ has been enlarged to a $(N, K + 2)$ matrix where the K first columns are zero except for the rows connected to the calibration parameter I_{*k} ($1 \leq k \leq K$) of the k th event. Now the parameter vector $\mathbf{x}$ reads $\mathbf{x}^T = (I_{*1}, \dots, I_{*K}, a, b)^T$.

The uncertainties in the estimated parameter $\mathbf{x}$ and in predicting a new intensity I for given predictor values I_* (or I_0), R , and h are connected with the covariance matrix $\mathbf{Cov}(\mathbf{x}; \mathbf{x})$ of the parameter estimates and the mean squared regression error $\bar{\sigma}$:

$$\mathbf{Cov}(\mathbf{x}, \mathbf{x}) = (\mathbf{A}^T \mathbf{W}^{-2} \mathbf{A})^{-1}, \quad \bar{\sigma}^2 = \frac{\|\mathbf{I} - \mathbf{A}\mathbf{x}\|^2}{N - M}. \quad (7)$$

For a specified level of certainty α , the confidence bounds $\mathbf{x}^c$ for the fitted parameters $\mathbf{x}$ are given by

$$x_i^c = x_i \pm t^{-1}\left(\frac{1-\alpha}{2}, N - M\right) \sqrt{\text{Cov}(x_i, x_i)}, \quad (8)$$

$i = 1, \dots, M \dots,$

where $t^{-1}(p; \nu)$ is the inverse of the cumulative t distribution for the corresponding probability p and ν degrees of freedom and $M = K + 2$ is the number of model parameters. For $\nu \geq 40$, as in our case, $t^{-1}(p; \nu)$ can be approximated by the inverse of the cumulative standard normal distribution of p . Then a certainty level of 68.3% ($\alpha = 0.683$) corresponds to the standard deviation (1σ) of normally distributed errors.

Of special interest in the case of the usage of the attenuation equation (2) is the uncertainty of a new intensity prediction I by the regression model in seismic hazard calculations. For given predictor values I_* , R , and h , this error can be expressed by

$$I_{\text{error}} = t^{-1}\left(\frac{1-\alpha}{2}, N - 2\right) \sqrt{\bar{\sigma}^2 + \mathbf{y}^T \mathbf{C}(\mathbf{x}, \mathbf{x}) \mathbf{y}}. \quad (9)$$

where the vector $\mathbf{y}$ is the Jacobian of equation (2), $\mathbf{y}^T = \left(\frac{\partial I}{\partial a}, \frac{\partial I}{\partial b}\right)^T$, and $\mathbf{C}(\mathbf{x}, \mathbf{x})$ denotes the $(2, 2)$ submatrix of $\mathbf{Cov}(\mathbf{x}, \mathbf{x})$ connected to the attenuation parameters a and b :

$$\mathbf{C}(\mathbf{x}, \mathbf{x}) = \begin{bmatrix} \text{Cov}(a, a) & \text{Cov}(a, b) \\ \text{Cov}(b, a) & \text{Cov}(b, b) \end{bmatrix}. \quad (10)$$

The calibration values ($I_{*1}, \dots, I_{*K}$) of the regressed events have no influence on the estimation of the uncertainty of a new intensity by equation (2).

Chi-Square Regression

This approach uses the mean radii R and their standard deviations $\sigma(R)$ of the intensity classes calculated from the epicentral distances of the IDP (intensity-level binning). We assume $\sigma(R)$ to be a global measure for the uncertainties of the IDP. Additionally, the standard errors $\sigma(h)$ of the source depths are taken into consideration; they can vary from event to event. This regression technique requires the complete range of used intensity classes. Therefore, we define only classes for intensities $I < I_0$ with a minimum number of 10 IDP. An example is given in Figure 4 for event number 8 in Table 1 (Saulgau).

The relation

$$\Delta I(\bar{R}, R) = a \log \frac{\sqrt{R^2 + h^2}}{\sqrt{\bar{R}^2 + h^2}} + b \left(\sqrt{R^2 + h^2} - \sqrt{\bar{R}^2 + h^2} \right), \quad (11)$$

which describes the difference $\Delta I(\bar{R}, R) = I(a, b, \bar{R}, h) - I(a, b, R, h)$ between two intensity classes with radii $\bar{R}$ and R , is more suitable than equation (2) for the estimation of the attenuation parameters a and b in the case of using intensity radii. This model type has the advantage that it is independent of any calibration term I_* or epicentral intensity I_0 . Furthermore, this form permits a regression without absolute intensity assignments to the mean radii themselves, that is, only the well-defined intensity-class difference $\Delta I(\bar{R}, R)$ is of importance.

Fitting equation (11) to n mean intensity radii R_i of a single earthquake with the consideration of the standard errors $\sigma(R_i)$ and $\sigma(h)$ requires the minimization of the chi-square merit function

$$\chi^2(a, b) = \sum_{i=1}^{n-1} \sum_{j=i+1}^n \frac{[f(R_i, R_j, h, a, b)]^2}{\sigma^2[f(R_i, R_j, h, a, b)]}, \quad (12)$$

where the summation runs over all $n(n-1)/2$ intensity class differences. f denotes the residuals of model (11) at mean radii $R = R_i$ and $\bar{R} = R_j$ of the associated intensity classes and fixed parameters a and b . These residuals are weighted with $\sigma^2(f)$, the variance of f with respect to the errors of R_i , R_j , and h . For independent data $\sigma^2(f)$, not

known in advance, can be approximated by

$$\sigma^2(f) \approx \left(\frac{\partial f}{\partial R_i} \right)^2 \sigma^2(R_i) + \left(\frac{\partial f}{\partial R_j} \right)^2 \sigma^2(R_j) + \left(\frac{\partial f}{\partial h} \right)^2 \sigma^2(h), \quad (13)$$

assuming that the intensity difference between two classes is exactly known, $\sigma[\Delta I(R_i, R_j)] = 0$.

For comparison with the matrix form of the WLQ of the Weighted Least-Square Regression section, we define

$$\mathbf{f} = \Delta \mathbf{I} - \mathbf{A} \mathbf{x}. \quad (14)$$

where $\mathbf{f}$ is the residuum vector with $n(n-1)/2$ components. The ordering of $\mathbf{f}$ can be arbitrary, dependent on the indices i and j of equation (12), but $\Delta \mathbf{I}$ and $\mathbf{A}$ have to be arranged in the same way. $\Delta \mathbf{I}$ is the vector of intensity-class differences, $\mathbf{A}$ is the appropriate $[n(n-1)/2, 2]$ design matrix with columns

$$A_{k,1} = \log \sqrt{\frac{R_i^2 + h^2}{R_j^2 + h^2}}, \quad A_{k,2} = \sqrt{R_i^2 + h^2} - \sqrt{R_j^2 + h^2}, \quad (15)$$

and $\mathbf{x}^T = (a, b)^T$ is the parameter vector to be estimated. The denominator of equation (12) can be understood as the diagonal entries of an $[n(n-1)/2, n(n-1)/2]$ weighting matrix. Then,

$$\chi^2(a, b) = \left\| \mathbf{W}^{-1}(\Delta \mathbf{I} - \mathbf{A} \mathbf{x}) \right\|^2. \quad (16)$$

For a number of earthquakes, K , the resulting merit function is the sum of all contributions:

$$\chi^2(a, b) = \sum_{k=1}^K \chi_k^2(a, b). \quad (17)$$

Hence, the chi-square minimization

$$\min_{a,b} \chi^2(a, b) = \min_{\mathbf{x}} \left\| \mathbf{W}^{-1}(\Delta \mathbf{I} - \mathbf{A} \mathbf{x}) \right\|^2 \quad (18)$$

is of the same form as equation (5). However, in contrast to the weighted least-squares equation, the weighting matrix $\mathbf{W}$ depends also on the unknown parameters $\mathbf{x}$. Therefore, equation (18) defines a nonlinear minimization problem that has to be solved numerically by an appropriate iterative procedure.

The uncertainties in the estimated parameter vector $\mathbf{x} = (a, b)^T$ are related to the probability distribution $\Delta \chi^2(\mathbf{x}) = \chi^2(\mathbf{x}) - \chi^2(\mathbf{x}^*)$, where $\mathbf{x}^*$ is the minimum solution of equation (18). The 68.3% region is characterized by the contour of $\Delta \chi^2(\mathbf{x}) \approx (\mathbf{x} - \mathbf{x}^*) \mathbf{Cov}(\mathbf{x}^*, \mathbf{x}^*) (\mathbf{x} - \mathbf{x}^*) = 2.3$. The covariance matrix cannot be expressed in analytical form

because of the nonlinearity of $\Delta\chi^2$ with respect to $\mathbf{x}$. Therefore, a Monte Carlo simulation is used to estimate reliable values of $\mathbf{Cov}(\mathbf{x}^*, \mathbf{x}^*)$. As will be shown subsequently the resulting empirical probability of $\mathbf{x}^*$ is normally distributed and equation (8) remains valid for the standard errors of a and b replacing the covariance matrix of the least-squares procedure by its chi-square analogue and setting the degree of freedom $M = 2$. The standard error of a new predicted intensity I by equation (9) requires the calculation of a mean squared regression error $\bar{\sigma}$. With the design matrix $\mathbf{A}$ and the intensity vector $\mathbf{I}$ from the Weighted Least-Squares Regression section

$$\sigma^2 = \frac{\|\mathbf{I} - \mathbf{A}\mathbf{z}\|^2}{N - 2}, \quad (19)$$

with $\mathbf{z}^T = (I_{*1}, \dots, I_{*K}, a, b)^T$. The involved individual calibration values I_{*k} for each event have to be estimated as

$$I_{*k} = \frac{1}{n_k} \sum_{i=1}^{n_k} \left[I_{ki} + a \log \sqrt{\frac{R_{ki}^2 + h_k^2}{h_k^2}} + b \left(\sqrt{R_{ki}^2 + h_k^2} - h_k \right) \right], \quad (20)$$

where I_{ki} , R_{ki} , and h_k are the appropriate input values of the k th event ($k = 1, \dots, K$) with n_k IDP.

Results and Discussion

We derived intensity-attenuation models for central Europe using two different regression techniques, a combined WLQ and a combined CHIQ, in terms of mean relations and probability distributions of their variability. WLQ was performed for all IDP with intensity assignment ≥ 3 from 31 events. The input data set was enlarged by the epicentral intensities I_0 . The integration of I_0 in the regression process is made without fixing I_* to I_0 , that is, I_0 does not have the function of a fixed calibration value. The calculated attenuation parameters are summarized in Table 2 together with the mean ~~squared~~ regression error $\bar{\sigma}$, the covariance matrix $\mathbf{C}$, and the mean and standard deviation of all differences $I_0 - I_*$.

It is notable that the attenuation parameters estimate

the epicentral intensity I_0 by I_* surprisingly well for all the four different source regions (Table 1). There are only three outliers, event number 8 with $I_0 - I_* = -0.45$, event number 13 with $I_0 - I_* = -0.61$, and event number 22 with $I_0 - I_* = 0.44$. Also the average misfit, separately calculated for the source regions, is sufficiently compatible with the overall regression error $\bar{\sigma}$ (Table 3). The regularity of the I_0 estimation and the consistent regional average misfit indicate that the attenuation of the four source areas can be sufficiently described by a joint model. This does not mean that regional relations that would better fit the data do not exist. However, in the frame of an accepted mean misfit of about 0.7 intensity units all regions show similar characteristics.

The uncertainties of a and b are visualized in Figure 5 by means of the 68% confidence regions of individual and joint parameter errors. The mean regression relation and its 68% confidence bounds for a predicted intensity (see equation 9) are shown in Figure 6 assuming a reference intensity $I_* = 7$ and a source depth $h = 15$ km. These bounds are mainly determined by the mean regression error $\bar{\sigma}$, whereas the influence of the distance and, therefore, the parameter errors expressed by the covariance $\mathbf{C}$ are negligible. This means that any parameter set a and b from the joint confidence region of Figure 5 predicts practically the same probability distribution of I .

The combined CHIQ was also applied to all IDP with intensity assignment ≥ 3 . Mean radii R and their standard deviation $\sigma(R)$ were calculated only for intensity classes $I < I_0$ with a minimum number of 10 IDP. As a consequence, three small events with very few IDP (numbers 6, 23, and 25 of Table 1) do not contribute to the regression. They do not have at least two complete intensity classes to build appropriate differences used in equation (11). A critical step in this approach is the estimation of the uncertainties $\sigma(h)$ of the source depths h . For the tectonic events, selected also with respect to reliable depth determinations, we set $\sigma(h) = 3$ km. The depth of the mining event (number 31) is much better known and $\sigma(h) = 0.1$ km can be set. The regressions parameter a and b and their probability distributions were calculated using Monte Carlo simulations. The minimization problem (equation 18) was solved using 50,000 training sets, which were generated assuming normally distributed mean radii R and source depths h . The resulting values for

Table 2

Estimated Regression Parameters

Method	a	b	$\bar{\sigma}^*$	$C(a; a)^\dagger$	$C(a; b)^\dagger$	$C(b; b)^\dagger$	$I_0 - I_*^z$
WLQ	2.80	1.26×10^{-3}	0.687	7.47×10^{-4}	-4.14×10^{-6}	3.23×10^{-8}	0.09 ± 0.22
CHIQ	2.95	2.52×10^{-3}	0.712	5.84×10^{-3}	-2.21×10^{-5}	1.14×10^{-7}	0.11 ± 0.27

* Mean ~~squared~~ regression error $\bar{\sigma}$.

† Covariance matrix $\mathbf{C}$ for the WLQ and CHIQ regression techniques.

‡ The mean and the standard deviation of all residuals between published epicentral intensities I_0 and their estimation I_* in the regression process.

Table 3

Average Misfit in the Four Different Source Regions for the Regression Solutions WLQ and CHIQ

Method	Region a	Region b	Region c	Region d
WLQ	0.637	0.713	0.668	0.717
CHIQ	0.643	0.748	0.688	0.802

a and b are compatible with a normal distribution, as was checked by the 5% significance test for the negative hypothesis.

The complete output of the chi-square regression is summarized in Table 2. The attenuation parameters a and b obviously differ from the weighted least-squares solution, but the mean squared regression error is only slightly higher. Also the mean and standard deviation of the residuals between I_0 and its estimation I_* are compatible. Larger differences $I_0 - I_*$ were obtained for four events (numbers 7 [0.47], 8 [0.51], 9 [0.74], and 31 [0.59]). All these residuals are positive and relate to the stronger events with $M_w \geq 5.0$ that were felt up to large distances. This could indicate that CHIQ overestimates b , which is responsible for the far-field attenuation. Whereas the weighted least-squares approach adequately considers the IDP near the epicenter, especially if I_0 is taken as an additional intensity point; the chi-square regression could underestimate these data because only classes with at least 10 IDP were used. The bulk of the covariances $\mathbf{C}$ are about 4-5 times higher compared with the weighted least-squares approach, which results in larger corresponding confidence regions, see Figure 7. However, derived relative values as the correlation coefficient

Figure 5. Sixty-eight percent confidence regions of individual (dark gray) and joint (light gray) errors of a and b (white cross) estimated by the weighted least-squares regression. The individual errors $\sigma(a)$ and $\sigma(b)$ (see equation 8) are connected with the projection of the dark gray region onto the a and b axis, respectively.

Figure 6. Mean attenuation relation (dark curve) from weighted least-squares regression and its 68% confidence region (gray area) for a predicted intensity assuming a reference intensity $I_* = 7$ and a source depth $h = 15$ km. White curves (only visible at great distances) mark the range for the impact of the parameter uncertainties with distance $R = 0$.

are nearly the same. $p \approx -0.85$ for both techniques demonstrating the strong correlation between a and b as stressed at the end of the Intensity Attenuation Model section.

For intensity-based hazard calculations, the probability distribution for predicted intensities is one of the most important parameters. This distance dependent distribution is shown in the form of the confidence bounds in Figure 8. Similar to WLQ this distribution is mainly con-

Figure 7. Sixty-eight percent confidence regions of individual (dark gray) and joint (light gray) errors of a and b (white cross) estimated by the CHIQ. The individual errors $\sigma(a)$ and $\sigma(b)$ (see equation 8) are connected with the projection of the dark gray region onto the a and b axis, respectively.

Figure 8. Mean attenuation relation (dark curve) from CHIQ and its 68% confidence region (gray area) for a predicted intensity assuming a reference intensity $I_* = 7$ and a source depth $h = 15$ km. White curves mark the range for the impact of the parameter uncertainties with distance $= 0$).

trolled by the mean regression error. The uncertainties of the parameter errors are of minor importance. Comparing the regional average misfit (Table 3), the CHIQ solution is also consistent for all different source regions. However, it should be kept in mind that this statement refers to the joint regression error as discussed previously for the WLQ solution.

There are remarkable differences between the two derived attenuation relations, even though the performance with respect to the observed intensities is nearly the same.

Figure 9. Attenuation relations of this study compared with the model of Bakun and Scotti (2006) and Ambraseys (1985). The relations were adjusted to fit the IDP of the Roermond 13 April 1992 earthquake. The squares mark the median radii of the respective intensity classes, and the horizontal bars visualize the standard deviation centered at the mean radii.

Particularly, the extrapolations to large distances ($R \geq 300$ km) differ by one unit of or more. This discrepancy cannot be resolved because there are no reliable IDP with intensity ≥ 3 at such distances, not even for the Roermond 13 April 1992 earthquake (Fig. 9), one of the strongest and best documented events of our study. Improvements can only be expected from data of considerably stronger earthquakes, say $M_w \geq 6$. Therefore, we have to regard the differences as intrinsic to the employed regression. WLQ uses the IDP in a direct way, whereas CHIQ homogenizes the data by a special intensity-level binning process. There are no sufficient reasons to prefer or reject either procedure.

To compare our relations with other attenuation models derived for the study area, we have rearranged the models into the form of equation (2). Equation (1) of Sponheuer (1960) ($k = 3$, $m = 1$) shows the best fit with data for $\alpha = 0.002$ (Meidow and Ahorner, 1994; Grünthal *et al.*, 1998), which leads to $a = km = 3$ and $b = k \log(e)\alpha = 0.0026$. This relation is very close to the CHIQ solution. The agreement is good over all distances. The model of Bakun and Scotti (2006) for the SCR of France neglects the parameter b , which results in a large attenuation value $a = 3.37$. The source calibration term I_* in equation (2) reads $I_* = 4.48 + 1.27M_w - 3.37 \log(h)$ in this case. Another SCR model of Europe, which is based on a sufficiently large data set, is that by Ambraseys (1985) for northwest Europe. It is not based on IDP but traditionally on iso-seismal radii and does not include any uncertainty estimates. The attenuation parameters of this model are $a = 2.85$ and $b = 0.0024$. It overestimates the intensity at the epicenter by a small constant value, $I_* = I_0 + 0.22$. Also this model fits ours quite well.

Figure 9 shows both relations in comparison with the WLQ and the CHIQ solutions adjusted to the Roermond event. The IDP are condensed to median radii in the manner of Bakun and Scotti (2006) and the standard deviations centered on the mean radii (CHIQ) for each intensity level. The differences between all four models are smaller than or equal to 0.7 intensity units, the uncertainty of a predicted value by our relations. Up to 250 km distance, the model of Bakun and Scotti (2006) approximately corresponds to the mean of WLQ and CHIQ. In this distance range the attenuation by the model of Ambraseys (1985) is obviously smaller, which originates in the constant parameter value of 0.22. At distances larger than 350 km the uncertainty of the far-field parameter b in all relations becomes obvious. There is hardly any chance to decrease this uncertainty for SCRs of Europe, where the probability of earthquakes with magnitudes of the largest events we have studied is very low. On the other hand, this uncertainty might already be intrinsic, that is, aleatory.

The variabilities in the derived mean relations WLQ

and CHIQ can be described by normal distributions centered at the mean relation with a variance $\sigma^2 = \bar{\sigma}^2 + \mathbf{y}^T \mathbf{C}(\mathbf{x}, \mathbf{x}) \mathbf{y}$ (see equation 9). The term $\bar{\sigma}^2$ is dominant because the impact of the distance and the covariance $\mathbf{C}$ is negligible. Often the approximation $\sigma = \bar{\sigma}$ can be made without loss of accuracy. The specification of the attenuation variability in terms of probability distributions is one of the advantages compared with former relations, making both attenuation models especially suitable to be applied in seismic hazard assessments. The two independent attenuation models can be regarded as an epistemic uncertainty in logic tree based hazard calculations.

This aspect of epistemic uncertainty can even be broadened. Because, as described previously, the SCR models for central Europe (this study), western Europe by Bakun and Scotti (2006), and northwestern Europe by Ambraseys (1985) behave similarly, we conclude that all these models fit the central European data. This suggests that the intensity attenuation is comparable for earthquakes throughout the SCR of Europe. Therefore, these four attenuation models could be used to represent the epistemic uncertainty for these areas. This would require converting the uncertainty representation in Bakun and Scotti (2006) into the form given for our WLQ and CHIQ models, whereas for the model of Ambraseys (1985) such an uncertainty quantification is missing and needs to be introduced.

Data and Resources

Most data used in this article came from published sources listed in the references. There are other data, which are now under preparation for publication. Currently these data are available from the authors upon request.

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