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The Attenuation of Seismic Intensity in Italy: A Bilinear Shape Indicates the Dominance of Deep Phases at Epicentral Distances Longer than 45 km

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Abstract The attenuation of seismic intensity with distance in Italy is analyzed by using felt intensity report data obtained from two comprehensive historical databases recently made available. The observed attenuation pattern that in the past was interpreted as a logarithmic or root (square or cubic) attenuation law shows quite clearly two different linear trends in the near and in the far field. At distances shorter than 45 km, the decrease of the intensity with distance is about one degree per 20 km, while at longer distances the slope is about one degree per 50 km. This is in agreement with some recent findings of realistic modeling of seismic ground motion that has been explained as the transition from upper-crust direct *Sg* phases to waves reflected at the Moho controlling the energy main release. The slope of the curve in the far field shows a regional dependence in agreement with recent works on the attenuation of *Pn* and *Sn* phases in Italy. If effective, this correlation might allow us to discriminate the contribution of crustal and subcrustal paths in seismic intensity attenuation studies.

Introduction

Over the last few decades, macroseismic intensity felt report data have been neglected by seismological research. However, presently, the wave field generated by a strong earthquake still cannot be completely described by source- and wave-propagation models, and in particular, the transmission of seismic waves in complex media cannot be predicted satisfactorily. Even the correlation between levels of ground shaking and damage is not yet well understood and represents a challenging development for modern seismic hazard assessment. Thus, the use of seismic intensity in seismic hazard studies may prove practical when compared to other observations, which are less strictly related to the damage phenomenology.

In most recent studies, seismic hazard is usually evaluated in terms of the level of the peak of ground acceleration (PGA). However in Italy, where most of the information available on past earthquakes is preinstrumental in nature, the same computations are also performed in terms of seismic intensity (Peruzza, 1996; Slejko, 1996). In these analyses, the attenuation model of a ground-shaking parameter (PGA, macroseismic intensity, Arias intensity, etc.) plays a fundamental role. In fact, it allows the prediction of the distance where a certain degree of shaking is expected to be exceeded for future earthquakes of a given magnitude (or epicentral intensity) and location. Moreover, nowadays, as seismic hazard is relevant to other communities, such as planners and insurers, variations in intensity (and therefore damage) as a function of distance and magnitude become increasingly important.

The literature proposed various formulas to model the attenuation of ground-shaking parameters with distance. In some cases they were derived on the basis of models of wave propagation, but in others they were purely empirical. In general, the fit of the model to real data is tested statistically, but often the principal obstacle to a good comprehension of the physics of the phenomena is the huge dispersion of the data. In fact, ground shaking is strongly affected not only by the size of the earthquake and by the characteristics of the seismic-wave propagation but also by soil response and topography. In the case of the seismic intensity even the characteristics of the human and natural environment and the arbitrariness of the intensity assignment process may add further uncertainty. In many cases, differences up to two or three degrees of intensity can be observed at settlements located a few kilometers from one another due to peculiar geotechnical and topographical characteristics of the sites. This scatter is also observed for PGA and other instrumental parameters.

Previous Efforts

Previous analysis on the modeling of the distance dependence of seismic intensity were guided by the obvious empirical evidence that the difference between epicentral and local intensity monotonically increases with distance while the slope of the curve describing its behavior monotonically decreases. Thus, in general, the functional relation was chosen among the class of continuous functions having a positive first derivative and a negative second derivative. Among them the most often used are the logarithm and the root functions.

One of the first attempts was due to Von Koveslighety (1906). He derived his well-known equation from geometrical considerations and the assumption that intensity is almost proportional to the logarithm of maximum acceleration

$$\Delta I = 3\log_{10}\left(\frac{D_i}{h}\right) + B(D_i - h),$$

where

$$D_i = \sqrt{D^2 + h^2}$$

and $\Delta I = I_o - I_i$ is the intensity difference between epicentral and local intensity, D is the epicentral distance in km, h is the focus depth in km, and B is a free parameter. Afterward, Blake (1941) simplified the formula, eliminating the linear term but letting the coefficient of the logarithm vary according to the differences in the absorption of the seismic energy among different areas

$$\Delta I = s\log_{10}\left(\frac{D_i}{h}\right)$$

where s is a free parameter.

Similar to the Von Koveslighety (1906) formula, Gupta and Nuttli (1976) proposed a more physically consistent approach. Assuming that the intensity difference is linearly correlated with the logarithm of the ratio between the amplitude and the period of seismic waves (A/T) they derived the equation

$$\Delta I = C_1 + C_2\,(k\Delta\log_{10}\mathrm{e} + \log_{10}\Delta)$$

where Δ is the distance in degrees, k(γ in the original article) is a coefficient of anelastic attenuation, while C_1 and C_2 are empirical constants. In this formula the linear term accounts for the anelastic dissipation, while the logarithmic one accounts for the geometrical spreading. The assumed values of k vary from 0.1/deg in the central United States to 0.6/deg in California (i.e., Nuttli, 1973).

For Italy, the Grandori *et al.* (1987) formula has to be mentioned because it was used for the most recent seismic hazard estimations in terms of intensity (Peruzza, 1996; Slejko, 1996):

$$\Delta I = \frac{1}{\ln\psi}\ln\left[1 + \frac{\psi - 1}{\psi_0}\left(\frac{R_i}{R_0} - 1\right)\right]$$

where R_i is the average radius of the i-th isoseismal, and ψ, ψ_0, and R_0 are free parameters.

The problem of choosing an empirical attenuation model among different alternatives requires a careful evaluation of the goodness of the fit of each model, taking into account the number of free parameters. As Enrico Fermi said: "with 4 free parameters one can fit an elephant". Thus, with a large enough number of parameters, it is possible to perfectly reproduce even the subtlest random fluctuations of any data set without describing the real physics of the problem. In line with this approach, Berardi *et al.* (1993) proposed a simple empirical attenuation model for Italy called the cubic root attenuation model (CRAM), which using only two free parameters, was shown to reproduce the behavior of intensity data with distance better than almost any other similar model

$$\Delta I = \alpha + \beta\sqrt[3]{D}$$

where D is the epicentral distance, and α and β are free parameters.

Up to now, there exists no satisfying physical justification of why the cubic root model fits the intensity data attenuation with distance better than other models, such as the square root and the logarithm, that in principle seem more physically justifiable. Indeed, the empirical approach does not exhaust the subject since in the lack of a clear physical explanation, one cannot be confident that the model that fits well a certain set of data represents a good choice also for a different sample. Thus our approach in this work will be first to empirically find the model that is able to reproduce the data at best and then to search for a reasonable explanation in terms of the seismic waves propagation theory and observation.

On the other hand, any theoretical approach conflicts with the qualitative nature of macroseismic intensity. This, in fact is an ordinal (ranking) index that classifies the severity of earthquake effects. Therefore its inclusion as a continuous variable in functional relationships with other parameters like magnitude, ground acceleration, and distance appears questionable (Rock, 1988, p. 65). In principle, intensity ought to be used only for rank comparisons among sites and/or earthquakes. Even the average of different values of intensity cannot be justified in terms of the intensity scale definition because the result of such operation is, in general, a rational number that cannot be strictly related to the descriptions of the effects of any grade of the scale.

However, more than one half of a century of experience of using seismic intensity in empirical relationships has demonstrated that maximum intensity shows a fairly good linear correlation with the magnitude of the earthquake (i.e., CPTI Working Group, 1999). On the other hand, as explicitly stated by the author of the Mercalli Cancani Sieberg (MCS) scale (Sieberg, 1931), intensity was originally designed and updated, according to Cancani (1904), with the objective of describing successive levels of effects somehow related to equal-spaced values of the logarithm of ground acceleration. On the problem of the linearity of the macroseismic scale, see also the discussion on the alleged need of a further degree in between degrees VI and VII of the European Macroseismic Scale 1998 (EMS98) scale in Grüntal (1998).

A possible meaning of the fractional part of a rational intensity value could be given both in terms of probability as well as of fuzzy set membership (Zadeh, 1965). In the first case, the fractional part can be seen as the probability that the intensity coincides with the highest of the two bounding integer degrees, while its one-complement would be the probability that the intensity coincides to the lowest one (i.e., 7.8 would mean that there is a probability $p = 0.8$ that the intensity is VIII, and a probability $p = 0.2$ that it is VII). This definition is consistent with the result of the average of a population of integer intensity values. In the second case, the fractional part could be the degree of membership of the intensity estimate to the set of highest bound intensities, while the complement to one could be the degree of membership to the set of the lowest bound ones. This last could be useful in light of fuzzy methods to estimate the intensity, as recently proposed by Ferrari *et al.* (1995) and Vannucci *et al.* (1999).

Nevertheless, on the basis of the above considerations, we believe it is reasonable to use in the following the concepts of average intensity (the arithmetic mean of intensity estimates at different sites located at the same epicentral distance) and of intensity difference (the difference between epicentral and local intensity) which, as mentioned previously, are both not fully justifiable in terms of the standard intensity definition.

Data Analysis

We reconsidered the problem of the attenuation of seismic intensity in Italy, in light of new data recently made available to the scientific community, which are the "Catalogo dei Forti Terremoti in Italia dal 461 a.C. al 1990" (CFTI) (Boschi *et al.,* 1997) and the "Database delle Osservazioni Macrosismiche" (DOM) (Monachesi and Stucchi, 1997). Both data sets tackle historical Italian earthquakes and include more than 30,000 MCS intensity observations. Since the intensity data of about 300 earthquakes are available in both of them, the authors of these data sets have compiled a consensus joint catalog (CPTI Working Group, 1999) where the study "that is preferable for seismic hazard computations" is chosen when two are available. The data set resulting from the application of these choices includes more than 50,000 MCS intensity observations, relative to about 1000 earthquakes from ancient times until the present.

To assure the homogeneity of the determination of the intensities and the derived source parameters, only the data coming from earthquakes that occurred in the last two centuries (from 1800 until 1990), including intensity estimates in at least 20 different localities, are used in our work. First of all, we computed the macroseismic epicenter and the epicentral intensity of each earthquake in our database. These parameters are not precisely defined in the literature (i.e., Cecic *et al.,* 1996). However, for events located in densely inhabited areas, as most of the Italian earthquakes are, they are commonly recognized as the maximum observed intensity excluding the anomalous amplifying sites and the barycenter of the localities with major effects, respectively. On the basis of these general concepts, Gasperini and Ferrari (1995, 1997) developed two computer algorithms to estimate macroseismic epicenter and epicentral intensity, which have proven to be very stable on a range of reasonable choices. A brief English description of the macroseismic location algorithm can be found in appendix 1 of Gasperini *et al.* (1999). The procedure to evaluate epicentral intensity can be summarized as follows: I_o is taken as equal to the maximum intensity I_{max} except when only one locality with intensity equal to I_{max} is reported and at least one other locality with a lower intensity value I is reported. In this case I_o is taken as the maximum between I and $I_{max} - 1$. The epicentral intensity estimated by this procedure is consistent with the integer (or at most half integer) definition of intensity and has shown a better correlation than I_{max} with instrumental magnitudes (Gasperini and Ferrari, 1995, 1997).

For every intensity estimate in our database (about 35,000), the epicentral distance and the difference between the epicentral and local intensity is computed. Since the wave path length in the vicinity of the source may be significantly affected by focal depth, for attenuation analysis it would be preferable to use the inclined distances from the surface to the source. Unfortunately, the instrumental depth of the seismic sources is unknown or unreliable for most of the earthquakes in our database. Moreover, since the instrumental hypocenter of an earthquake, being by definition the point where the fracture originates, may not coincide with the radiation centroid of the released seismic-wave energy, the use of instrumental depth in the macroseismic attenuation formula might even be misleading. On the other hand, the focus of the great majority of the Italian damaging earthquakes, based on recent instrumental event data, is shallower than 20 km, and therefore we can confidently assume 10 km to be an average depth of the aforementioned centroid for all the earthquakes. Although this choice represents a raw simplification of the problem, it can be useful to reduce the distortion of the attenuation curves due to focal depth. After the application of this correction, the residual deviations should be largely smaller than the location and depth uncertainties and be confined to the smallest epicentral distances. The application of this procedure also implies that in the resulting data set, we have no data at distances shorter than 10 km.

Another kind of distortion can be induced by the source extent. In fact, this may cause significant deviations from the circularly symmetric pattern implicitly assumed in attenuation studies. In particular for large earthquakes, the attenuation of intensity with distance from the epicenter may appear lower than the real one in the direction along the fault strike. On the other hand, since the known reliable fault-plane solutions of strong Italian earthquakes mostly indicate dip-slip mechanisms, we could expect this asymmetry to be at least partially compensated by the wave radiation pattern, which has its maximum in the direction perpendicular to the fault strike. Deviations from the circular symmetry up to a distance of half of a source length were found statistically significant by Gasperini *et al.* (1999) for some of the largest earthquakes of our database. Thus, to reduce this distortion, we excluded from our computations all localities at epicentral distances shorter than half the subsurface rupture length (RLD), computed from magnitude by Wells and Coppersmith (1994) formulas for all types of sources.

All the pairs of distances and intensity differences available in our database have been plotted in Figure 1. It is evident that the dispersion of data is very large and that they are not uniformly distributed, as they are denser at shorter distances than at longer ones. In order to provide a clearer representation of the behavior of the intensity attenuation with distance, we computed the arithmetic average of the intensity difference and the corresponding 95% confidence intervals, within intervals of hypocentral distance of 5 km of width (Figure 2). On average there are more than 1000 data points in each interval for distances below 70 km, about 100 points around 200 km, and less than 10 points at 350 km and more. In Figure 2 only the averages with at least 10 data points are plotted.

We can see that the averages show a clear monotonic trend very easy to identify visually. In particular, in the first part, the curve is linear up to about 40–50 km, where the slope decreases markedly. The second part is also almost linear and shows a quite constant slope up to about 150–200 km where the slope decreases again. At this point, due to the scarce number of data, the confidence intervals became very large and thus the visual identification of the trend becomes difficult. This quite regular multilinear shape is somewhat surprising since the data come from earthquakes with very different epicentral intensity and location.

Figure 3 supports the appropriateness of the use of intensity as a continuous variable. We show the stacked histogram of the intensity difference residuals of data reported in Figure 1 with respect to their respective means in Figure 2, for all the 5-km intervals. We can see that the resulting distribution is not very far from a Gaussian curve, with a standard deviation of 1.15 intensity degrees, having moderate skewness and kurtosis.

Attenuation Model Fitting

In order to build an attenuation model that follows the observed behavior of the data of Figure 2, we fitted, by least-

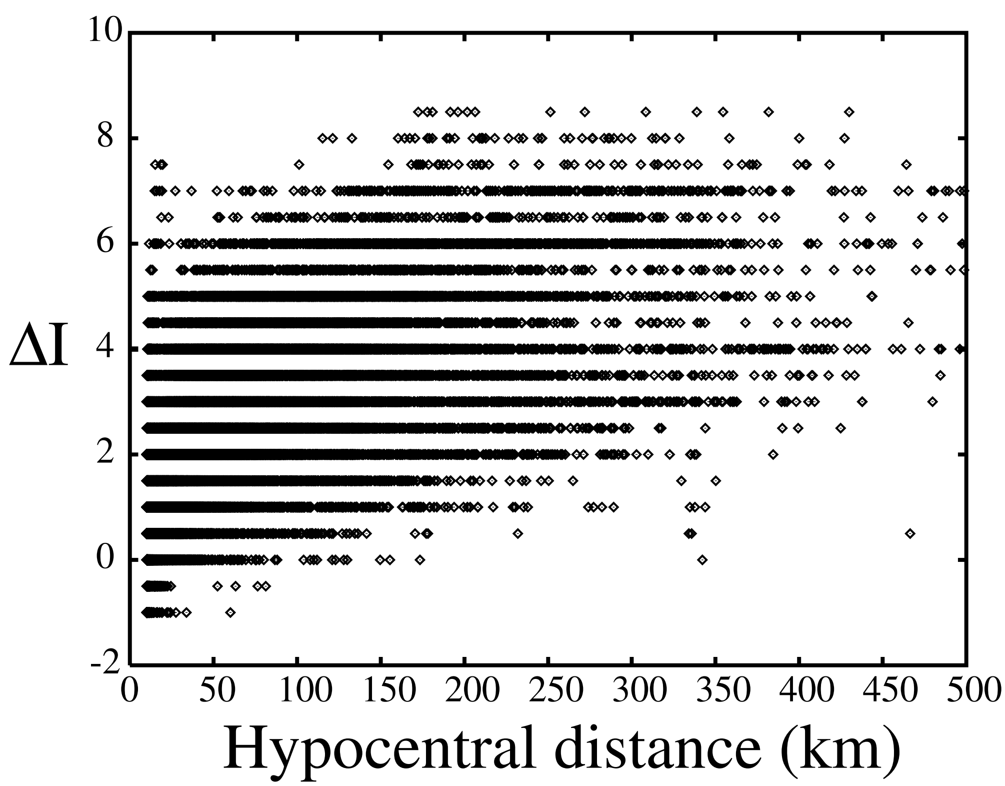

Figure 1. Differences between hypocentral and local intensity (ΔI) as a function of distance for Italian intensity felt report database. In total about 35,000 observations concerning 400 earthquakes from 1800 until now.

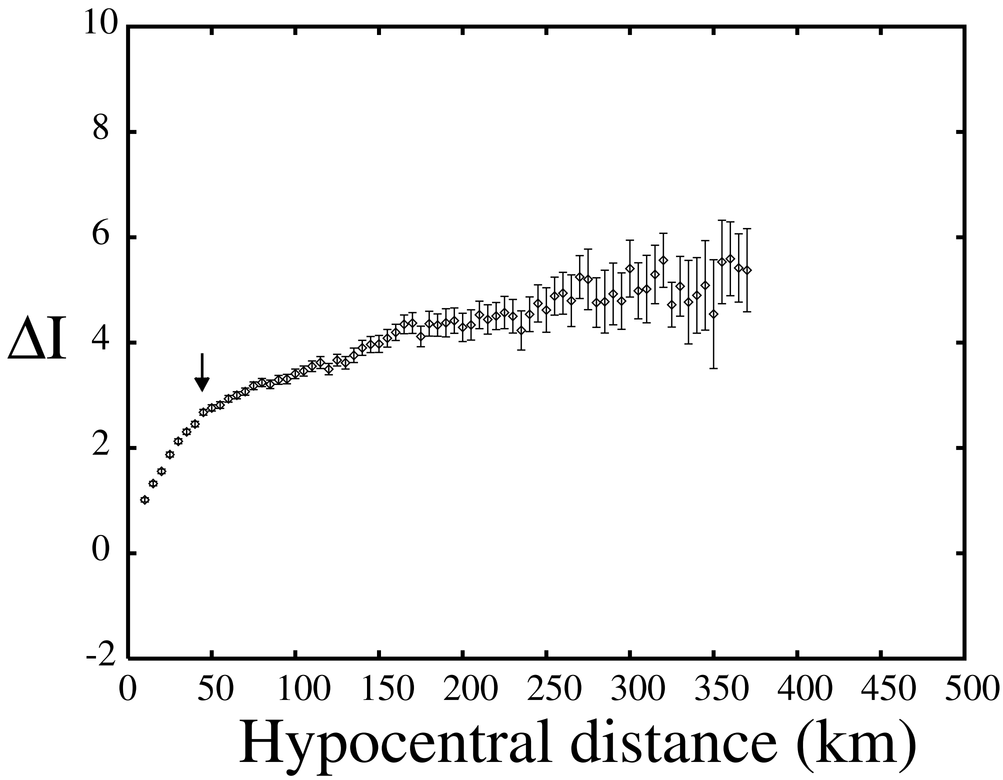

Figure 2. Average intensity differences and relative 95% confidence intervals for the same data set of Figure 1. Only the intervals with at least 10 occurrences are reported.

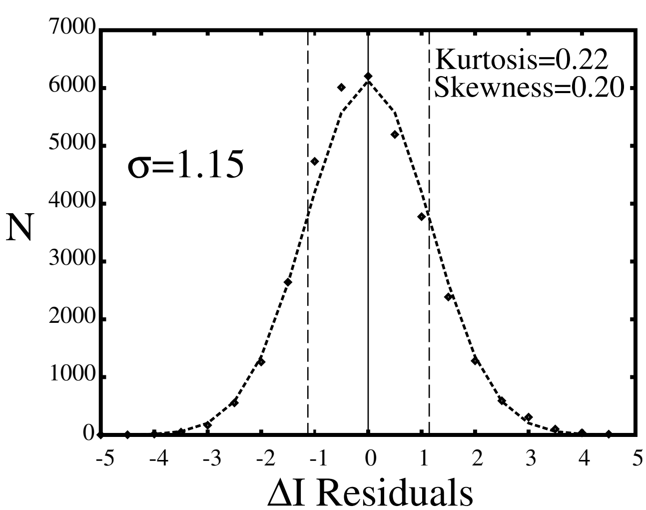

Figure 3. Histogram of the intensity difference residuals with respect to the mean for all analyzed 5-km intervals of Figure 2. Vertical dashed lines indicate the 1σ confidence interval.

squares, a multilinear (piece-wise linear) model on the grouped data set. After a minimal adjustment, we chose as a change point sample no. 8 (at a distance of 45 km). The first part (from sample 1 to 8) was then fitted with a straight line with slope and intercept, while for the second part only the slope was computed:

$$\Delta I = \begin{cases} \alpha + \beta \cdot D & (D \leq 45) \\ \alpha + \beta \cdot 45 + \gamma \cdot (D - 45) & (D > 45) \end{cases}$$

where D is the epicentral distance in km, ΔI is the intensity difference, and α, β, and γ are free parameters. Since we arbitrarily established even the position of the slope change, the total number of free parameters for our model is four.

Another important point to consider is the completeness of the intensity data at the lower end of the intensity distribution. It is obvious that the value of the mean intensity difference can be biased at large epicentral distances due to the possible omission of intensity observations below the limit of widespread perceptibility by the population (around degree IV). In fact, at distances where the average intensity is below this value, we can suppose that a certain amount of data regarding localities with low intensity could be lost, in particular for small settlements, due to the difficulty to record the feelings of peculiar categories of people (in bed, inside buildings, at highest floors, etc.) or more simply due to the lower effort made by macroseismic researchers to recover this kind of data. For various reasons (geology, building characteristics, etc), the sites showing an intensity higher than average are more likely to be reported by macroseismic surveys; we could observe an apparent average intensity difference ($I_o - I$) definitely smaller than the real one, and thus an apparent attenuation of the intensity with distance lower than the real one.

The evaluation of the true completeness of the macroseismic reports is quite a hard task that in case could be accomplished only with the help of the researcher who compiled the reports. Therefore, we decided to overcome this problem simply by adopting cautious criteria to choose data to include in our sample. We proceeded as follows: first, we fitted the raw intensity data with our attenuation model and then repeated the computations, however excluding the data located at distances where the intensity predicted by the model lies below IV. Since this procedure actually modifies the average intensities, the operation requires a certain number of iterations up to reach a satisfying agreement between the *a priori* and *a posteriori* attenuation models. A side effect of this decimation (the total number of remaining data is about 20,000) is that the maximum distance to reliably compute the average intensities is reduced to about 180 km. Therefore, below we will only consider the range of distances from 10 to 180 km for our computations.

Table 1 shows the values of the estimated model parameters with their standard errors. We also included the inverses of the slopes (in km/degree) since their meaning is more immediately perceived. The first part shows a slope of about 18 km/degree, while the second shows a slope of about 46 km/degree. If the aforementioned data selection procedure is not applied, the intensity averages in the first part do not change significantly, while in the second part the slope decreases markedly (from 46 to about 82 km/degree), thus indicating that a significant underestimation of the intensity difference might be due to the loss of low intensities not reported by macroseismic surveys. This also indicates that all the intensity attenuation formulas that are fitted, without taking into account this source of bias, may compute, at large distances, an attenuation lower than the real one and predict higher intensities than actually observed on average.

The goodness of the fit of our attenuation model can be verified in Figure 4 (solid line), but we also report in Table 1 the variation coefficient R^2 (which indicates the rate of the total variance of data explained by the model), the total standard deviation of the model *s*, and the *F*-value of the analysis of variance (ANOVA) test of goodness of fit F (good) (Davis, 1986) with the corresponding numerator, n_a, and denominator, n_d, degrees of freedom and the resulting confidence level, *p*. On the basis of the values of these statistical estimators, the fit results to be highly significant.

In order to verify whether this model is the best on a range of reasonable alternatives we also tested the fit of our data set with other attenuation models. We concentrated our analysis to simple models of the type

$$\Delta I = \alpha + \beta \sqrt[n]{D},$$

where D is epicentral distance in km, ΔI is the intensity difference (see above), $n = 2,3,4$, and α and β are free parameters. We also analyzed the simple logarithmic law (similar to Blake's formula),

$$\Delta I = \alpha + \beta \log_{10}(D)$$

and a formula functionally equivalent to the expression proposed by Gupta and Nuttli (1976),

Table 1
Data Fitting and Goodness of Fit Parameters

Model	α	β $1/\beta$	γ $1/\gamma$	R^2	s	F(good) F(best) F(log1)	n_n	n_d	p
Bilinear	0.52 ± 0.01	0.056 ± 0.003	0.0217 ± 0.0005	0.99	0.11	4107.	1	31	1.00
		17.9 ± 0.9	46.1 ± 1.2			16.2	2	31	1.00
						21.1	1	31	1.00
CRAM	−2.06 ± 0.14	1.38 ± 0.03		0.98	0.17	1861.	1	33	1.00
SQ root	−2.83 ± 0.10	0.46 ± 0.01		0.98	0.16	2126.	1	33	1.00
FT root	−3.83 ± 0.20	2.60 ± 0.06		0.98	0.19	1529.	1	33	1.00
Log	−3.67 ± 0.29	4.05 ± 0.15		0.96	0.28	712.	1	33	1.00
Log + lin	−1.12 ± 0.32	2.01 ± 0.24	0.0137 ± 0.0015	0.99	0.15	2588.	1	32	1.00

Bilinear, cubic root (CRAM), square root (SQ root), fourth root (FT root), logarithm (Log) and the Gupta and Nuttli (1976) type (Log + lin) attenuation models for epicentral distances from 0 to 180 km. R^2 indicates the coefficient of variation and s indicates the total standard deviation of the model. F(good) is the Fisher statistics for the analysis of variance test of goodness of fit, n_a and n_d are the numerator and denominator degrees of freedom respectively, p is the estimated confidence level. F(best) indicates the Fisher statistics for added terms of the bilinear model with respect to the best of the simple two parameter formulas while F(logl) is the value of the same statistics with respect to logarithmic + linear formula.

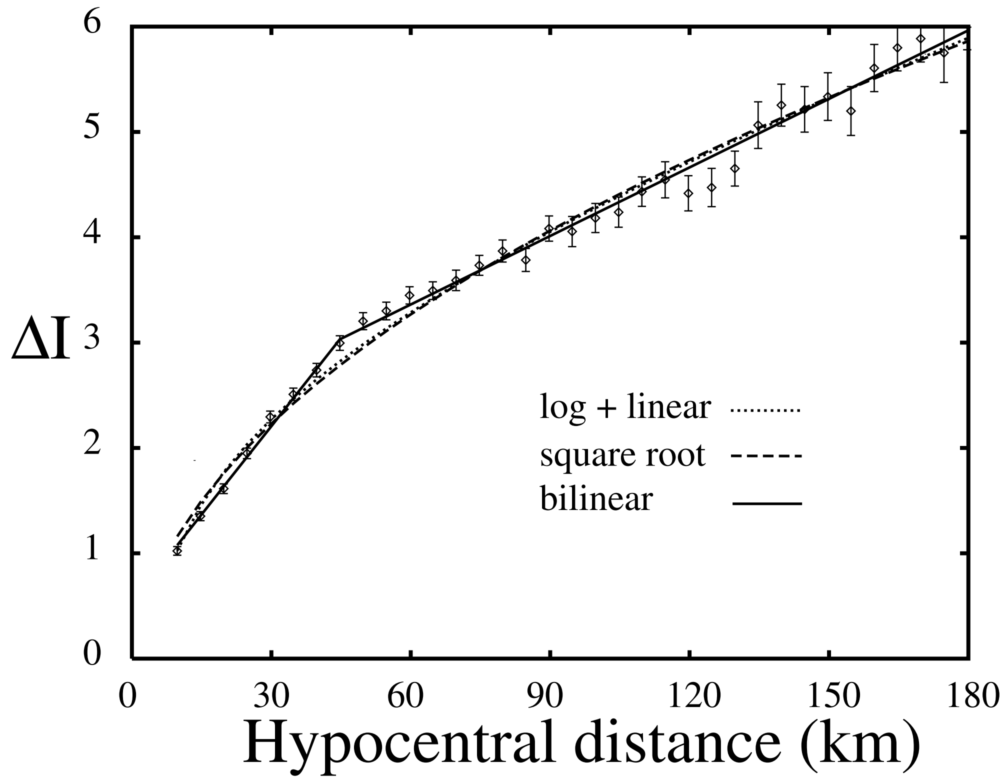

Figure 4. Data fitting, for the whole intensity data set, of the bilinear (solid), square root (dashed), and logarithmic + linear (dotted) attenuation models.

$$\Delta I = \alpha + \beta \log_{10}(D) + \gamma D,$$

where, however, we let both the coefficients of log(D) and D vary independently. With this formulation the coefficient of anelastic attenuation (in deg^{-1}) is

$$k = \frac{\gamma}{\beta \log_{10}\mathrm{e}} \frac{\pi a}{180},$$

where a is the Earth radius and e is the natural logarithm base.

The results of the fit with these models are reported in Table 1, while the curves for the best root models ($n = 2$, dashed), and the mixed logarithmic-linear model (dotted) are plotted in Figure 4. For all the models, the fit is highly significant, but even for the best one, both R^2 and s are worse than the bilinear one. Since it includes more free parameters than other models we performed the ANOVA analysis to test the significance of added term. The resulting statistics F(best) for the best of the two parameters models and F(logl) for the logarithmic-linear one are reported in Table 1 together with the numerator and denominator degrees of freedom. In both cases, the confidence levels (p) are greater than 0.995, thus indicating that the bilinear model represents a significant improvement with respect to all other simpler models.

In Figure 5, we reported the ΔI residuals (observed—predicted) for the logarithmic, the logarithmic-linear, the square root, and the bilinear models. Moderate oscillations are present in the bilinear model, but, for almost all the data points, the residuals are smaller than the estimated 95% confidence intervals. On the contrary, systematic deviations of the residuals can be observed for all others. Besides the bilinear model even the composite logarithmic-linear model, which fits better with data, is not able to well reproduce the observed behavior at distances from 15 to 30 km and from 45 to 60 km.

In Figure 5, we can also note that even the bilinear model fails to predict the data from 120 to 130 km where indeed all the models overestimate the observed intensity difference. Due to the wide error bounds, we cannot exclude that this deviation is due to the chance. However we could argue that the high intensities that cause this attenuation anomaly could be somehow related to the positive interference of direct and Moho refracted phases, this distance range being very close to the corresponding crossover distance.

A possible objection that might be adduced against a broader applicability and physical significance of the bilinear model is that the observed abrupt slope change can be induced by a nonlinearity of the scale. In fact, since the equidistance of the degrees of the intensity scale cannot be demonstrated, the steepest slope in the near field could be due to a higher attenuation of highest degrees of the scale with

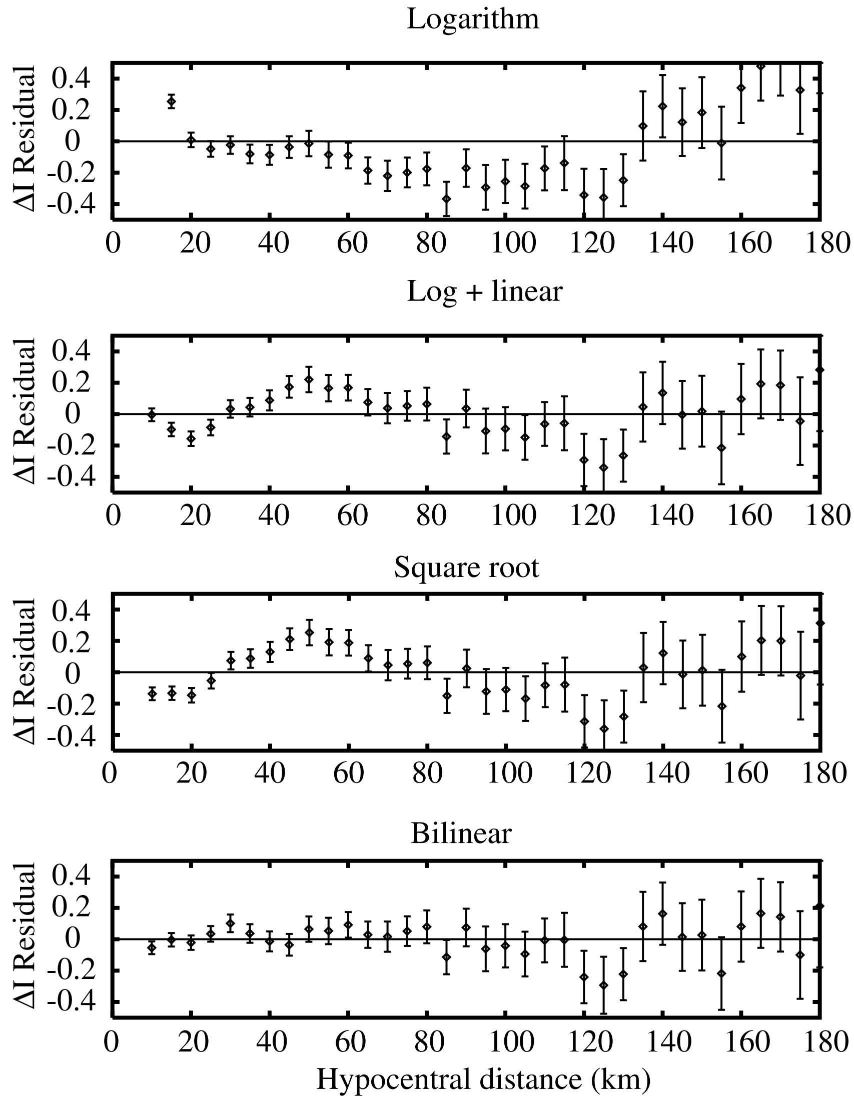

Figure 5. Observed-computed residuals resulting from the application of logarithm, square root, logarithm + linear, and bilinear attenuation models.

respect to the lowest ones. We tested this hypothesis selecting the earthquakes by epicentral intensity. In Table 2 and in Figure 6, we show the results of these computations for the set of earthquakes with I_0, values respectively larger than (Fig. 6a) and equal to or lower than VIII (Fig. 6b). We chose this threshold in order to subdivide the cases in two sets with almost the same number of felt report data (about 10,000 in each). For both sets, the bilinear model still shows a better performance than all other simple models (note that for the set with $I_0 \leq$ VIII, the maximum distance allowed by the cut of low intensities is reduced to 90 km). We can see that both curves shows a clear change point at about 40–50 km from the source and have quite similar slopes. Saturation and scaling biases seem to be absent or negligible in our intensity data set. We can thus confidently reject the hypothesis that the slope change is due to a nonlinearity of the intensity scale, and we can assert that this behavior is actually a peculiar characteristic of the propagation of the seismic energy in the Italian crust and lithosphere.

The only significant difference between the two fitted curves concerns the value of the intercept that is about half a degree higher for the $I_0 >$ VIII set. This difference is clearly highlighted in Figure 7, where the data and the fitted bilinear models for the two separate sets as well as the whole set are shown. We can see that the curves for the former two sets are almost parallel with the high intensity one that has shifted higher with about half an intensity degree. We can argue that, at least in part, this shift could be due to a small bias in the computation of I_0. In fact, due to the characteristics of the simple algorithm we used, I_0 is more likely to be overestimated for large earthquakes, as these have a larger number of felt reports in the mesoseismal area than small earthquakes. However, the curve fitted with the whole data set lies about in the middle between the other two and can

Table 2
Data Fitting and Goodness of Fit Parameters of Different Attenuation Parameters*

Model	α	β $1/\beta$	γ $1/\gamma$	R^2	s	F(good) F(best) F(log1)	n_n	n_d	p
			I > VIII						
Bilinear	1.05 ± 0.12	0.047 ± 0.003	0.0203 ± 0.0006	0.99	0.13	2560.	1	31	1.00
		21.2 ± 1.4	49.2 ± 1.5			12.2	2	31	1.00
						14.0	1	31	1.00
CRAM	−1.40 ± 0.16	1.25 ± 0.04		0.97	0.20	1198.	1	33	1.00
SQ root	0.20 ± 0.10	0.416 ± 0.011		0.98	0.18	1510.	1	33	1.00
FT root	−3.00 ± 0.23	2.35 ± 0.07		0.97	0.22	996.	1	33	1.00
Log	−2.83 ± 0.31	3.67 ± 0.16		0.94	0.29	521.	1	33	1.00
Log + lin	−0.16 ± 0.35	1.52 ± 0.26	0.0144 ± 0.0016	0.98	0.16	1867.	1	32	1.00
			I ≤ VIII						
Bilinear	0.37 ± 0.05	0.055 ± 0.002	0.0184 ± 0.0012	1.00	0.05	3470.	1	13	1.00
		18.2 ± 0.5	54.3 ± 3.5			11.1	2	13	1.00
						22.5	1	13	1.00
CRAM	−1.79 ± 0.13	1.24 ± 0.04		0.99	0.10	1180.	1	15	1.00
SQ root	−0.42 ± 0.12	0.45 ± 0.02		0.98	0.12	691.	1	15	1.00
FT root	−3.15 ± 0.15	2.24 ± 0.06		0.99	0.09	1466.	1	15	1.00
Log	−2.45 ± 0.14	3.13 ± 0.09		0.99	0.09	1299.	1	15	1.00
Log + lin	−2.06 ± 0.33	2.75 ± 0.30	0.004 ± 0.003	0.99	0.09	1457.	1	14	1.00

*As in Table 1, but with selection of the epicentral intensity.

therefore be confidently used to infer the average properties of the attenuation of seismic intensity in Italy.

It is also interesting to note that, at odds with previous findings, in our computations the CRAM is never the best among the two-parameter models. We have noticed instead (but we do not report the computations here) that, if the low intensity decimation procedure is not applied to the data, the best performance of the CRAM model is indeed confirmed. As a result, we can argue that the Berardi *et al.* (1989) results, as well as results of many previous studies on seismic intensity attenuation, could have been biased significantly by the incompleteness of low intensities in the far field. We also note that for two of the three analyzed data sets, the simple logarithmic law results to be the worst of the two-parameter models. This would allow us to definitely conclude that this model, which is considered by some as the most appropriate and physically grounded to describe the behavior of the logarithm of the PGA with distance (i.e., Sabetta and Pugliese, 1987), is actually unsuitable to model seismic intensity attenuation in Italy.

Geophysical Inferences

Following the physical approach of Gupta and Nuttli (1976), the very good fit of intensity data with a linear law in the near field seems to indicate that, in Italy, the prevailing physical mechanism controlling the amplitude of seismic waves is actually the anelastic attenuation. At distances shorter than 45 km, the logarithmic contribution of the geometrical spreading appears, in fact, almost negligible. In this framework the second linear section with slighter slope still indicates a prevailing anelastic attenuation mechanism but with a lower dissipation. Even the very high value ($k = 1.7$) of the coefficient of anelastic attenuation as can be computed by the previous expression indicates an anomalously high attenuation of the Italian area when compared, for example, with California ($k = 0.6$) and the central United States ($k = 0.1$) (i.e., Nuttli, 1973).

The prevalence of anelastic attenuation with respect to geometrical spreading can be related to the existence, in the Italian area, of a highly fractured thrust belt in the central and northern Apennines and of a high heat flow extensional basin in the Tyrrhenian Sea and surrounding regions (see Fig. 8). The former may induce multiple ray path scattering of seismic waves that could reduce the effect of geometrical spreading even in the vicinity of the source, while the latter may be responsible for a particularly high anelastic dissipation.

The observed abrupt change of the slope of the attenuation curve at about 40–50 km from the epicenter was already predicted for central Italy by Fäh and Panza (1994) on the basis of the numerical simulation of the attenuation of the PGA and of the energy integral *W* that appears in the definition of the Arias intensity (Arias, 1970). They obtained the curves of Figure 9 by a hybrid technique that combines the propagation of seismic waves from the source to the bedrock in the vicinity of the observing site by normal mode summation, and the finite-difference computation of the actual motion in the vicinity of the receiver. Fäh and Panza (1994) interpreted this slope change, according to Suhadolc and Chiaruttini (1985), as the transition from direct *Sg* phase to one where several *S*-wave phases mostly reflected at the Moho gradually become part of the *Lg* wave train.

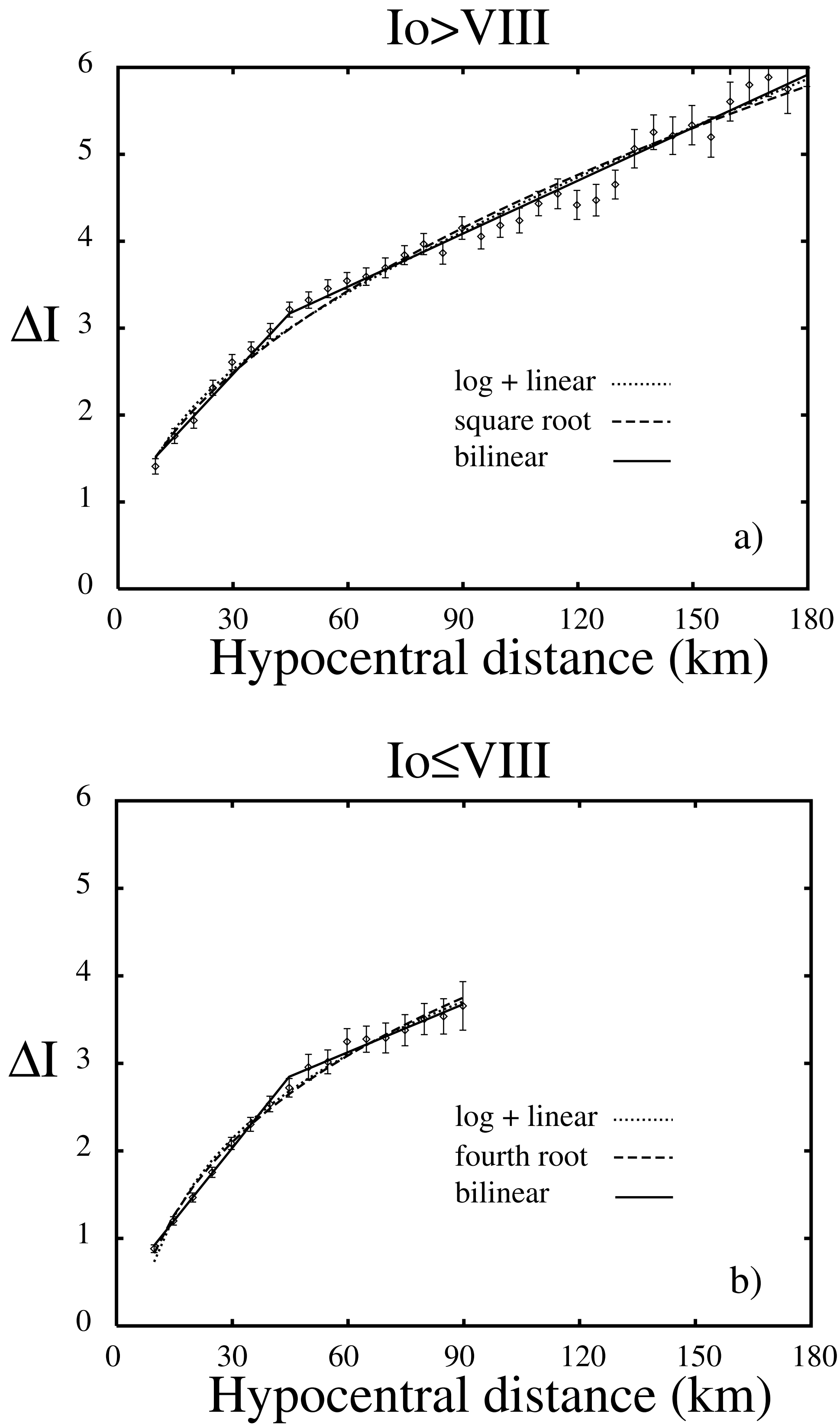

Figure 6. Data fitting of the different attenuation models, as in Figure 4 but with selections of the epicentral intensity.

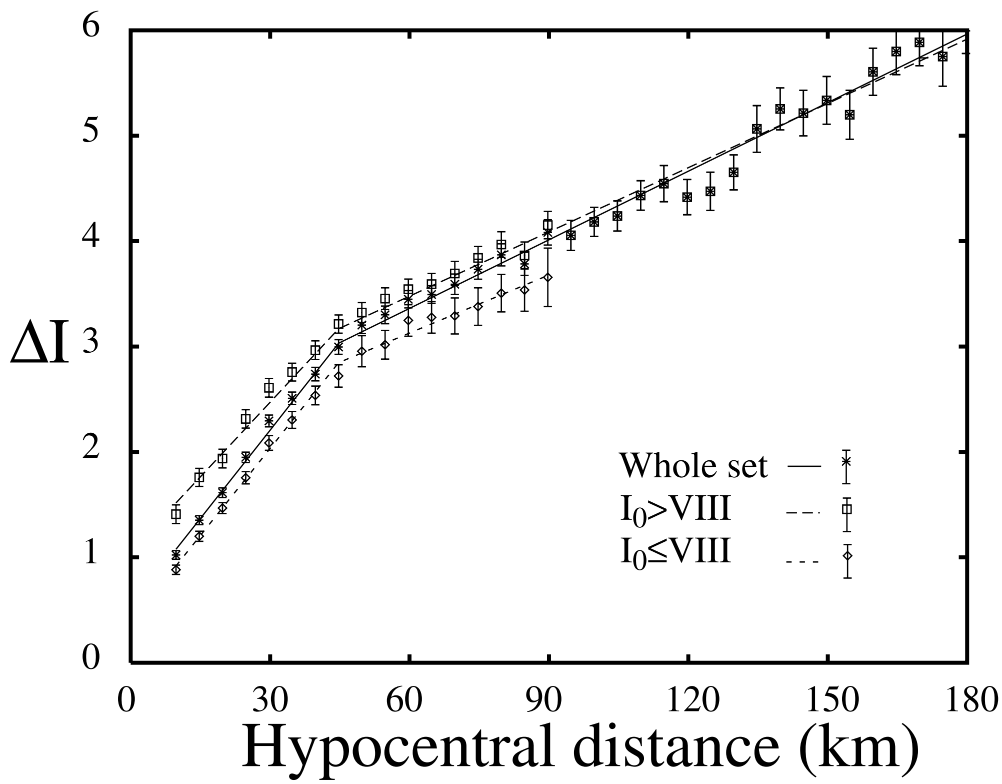

Figure 7. Comparison of the intensity attenuation data fitting for the full database (solid line), the set with $I_0 >$ VIII (dashed) and the set with $I_0 \geqq$ VIII (dotted). The observed averages of intensity differences are reported with 95% error bars. The symbols are as follows: * for the full set, □ for $I_0 >$ VIII, and a ◇ for $I_0 \leq$ VIII.

An anomalously low attenuation of ground-motion amplitudes at distances between 50 and 100 km from the source of the 1989 Loma Prieta earthquake was observed by several investigators (Somerville and Yoshimura, 1990; Fletcher and Boatwright, 1991; McGarr *et al.,* 1991). Somerville and Yoshimura (1991) interpreted this as the effect of critical reflections from the base of the crust that they identified on the basis of the arrival times and phase velocity, and by comparison with simulated accelerograms. The amplitude of these phases would be large and would occur at relatively close range because of the deep focal depth of the earthquake and the strong velocity gradient at the base of the crust (ibidem). Further analysis of four aftershocks of the Loma Prieta earthquake by McGarr *et al.* (1991) confirmed that the *SmS* phase might cause the maximum ground motion of the *S*-wave train at distance longer than critical (43 km).

In Italy, the low attenuation at distances longer than 45 km could not be justified by the deep focal depth of damaging sources (see above), while a rapid increase of the *S* waves quality factor with depth would indeed well explain the observed pattern. This increase would cause the lower crust act like a wave guide that transmits the seismic energy of reflected waves with higher efficiency than the direct waves in the upper crust. However, a clear transition between upper and lower crust layers that justifies the hypothesis of a low attenuation channel was not evidenced in Italy by 1D seismic velocity analysis (Chiarabba and Frepoli, 1997). Another peculiarity in our data is that the low attenuation distance range is not limited as in California to the interval from 50 to 100 km but extends with almost the same characteristic up to at least 180 km.

A different interpretation of this shape of the attenuation curve can be inferred in light of the results of recent works by Mele *et al.* (1996, 1997) on compressional and shear refracted wave attenuation in Italy. Analyzing the seismograms of Ionian Greece earthquakes at stations located in northern and central Italy, they found a very high attenuation zone beneath the central and southern Apennines (see Fig. 8) where it would seem that the *Sg* waves do not to travel enough distance to generate the *Lg* waves. The latter, in fact, disappear from the seismograms. In this area, which also extends beneath the Southern Tyrrhenian Sea, the refracted phases instead are propagated, although considerably attenuated.

The existence of a high-attenuating (low velocity) zone in the Apennines bounded on the northeast by a low-atten-

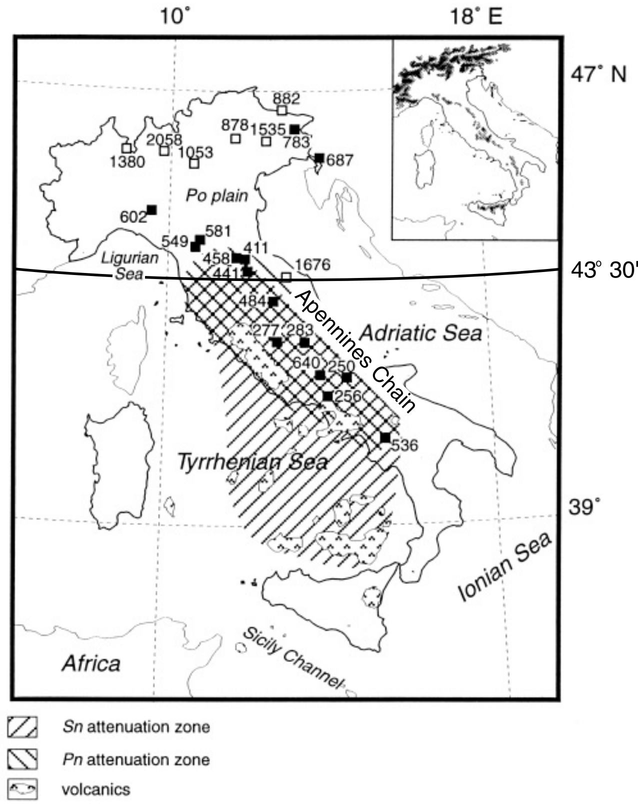

Figure 8. Map of Italy showing the *Sn* attenuation zone inferred by Mele *et al.* (1997). The uppermost mantle *Q* values, estimated by Mele *et al.* (1996), from *Pn* spectral ratios and the inferred high attenuation zone are also shown.

uating (high velocity) zone is also confirmed by recent travel-time tomography images of Italy by Piromallo and Morelli (1997) and Di Stefano *et al.* (1999).

Since most of the Italian earthquake sources included in our macroseismic database are located just along the Apennines chain, the aforementioned evidence could mean that the second part of the intensity attenuation curve shown above is actually controlled by subcrustal phases. Under this hypothesis the lower attenuation of seismic waves in the upper layer of the mantle with respect to the crust could indeed well explain the slighter slope of the second part of the attenuation curve.

This interpretation is not in disagreement with the obvious evidence that the crossover distance between *Sg* and *Sn* in the Italian region (with a crustal thickness ranging from 20 to 35 km) is about 120–150 km. In fact, the key point here is not which phase is the fastest but instead which phase is the strongest. Thus, the position of the change point at 45 km from the seismic source would be determined only by the interplay of the different attenuation of the crust rocks compared to the upper mantle.

In order to further investigate this hypothesis, we performed the same computations shown previously, but we selected the geographic region of the sites where intensity was observed. In Figure 10a and b the attenuation curves are shown for the sites located north and south of latitude 43°30′, respectively (see Fig. 8). This latitude indicates approximately the boundary between the refracted waves attenuating zone in central Italy (values of *Q* parameter for *Pn* phases below 300) and the less attenuating area in the north (values of *Q* above 400 up to 1500). In both cases (Table 3) the bilinear model provides a better fit to the data than the other simpler models.

We can note that the slope of the first part of the attenuation curve (up to 45 km) is similar for the two regions (from 16 to 19 km per degree), while it is quite different in the second part, where it is significantly slighter for the northern region (about 73 km/degree) than for the southern one (42 km/degree). The correspondence with the evidence, pointed out by Mele *et al.* (1996, 1997), of a decidedly lower attenuation of *Pn* and *Sn* phases beneath the Po Plain and the Adriatic Sea with respect to the Apennines is quite good. Although the hypothesis that at distances longer than 40–50 km the transmission of seismic energy is actually controlled by refracted waves is supported by these findings, it however

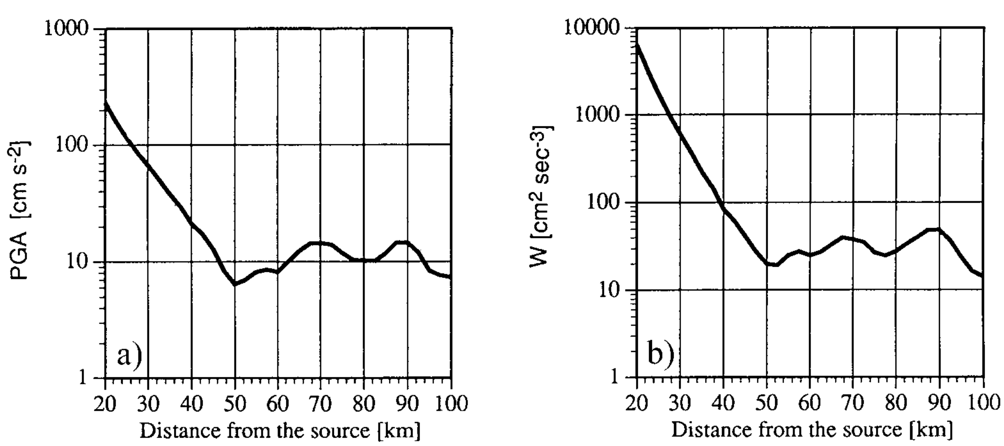

Figure 9. Simulated attenuation of (a) PGA and energy integral (b) W at bedrock with distance according to Fäh and Panza (1994).

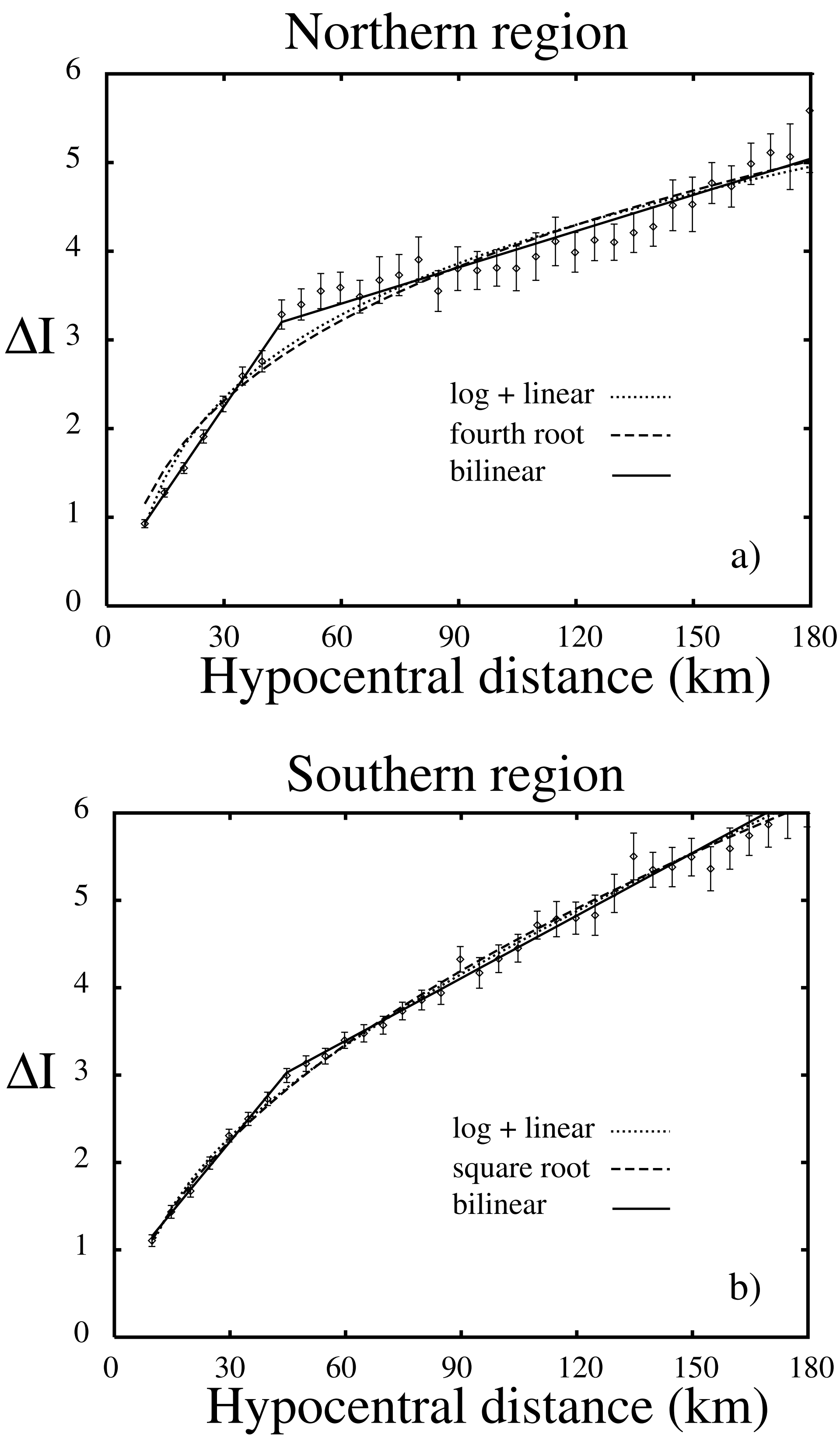

Figure 10. Data fitting of different attenuation models as in Figure 4 but for selecting the sites located (a) north and (b) south of latitude 43°30′.

Table 3

Data Fitting and Goodness of Fit Parameters of the Attenuation Model for Northern and Southern Regions*

Region	α	β / $1/\beta$	γ / $1/\gamma$	R^2	s	F(good) / F(best) / F(log1)	n_n	n_d	p
Northern	0.30 ± 0.15	0.065 ± 0.004	0.0136 ± 0.0008	0.97	0.17	1231.	1	31	1.00
		15.5 ± 0.9	73.5 ± 4.3			17.2	2	31	1.00
						32.4	1	31	1.00
Southern	0.63 ± 0.11	0.054 ± 0.003	0.0239 ± 0.0006	0.99	0.12	4174.	1	31	1.00
		18.7 ± 1.0	41.8 ± 1.0			4.6	2	31	0.98
						4.0	1	31	0.94

*Northern, north of latitude 43°30′; southern, south of latitude 43°30′.

requires further investigations in order to be confirmed. In fact, due to the critical nature of head waves, it is quite surprising that these phases can be responsible for significant effects on the natural and human environment. As a matter of fact, the very efficient propagation ($Q \approx 4000$) of *Sn* phases, observed by Isacks and Stephens (1975) in the suboceanic lithosphere, could suggest that channeled waves, propagating in the upper mantle above the low-velocity zone and with a group velocity extrema near the *Sn* velocity, might carry enough energy to explain the observed effects.

Summary

The analysis of the Italian intensity database showed that the attenuation of seismic intensity is well described by a bilinear (piece-wise linear) function with a clear slope change at about 45 km from the epicenter. The analysis of variance statistical test demonstrated that the bilinear model fits the data better than any other simple attenuation formulas. A very similar slope change at about 50 km from the source was already observed for PGA, on occasion of the Loma Prieta earthquake (Somerville and Yoshimura, 1990; Fletcher and Boatwright, 1991; McGarr *et al.,* 1991) and was predicted for central Italy, by numerical simulation of the attenuation of the PGA and the Arias intensity (Fäh and Panza, 1994).

In most cases, the square root model proved to be the simple two-parameters formula which fit better than others with the different data sets. The cubic root model (CRAM) is the best only when data with intensity below the completeness threshold are used, while the logarithmic model is the worst in most cases. A three-parameter model of the type proposed by Gupta and Nuttli (1976), including both a logarithmic and a linear term, showed better agreement than two-parameter models but worse agreement than the bilinear.

The use of the MCS intensity as a continuous linear parameter appears reasonably appropriate, as the observed attenuation trends look similar for both high and moderate intensity earthquakes. Moreover the intensity difference residuals at various epicentral distances closely follow a normal distribution (with standard deviation of 1.15 intensity degrees).

The linear behavior of the intensity decay with distance both in the near and in the far field indicate that in Italy, the anelastic attenuation definitely prevails over geometrical spreading. This may be interpreted as the effect of a high heat flow and of a highly fractured crust that induce both an anomalously high dissipation of wave energy and a marked multipath scattering of seismic phases.

The slope of the attenuation curve in the first 45 km from the epicenter is quite uniform (16 to 19 km/degree) in two different macroareas of the Italian peninsula. However, for longer epicentral distances, a lower attenuation in the northern region (about 75 km/degree) than the southern one (about 42 km/degree) was found. This agrees with recent results on attenuation of *Pn* and *Sn* waves in Italy by Mele *et al.* (1996; 1997). The correspondence between seismic intensity and refracted wave attenuation properties could indicate a subcrustal path of the phases controlling the radiation of seismic energy at distances longer than 45 km. Further investigations are required to demonstrate the validity of this hypothesis. These investigations should include fitting a bilinear attenuation law with instrumentally measured ground-motion parameters like the logarithm of PGA or the Arias intensity, or simulation of seismic phase amplitudes by ray tracing codes, using appropriate values of the attenuation parameters.

Even though this quite strong interpretation is not accepted, it can be reasonably asserted that the path of waves controlling the second part of the curve is definitely deeper than the path corresponding to the first part, as the differences in the attenuation between the uppermost and lowermost crustal layers may as well explain the observed piecewise linear behavior. Even the agreement of the observed differences between the northern and southern regions with the works by Mele *et al.* (1996, 1997) can be interpreted in this framework, assuming that a strict connection does exist between the physical and compositional properties of the lower crust and upper mantle in the different zones.

The bilinear form of the intensity attenuation curve allows inferring the lateral variation of attenuation properties in Italy by a simple 2D linear inversion of intensity data (currently in preparation). In light of the previous interpretations, this would map the attenuation heterogeneities, discriminating the contribution of two lithospheric layers with different depth. A more detailed and physically meaningful description of the attenuation of seismic intensity could be useful to better constrain the hazard estimates in Italy as well as the methods for the inversion of seismic parameters of historical earthquakes that some authors (Bakun and Wentworth, 1997; Gasperini *et al.,* 1999) recently developed.

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