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Attenuation of Macroseismic Intensity: A New Relation for the Marmara Sea Region, Northwest Turkey

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Abstract Prediction equations for macroseismic intensity are the backbone of seismic hazard assessment, of source parameter estimation, and of shake map generation in cases where an output in terms of intensity is desired. This is especially required when a direct relation to the damage associated with ground shaking is of interest or if ground shaking estimates will be used for informing nonseismologists such as emergency response teams or the general public. In the current study we derive ground-motion prediction equations for macroseismic intensity valid for the Marmara Sea region, northwest Turkey. The relations have a physical basis and are easy to implement for the user. In one relation, the finite extent of the fault rupture is accounted for by defining distance as the Joyner-Boore distance leading to the relation

$$I_S = 0.376 M_w + 5.913 - 2.656 \log \sqrt{\frac{R_{JB}^2 + h^2}{h^2}} - 0.0020 \left(\sqrt{R_{JB}^2 + h^2} - h \right).$$

where M_w is the moment magnitude, R_{JB} is the Joyner-Boore distance, and h is the hypocentral depth. Furthermore, a relation based on the epicentral distance (R_{epi}) is derived for application in cases where the extent of the fault plane is unknown:

$$I_S = 0.793 M_w + 3.417 - 2.157 \log \sqrt{\frac{R_{epi}^2 + h^2}{h^2}} - 0.0065 \left(\sqrt{R_{epi}^2 + h^2} - h \right).$$

The relations are valid for the ranges $5 \leq I \leq 10$, $5.9 \leq M_w \leq 7.4$, and $R \leq 350$ km. It is shown that inclusion of the rupture dimensions leads to an improvement in the ability of the relation to fit observations in the near field for large earthquakes. Comparison to already existing intensity prediction equations for the region shows that the new relations provide better estimates of the macroseismic intensity distribution, especially in the region near the rupturing fault plane.

Introduction

When generating shake maps with the purpose of earthquake early warning or rapid earthquake response, an essential parameter is the attenuation of seismic waves in the area of interest. Such ground-motion prediction equations (GMPEs) are also of crucial importance for seismic hazard assessment. GMPEs are traditionally given in terms of recorded ground-motion parameters, for example, peak ground acceleration, peak ground velocity, or response spectral acceleration based on recorded strong-motion data (e.g., see Joyner and Boore [1993]; Campbell [1997] and other articles in the same issue; Ambraseys and Douglas [2003]). An important element connected to such relations is the associated uncertainty caused by, for

example, spread in the data, assumption of the functional form of the relation or of the source being a point source, or lack of knowledge of earthquake faulting parameters. When studying the damage potential of large earthquakes, GMPEs based on recorded ground motions have two drawbacks. First, the availability of strong-motion recordings is limited and therefore one is often forced to apply GMPEs based on recordings from different areas with similar tectonics. Second, there is no straightforward way to associate the recorded ground motions with damage, which is a complex function of ground-motion level, duration, local site conditions, and building vulnerability.

As an alternative, to overcome these problems, ground-motion attenuation can be expressed in terms of macroseismic intensity. Intensities have the major advantage of

much better availability, as data are dependent on the availability of people and a built environment rather than on instrumentation and therefore can be sampled closer and as far back in time as historical records allow. Furthermore, the macroseismic intensity is assigned based on the observed ground shaking and damage, and thereby it can be directly related to the damage potential of future earthquakes. Another advantage is that intensity data are easily understandable for nonseismologists and easily convertible for risk management teams.

In the present study we derive a new attenuation model for macroseismic intensity and apply it for the Marmara Sea region, northwest Turkey. Here, especially the city of Istanbul is under a significant seismic hazard and potential seismic risk due to the likely rupture of a 100-150 km long segment along the North Anatolian fault zone (NAFZ) just south of the city within the lifetime of the present city environment (e.g., Parsons, 2004). The capacity of the NAFZ for generating large earthquakes was latest manifested by the occurrence of the 1999 M_w 7.4 Izmit earthquake. This event caused damage over an extended region around the rupturing fault plane and lead to the loss of more than 18,000 lives (Barka *et al.*, 2002).

Macroseismic intensity prediction equations have previously been derived by Erdik and Eren (1983) and Erdik *et al.* (1985), valid for the NAFZ in general. These relations give the Medvedev-Sponheuer-Kárník (MSK-64) intensity as a function of M_S and the natural logarithm of the rupture distance. More recently, Böse (2006) derived a GMPE for macroseismic intensity in the Marmara Sea region based purely on simulated ground motions that are converted into intensity following Sokolov (2002). Ambraseys (2001) presents a relation giving M_S as a function of intensity and distance. This relation is based on Greek and western Turkish earthquakes and is valid for far-field conditions.

The attenuation model derived in this study takes into account the finite extent of the fault plane and represents site intensities as a function of fault distance, event depth, and moment magnitude. Additionally, we derive a model under a point-source assumption for application in cases where the extent of faulting is not known. It is aimed at deriving simple, physically based relations that are easy to implement for the user. We base our relations on available macroseismic information for significant earthquakes in the region.

Regression Method

The regression for intensity prediction equations is based on the least-squares regression method of Stromeyer and Grünthal (2009) for the well-established and physically based attenuation model for point sources (Spon-

heuer [1960]; later adopted by, e.g., Kárník [1969] and Howell and Schulz [1975]):

$$I = I_0 - a \log \sqrt{\frac{R^2 + h^2}{h^2}} - b \left(\sqrt{R^2 + h^2} - h \right). \quad (1)$$

In this expression, I_0 is the epicentral intensity, R is the epicentral distance, and h is the focal depth, usually taken as the hypocenter depth. The term $\log$ refers to the decadic (base 10) logarithm. The first term $a \log \sqrt{(R^2 + h^2)/h^2}$ describes the geometrical spreading (having its main effect at short distances) and the second term $b(\sqrt{R^2 + h^2} - h)$ represents the energy absorption (most significant at larger distances).

The epicentral intensity can be described by a regression model between I_0 , moment magnitude M_w , and depth h (Stromeyer *et al.*, 2004):

$$I_0 = cM_w + d \log(h) + e. \quad (2)$$

Combining (1) and (2) leads to the following model for the macroseismic site intensity:

$$I = cM_w + d \log(h) + e - a \log \sqrt{\frac{R^2 + h^2}{h^2}} - b \left(\sqrt{R^2 + h^2} - h \right). \quad (3)$$

This expression is in many respects comparable with the common type of strong-motion intensity predicting equations consisting of a term giving the epicentral ground-motion level, a linear distance term, and a logarithmic distance term (e.g., Joyner and Boore, 1993). For large earthquakes, the point-source assumption fails and the finiteness of the fault must be accounted for. In this study, this is included by defining the distance R as the Joyner-Boore distance (i.e., the shortest distance to the surface projection of the fault plane) instead of as the epicentral distance. The functional form of our relation does not need to be adjusted as we do indeed expect the previously described decay with distance and just extend the epicentral intensity to cover the entire surface projection of the fault plane. In this way we derive a relation that is symmetric around the rupturing fault plane. Our relation does not account for site effects explicitly and, hence, site effects are included in the uncertainty related to the relation.

Input data for the regression is a collection of intensity data points (IDP) describing the intensity at a given location. Usually intensity datasets are characterized by great variety in the number of observations of the different intensity levels. In most cases the highest intensities are undersampled as these are restricted to a relatively small area compared to the lower, more distributed intensities.

To avoid bias in the data due to such effects, a weighting scheme has been applied where each intensity class (integer intensity level) has been assigned the same weight in the regression, regardless of the number of observations within the class. Therefore, the determination of the regression parameters a ; b ; ..., e leads to the weighted least-squares problem

$$\min_{\mathbf{x}} \left\| \mathbf{W}^{-1}(\mathbf{I} - \mathbf{A}\mathbf{x}) \right\|, \quad (4)$$

where $\mathbf{I} = I_i$ ($i = 1, \dots, n$) is a vector of n IDP, $\mathbf{A}$ is an $(n \times 5)$ design matrix, $\mathbf{W}$ is an $(n \times n)$ weighting matrix with only diagonal entries, and $\mathbf{x} = (c, d, e, -a, -b)$ is the parameter vector to be estimated. The values of the diagonal elements of $\mathbf{W}$ are chosen in such a way that (1) they are equal for all data 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, which does not influence the regression solution $\mathbf{x}$ but is important for estimating uncertainties for a new intensity predicted by the model. The natural way to overcome this problem is a rescaling of $\mathbf{W}$ in such a way that the mean weighted and unweighted residuals are equal:

$$\mathbf{W}_{\text{new}} = \mathbf{W}_{\text{old}} \frac{\left\| \mathbf{W}_{\text{old}}^{-1}(\mathbf{I} - \mathbf{A}\mathbf{x}) \right\|}{\left\| (\mathbf{I} - \mathbf{A}\mathbf{x}) \right\|} \quad (5)$$

For a combined regression of k events $\mathbf{I}$, $\mathbf{A}$, and $\mathbf{W}$ are the stacked versions of individual terms.

The uncertainties in the estimated parameters $\mathbf{x}$ and in predicting a new intensity I for given predictor values M_w , R , and h are connected with the covariance matrix $\mathbf{C}$ of the parameter estimates

$$\mathbf{C} = (\mathbf{A}^T \mathbf{W}^{-2} \mathbf{A})^{-1} \quad (6)$$

and the mean squared regression error (making use of equation 5)

$$\sigma^2 = \frac{\left\| \mathbf{W}^{-1}(\mathbf{I} - \mathbf{A}\mathbf{x}) \right\|^2}{n - m} = \frac{\left\| \mathbf{I} - \mathbf{A}\mathbf{x} \right\|^2}{n - m}, \quad (7)$$

where m is the dimension of $\mathbf{x}$ (the number of model parameters). For a specified level of certainty α , the confidence bounds $\mathbf{x}^c$ for the fitted parameters $\mathbf{x}$ are given by

$$\mathbf{x}^c = \mathbf{x} \pm t^{-1}[(1 - \alpha)/2, n - m] \sqrt{\text{diag}(\mathbf{C})}, \quad (8)$$

where $t^{-1}(p, \nu)$ is the inverse of the cumulative t distribution for the corresponding probability p and ν degrees of freedom. For $\nu \geq 40$, $t^{-1}(p, \nu) \approx N^{-1}(p)$, the inverse of the cumulative standard normal distribution at p . In this case

a certainty level of 68.3% ($\alpha = 0.683$) corresponds to the standard deviation (1σ) of normally distributed errors.

Much more interesting in this study is the error of a new intensity prediction I of the estimated model. For given predictor values M_w , R , and h , this can be expressed by

$$I_{\text{error}} = t^{-1}[(1 - \alpha)/2, n - m] \sqrt{\sigma^2 + \mathbf{y}^T \mathbf{C} \mathbf{y}}, \quad (9)$$

where $\mathbf{y}$ is the Jacobian of equation (3) with respect to the model parameters at the predictor values:

$$\begin{aligned} \mathbf{y}^T &= \frac{\partial I}{\partial \mathbf{x}} \\ &= \left[M_w, \log(h), 1, \log \sqrt{\frac{R^2 + h^2}{h^2}}, \sqrt{R^2 + h^2} - h \right]. \end{aligned} \quad (10)$$

Earthquake Sources in the Marmara Sea Region

The described methodology has been applied for the Marmara Sea region of northwestern Turkey, which is a seismically active region having experienced many large earthquakes in the past. The dominating tectonic feature in the region is the NAFZ, which is an approximately 1200 km long fault zone passing through northern Turkey, accommodating the westward movement of the Anatolian Block with respect to the Eurasian plate as a consequence of the African-Eurasian collision. The Marmara Sea was probably developed as a pull-apart basin along the NAFZ, causing an increased complexity with the fault zone splitting into two main branches (e.g., Sengör *et al.*, 2005; see Fig. 1 for the most important faults in the region).

One branch (the southern branch) continues south of the Marmara Sea whereas the other (the northern branch) extends further north, under the sea. Some authors argue for a third branch striking through the eastern part of the central Marmara Sea (e.g., Okay *et al.*, 2000) whereas others find no evidence for this (e.g., Imren *et al.*, 2001). Based on Global Positioning System displacement vectors, it has been shown that the main part of the strain accumulation due to the 22 ± 3 mm/yr plate motion takes place along the fault segment in the northern Marmara Sea (Straub *et al.*, 1997; Meade *et al.*, 2002), and this is therefore the most likely segment to break in a future large earthquake.

Throughout the historical record, there are several examples of significant earthquakes in the Marmara Sea, some of which have caused great damage in Istanbul. The most recent ruptures of the northern strand of the NAFZ in the Marmara Sea are the 1509 M_S 7.2 and the 1766 M_S 7.1 and M_S 7.4 earthquakes. More recently, a smaller (M_S 6.4)

Figure 1. Surface projections of the fault planes used for the studied earthquakes. Faults in the region are shown as gray lines (redrawn from Okay *et al.* [2000] and Saroglu *et al.* [1992]).

earthquake ruptured in the eastern part of the Marmara Sea in 1963. East and west of the Marmara Sea, recent large ruptures have occurred with the 1912 M_s 7.3 Ganos earthquake to the west and the 1999 M_s 7.4 Izmit earthquake to the east (Ambraseys and Jackson, 2000). The general style of faulting along the NAFZ is right-lateral strike-slip faulting, but deviations from this occur in connection with changes in fault orientation. For example, earthquakes in the eastern Marmara Sea such as the 1963 event usually have normal or oblique normal mechanisms (e.g., Sato *et al.*, 2004).

During the last century there has been a westward migration of large, destructive earthquakes along the NAFZ with the most recent events occurring in Izmit and Duzce in 1999 (e.g., Barka *et al.*, 2002). Following these large events, there has been an increase in the Coulomb stress along the Marmara Sea segment (Hubert-Ferrari *et al.*, 2000), bringing this segment closer to rupture. From this observation, combined with recurrence relationships based on the earthquake history in the Marmara Sea, the probability of an M 7+ earthquake in the Marmara Sea within the next 30 yr has been calculated to be in the range of 35%-70% (Parsons, 2004).

Macroseismic Intensity Data

We have collected a dataset of macroseismic intensities for the study area based on available sources. A general problem when working with macroseismic intensity data is that intensities are assigned relative to various scales and with, in certain cases, a considerable portion of personal judgment. The personal judgment has generally a much greater impact on the assigned intensity than the use of the different 12 degrees scales (Musson *et al.*, 2006). The European

Macroseismic Scale (EMS-98; Grünthal, 1998), which provides detailed guidelines for the assignment of macroseismic intensities taking into account building vulnerability, was developed in an attempt to overcome this problem. The EMS-98 has been in use since it was first introduced as an update to the MSK-64 scale in 1993. The EMS-98 was developed to avoid at all costs that conversions between MSK-64 and Modified Mercalli (MM) intensities would have to be applied. In this study, we follow Musson *et al.* (2006) and can therefore assume that the three intensity scales are consistent in the intensity range used here.

The macroseismic intensity data available for the Marmara Sea region consist of a number of isoseismal maps for selected large earthquakes. Most of these are collected by Eyidogan *et al.* (1991) in terms of MM intensity, covering the time interval 1900-1988. In addition, an isoseismal map for the 1999 Izmit earthquake is available from Özmen (2000). We include data for earthquakes in the Marmara Sea region (26° - 31° E, 39.5° - 41.5° N) for which a minimum of four intensity levels are available as isoseismal contours, which is the case for seven events in the time period 1912-1999. An event just east (but outside) of the study area on 26 May 1957 has not been included. The intensity distribution of this event is significantly different from what is observed for the remaining events in the Marmara Sea area, and as the event is furthermore located so far east that tectonic differences may be the explanation for this discrepancy, the event has been left out.

Unfortunately, none of the original datasets that have formed the basis for the isoseismal maps used are available today. The only available information for the region in terms of IDP is a collection of 61 observations for the 1999 Izmit earthquake, which has been published by Mucciarelli *et al.* (2002). A comparison between the IDP

Figure 2. Comparison of the intensity versus distance distribution of the datasets of Özmen (2000) and Muciarrelli *et al.* (2002) for the 1999 Izmit earthquake. The intensity map of Özmen (2000) is represented by data (gray circles) extracted from a regular grid covering the map (see text for details). The IDP of Muciarrelli *et al.* (2002) are

of Muciarrelli *et al.* (2002) and the intensity map of Özmen (2000), as shown in Figure 2, indicates a much stronger attenuation of the intensity in the dataset of Muciarrelli *et al.* (2002). There can be several reasons for this discrepancy. Firstly, we expect the intensity assignments to be associated with a natural variability as well as uncertainties in the intensity assignments (for Italian earthquakes we have estimated this variability to be of the order of 1 intensity unit [M. B. Sørensen, D. Stromeyer, G. Grünthal, unpublished manuscript, 2008] and we expect a similar value here). Secondly, within a given intensity contour we expect also to see subregions with damage less than the class of the intensity contour. In this respect, a single local intensity assignment may not represent the picture on a regional scale. As the dataset of Muciarrelli *et al.* (2002) covers only 61 locations and because all other macroseismic information is available as isoseismal maps, it has been chosen to derive the attenuation relation based only on the isoseismal map of Özmen *et al.* (2000) in order to have a uniform dataset. An over-

view of the studied earthquakes and the macroseismic data is given in Table 1.

The problem of lacking intensity assignments is well known and arises when working with macroseismic information in most regions in the world. Following a large earthquake it has been, and in some places still is, common practice to use intensity assignments for drawing isoseismal maps and then discard the originally assigned intensity values. Our recommendation is to always publish the raw intensity data and to preferably use IDP when studying such data in retrospect. However, the vast amount of data available for longer time spans only in terms of isoseismal maps calls for a method to treat such data causing the minimum amount of bias. In the following we give our suggestion for how to convert isoseismal contours into IDP.

We convert the digitized isoseismal maps into IDP by covering the map area with a fine grid (2 km grid spacing). Each grid point is assigned the intensity value of the contour containing the point. Grid points located offshore or outside the intensity contours of the isoseismal map are not included in the study as we do not expect that any intensity observations have been available in these regions for drawing the map. In this approach we maintain the discrete nature of the intensity assignments, which we find important because assigning noninteger intensity values through some interpolation procedure would lead to data points that do not make sense in relation to the definition of macroseismic intensity and furthermore indicate a nonexistent level of precision in the data. We are aware that there is currently an increased use of so-called instrumental intensities that are converted from recorded ground shaking and can take noninteger values, but these are not to be confused with macroseismic intensities, which are assigned based on felt reports and damage observations. We assume only that observation points are homogeneously distributed over the study area and, furthermore, apply a weighting scheme reducing the influence of large contours (low intensities) with respect to the smaller ones. In comparison to the true observations this approach is expected to smooth the data rather than to bias it, as we would expect the contours to be drawn in such a way that there is a comparable number of too low and too high intensities within a given contour.

In the GMPE in equation (3), distance (R), event depth (h), and moment magnitude (M_w) must be input for each event/IDP pair. It has therefore been necessary to collect basic source parameters for the studied earthquakes. For most of the events, the information is limited, and different approaches have been followed depending on the available information. A summary of the source parameters is given in Table 2 and fault locations are shown in Figure 1. The details of assigning the parameters are described in the following text.

Table 1

Macroseismic Data Included in This Study

Year	Date (mm/dd)	I_0	$I_{\min}$	$I_{\max}$	Number of IDP
1912	08/09	10-11	5	9	27,015
1935	01/04	9	5	8	4375
1953	03/18	9-10	5	8	21,720
1963	09/18	8	5	8	14,985
1964	10/06	8-9	5	9	16,308
1967	07/22	10	5	9	22,302
1999	08/17	10-11*	5	10	14,490

I_0 is the epicentral intensity as given by Eyidogan *et al.* (1991) (in most cases the presence of two values is due to results of various authors being presented). $I_{\min}$ and $I_{\max}$ are the minimum and maximum intensity contour levels in the isoseismal maps. Number of IDP is the number of intensity points available on a regular grid (see text).

Table 2

Source Parameters of the Studied Earthquakes

Year [*]	Date [*] (mm/dd)	Time [*] (UTC)	Longitude (°)	Latitude (°)	<i>h</i> (km)	M_w [*]	M_S [*]	Strike (°)	Dip (°)
1912	08/09	01.28	27.10 [†]	40.67 [†]	10 [‡]	7.3	7.3	-	-
1935	01/04	14.41	27.51 [§]	40.64 [§]	5 ^{‡‡}	6.4	6.4	45	45
1953	03/18	19.06	27.40 [#]	40.00 [#]	10 ^{§§}	7.0	7.0	60 [#]	90 [#]
1963	09/18	16.58	28.95 [*]	40.70 [*]	15 [*]	5.9	6.4	304 [#]	56 [#]
1964	10/06	14.31	27.90 ^{**}	40.05 ^{**}	14 [#]	6.3	6.8	100 [#]	40 [#]
1967	07/22	16.57	30.69 [#]	40.67 [#]	12 [#]	7.2	7.2	275 [#]	88 [#]
1999	08/17	00.01	29.97 ^{††}	40.76 ^{††}	17	7.4	7.4	-	-

^{*} After Ambraseys (2001). M_w is calculated from $\log(M_0)$ using Kanamori's (1977) definition.[†] Central point of fault plane digitized after Armijo *et al.* (2005).[‡] Estimated, shallow event in standard depth of 10 km.[§] Altinok and Alpar (2006).^{||} Estimated based on descriptions in Altinok and Alpar (2006).[#] Taymaz *et al.* (1991).^{**} Estimated from isoseismal lines (see text).^{††} Gülen *et al.* (2002).^{‡‡} Estimated, very shallow event (see text).^{§§} Eyidogan (1988).^{|||} Li *et al.* (2002).

In general, the events of this study can be separated into two groups where the fault plane has been determined based on either (1) surface rupture or (2) magnitude and event strike/dip. The largest events of this study, the 1912 Ganos earthquake and the 1999 Izmit earthquake belong to the group 1. The offshore extent of the 1912 earthquake rupture has been debated (e.g., Ambraseys and Finkel, 1987; Altinok *et al.*, 2003; Altunel *et al.*, 2004; Armijo *et al.*, 2005), but there is increasing evidence for a significant offshore rupture in the Marmara Sea. We adopt the fault rupture mapped by Armijo *et al.* (2005). The 1999 event is well studied and we represent the fault by a number of segments as mapped by Gülen *et al.* (2002). For both these events we assume a vertically dipping fault. We have no exact knowledge about the hypocentral depth of the 1912 event, which occurred at shallow depth and therefore assume a depth of 10 km as is also suggested by Ambraseys and Finkel (1987). For the 1999 event, Li *et al.* (2002) present depth estimates from four agencies from which we use an average value of 17 km.

For the remaining five earthquakes, the location of the fault plane has been determined based on the hypocenter location, the strike and dip of the fault plane, and the earthquake magnitude. The hypocenters have mostly been taken from the published literature (see Table 2). For the event in 1935, no depth estimate has been published and the depth was fixed at 5 km, assuming a shallow depth due to its small magnitude. For the 1964 event, the published epicenter location (40.30°N, 28.23°E, Taymaz *et al.*, 1991) is shifted to the northeast relative to the maximum intensities. As we are seeking a symmetric relation around the fault plane and the location is furthermore associated with significant uncertainty at this time where seismic networks were limited, it has been chosen to move the epicenter to the location of maximum intensity,

maintaining depth and fault orientation given by Taymaz *et al.* (1991). Strike and dip estimates have been published by Taymaz *et al.* (1991) for all events except the 1935 earthquake. This event is described by Altinok and Alpar (2006) as occurring along the northern margin of the Marmara Island. Magnitude estimates have been based on the information of Ambraseys (2001). He gives estimates of M_S and $\log(M_0)$ (from which M_w is calculated) for all the events and it has been chosen to use this information, even if more detailed studies are available, to have a consistent magnitude estimate. For a given event, the hypocenter is located in the middle of the fault plane with orientation as given by the strike and the dip. The extent of the fault plane is calculated based on Wells and Coppersmith's (1994) relations between rupture length or width and magnitude for a general focal mechanism, which are based on a global dataset. Here it should be mentioned that Ambraseys and Jackson (1998) have presented regional relations between magnitude and rupture length for the eastern Mediterranean region. Unfortunately their relations provide only information about fault length whereas we need also to derive the fault width. By using Wells and Coppersmith's (1994) relations we obtain consistent values for length and width that are based on the same dataset. As Ambraseys and Jackson (1998) furthermore find their relations to be very similar to the ones of Wells and Coppersmith (1994), we prefer to use their relations in this study.

It is evident that the source parameters in Table 2 are associated with some uncertainties, which may influence the regression result. This issue has been studied in detail by M. B. Sorensen, D. Stromeyer and G. Grünthal (unpublished manuscript, 2008) through a Monte Carlo approach where 1 million regressions were performed with source parameters sampled from within given uncertainty ranges of the parameters. The results of this study showed

that the effect of such uncertainties on the derived relation and its prediction error for a new intensity estimate is negligible and that the prediction error is mainly controlled by the uncertainties in the IDP themselves. In this respect, the prediction error is mainly determined by the mean prediction error σ (see equation 9), which for reliable models can be understood as a first estimate of the uncertainty of the used IDP and should have a value between 0.5 and 1 intensity units. Based on this experience, we choose not to include errors in the earthquake source parameters in this study.

The final dataset consists of 121,195 IDP covering the

intensity range 5-10, the magnitude range M_w 5.9-7.4, and a distance range of $R \leq 350$ km. These values also represent the limitations of the derived GMPEs as it cannot be assumed that simple extrapolation outside these bounds will be successful.

Derived Intensity Prediction Equation

Regressions were performed first for each earthquake individually to see how well the individual events could be fit by relation (3). For the case of one single earth-

Figure 3. Comparison of observed intensities for the 1912 event with theoretical predictions from relations (13) (left-hand panels) and (14) (right-hand panels). Upper panels: predicted intensity contours (black) in a map view compared to the observed isoseismal lines (gray scale). The intensity contours are drawn for the midpoint value between two integer intensities (e.g., for half intensities). The surface trace of the fault plane is shown as a white line. Lower panels: intensity versus distance plots comparing the grided intensities based on the isoseismal map (circles) with the predicted intensities (solid curve) together with the 68.3% confidence bounds (dashed curves) corresponding to one standard deviation of normally distributed errors.

quake, the terms $cM_w + d \log_{10}(h) + e$ describing the source are inseparable and only their sum c^*M_w can be resolved. In this case, regressions can be performed for the relation (I_S indicating site intensity):

$$I_S = x_1 M_w + x_2 \log \sqrt{\frac{R^2 + h^2}{h^2}} + x_3 (\sqrt{R^2 + h^2} - h). \quad (11)$$

As this test showed that a relatively good fit could be obtained for all earthquakes, a joint regression was performed based on the seven events. As can be seen in Table 2, the depth variation among the individual events is small, and as some event depths are furthermore associated with significant uncertainty, it was chosen to exclude the $d \log(h)$ expression in the source term from the regression model to

keep the problem as simple as possible. This in practice means that an average depth effect on the epicentral intensity is included in the constant term e . The regression is then performed for a relation of the form:

$$I_S = x_1 M_w + x_2 + x_3 \log \sqrt{\frac{R^2 + h^2}{h^2}} + x_4 (\sqrt{R^2 + h^2} - h). \quad (12)$$

When introducing the Joyner-Boore distance (R_{JB}), the following intensity prediction equation in the Marmara Sea region was obtained:

$$I_S = 0.376 M_w + 5.913 - 2.656 \log \sqrt{\frac{R_{JB}^2 + h^2}{h^2}} - 0.0020 (\sqrt{R_{JB}^2 + h^2} - h). \quad (13)$$

Figure 4. Comparison of observed intensities for the 1935 event with theoretical predictions from relations (13) (left-hand panels) and (14) (right-hand panels). Upper panels: predicted intensity contours (black) in a map view compared to the observed isoseismal lines (gray scale). The intensity contours are drawn for the midpoint value between two integer intensities (e.g., for half intensities). The surface projection of the fault is shown as a gray rectangle. Lower panels: intensity versus distance plots comparing the grided intensities based on the isoseismal map (circles) with the predicted intensities (solid curve) together with the 68.3% confidence bounds (dashed curves) corresponding to one standard deviation of normally distributed errors.

The mean regression error is relatively small, $\sigma = 0.672$, and the dependency of the prediction error I_{error} on the predictor values M_w , R_{JB} , and h , as listed in the covariance matrix $\mathbf{C}$, is negligible in comparison to the regression error. Therefore, I_{error} is approximately equal to σ ; $I_{\text{error}} \approx 0.7$.

In some applications (e.g., in shake map generation for early warning purposes or when studying extensive earthquake catalogs), the extent of the fault plane is not known. In such cases it is better to use a less precise GMPE derived for epicentral distance (R_{epi}) than to simply enter the epicentral distance in an R_{JB} -based relation. For this reason, and to be able to quantify the improvement obtained by accounting for rupture dimensions, we

derive an additional relation for epicentral intensity:

$$I_S = 0.793 M_w + 3.417 - 2.157 \log \sqrt{\frac{R_{\text{epi}}^2 + h^2}{h^2}} - 0.0065 \left(\sqrt{R_{\text{epi}}^2 + h^2} - h \right). \quad (14)$$

The mean regression error for this relation is 0.742 indicating, as expected, that a worse fit is obtained for the epicentral distance than for the Joyner-Boore distance, but also that reasonable intensity estimates can be obtained with this relation.

Figures 3-9 show the performance of the relations in comparison with the observed intensity data. For each event and relation the intensity contours are shown in a

Figure 5. Comparison of observed intensities for the 1953 event with theoretical predictions from relations (13) (left-hand panels) and (14) (right-hand panels). Upper panels: predicted intensity contours (black) in a map view compared to the observed isoseismal lines (gray scale). The intensity contours are drawn for the midpoint value between two integer intensities (e.g., for half intensities). The surface trace of the fault plane is shown as a white line. Lower panels: intensity versus distance plots comparing the grided intensities based on the isoseismal map (circles) with the predicted intensities (solid curve) together with the 68.3% confidence bounds (dashed curves) corresponding to one standard deviation of normally distributed errors.

Figure 6. Comparison of observed intensities for the 1963 event with theoretical predictions from relations (13) (left-hand panels) and (14) (right-hand panels). Upper panels: predicted intensity contours (black) in a map view compared to the observed isoseismal lines (gray scale). The intensity contours are drawn for the midpoint value between two integer intensities (e.g., for half intensities). The surface projection of the fault is shown as a gray rectangle. Lower panels: intensity versus distance plots comparing the grided intensities based on the isoseismal map (circles) with the predicted intensities (solid curve) together with the 68.3% confidence bounds (dashed curves) corresponding to one standard deviation of normally distributed errors.

map view compared to the observed isoseismal lines. Intensity versus distance plots comparing the grided intensities based on the isoseismal map (circles) with the intensities predicted from equation (13) or (14) (solid curve) together with the 68.3% confidence bounds (dotted curves) corresponding to one standard deviation of normally distributed errors are also shown. It is seen that for all events, the predicted intensities are within the range of the observed data. The intensity maps in Figures 3-9 show that an important reason for the misfit between the observed and predicted intensity is due to the symmetric distribution of the predicted intensities around either the epicenter or the fault plane. In reality, intensities do not follow such isotropic distributions. It is, though, evident

that the inclusion of the dimensions of the rupturing fault plane by using the Joyner-Boore distance instead of the epicentral distance provides an improvement in the shape of the intensity curves, especially for the larger events. This is most evident for the near-field intensities, whereas the two distance measures seem to be equally appropriate at some distance from the fault plane. It is also clear from Figures 3-9 that the average trend of the intensity decay is well reproduced by our relations.

Because of the limited distance range represented by the dataset, there is a strong trade-off between the parameters x_3 and x_4 in equation (12). This trade-off is represented by the corresponding correlation coefficient r_{34} and similarly for the parameters x_1 and x_2 , r_{12} , where

Figure 7. Comparison of observed intensities for the 1964 event with theoretical predictions from relations (13) (left-hand panels) and (14) (right-hand panels). Upper panels: predicted intensity contours (black) in a map view compared to the observed isoseismal lines (gray scale). The intensity contours are drawn for the midpoint value between two integer intensities (e.g., for half intensities). The surface projection of the fault is shown as a gray rectangle. Lower panels: intensity versus distance plots comparing the grided intensities based on the isoseismal map (circles) with the predicted intensities (solid curve) together with the 68.3% confidence bounds (dashed curves) corresponding to one standard deviation of normally distributed errors.

the magnitude range limits the resolution. For the Joyner-Boore distance, the correlation coefficients are (similar values are obtained for the epicentral distance)

$$r_{12} = \frac{C_{12}}{\sqrt{C_{11}C_{22}}} = -0.991 \quad \text{and} \quad (15)$$

$$r_{34} = \frac{C_{34}}{\sqrt{C_{33}C_{44}}} = -0.887.$$

For our application this trade-off is not a problem as long as the relations are only applied within the distance and magnitude ranges specified previously. Extrapolating to other magnitudes or distances, however, can lead to in-

creased uncertainties in the estimated intensities.

Discussion

The GMPEs derived in the previous section can be used for estimating ground shaking either in seismic hazard assessment or for early warning purposes. In this respect, however, it is important to keep in mind that equation (12) provides a continuous representation of a discrete parameter. Based on the definitions of intensity scales it only makes sense to represent intensities as integer values. Our suggestion for dealing with situations where integer values are needed is to apply a simple rounding

Figure 8. Comparison of observed intensities for the 1967 event with theoretical predictions from relations (13) (left-hand panels) and (14) (right-hand panels). Upper panels: predicted intensity contours (black) in a map view compared to the observed isoseismal lines (gray scale). The intensity contours are drawn for the midpoint value between two integer intensities (e.g., for half intensities). The surface trace of the fault plane is shown as a white line. Lower panels: intensity versus distance plots comparing the grided intensities based on the isoseismal map (circles) with the predicted intensities (solid curve) together with the 68.3% confidence bounds (dashed curves) corresponding to one standard deviation of normally distributed errors.

scheme to the assigned intensities such that, for example, intensities in the interval $4.50 \leq I \leq 5.49$ are all assigned an intensity value of $I = 5$. This approach has also been followed in Figures 3-9. Here it is important to keep in mind the difference between calculated and assigned intensities. When assigning intensity values based on macroseismic observations, uncertain observations that can be associated with either of two integer intensity values (e.g., 5 and 6) will usually be assigned the lower intensity value (5) or both values (5-6) (Grünthal, 1998).

The error in a new intensity estimate using our relations is of the order of 0.7 intensity units. This relatively large error level is to a large extent due to the nature of intensity assignments that have some associated variability, but is also caused by variations in, for example, local site effects. The advantages of good data availability also for historical earthquakes and the direct relation to earthquake damage, however, makes macroseismic intensity a useful ground-motion measure despite the associated uncertainties.

We compare our relations (13) and (14) to the already existing intensity prediction equations for the North Anatolian fault. These relations are listed in Table 3. The relations of Erdik and Eren (1983) and Erdik *et al.* (1985) are simple functions of surface wave magnitude and the natural logarithm of rupture distance. The relation of Böse (2006) is a more complex function of Joyner-Boore distance, moment magnitude, and a correction factor c_B , which depends on site class and magnitude. The correction factor (with values less than 0.06) has not been included in this comparison. As the Ambraseys (2001) relation is valid only for far-field conditions, we make no comparison to this relation.

The comparison is based on the mean regression error (equation 7) with different design matrices A and number of parameters m . The results for our relations and the relations in Table 3, when applying the relations to the earthquakes included in this study, are presented in Table 4.

Table 4 confirms the observation that a similar fit is obtained with our two relations for the small events, whe-

Figure 9. Comparison of observed intensities for the 1999 event with theoretical predictions from relations (13) (left-hand panels) and (14) (right-hand panels). Upper panels: predicted intensity contours (black) in a map view compared to the observed isoseismal lines (gray scale). The intensity contours are drawn for the midpoint value between two integer intensities (e.g., for half intensities). The surface trace of the fault plane is shown as a white line. Lower panels: intensity versus distance plots comparing the grided intensities based on the isoseismal map (circles) with the predicted intensities (solid curve) together with the 68.3% confidence bounds (dashed curves) corresponding to one standard deviation of normally distributed errors

reas the Joyner-Boore distance-based relation can fit the observations for the larger events better. As was observed in Figures 3-9, this is especially important in the region near the fault plane where the strongest ground shaking occurs. Therefore, we recommend the use of the Joyner-Boore-based relation in all applications and especially when estimating near-field ground shaking. It is, however, always better to use an epicentral distance-based relation than to simply enter the epicentral distance in a relation derived from another distance measure.

In comparison to the previously published relations, it is seen that for most events our relations provide a better

fit to the observations. This is partly to expect as our relations are derived based on the data to which we compare; however, the improvement is expected to also originate from the appropriate functional form used. Our functional form is more complex than for the relations of Erdik and Eren (1983) and Erdik *et al.* (1985), taking into account anelastic attenuation and geometrical spreading in separate terms and also including the event depth. Especially problematic are the predictions of the Erdik and Eren (1983) and Erdik *et al.* (1985) relations near the epicenter where the $\ln(R)$ term increases toward infinity. This leads to highly overestimated epicentral intensities.

Table 3

Previously Published Prediction Equations for Macroseismic Intensity for the North Anatolian Fault

Author	Relation
Erdik and Eren (1983)	$I_{MSK} = 0.34 + 1.54M_S - 1.24 \ln(R)$
Erdik <i>et al.</i> (1985)	$I_{MSK} = -3.92 + 2.08M_S - 0.98 \ln(R)$
Böse (2006)	$\ln(I) = 0.8089 + 0.2317M_w - 0.1073 \ln(R_{JB} + 0.6M_w) - 0.0052R_{JB} + c_B$

Table 4
Comparison of Mean Squared Prediction Error for Relations (13), (14), and Previously Published Intensity Prediction Equations

Earthquake	This Study (Joyner-Boore)	This Study (Epicentral)	Erdik and Eren (1983)	Erdik <i>et al.</i> (1985)	Böse (2006)
1912	0.646	0.606	0.615	0.971	2.123
1935	0.369	0.345	0.497	0.354	1.036
1953	0.600	0.747	0.635	0.759	2.079
1963	0.642	0.617	1.016	0.656	2.350
1964	0.563	0.570	0.577	0.462	2.092
1967	0.673	0.901	0.566	1.105	1.934
1999	0.970	1.017	0.744	1.412	1.113

As most damage is associated with the highest intensity levels, this is problematic and makes these relations less suitable for ground-motion estimation in the near field of an earthquake. The relation of Böse (2006) is much more complex but based on simulated data. This relation performs well for the 1999 earthquake but has problems reproducing the macroseismic fields of the other earthquakes considered here. Furthermore, it is more complicated to implement due to the constant c_B for which the soil class must be known. In conclusion, even though the performance of the previous relations at some distance from the fault may be relatively good and will lead to realistic estimates of the ground motion, our relations provide an improved estimate of the macroseismic field due to a strong earthquake in the Marmara Sea area.

Conclusions

GMPEs for macroseismic intensity are useful for shake map generation and seismic hazard assessment when an output is required that is directly associated with the damage caused by an earthquake. In the present study, such relations have been derived for the Marmara Sea region, northwest Turkey, which is under a significant seismic hazard due to an expected large earthquake along the North Anatolian fault. The derived GMPEs are based on intensity maps from seven large earthquakes, which occurred during the last century. One relation takes into account the finite extent of the fault plane whereas another is derived based on a pointsource assumption for application in cases where fault dimensions are not known. The relations can be used to predict the distribution of macroseismic intensity based on moment magnitude, event depth, and fault distance. Uncertainties have been estimated based on the regression error, and a new intensity value for an earthquake in the Marmara Sea region can be given with an uncertainty of ca. 0.7. In this respect, the relations can be used to obtain reliable estimates of the intensity distribution due to large earthquakes in the Marmara Sea region, which is important for seismic hazard assessment and risk mitigation measures.

Data and Resources

All data used in this article came from published sources listed in the references. Figure 1 was drawn using the Generic Mapping Tools (GMT; Wessel and Smith, 1998).

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