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S-Wave Attenuation Characteristics beneath the Vrancea Region in Romania: New Insights from the Inversion of Ground-Motion Spectra

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Abstract The *S*-wave attenuation characteristics beneath the Vrancea region in Romania are analyzed from the spectra (frequency range 0.5–20 Hz) of more than 850 recordings at 43 accelerometric stations of 55 intermediate-depth earthquakes (M 4.0–7.1) that occurred in the Vrancea seismogenic zone. The method commonly chosen for this type of investigation in the case of crustal earthquakes is the generalized inversion technique (GIT) (e.g., Andrews, 1986; Castro *et al.*, 1990). Yet the Vrancea dataset is entirely different from common crustal datasets. Because of the strong clustering of the hypocenters within a very small focal volume, there are only few crossing ray paths from sources to receivers. As a consequence, inhomogeneities in the attenuation properties are not averaged out, which leads to unphysical results if the standard GIT approach is adopted. The problem is discussed qualitatively by performing tests with synthetic data and solved quantitatively by adapting the GIT technique in view of these peculiarities. With the optimally adapted inversion scheme, it is possible to unravel differences in the attenuation characteristics between two (or more) sets of stations. The results show that the attenuation of seismic waves is roughly comparable in the low frequency range ($<$4–5 Hz) but stronger by up to an order of magnitude at higher frequencies within the Carpathian mountain arc as compared with the foreland area. Modeling this strongly frequency-dependent lateral variation of seismic attenuation by a significantly lower Q beneath Vrancea (1) provides a very good fit of observed strong-motion characteristics, (2) sheds new light on the distribution of intensities of the previous strong earthquakes, (3) will have strong implications for future hazard assessment, and (4) is fully compatible with structural models from deep seismic sounding, tomography, and teleseismic attenuation.

Introduction

The high level of seismic hazard for Romania and especially for the city of Bucharest is almost entirely due to the intermediate-depth seismicity within the Vrancea seismogenic zone (e.g., Lungu *et al.*, 1999; Sokolov *et al.*, 2004). During the last century, four large Vrancea earthquakes with moment magnitudes higher than 6.5 occurred here. Of these events, the 10 November 1940 (M_w 7.7) and 4 March 1977 (M_w 7.4) inflicted serious consequences on Romanian territory. More than 1500 fatalities resulted from the latter earthquake throughout Romania, with about 1400 of these in the capital Bucharest, where 32 tall buildings collapsed (Cioflan *et al.*, 2004). The Vrancea earthquakes show the peculiarity that the hypocenters are clustered in a very narrow source volume. The epicentral area is limited to approximately 30 × 70 km^2 and the depth range extends from around 80 to 200 km (Fig. 1). Furthermore, there is a seismic gap at 40–70 km depth separating the intermediate-depth from the less pronounced crustal seismicity (Fuchs *et al.*, 1979).

The present tectonic setting of the area under investigation is considered to be the result of a retreating subduction zone with subsequent soft continental collision and slab rollback (Sperner *et al.*, 2001). The continent-continent collision formed the Carpathian mountain arc (Fig. 1). The Vrancea seismogenic zone is located in the bending zone of the Carpathian arc and is bounded by the Transylvanian and Focsani basins in the northwest and southeast, respectively. The strong seismicity at intermediate depth in the Vrancea region is viewed to be related to the break-off of the subducted lithosphere, in a slab segment not yet completely detached. This interpretation is not only compatible with the observed focal mechanisms, indicating a thrust regime with horizontal compression and vertical extension (e.g., Constantinescu and Enescu, 1964; Oncescu and Bonjer, 1997), but it is also supported by the regional seismic tomography results of Martin *et al.* (2006), which image the slab and show that the seismicity is confined to the slab.

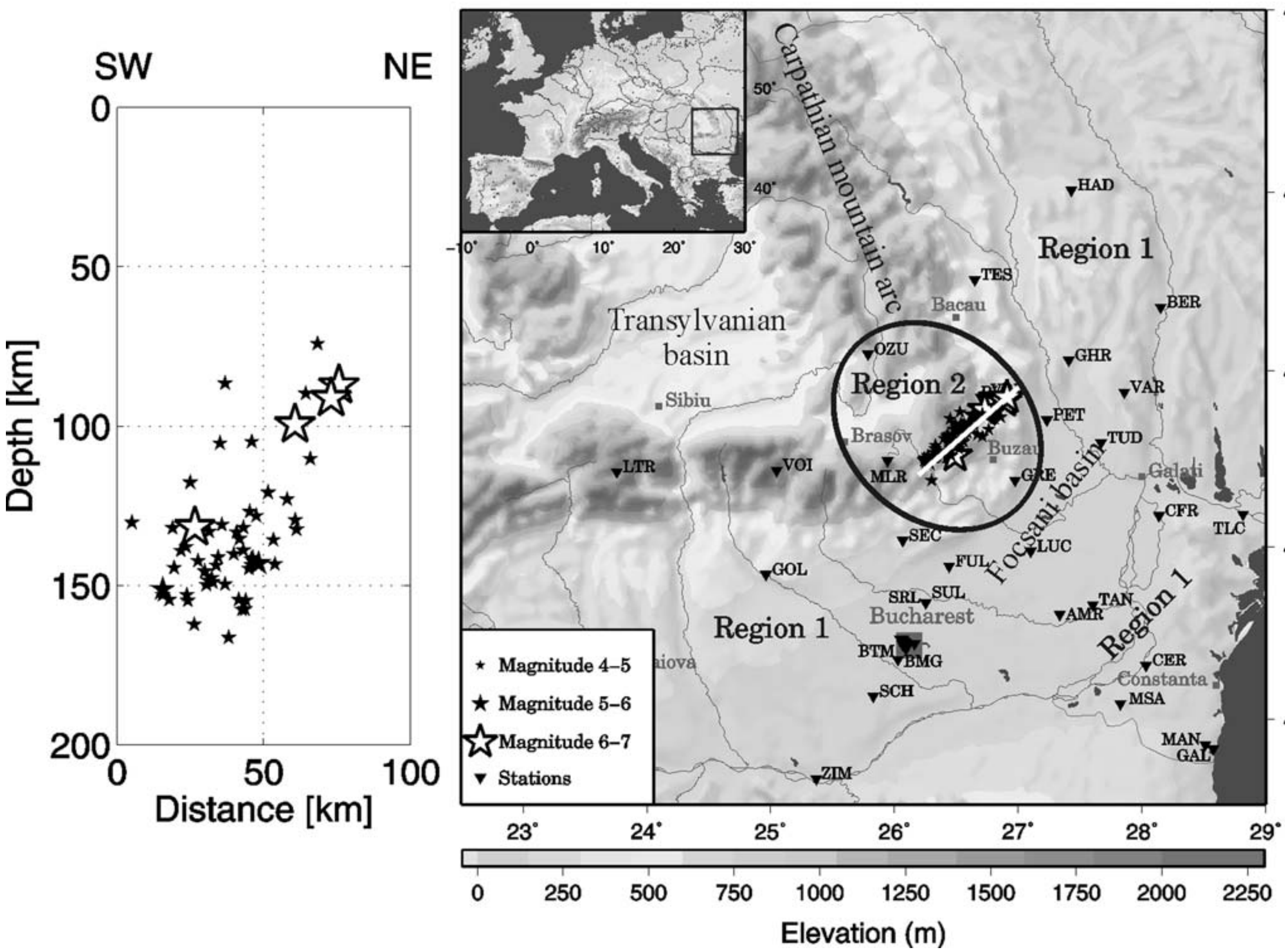

Figure 1. Map of Romania and source-station configuration. The epicenters of the utilized earthquakes are depicted by stars and the stations by inverse triangles. In total, 55 earthquakes and 43 stations have been used. On the left, a southwest–northeast vertical cross section through the epicentral area is shown (white line on the map). Region 1 is given by the forearc area, whereas region 2 includes stations in and around the epicentral area. More details on the classification of the stations in region 1 and 2 can be found in the text.

Seismic attenuation (whose anelastic part is commonly characterized by the quality factor Q) and its frequency dependence can provide valuable information on the underground structure, and they are important quantities in view of seismic hazard assessment. In fact, they strongly influence the distribution of ground-motion parameters and are needed for the simulation of realistic strong-motion time histories, for instance, when using the stochastic technique (e.g., Boore, 2003). Previous studies have been dealing with the spectral attenuation characteristics beneath Vrancea (Popa *et al.*, 2005; Russo *et al.*, 2005, Sudhaus and Ritter, 2005; Radulian *et al.*, 2006, Weidle *et al.*, 2007) and their results will be discussed in conjunction with those of this study. However, no systematic analysis of the spectral attenuation characteristics over a large frequency band using a rigorous inversion procedure has been performed for the Vrancea area as yet, and the main purpose of this study is to fill this gap using the largest amount of accelerometric data available. The analyzed frequency range extends from 0.5 to 20 Hz.

For attenuation studies with datasets from crustal earthquakes, a commonly applied method is the generalized inversion technique (GIT) (Castro *et al.*, 1990; Parolai *et al.*, 2000, 2004; Bindi *et al.*, 2006), which can be used to separate source, path (i.e., attenuation), and site effects contributing to the Fourier amplitude spectra of ground motion. In the case of Vrancea earthquakes, however, the geometry of the dataset is entirely different from the common crustal case. As the sources are all clustered in approximately the same location (Fig. 1), there are few crossing ray paths between different source-station pairs, and all rays from the different sources to a given station are almost identical. Whereas for usual datasets of crustal earthquakes, due to many ray crossings, differences in attenuation along different ray paths can be viewed to average out and the determination of an average attenuation curve with distance at each analyzed frequency makes sense, this is not the case for Vrancea earthquakes, as we show in the following. Furthermore, due to the depth range of these events, the lowest hypocentral distance in the dataset is about 90 km. These problems are inherent to the dataset and cannot be avoided in any study of seismic attenuation for the Vrancea case. However, we show in this article that with some modifications in the GIT inversion scheme, these difficulties can be overcome to a large extent. Hence this study contributes both to the further development of the GIT approach and provides new insights into the attenuation of seismic waves beneath Vrancea.

We first present results obtained by applying the standard GIT procedure to the dataset. Synthetic tests are carried out to show that the standard procedure is inefficient in providing reliable attenuation estimates when lateral variability is expected in the attenuation characteristics. After a modification of the GIT inversion scheme to account for this variability, the attenuation characteristics for two separate regions (within the Carpathian arc and around it) are derived in one common inversion.

Dataset

We analyze seismograms from 52 earthquakes recorded by 43 stations (Fig. 1) of the accelerometer network deployed in Romania since 1997 in the framework of the Collaborative Research Center (CRC) 461: "Strong Earthquakes: A Challenge for Geosciences and Civil Engineering" of the University of Karlsruhe in collaboration with the National Institute for Earth Physics (NIEP) in Bucharest. The stations consist of three-component accelerometers (Episensors) equipped with Kinemetrics K2-dataloggers with Global Positioning System (GPS) timing. Fifteen of the stations are deployed within the city limits of Bucharest, 10 of which are used in this study (these stations are not all labeled on the map in Fig. 1, due their closeness to each other). The sampling rate of the data from this network is 200 samples/sec. The records were corrected for instrumental response and band-pass filtered between 0.1 and 50 Hz using standard Kinemetrics software.

The magnitude range of the Vrancea earthquakes recorded with acceptable signal-to-noise ratio (SNR) with this network ranges from 4.0 to 5.8. This dataset was complemented with several strong-motion recordings from the three large Vrancea earthquakes that occurred on 30 August 1986 (M_w 7.1, eight records), 30 May 1990 (M_w 6.9, eight records) and 31 May 1990 (M_w 6.4, three records). These are analog SMA-1 recordings obtained by a network operated by the National Institute for Earth Physics (Oncescu, Bonjer, *et al.*, 1999). These data have been scanned and digitized at NIEP. Following a standard procedure (Lee and Trifunac, 1979), baseline correction, instrument removal, and Ormsby filtering between 0.125–24 Hz were performed. The digitized recordings have a sampling rate of 100 samples/sec.

Thus, in total, more than 850 records from 55 Vrancea earthquakes recorded at 43 sites are used in this study. The hypocentral coordinates were extracted from the ROMPLUS catalog (Oncescu, Marza, *et al.*, 1999), and the depth range of the events (Fig. 1) ranges from 70 to 180 km. An example for a three-component recording is shown in Figure 2. Because of the near-vertical incidence, the *P* wave on the vertical (*Z*) component depicts large amplitudes, with maximum energy in the frequency range around 8–10 Hz. Very little *S*-wave energy is seen on the *Z* component, and in many cases, it is entirely missing. At some stations located in the forearc region, such as GRE shown in Figure 2, strong high-frequency arrivals can occasionally be found on the *Z* component around 5–8 sec before the *S*-wave arrival on the horizontal components. These may be related to *S*-*P* conversions at the base of the deep Focsani basin (with depths up to 22 km, Hauser *et al.*, 2007) located on the forearc side of the mountain range. As a result of the missing *S*-wave energy, the *Z* component cannot, of course, be used to derive *S*-wave attenuation properties. The observations mentioned previously are, however, worth further investigation and have important consequences with respect to the estimation of site amplification effects using the H/V spectral ratio method, as we discuss elsewhere (Oth, 2007; Oth *et al.*, 2009).

In the following, the *S*-wave portions of the horizontal component accelerograms are used. Time windows starting 1 sec before the *S*-wave onset and ending when 80% of the total energy of the record are reached were extracted and tapered with a 5% cosine window. Typical window lengths range between 5 and 15 sec. In the few cases where the determined window was longer than 20 sec, it was fixed to a maximum duration of 20 sec to avoid having too much coda energy in the analyzed time windows. The Fourier amplitude spectra (FAS) were computed for each window and smoothed around 30 frequency points equidistant on logarithmic scale between 0.5 and 20 Hz using the windowing function proposed by Konno and Ohmachi (1998), with $b = 20$. Noise spectra were computed from preevent noise windows of equal length as the considered signal windows, and at each

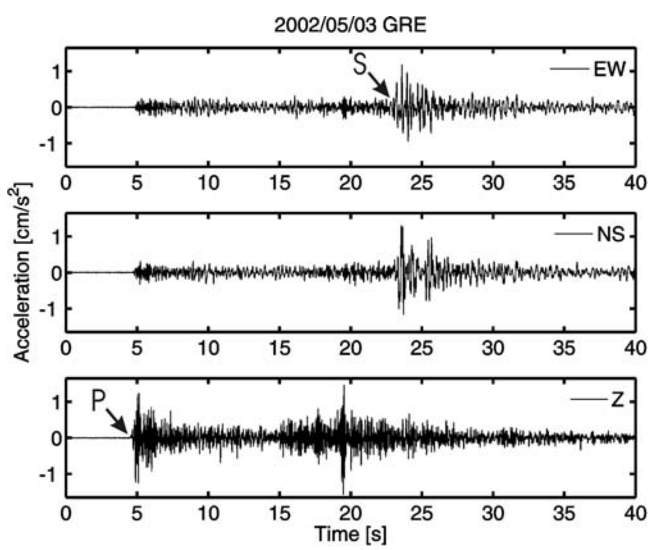

Figure 2. Example for the three-component recordings at station GRE. The accelerograms shown were obtained from a M_w 4.6 event in 162 km depth.

of the chosen 30 frequencies, only data points with an SNR higher than three were included in the final dataset. Below 0.5 Hz, the SNR was generally not acceptable for the smaller earthquakes (mostly smaller than 1.5–2). As a result, the database is slightly sparser at the lowest frequencies. Finally, the root-mean-square (rms) average of the two horizontal component spectra was computed. Figure 3 shows several examples of the chosen time windows and the respective spectra.

Methodology and Synthetic Data Tests

The generalized inversion technique (GIT) (Andrews, 1986; Castro *et al.*, 1990) is a well-studied method to derive frequency-dependent attenuation characteristics as well as source and site spectra, especially in the case of crustal earthquakes (e.g., Parolai *et al.*, 2000, 2004; Bindi *et al.*, 2006). There are essentially two ways to tackle the inversion: either a parametric approach is adopted (i.e., the attenuation along the travel path is assumed to follow a particular functional form, e.g., Salazar *et al.*, 2007) and the inversion is performed in one step to obtain source and site spectra and a $Q(f)$-model, or attenuation is only constrained to be a smooth decaying function with distance, which is the nonparametric approach introduced by Castro *et al.* (1990). With the parametric approach, it is a common problem that negative $Q(f)$-values are obtained (e.g., Castro *et al.*, 1990; Salazar *et al.*, 2007). Such results can, for instance, be due to the assumptions regarding the geometrical spreading, as discussed by Castro *et al.* (1990), or to the simple fact that the seismic attenuation in a given region cannot be appropriately described with the commonly utilized simple parameterizations.

Therefore, we follow the nonparametric approach, which splits the inversion into two steps: first, the dependence of the spectral amplitudes $U(f,r)$ at frequency f on distance may be written as:

$$U_{ij}(f, r_{ij}) = A(f, r_{ij})\hat{S}_j(f), \tag{1}$$

where $U_{ij}(f, r_{ij})$ is the spectral amplitude (acceleration) at the ith station resulting from the jth earthquake, r_{ij} is the source-site distance, $A(f, r_{ij})$ is the function describing seismic attenuation with distance, and $\hat{S}_j(f)$ is a scaling factor depending on the size of the jth event. $A(f, r_{ij})$ is not supposed to have any specific shape and implicitly contains all effects leading to attenuation along the travel path (geometrical spreading, anelasticity, scattering, etc.). Based on the idea that these properties vary slowly, $A(f, r_{ij})$ is constrained to be a smooth function of distance (i.e., with a small second

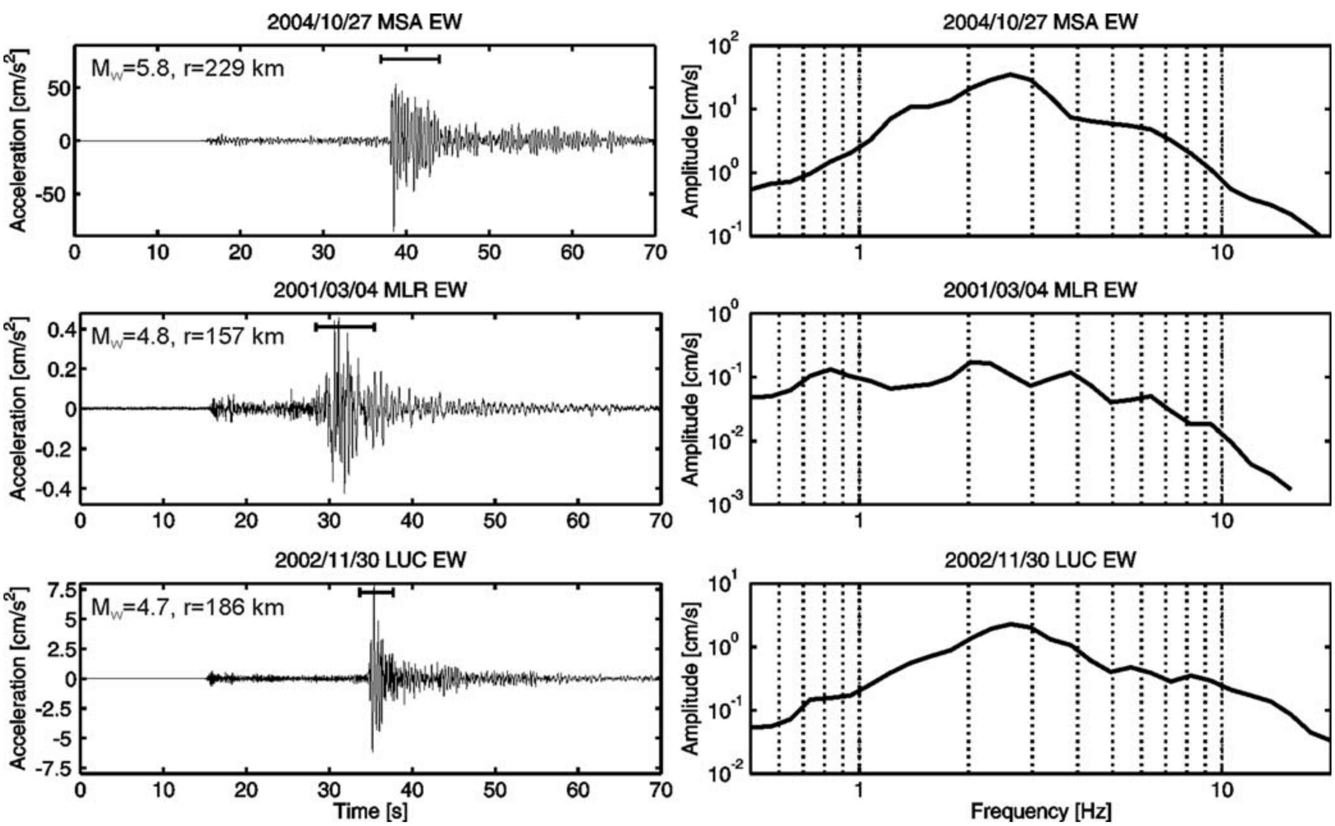

Figure 3. Left panel: example records (east–west components) and *S*-wave windows (marked by the bar above the time history) used for the computation of Fourier amplitude spectra. The origin of the time axis does not correspond to the origin time of the respective event, but it has been adjusted for displaying purposes. The magnitude and hypocentral distance are given in the upper left corner. Right panel: smoothed Fourier amplitude spectra of the selected *S*-wave windows.

derivative, Castro *et al.*, 1990) and to take the value $A(f, r_0) = 1$ at some reference distance r_0.

The representation given by equation (1) does not include a factor related to the site, and hence the site effects are necessarily absorbed both in $A(f, r_{ij})$ and $\hat{S}_j(f)$. $\hat{S}_j(f)$ contains, in fact, an average value of the site amplifications of all stations that recorded the *j*th earthquake. Thus, $\hat{S}_j(f)$ is not the true source spectrum of the *j*th event. The smoothness constraint mentioned previously is implemented by a weighting factor (see, e.g., Castro *et al.*, 1990 and the following section) and is supposed to suppress rapid undulations in the attenuation function. This weighting factor must be reasonably chosen in order to suppress site-related effects and yet preserve variations of the attenuation characteristics with distance.

After the attenuation functions have been determined, a subsequent inversion allows separating source and site effects (Andrews, 1986):

$$R_{ij}(f) = S_j(f) Z_i(f). \qquad (2)$$

Herein, $R_{ij}(f) = U_{ij}(f, r_{ij})/A(f, r_{ij})$ are the residual spectral amplitudes after correction of the attenuation functions, $S_j(f)$ is the source spectrum of the *j*th earthquake, and $Z_i(f)$ the site amplification function at station *i*. In this article, we focus on the attenuation characteristics $A(f, r_{ij})$ resulting from the first step, and the site amplification and source spectra obtained from the second inversion step are discussed elsewhere (Oth, 2007; Oth *et al.*, 2009).

Equation (1) can be easily linearized by taking the logarithm:

$$\log_{10} U_{ij}(f, r_{ij}) = \log_{10} A(f, r_{ij}) + \log_{10} \hat{S}_j(f). \qquad (3)$$

Equation (3) represents a linear system of the form $\mathbf{Ax} = \mathbf{b}$, where $\mathbf{b}$ is the data vector containing the logarithmic spectral amplitudes, $\mathbf{x}$ is the vector containing the model parameters, and $\mathbf{A}$ is the system matrix relating the two of them. Practically, the distance range of the data (Fig. 4) is divided into N_D bins, in our case 10 km wide, and the value of $A(f, r_{ij})$ is computed in each bin. In matrix formulation, equation (3) takes the form:

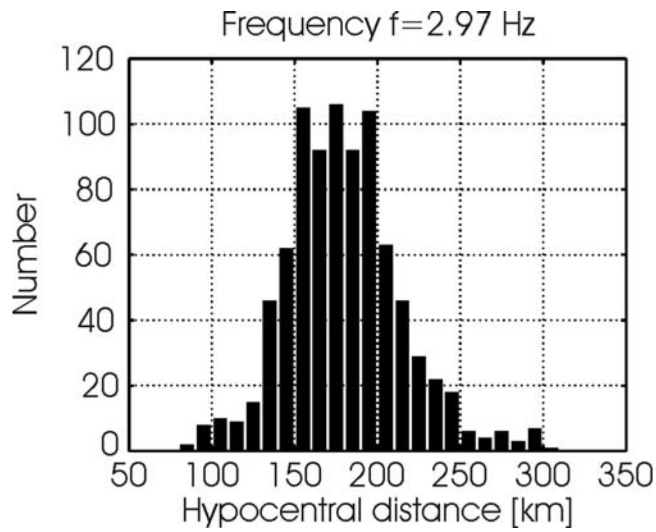

Figure 4. Example of the distribution of hypocentral distances in the dataset at one selected frequency. At lower frequencies, the database is slightly sparser due to signal-to-noise ratio constraints.

The left part of the system matrix contains the factors related to the attenuation parameters, whereas the right-hand side reflects those related to the source terms. The weighting factor w_1 is used to impose $A(f, r_0) = 1$ at the reference distance and w_2 is the factor determining the degree of smoothness of the solution, as discussed previously. The reference distance is set to 90 km, which is the smallest hypocentral distance in our dataset. This very large reference distance is problematic, because setting $A(f, r_0) = 1$ at $r_0 = 90$ km is, in fact, equal to assuming that there is no attenuation over that distance. Hence there is a cumulative attenuation effect over these 90 km that it is impossible to resolve and therefore, the *Q*-models derived later in this article from the slopes of the attenuation functions only reflect the attenuation characteristic over the remaining part of the travel path (i.e., more than 90 km away from the source). However, this cumulative effect plays a crucial role in the modification

$$\begin{pmatrix} \log_{10} U_{11} \\ \cdot \\ \cdot \\ \log_{10} U_{ij} \\ \vdots \\ 0 \\ 0 \\ 0 \\ \cdot \end{pmatrix} = \underbrace{\begin{pmatrix} 1 & 0 & 0 & \cdot & \ldots \\ 0 & 1 & 0 & \cdot & \ldots \\ \cdot & \cdot & \cdot & \cdot & \ldots \\ 1 & 0 & 0 & \cdot & \ldots \\ \vdots & \vdots & \vdots & \vdots & \vdots \\ w_1 & 0 & 0 & \cdot & \ldots \\ -w_2/2 & w_2 & -w_2/2 & \cdot & \ldots \\ 0 & -w_2/2 & w_2 & -w_2/2 & \ldots \\ \cdot & \cdot & \cdot & \cdot & \ldots \end{pmatrix}}_{\text{attenuation parameters}} \underbrace{\begin{matrix} 1 & 0 & 0 & \cdot & \cdots \\ 1 & 0 & 0 & \cdot & \cdots \\ \cdot & \cdot & \cdot & \cdot & \cdots \\ 0 & 1 & 0 & \cdot & \cdots \\ \vdots & \vdots & \vdots & \vdots & \vdots \\ \cdot & \cdot & \cdot & \cdot & \cdots \\ \cdot & \cdot & \cdot & \cdot & \cdots \\ \cdot & \cdot & \cdot & \cdot & \cdots \\ \cdot & \cdot & \cdot & \cdot & \cdots \end{matrix}}_{\text{sources}} \cdot \begin{pmatrix} \log_{10} A_1 \\ \cdot \\ \cdot \\ \cdot \\ \log_{10} A_{N_D} \\ \log_{10} \hat{S}_1 \\ \cdot \\ \cdot \\ \log_{10} \hat{S}_{N_E} \end{pmatrix}. \qquad (4)$$

of the GIT scheme discussed later in order to account for lateral variability of attenuation characteristics beneath Vrancea.

At each of the 30 selected frequencies, the least-squares solution $\mathbf{x} = (\mathbf{A}^T\mathbf{A})^{-1}\mathbf{A}^T\mathbf{b}$ (if the system is numerically stable) can be computed by means of an appropriate inversion scheme (e.g., singular value decomposition, which we use in this work, Menke, 1989). In order to assess the stability of the inversion results, we perform 200 bootstrap inversions (Efron and Tibshirani, 1994) at each frequency point following the procedure explained in Parolai *et al.* (2000, 2004). Given an original dataset of N_{rec} data points, this technique works by repeated inversions of datasets obtained by randomly resampling the original one. From the 200 bootstrap samples, the mean and standard error are calculated for each model parameter.

First Results: Failure of the Standard Approach

In a first attempt, we applied the standard nonparametric GIT to derive $A(f, r)$ as described previously. The results at four selected frequencies are shown in Figure 5. At low frequencies (upper panel), the attenuation functions are more or less monotonically decaying, with a flat portion in the distance range 150–200 km. At frequencies higher than about 4–5 Hz, a bump starts to appear in this distance interval, the effect getting stronger and stronger with increasing frequency (bottom panel). Even though it is not unusual that an attenuation function flattens or even gets sloped upward

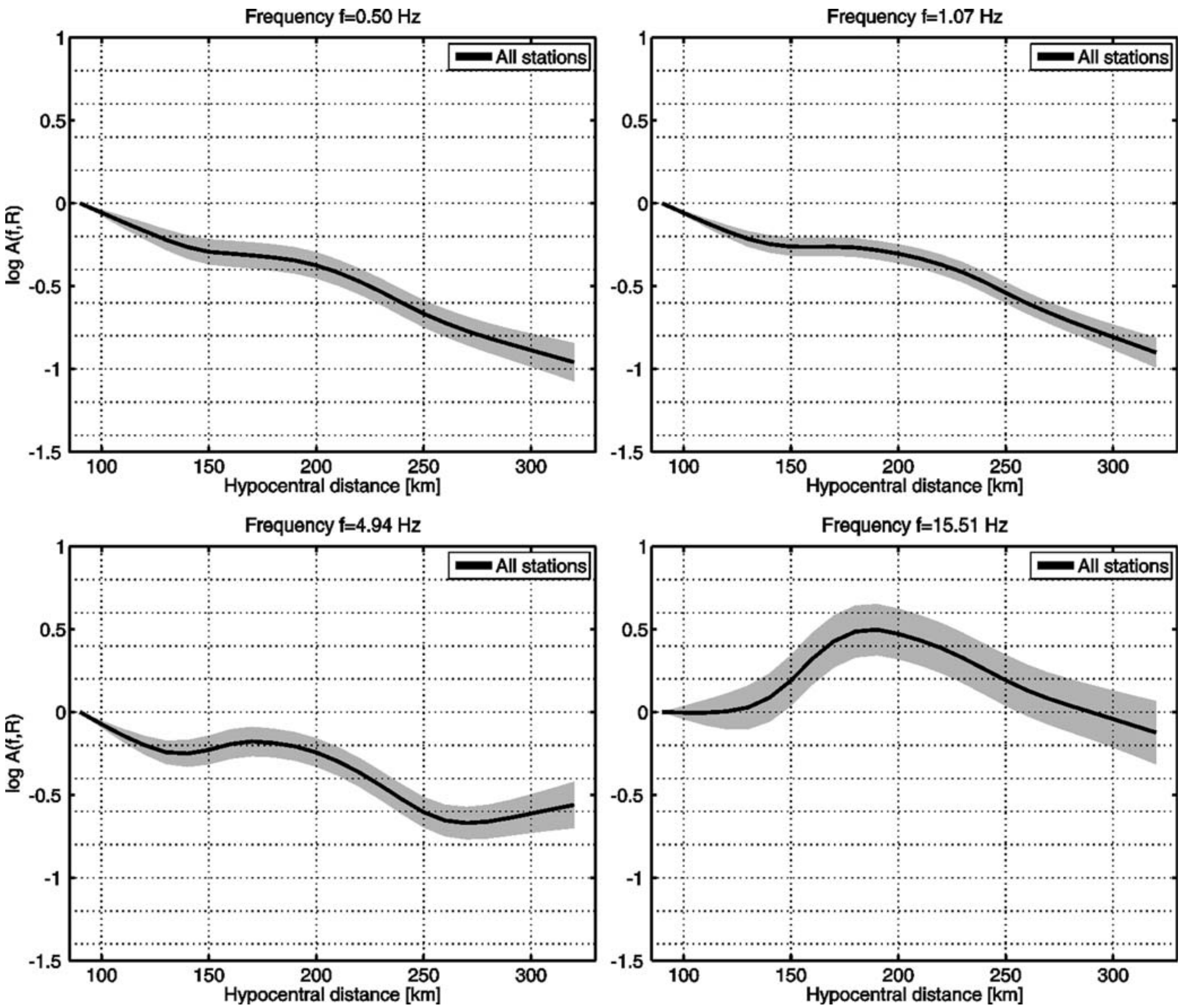

Figure 5. Attenuation functions (at four selected frequencies) obtained by fitting one function to the entire dataset at each frequency point. Black line: mean of 200 bootstrap samples. Gray shaded area: mean ± one standard deviation.

in certain distance ranges due to, for example, critically reflected arrivals from the Moho in the case of crustal earthquakes (Bindi *et al.*, 2006), both the shape and the amplitude of the bump in the attenuation functions at high frequencies obtained here are unphysical. For instance, the curve at a frequency of 15.51 Hz in Figure 5 does not show any decay, with values of $A(f, r) > 1$ everywhere except at the largest hypocentral distances.

Visual inspection of the spectral amplitudes versus distance reveals that the stations MLR, SIR, VRI, PLO, GRE, and OZU, all located directly within the Carpathian arc bending zone (VRI, PLO, and SIR are covered by the stars marking the epicenters in Fig. 1) or close to it, show systematically much lower amplitudes at high frequencies than the other stations of the network (Fig. 6). The difference reaches about one order of magnitude for a given distance at frequencies higher than 8–10 Hz. Therefore, these six stations are herein after regrouped into region 2, while the rest of the stations, almost all of them located in the forearc area, are regrouped in region 1 (Fig. 1). With the introduction of these two regions, radial variations of ground motion (i.e., away from the epicenter) are taken into account, as region 2 is more or less restricted to the epicentral area. In order to also unravel azimuthal variations, a denser network with enough stations in all azimuth ranges around the epicenters would be required.

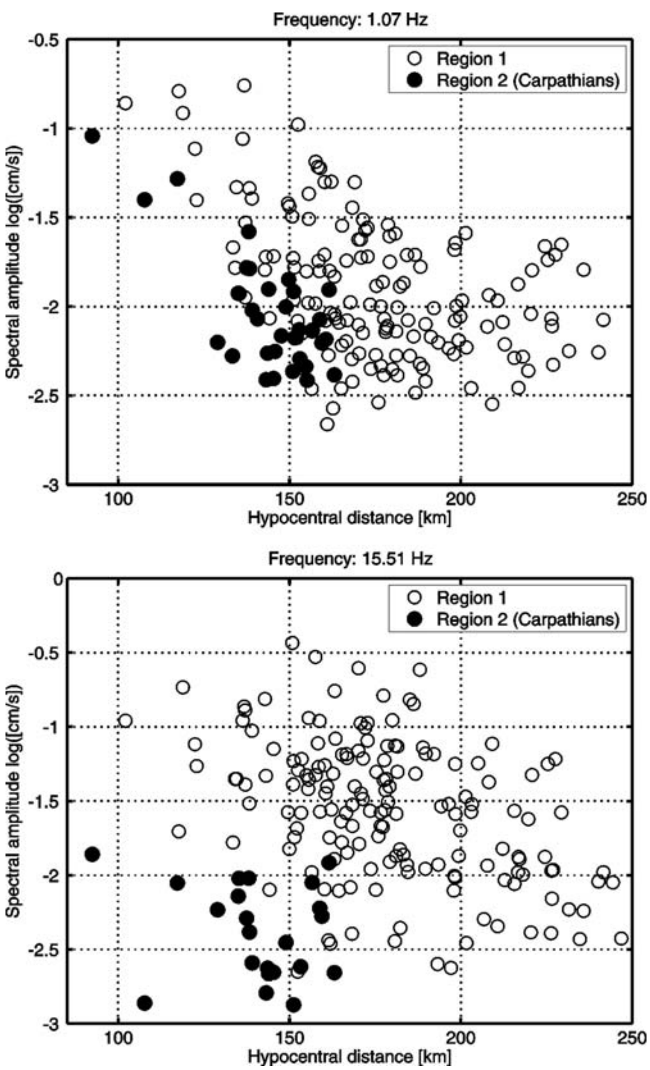

Figure 6. Spectral amplitudes observed at low frequencies (example at ≈1 Hz, top) and high frequencies (example at ≈15.5 Hz, bottom). Note that, with increasing frequency, the spectral amplitudes in region 2 get systematically lower than the ones in region 1 by up to one order of magnitude at similar distances.

This remarkable difference in spectral amplitudes could, in principle, be explained either by a source, site, or attenuation effect. Regarding the source, directivity is an improbable cause, as the effect also appears for small earthquakes, for which this source effect is rather unimportant due to the very small source size of the smaller Vrancea earthquakes and the considered distances (Oth *et al.*, 2007). As the effect appears at high frequencies, the radiation pattern cannot lead to these systematic differences. If the radiation pattern were the cause, one should rather expect to see systematic differences more clearly at low frequencies, as the influence of the radiation pattern decreases with increasing frequency (e.g., Takenaka *et al.*, 2003; Castro *et al.*, 2006).

The possibility that these large differences in amplitude result from different site and basin effects is also unlikely, because except GRE, all of the sites in region 2 are classified as rock stations (Sokolov *et al.*, 2005). If site conditions were the cause, one would expect that a station such as SEC or FUL, located in the deep Focsani basin (Hauser *et al.*, 2007), should show large amplification at some low resonance frequency (as the basin is very deep, see, e.g., Bard, 1999) and large attenuation at high frequencies due to the strong damping in the sediments. However, the opposite of this expected effect is observed.

Hence, a strong difference of attenuation along the travel path is the preferred explanation for these results. In order to verify this hypothesis, we will now summarize the results of two synthetic data tests.

Synthetic Data Tests

The first test is intended to verify whether the bump is a simple problem related to the geometry of the dataset. Therefore, no inhomogeneities in attenuation are considered when generating the synthetic data, using a homogeneous half-space Q-model (model A). With the second test, we investigate whether the bump observed before can be reproduced under the assumption of a variation of Q between regions 1 and 2 (model B).

Both synthetic datasets have exactly the same number of stations and sources as well as the same geometry (i.e., source-station configuration) as the real data. The synthetic acceleration spectra are computed following the standard spectral model as used for stochastic point source simulations (Boore, 1983, 2003):

$$U(M_0, f, r) = (2\pi f)^2 C S(M_0, f) P(f, r) Z(f), \quad (5)$$

with

$$C = \frac{\Re^{\theta\varphi} V F}{4\pi\rho v_S^3 r_0} \quad \text{and} \quad P(f, r) = \frac{1}{r}\exp\left(-\frac{\pi f r}{Q(f) v_S}\right). \quad (6)$$

In equation (5), $U(M_0, f, r)$ denotes the acceleration spectrum, $S(M_0, f)$ the source spectrum, $P(f, r)$ the propagation path effect [i.e., equivalent to $A(f, r)$], and $Z(f)$ the site amplification function. Furthermore, M_0 is the seismic moment, r the hypocentral distance, $\Re^{\theta\varphi} = 0.6$ the average S-wave radiation pattern, $F = 2$ the free surface amplification, and $V = 1/\sqrt{2}$ accounts for the splitting of shear-wave energy into two horizontal components. $\rho = 3.2$ g/cm^3 is the density, and the average shear-wave velocity is set to $v_S = 4.5$ km/sec, which is an acceptable approximation following the seismic tomography results of Martin *et al.* (2006).

The two $Q(f)$-models used are listed in Table 1. For model A, a homogeneous half-space is assumed, whereas for model B, attenuation along travel paths to stations in region 2 is stronger (i.e., Q is lower) than along paths ending in region 1. The frequency dependence is chosen such that the difference in attenuation increases with increasing frequency, in agreement with the observations regarding the spectral amplitudes. These $Q(f)$-models are arbitrarily chosen but are close to models used by other authors (e.g., Sokolov *et al.*, 2005). Because the purpose of these synthetic tests is only to show that lateral variations of the frequency-dependent attenuation characteristics can indeed explain the shape of the attenuation functions but not to derive quantitative results from them (this is done by modification of the original inversion scheme discussed later), this degree of arbitrariness is justified.

The source spectra are modeled following the classical Brune (1970, 1971) model:

$$S(M_0, f) = M_0\left[1 + \frac{f^2}{f_C^2}\right]^{-1}, \quad (7)$$

Table 1
Q-Models Used in the Synthetic Data Tests

Model A	$Q_A(f) = 100f^{0.8}$	
Model B	Region 1 $Q_{B,1}(f) = 150f^{0.8}$	Region 2 $Q_{B,2}(f) = 100f^{0.5}$

Model A assumes a homogeneous half-space, while in model B, seismic attenuation is stronger (i.e., Q is lower) in region 2 than in region 1, the difference between the two regions getting larger with increasing frequency (i.e., different frequency dependence).

where f_c is the corner frequency of the spectrum. The corner frequencies of the synthetic events are randomly chosen, yet decreasing with increasing magnitude. For different magnitude bins, the bounds in between which f_c is chosen randomly are different. For instance, f_c is given in the interval 5–10 Hz for $4 \leq M_w < 4.5$ and 2–6 Hz for $4.5 \leq M_w < 5$. The ranges are deliberately chosen this high, as high corner frequencies are expected for Vrancea earthquakes (Oncescu, 1989; Oth *et al.*, 2007).

For each of the two $Q(f)$-models, site effects in equation (5) were once modeled using the H/V ratios (Lermo and Chávez-García, 1993) and once simply neglected in order to assess the influence of site amplification on the shape of the resulting attenuation curves. Here we show the results obtained including the site effect estimate, but the bump in the attenuation curves discussed later in the article proved to be a stable feature, no matter if an estimate of site amplification is included in the synthetic spectra or not. Finally, normally distributed random noise with a standard deviation of 10% of the respective data values is added to the synthetic spectra.

The resulting attenuation curves $A(f, r)$ are shown for two frequencies in Figure 7a,b for model A and B, respectively. For model A, the attenuation functions are also monotonically decaying at high frequencies, which clearly shows that the spurious results obtained with the real dataset are not related to the geometry of the dataset. Apart from slight deviations from the theoretical model (related to the assumed site effects, as these deviations are not present when the site effect free synthetics are used) at few frequencies, the attenuation model can be acceptably well reproduced by the inverted solution. The attenuation functions obtained using the attenuation model B, however, depict a striking similarity with the results shown in Figure 5. Instead of obtaining attenuation functions in between the two attenuation models for region 1 and 2, a strong bump is obtained. This outcome strongly suggests that the variation in spectral amplitudes between region 1 and 2 is the result of inhomogeneities in whole path attenuation not correctly taken into account by the standard GIT procedure. From the histograms of the distribution of data points with hypocentral distance shown in Figure 8, one can see that the stations in region 2 have a strong influence at smaller hypocentral distances, with a maximum number of data around 150 km. With increasing distance, their influence is constantly reduced until they disappear for distances higher than 180 km. Because, at high frequencies, the spectral amplitudes are much larger in region 1 even for large distances (see Fig. 6, lower panel) as compared with those in region 2, the bump in the attenuation functions is unavoidable if a single attenuation function is fit to the data at each frequency point.

In order to account for lateral variability in the attenuation characteristics between the two regions, the inversion scheme presented in equations (3) and (4) is adapted to this new situation and separate attenuation functions for the two regions are derived in a single inversion.

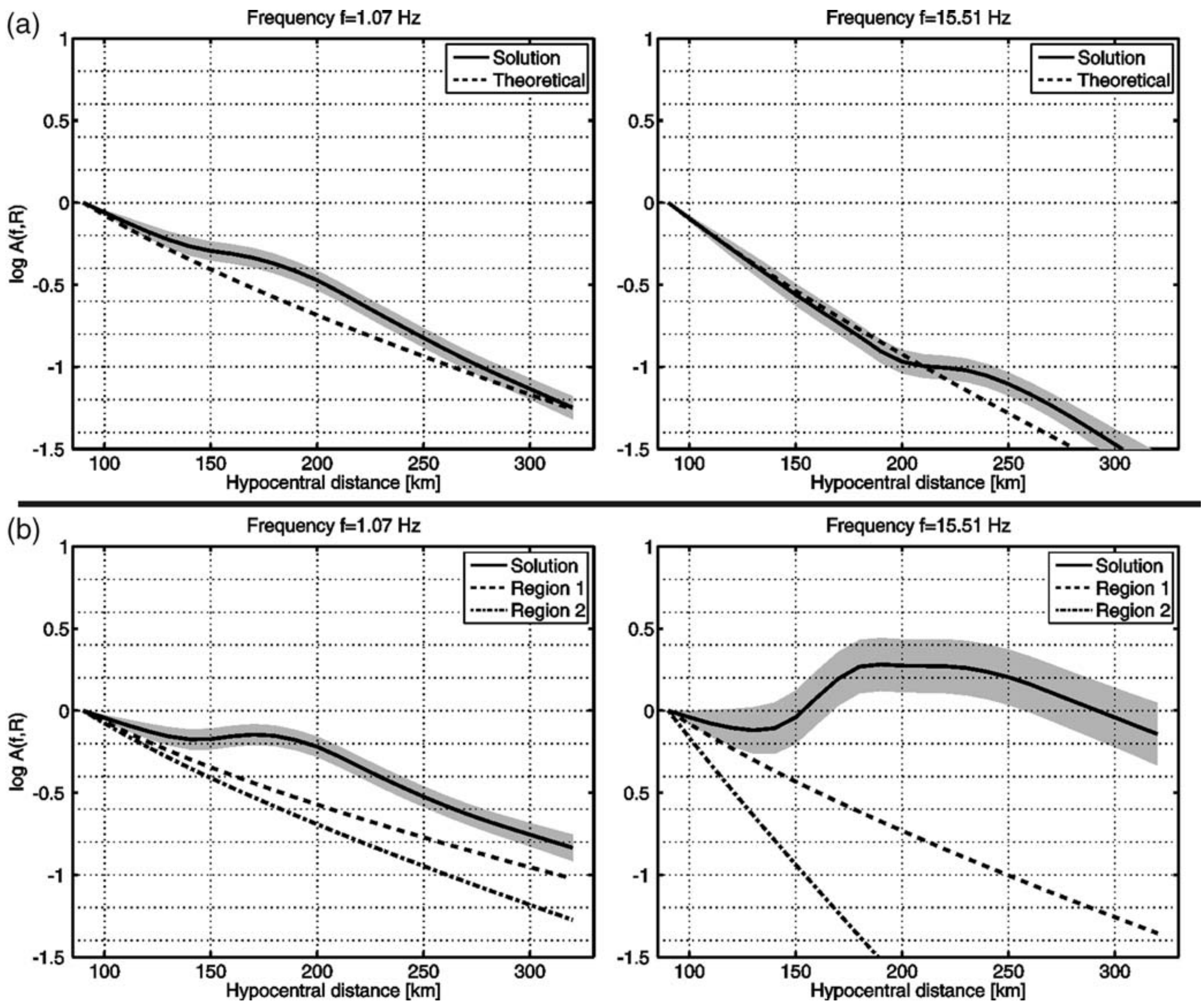

Figure 7. Attenuation functions derived from the synthetic data described in the text at two frequency points. Upper panel: model *A*. Lower panel: model *B*. The black line shows the mean of 200 bootstrap samples, while the gray shaded area delimits the region within one standard deviation. The theoretical models used in the computation of the synthetics are shown as dashed (model *A*) and dashed/dash-dotted lines (model *B*, for the two regions).

The Adapted Inversion Scheme

The adapted inversion scheme can be written as:

$$\log_{10} U_{ij}(f, r_{ij}) = p_1 \log_{10} A_1(f, r_{ij}) + p_2 \log_{10} A_2(f, r_{ij}) + \log_{10} \hat{S}_j(f), \quad (8)$$

where p_1 equals 1 if the considered station is located in region 1 and otherwise zero and p_2 is defined the opposite way. $A_1(f, r)$ is discretized into N_{D1} while $A_2(f, r)$ is discretized into N_{D2} equally sized bins (10 km in both regions), with $N_{D2} < N_{D1}$ in the case analyzed here (as the maximum hypocentral distance for region 2 is approximately 180 km). For both attenuation functions, the reference distance is set to $r_0 = 90$ km. Regarding the constraints, we impose $A_1(f, r_0) = 1$ whereas the origin [i.e., the value of $A_2(f, r_0)$] of $A_2(f, r)$ is free. Constraining $A_2(f, r_0) = 1$ and leaving the origin of $A_1(f, r)$ free did not modify the results discussed later in the article in any way. These constraints allow accounting for the fact that not only the decay with distance, but also the general level of the spectral amplitudes may be different between the two regions, which, as shown previously, is clearly the case for Vrancea earthquakes. In other words, the shift of the origin between the two attenuation functions reflects the (cumulative) difference in attenuation over the distance r_0 for both sets of travel paths defined by regions 1 and 2. Both attenuation functions are, of course, also constrained to be smooth functions of distance.

The system matrix **A** as given in equation (4) then takes the following form:

$$\left(\underbrace{\begin{matrix} 1 & 0 & 0 & \cdot & \cdots \\ 0 & 0 & 0 & \cdot & \cdots \\ . & . & . & . & \cdots \\ 0 & 1 & 0 & \cdot & \cdots \\ \vdots & \vdots & \vdots & \vdots & \vdots \\ w_1 & 0 & 0 & \cdot & \cdots \\ -w_2/2 & w_2 & -w_2/2 & 0 & \cdots \\ 0 & -w_2/2 & w_2 & -w_2/2 & \cdots \\ . & . & . & . & \cdots \\ 0 & 0 & 0 & 0 & \cdots \\ 0 & 0 & 0 & 0 & \cdots \\ . & . & . & . & \cdots \end{matrix}}_{\text{attenuation parameters region 1}} \quad \underbrace{\begin{matrix} 0 & 0 & 0 & \cdot & \cdots \\ 1 & 0 & 0 & \cdot & \cdots \\ . & . & . & . & \cdots \\ 0 & 0 & 0 & \cdot & \cdots \\ \vdots & \vdots & \vdots & \vdots & \vdots \\ 0 & 0 & 0 & \cdot & \cdots \\ 0 & 0 & 0 & \cdot & \cdots \\ 0 & 0 & 0 & \cdot & \cdots \\ . & . & . & . & \cdots \\ -w_2/2 & w_2 & -w_2/2 & 0 & \cdots \\ 0 & -w_2/2 & w_2 & -w_2/2 & \cdots \\ . & . & . & . & \cdots \end{matrix}}_{\text{attenuation parameter region 2}} \quad \underbrace{\begin{matrix} 1 & 0 & 0 & \cdots \\ 1 & 0 & 0 & \cdots \\ . & . & . & \cdots \\ 0 & 1 & 0 & \cdots \\ \vdots & \vdots & \vdots & \vdots \\ . & . & . & \cdots \\ . & . & . & \cdots \\ . & . & . & \cdots \\ . & . & . & \cdots \\ . & . & . & \cdots \\ . & . & . & \cdots \\ . & . & . & \cdots \end{matrix}}_{\text{source}}\right), \tag{9}$$

and the new solution vector contains model parameters for both attenuation functions:

$$\mathbf{x} = (\underbrace{\log_{10} A_{1,1} \quad \ldots \quad \log_{10} A_{1,N_{D1}}}_{\text{attenuation parameters region 1}} \quad \underbrace{\log_{10} A_{2,1} \quad \ldots \quad \log_{10} A_{2,N_{D2}}}_{\text{attenuation parameter region 2}} \quad \underbrace{\log_{10} \hat{S}_1 \quad \ldots \quad \log_{10} \hat{S}_{N_E}}_{\text{source}})^T. \tag{10}$$

As for the original scheme, the source scaling factors $\log_{10} \hat{S}_j(f)$ include in this case the (logarithmic) average of the site amplification of the stations in region 1 that recorded the event, as the origin of $A_1(f, r)$ is fixed. Suppose that all earthquakes are recorded by all stations in region 1 and all stations in region 2 (which is, of course, not the case in reality). If now the average site amplification of the stations in region 1 differs from the one in region 2, this difference is also mapped into the offset of the two attenuation functions. However, as each event is recorded by a different combination of stations in region 1 and in region 2, the relation of the average site amplifications between the two regions is different for each earthquake, and thus, the effect will most likely average out. Therefore, it is a reasonable assumption to suppose that the shift of the origin of $A_2(f, r)$ relative to $A_1(f, r)$ is really entirely determined by the difference in attenuation along the paths.

This adapted inversion scheme has been applied to the synthetic dataset described previously in order to assess its performance. The different Q-models given in Table 1 for the two regions (see Figure 1) and especially also their different frequency dependences could be satisfactorily retrieved,

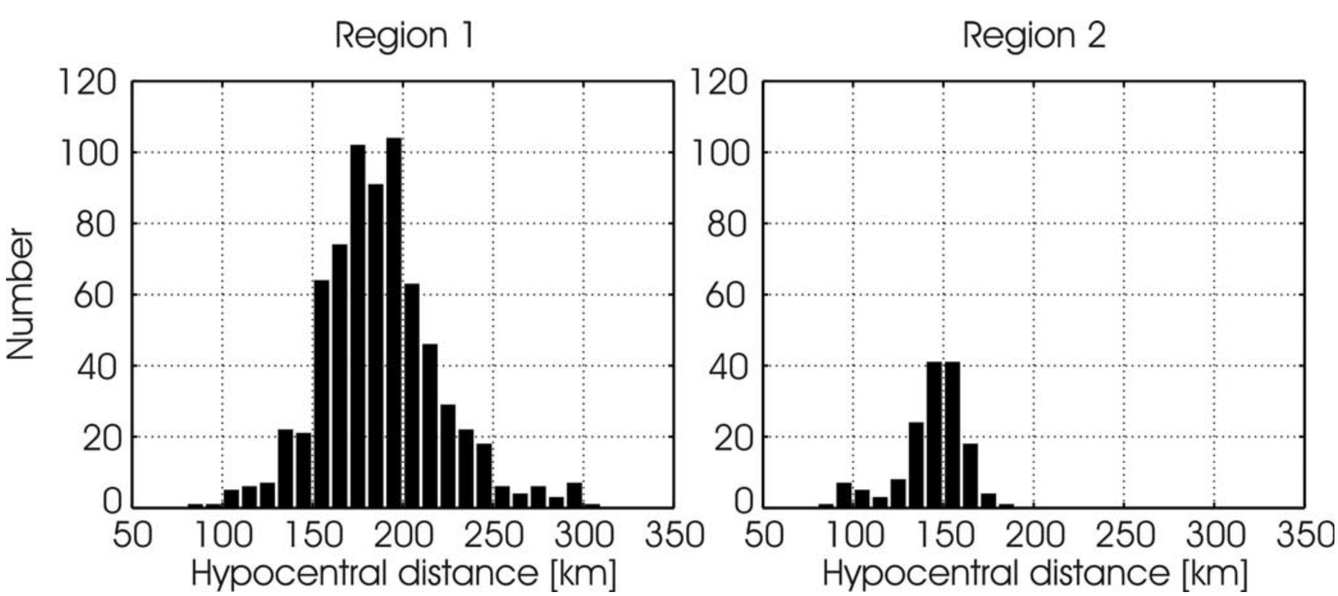

Figure 8. Example of the distribution of hypocentral distances in the dataset at one selected frequency (same as in Fig. 4) separated in region 1 (left) and 2 (right).

which shows that this scheme is indeed well suited to unravel lateral heterogeneities in attenuation characteristics between different sets of stations.

Results and Discussion

The attenuation functions resulting from the application of the modified scheme to the real data are shown for four selected frequencies in Figure 9. At low frequencies (upper panels), the origin of $A_2(f, r)$ is almost identical to the one of $A_1(f, r)$ (the latter one being constrained) and the slope of $A_2(f, r)$ is slightly larger. As the frequency increases (bottom panels), the origin of $A_2(f, r)$ is progressively shifted downward, with a maximum difference of about one order of magnitude at 15.51 Hz. The slopes of the functions are, however, very similar.

Note that both attenuation functions are now monotonically decaying, in contrast to the results obtained by fitting a single attenuation function to the entire dataset (Fig. 5). For each of the two attenuation functions, $Q(f)$ can be evaluated by fitting a parametric model to $A(f, r)$:

$$A(f, r) = G(r) \exp\left(-\frac{\pi f r}{Q(f) v_S}\right), \tag{11}$$

where $G(r)$ is the geometrical spreading function. Because of the large depth of Vrancea earthquakes, surface waves are expected to play only a minor role in the geometrical spreading term for the frequency range of interest. Therefore, the geometrical spreading function is chosen as $G(r) = r_0/r$, with r_0 being the reference distance of 90 km. Taking the logarithm of equation (11) and rearranging leads to:

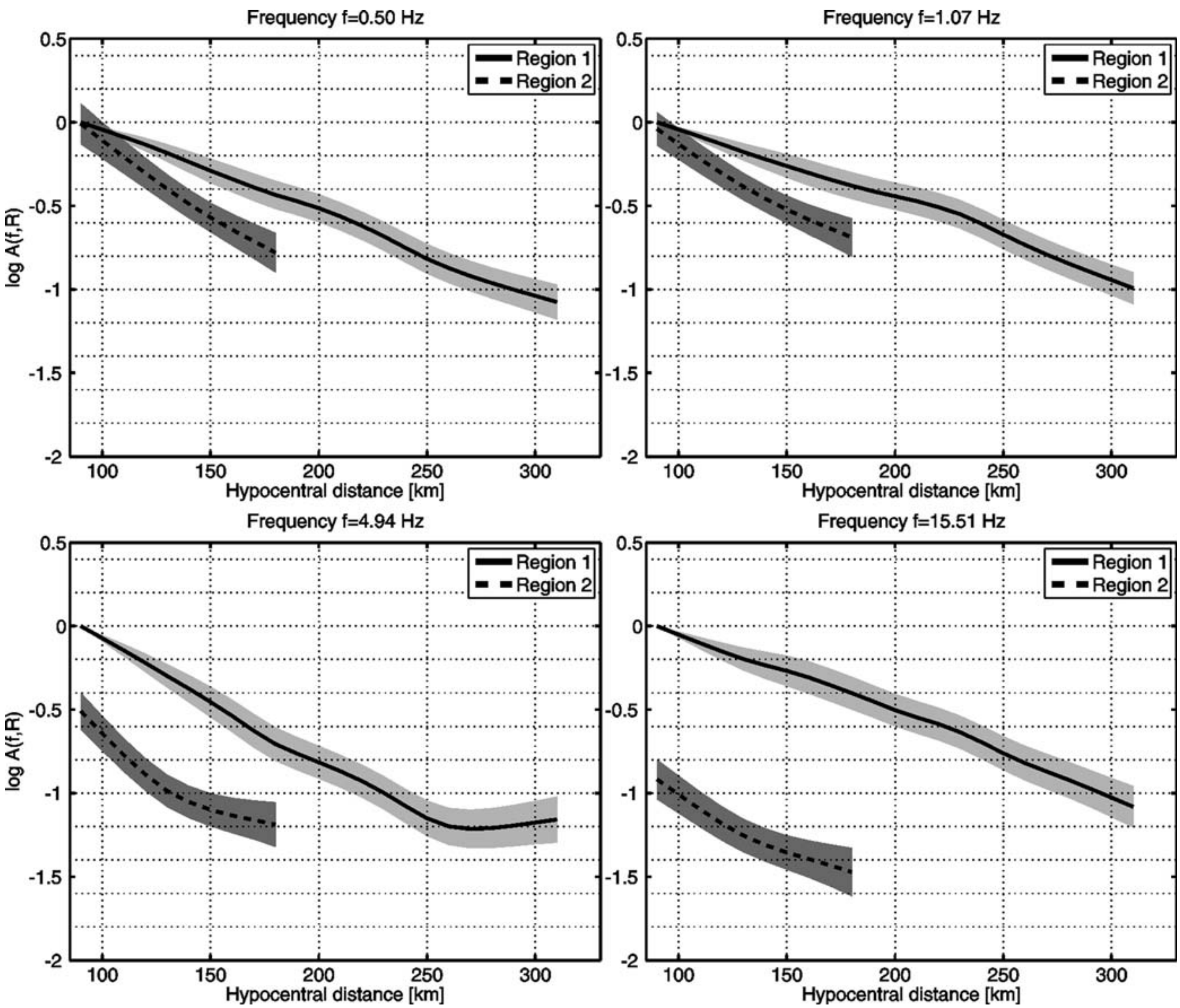

Figure 9. Attenuation functions (at four selected frequencies) obtained by fitting two separate functions for region 1 (continuous line) and 2 (dashed line) to the dataset at each frequency point using the modified inversion scheme. Gray shaded area: mean ± one standard deviation.

$$\log_{10} A(f,r) - \log_{10} G(r) = -\frac{\pi f}{2.3 Q(f) v_S} r. \quad (12)$$

The factor 2.3 in the denominator comes from $\log_{10}(e)$, which does not appear if the natural logarithm is used. Thus, by correcting the attenuation functions for geometrical spreading and plotting versus distance, $Q(f)$ can be evaluated from the slope of a linear least-squares fit. An example for this procedure is shown in Figure 10 (top).

With this procedure, however, only the slope of the attenuation function can be accounted for. Therefore, before fitting equation (12) to $A_2(f,r)$, we normalize $A_2(f,r)$ to equal 1 at $r = r_0$. The offset between the origins of the two functions [i.e., the difference between $A_1(f, r_0)$ and $A_2(f, r_0)$] corresponds to a cumulative difference in attenuation over $r_0 = 90$ km.

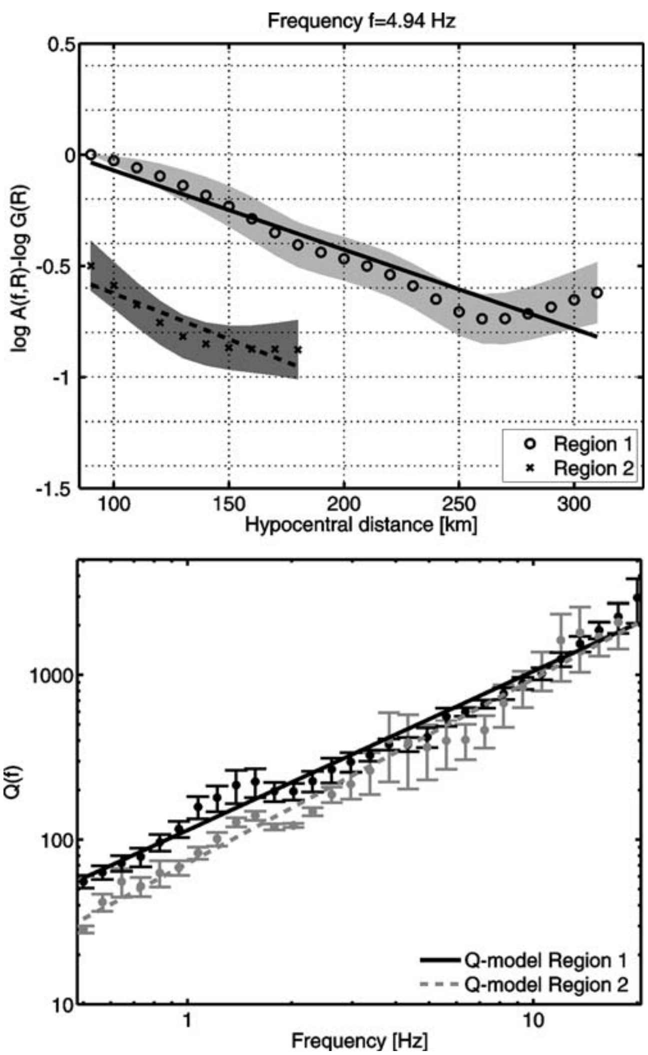

Figure 10. Top: attenuation functions corrected for geometrical spreading [$\log A(f,r) - \log G(r)$ versus r] and fitted straight line (least-squares fit) at one frequency point. Both functions (i.e., for region 1 and 2) are normalized to zero (in logarithm) at the reference distance r_0, but offset with 0.5 in the figure for displaying purposes. Bottom: derived $Q(f)$-models for region 1 and 2 by evaluating the slope of the fitted lines at each frequency.

After $Q(f)$ has been determined at all 30 frequencies for both regions, we fit a model of the form $Q(f) = Q_0 f^N$ to the determined values (Fig. 10, bottom, the vertical error bars indicate the regressional error of fit from the linear regression performed at each frequency as described previously) and obtain:

$$Q(f) = 114 f^{0.96} \quad \text{for region 1}, \qquad Q(f) = 72 f^{1.12} \quad \text{for region 2}. \quad (13)$$

Hence, $Q(f)$ follows a roughly linear trend with frequency in both regions. Surprisingly at first sight, these two $Q(f)$-models are almost identical. Remember, however, that these models are derived from the slope of the attenuation functions alone, not including the offset at the reference distance. A physical interpretation for this offset is provided in the following.

Several other researchers already focused their attention on the seismic attenuation in the Carpathians and surroundings. Based on a qualitative discussion of a small set of seismograms, Popa *et al.* (2005) and Radulian *et al.* (2006) presented indications that strong lateral variability of upper-mantle attenuation characteristics must exist in order to be able to explain the large amplitude differences within the Carpathians and the foreland at frequencies higher than 1 Hz. Russo *et al.* (2005), with data from the same network as used here, computed the ratios between the spectra of *S* and *P* waves to derive differential Q_S estimates. They obtained average Q_S-values around 100–250 for stations within the Carpathian arc and up to 1000 for stations in the foreland, reaching thus similar conclusions as Popa *et al.* (2005) and Radulian *et al.* (2006). Their analysis technique, however, makes strong assumptions both regarding the source (identical source spectra for *P* and *S* waves) and site (equality of site amplification on the horizontal component of *S* and vertical component of *P* waves) contributions to the ground-motion spectra. Sudhaus and Ritter (2005) and Weidle *et al.* (2007) used teleseismic data to derive t^* measurements at low frequencies (< 1–1.5 Hz) and also found significant lateral variations of attenuation.

In general, the results of the aforementioned studies are in good agreement with those presented in this article. This study represents the first systematic analysis of the frequency dependence of seismic attenuation beneath Vrancea and its lateral variations. At low frequencies, the difference between the two regions is not extremely striking in our study. There is no offset between the origins of the two attenuation functions and the slope of the function in region 2 is somewhat larger. This observation is consistent with the results of Weidle *et al.* (2007), who found the largest t^*-values (i.e., the largest attenuation) at stations located in the epicentral zone of Vrancea earthquakes. At high frequencies, the variation in attenuation between region 1 and 2 becomes larger and larger, quantified through the offset of the attenuation functions origins. The slopes of the functions, however, are quite similar, leading to the determination of roughly identical $Q(f)$-models.

An attempt to explain this observation is illustrated in Figure 11. As can be seen from the distribution of the earthquake hypocenters (Fig. 1), they are all located approximately directly beneath the stations in region 2. Thus, in order to increase the distance in $A_2(f, r)$, the source depth has to be increased. The difference in travel path for two data points at different distances in region 2 are hence related to the path traveled from the deeper event's hypocenter up to the one of the shallower one. As schematically indicated in Figure 11, within the gray shaded area, all the rays travel more or less the same path. In conclusion, $A_2(f, r)$ only samples the lower part of the travel path, whereas the large offset observed between $A_1(f, r_0)$ and $A_2(f, r_0)$ is related to some strongly attenuating region somewhere in the gray shaded area, approximately between the shallowest event's hypocenter and the surface.

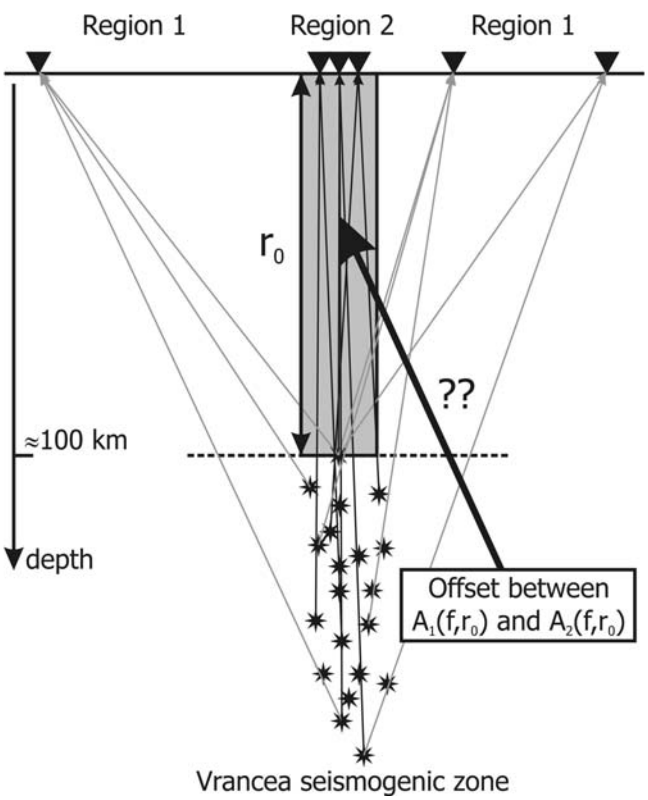

Figure 11. Sketch depicting the ray paths from sources to stations (orientation roughly soutwhest–northeast). For the stations in region 2, the rays from different earthquakes all travel the same path through the gray shaded region, which is defined by the depth of the shallowest event in the dataset. Increasing the distance r in the attenuation function $A_2(f, r)$ means to look mainly at data from deeper earthquakes and therefore, $A_2(f, r)$ only samples the hypocentral zone. The offset between the origins of the two attenuation functions, $A_1(f, r_0)$ and $A_2(f, r_0)$, is related to a process occurring somewhere in the gray shaded zone.

The seismic tomography results presented by Martin *et al.* (2006) and the outcome of the seismic refraction studies (Hauser *et al.*, 2001, 2007) provide indications on the reasons for these strong lateral variations in seismic attenuation. The seismic refraction data (Hauser *et al.*, 2001) suggest the presence of a low-velocity zone at a depth of 47–55 km beneath the Vrancea region, which, as Sperner *et al.* (2005) note, coincides quite well with the observed seismic gap between 40 and 70 km depth already discussed by Fuchs *et al.* (1979). Here, the slab seems to be mechanically decoupled (or only weakly coupled) from the crust through a weak zone. This zone is commonly interpreted as the place where slab detachment currently takes place. This area of weak coupling could be an explanation for the lower spectral amplitudes in the vicinity of the epicentral area (region 2) and the offset between the attenuation functions for the two regions. Russo *et al.* (2005) invoke the existence of thermal anomalies influencing the travel paths to the stations SIR and MLR in view of recent volcanism at several sites near these stations. Ismail-Zadeh *et al.* (2005), following the work of Demetrescu and Andreescu (1994), show that indeed very high temperatures might persist in the subcrustal mantle (at a depth of around 50 km) beneath the mountain arc.

Beneath the Transylvanian basin (i.e., behind the mountain arc), the tomography (Martin *et al.*, 2006) clearly reveals a very strong low-velocity anomaly in the depth range 35–110 km, which is interpreted as the result of slab rollback and the delamination of continental lithosphere. Additionally, Martin *et al.* (2006) invoke the possible presence of partial melts at depth of approximately 100 km. The strong contrast between the high-velocity body (i.e., the slab with low attenuation), which influences the waves traveling to stations in the foreland (see tomography results, Martin *et al.*, 2006) and the very strong low-velocity anomaly behind the mountain range is proposed by Popa *et al.* (2005) to explain the observed variations in attenuation, especially in the Transylvanian basin. We can neither confirm nor reject this interpretation regarding sites behind the mountain arc with this dataset, as no such location is included in the database studied (station OZU is the only site that is close to the Transylvanian basin, although still in the mountain arc).

Yet an interesting observation in that direction is the fact that no single recording from stations DRG (46.79° N, 22.71° E) and MED (46.14° N, 24.37° E), located farthest to the northwest from the Vrancea region (thus far behind the mountain arc), was usable for this analysis, hinting for strong signal attenuation. Although for smaller events the installation at noisy sites and the large hypocentral distances might provide an explanation for this evidence, they cannot provide a convincing explanation when also for the strongest events no clear S-wave signal can be detected. For instance, the recordings at DRG and MED (due to their very low amplitudes, these two stations would have been classified into region 2 if they would have been usable for the analysis) of the 27 October 2004 (M_w 5.8) event show a clear P onset, but no S wave is detectable, neither on the horizontal nor on the vertical components, and the maximum amplitude is approximately 0.2 cm/sec^2 (hypocentral distance $\approx$340 km) for DRG and 2.5 cm/sec^2 for MED (hypocentral distance $\approx$215 km). At station ZIM (in region 1, at hypocentral distance $\approx$280 km), the *S* wave is very clear, with a maximum amplitude of around 30 cm/sec^2, which is more than one order of magnitude larger than at MED (which is even at closer distance to the source) and more than two orders of magnitude larger than at DRG. A very strong attenuation for travel paths behind the mountain arc might explain this peculiarity. Thus, the observed inhomogeneity in attenuation is rather explainable by strong upper-mantle heterogeneity than crustal features. These variations in seismic attenuation between the foreland (region 1) on one hand and the mountain range (region 2) and backarc area on the other also have important implications for the seismic hazard of Romania. As Popa *et al.* (2005) also note, the difference in attenuation is most relevant at high frequencies. This view is clearly confirmed by this study with a much larger dataset, and the main advantage of the results presented here is that the variation in attenuation has been quantitatively analyzed by means of an inversion scheme. At low frequencies, the spectral amplitudes of ground motion have similar values in regions 1 and 2. Thus, for structures sensitive to low frequencies, such as high-rise buildings or long-span bridges, the risk is similar in the mountain range (and behind) and in the foreland, depending, of course, on the distance from the source region. At high frequencies, this situation changes drastically. Within and behind the mountain arc, the risk, for example, for buildings with roughly less than three floors seems to be much smaller as compared with the foreland, as the high-frequency energy is efficiently attenuated for the former region.

In view of the intensity maps from Vrancea earthquakes (the intensity maps of the large 1977 and 1986 events are shown in Oth *et al.* [2007], Fig. 7), the observed variations in attenuation might be the key to understand the overall shape of the isoseismals. Apart from the small-scale structure within the high-intensity (VII and VIII) areas (which are discussed in terms of source physics in Oth *et al.* [2007]), it is striking that the isoseismals separating the high and low intensities almost exactly follow the shape of the mountain arc. Within the Carpathians as well as behind them, there are only few buildings sensitive to low frequencies (as Popa *et al.*, 2005, note, the largest amount of tall buildings is found in Bucharest), and the structures sensitive to higher frequencies are less affected due to the strong attenuation of high-frequency seismic waves. Thus, the overall shape of the macroseismic intensity pattern can be explained by the combination of a less endangered inventory of buildings (with respect to low frequencies) and strong attenuation of high-frequency energy within and behind the mountain arc as compared with the foreland.

It should be emphasized that the attenuation functions $A(f, r)$ contain all effects leading to seismic-wave attenuation and that we cannot distinguish between the contributions of anelastic and scattering attenuation. As Furumura and Kennett (2005) show, internal heterogeneities within the subducted plate and particularly their shape play an important role in guiding high-frequency seismic signals. Weidle *et al.* (2007) show from anelastic attenuation modeling that it is not possible to explain their t^* observations from teleseismic earthquakes solely by anelasticity mechanisms. An especially strong indication for the influence of scattering effects in the attenuation of seismic waves is its strong frequency dependence (Furumura and Kennett, 2005). However, we cannot definitely conclude on these issues and further investigations are required in order to clarify the role of scattering attenuation as compared with intrinsic attenuation.

The results for seismic attenuation presented in this article, together with the site amplification functions and source spectra discussed in Oth (2007) and Oth *et al.* (2009), can also be used in the computation of ground-motion simulations, for instance, using the stochastic technique (Boore, 1983, 2003), and hence are directly usable for seismic hazard assessment studies in the area. In such simulations, attenuation can be parameterized by using the *Q*-models derived for regions 1 and 2 and a set of tabulated values for the offset between the attenuation functions over the frequency range of interest.

Intermediate-depth seismicity in continental lithosphere is not a frequent phenomenon, but it occurs apart from the Vrancea region also in the Hindu Kush, the Pamirs, and the Bucaramanga area (e.g., Frohlich *et al.*, 1995). If the tectonic settings of these regions are similar to the Vrancea area, the available attenuation and source scaling relations should be reviewed. We discussed potential reasons for the peculiar attenuation characteristics in the Vrancea region and attribute them to strong lateral variations of *Q* in the upper mantle, which in turn can be associated with an ongoing slab detachment process. From this perspective, it might be worthwhile to reexamine attenuation and source scaling relations in areas where this process is believed to occur (Wortel and Spakman, 2000).

Conclusions

Seismic attenuation in the Carpathians and the surrounding areas depicts strong lateral variations, with very strong attenuation in the epicentral area (and, based on the findings of other authors mentioned previously, probably also behind the mountain arc in the Transylvanian basin) and much lower attenuation in the foreland. Moreover, the variability strongly increases with increasing frequency. From synthetic data tests, we were able to show that the shape of the attenuation functions obtained applying the standard GIT approach can be well explained by such lateral, frequency-dependent variability of attenuation characteristics between the bending zone of the Carpathian arc and the foreland.

With a modification of the inversion scheme, it is possible to simultaneously invert for the attenuation properties of two (or more) regions in one single inversion, allowing us to quantify the lateral inhomogeneity at each analyzed frequency. At high frequencies (> 4–5 Hz), there is approximately one order of magnitude difference in attenuation between the mountain range and the foreland, whereas at lower frequencies, the attenuation characteristics are rather similar. Possible physical explanations for these results involve the degree of coupling between the slab and the overlying crust (Sperner *et al.*, 2005), strong temperature variations (Ismail-Zadeh *et al.*, 2005), as well as scattering phenomena within the subducted lithosphere (Furumura and Kennett, 2005).

Previous attempts of hazard assessment focused on the azimuthal dependence of ground motion (Lungu *et al.*, 1999; Sokolov *et al.*, 2004). This study demonstrates that the radial variation is equally important: it allows us to understand the intensity distribution of previous strong earthquakes and should be incorporated in future hazard assessment.

Data and Resources

The seismograms used in this study were collected by instruments of the strong-motion network deployed in Romania since 1997 in the framework of the Collaborative Research Center (CRC) 461 "Strong Earthquakes: A Challenge for Geosciences and Civil Engineering" of the University of Karlsruhe in collaboration with the National Institute for Earth Physics (NIEP) in Bucharest. The recordings from the three large Vrancea events that occurred in 1986 and 1990 were obtained from an analog SMA-1 network and digitized at NIEP. The data may be obtained from the authors upon request. The hypocentral coordinates of the analyzed events have been taken from the ROMPLUS catalog (Oncescu, Marza, *et al.*, 1999), which is continuously updated as new earthquakes occur and is available at http://www.infp.ro/catal.php.

Acknowledgments

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