S-waves attenuation and separation of scattering and intrinsic absorption of seismic energy in southeastern Sicily (Italy)

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INTRODUCTION

Modern assessment of seismic hazard and information on earthquake source parameters in a given region requires a good knowledge of attenuation and properties of the medium. Especially, information on high-frequency seismic wave attenuation in the lithosphere is of particular interest (Yoshimoto et al. 1993). In southeastern Sicily, the attenuation of seismic waves has been studied by using different data set and methods (e.g. Castro et al. 1993; Giampiccolo et al. 2002, 2003, 2004; de Lorenzo et al. 2004). Castro et al. (1993) first obtained a frequency-dependent quality factor of coda waves ($Q_c$) and calculated the average attenuation law for the area. Giampiccolo et al. (2002, 2004) investigated in more detail both the frequency and lapse time dependence of $Q_c$ by using different data sets. Their results show a good agreement with those obtained by Castro et al. (1993). Moreover, a clear increase of $Q_c$ with lapse time and, therefore, with depth was observed. Giampiccolo et al. (2003) and de Lorenzo et al. (2004) estimated the quality factor of P-waves ($Q_p$) by applying the pulse-broadening method (Wu & Lees 1996) whereas Giampiccolo et al. (2003) estimated the attenuation of S waves ($Q_{s1}$) by applying the frequency decay method (see Bianco et al. 1999) in the low- (below the corner frequency) and high- (above the corner frequency and below the cut-off filter) frequency ranges. The obtained results suggested that attenuation at higher frequencies is less pronounced than at lower ones. However, the detailed frequency-dependent nature of $Q_s$ and $Q_c$ was not resolved yet by the above quoted studies.

It is worth stressing that attenuation estimated from direct S waves contains the combined effects of scattering and intrinsic loss. Scattering attenuation is described by the quality factor $Q_{s1}$ and is due to the presence of inhomogeneities. Therefore, it depends on the spatial structure of the heterogeneities in the medium and on the size of the velocity and density fluctuations. Intrinsic absorption $Q_{s1}$ is caused by the anelasticity of the medium and depends on viscous processes between the rock matrix and liquid inclusions, such as pore fluids, and on movements of dislocations through the mineral grains (Goric & Muller 1987). Quantifying the relative contribution of scattering and intrinsic attenuation has been a subject of considerable interest among seismologists and different methods have been developed (e.g. Wu 1985; Hoshiba et al. 1991; Wennerberg 1993).

In the present paper we will estimate the quality factor of S waves ($Q_s$) in the lithosphere beneath southeastern Sicily and clarify its frequency dependence by means of the coda-normalization method (Aki 1980), applied in the frequency range 1.5–15 Hz. We will also attempt by applying the multiple lapse time window analysis (MLTWA), under the hypothesis of multiple isotropic scattering with uniform distribution of scatterers. Intrinsic absorption dominates over scattering in the attenuation process at high frequencies (above 3 Hz). Below 3 Hz scattering is the predominant attenuation effect in the region, at the scale length of these frequencies. However, some discrepancies have been observed between the theoretical model and the observations. This indicates that models with depth-dependent velocity structure and/or non isotropic scattering should be taken into account in order to remove ambiguities in the interpretation of the results.

SUMMARY

Seismic wave attenuation in southeastern Sicily was investigated by using a data set of about 180 local earthquakes ($1.5 \leq M_L \leq 4.6$) recorded in the period 1994–2003. We first estimated the quality factor of S waves ($Q_{s1}$) and clarified its frequency dependence by means of the coda-normalization method, applied in the frequency range 1.5–15 Hz. The average $Q_s$ as function of frequency is given by $Q_s = 49 f^{0.88}$. A detailed separation of scattering attenuation ($Q_{s1}^{-1}$) from intrinsic absorption ($Q_{s1}^{-1}$) was also attempted by applying the multiple lapse time window analysis (MLTWA), under the hypothesis of multiple isotropic scattering with uniform distribution of scatterers. Intrinsic absorption dominates over scattering in the attenuation process at high frequencies (above 3 Hz). Below 3 Hz scattering is the predominant attenuation effect in the region, at the scale length of these frequencies. However, some discrepancies have been observed between the theoretical model and the observations. This indicates that models with depth-dependent velocity structure and/or non isotropic scattering should be taken into account in order to remove ambiguities in the interpretation of the results.

Key words: absorption, attenuation, coda-normalization method, multiple lapse time window analysis (MLTWA), scattering, southeastern Sicily.
Under the assumptions of multiple and isotropic scattering and uniform distribution of scatterers, two attenuation parameters will be calculated: the seismic albedo \( B_0 \), defined as the dimensionless ratio of the scattering loss to total attenuation \( (B_0 = Q^{-1}_{sc}/Q^{-1}_T) \) and the inverse of the extinction length \( L_e \), that is the inverse of the distance (in kilometres) over which the primary S-wave energy is decreased by \( e^{-1} \). \( B_0 \) ranges between 0 and 1 and was proposed by Wu (1985) to describe the proportions of energy loss dominated by intrinsic attenuation \( (B_0 < 0.5) \) or scattering attenuation \( (B_0 > 0.5) \). The estimated scattering \( (Q_{s}^{-1}) \) and intrinsic \( (Q_{i}^{-1}) \) attenuation mechanisms in the frequency range 1.5–15 Hz will be discussed and compared with previous results obtained by Giampiccolo et al. (2004). Estimates of total attenuation \( Q_T^{-1} \) will be compared with the coda-Q values \( (Q_{Coh}) \) obtained by Giampiccolo et al. (2004) with the expected coda-Q \( (Q_{Cexp}) \) calculated in this study by using Hoshiba (1991) relationship. Finally, since the MLTWA technique has been widely applied to several areas in the world (e.g. Mayeda et al. 1992; Hoshiba 1993; Akinci et al. 1995; Pujades et al. 1997; Akinci & Eydo˘gan 2000; Bianco et al. 2002; Ugalde et al. 2002; Bianco et al. 2005), differences and analogies observed among southeastern Sicily and other tectonic settings will be discussed.

**TECTONIC SETTING AND DATA**

Southeastern Sicily is one of the Italian regions with the highest seismic hazard. The main structural domain is represented by the Hyblean Foreland (Fig. 1) that is the northern margin of the African continental crust, which was weakly deformed during alpine orogenesis and continues to exhibit moderate uplift and overall extensional tectonics (Adam et al. 2000). It is represented by a NE–SW trending horst characterized by a sequence of more than 6 km of Meso–Cenozoic carbonate rocks (Bianchi et al. 1987) and thinner marls levels with intercalated Plio–Pleistocene mafic volcanics. The tectonic setting of this area is mainly characterized by a series of NW–SE and NNW–SSW oriented horst and graben, linked to the Hybleo–Maltese escarpment and connected with the lower Pleistocene alternating tectonic movements of uplift and lowering (Bianca et al. 1999). The Hybleo–Maltese escarpment is a NNW–SSE striking normal fault, which extends for more than 300 km from North Africa to Eastern Sicily and represents the morphological evidence of a lithospheric fault zone separating the 23-km-thick continental crust of the African platform from the 13-km-thick oceanic crust of Ionian Sea domain (Scandone et al. 1981; Reuther et al. 1993; Adam et al. 2000). Secondary faults (e.g. the Avola fault, the Rosolini-Pozzallo System, the Scordia-Lentini Graben) strike along an overall NE–SW and NNE–SSW direction and represent the southern branch of the Siculo-Calabrian Rift (Tortorici et al. 1995; Monaco et al. 1997), an active rift zone extending from southern Italy to the Ionian shore of Sicily for a total length of about 370 km.

Several strong earthquakes occurred in southeastern Sicily during the last 1000 yr (e.g. 1169 February 4 and 1693 January 11 events). The macroseismic magnitude for these events was estimated at about \( M = 7 \) or higher (Boschi et al. 1997). Four other earthquakes have

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**Figure 1.** Principal structural elements in Southeastern Sicily (after Giampiccolo et al. 2003). The star indicates the epicentral location of the \( M_L 5.4 \) earthquake occurred on 1990 December 13.
exceeded magnitude 5.8: the 1125 July 7, the 1542 December 10, the 1693 January 9 and the 1818 February 20 events (Azzaro & Barbano 2000). The strongest ($M_L = 5.4$) recent earthquake occurred on 1990 December 13 (Fig. 1), about 10 km offshore (Amato et al. 1995). Recently, Musumeci et al. (2005) have evidenced that the northern, the eastern and the western margins of the Hyblean Foreland are the zones of major seismic energy release. They also showed that the present-day direction of maximum horizontal compression in the region is trending NNW–SSE to NW–SE. These results suggest that seismic activity is associated with the still active Africa–Eurasia plate convergence, as well as with other local stress factors.

The Southeastern Sicily Seismic Network (hereafter SESSN; Fig. 2), run by the Istituto Nazionale di Geofisica e Vulcanologia, has been operating since 1994. It consists of nine digital three-component stations, each equipped with short period Mark L4–3D seismometers having a natural frequency of 1.5 Hz and a damping of ca. 60 per cent of critical. The data are sampled with a sampling frequency of 125 Hz and the instrumental response of the whole data acquisition system is almost flat in the range 1.5–51 Hz. The stations are deployed on different kinds of geologic outcrops. Stations SR5, SR6, SR7 and SR8 are located on Meso–Cenozoic carbonate deposits; stations SR3, SR4 and SR9 are located on volcanoclastic deposits; station SR1 is located on marly clays and sands belonging to Quaternary deposits. Stations SR2, SR4 and SR3 are deployed on different kinds of geologic outcrops. Stations SR5, SR6, SR7 and SR8 are located on Meso–Cenozoic carbonate deposits; stations SR3, SR4 and SR9 are located on volcanoclastic rocks; station SR1 is located on marly clays and sands belonging to the Gela-Catania foredeep domain; station SR2 is located on recent Quaternary deposits.

The data set used in this study consists of about 180 local events with magnitudes $M_L$ ranging between 1.5 and 4.6, recorded in the period 1994–2003. The local magnitude $M_L$ is routinely computed as reported in Di Grazia et al. (2001). The earthquake locations (Fig. 2) have been selected from the SESSN database and are affected by uncertainty of about 2 km, both in epicentral coordinates and focal depths. The area covered by the earthquake data set encompasses the eastern coastal range of the Hyblean Plateau and parts of the adjacent offshore. Focal depths of the events mainly range from 3 to 30 km and hypocentral distances are less than 100 km.

**The MLTWA Technique**

The basis of the MLTWA technique (Hoshiba et al. 1991), is to measure the seismic wave energy as a function of distance and frequency for three consecutive time windows defined on the seismogram of a local event, starting from the S-wave onset. The window length is chosen in such a way that the first window contains a significant contribution of the direct S-wave energy and the last two windows mainly contain the contribution of the scattered energy.

The integrated seismic energy for the three time windows is calculated by measuring the $\text{rms}$ amplitudes over bandpass filtered seismograms. Each integral is normalized according to the coda-normalization method (Aki 1980), to correct for different sources and site amplifications:

$$E_i(f, r_w) = \left[ \frac{\langle c_h(f, r_w) \rangle}{\langle c_{\text{coda}}(f, t_{\text{ref}}) \rangle} \right]$$

where $\langle c_{\text{coda}}(f, t_{\text{ref}}) \rangle$ is the energy of the coda observed at the reference time $t_{\text{ref}}$ selected at a lapse time at least twice the S-wave traveltime, $c_h(f, r_w)$ is the $\text{rms}$ amplitude calculated as a function of frequency $f$ and distance $r_w$, integrated for three consecutive time windows and $E_i(f, r_w)$ represent the normalized observed seismic energies for the central frequency $f$ and the $i$th time window. (e.g. Hoshiba 1993). Finally, a correction for the geometrical spreading is performed by multiplying each integrated energy by the factor $4\pi r^2$. A comparison between observed integrated energies and theoretical curves as a function of the hypocentral distance, leads to a least-squares estimate of $B_0$ and $L^{-1}_c$. Under the assumption of multiple isotropic scattering and uniform distribution of scatterers, the theoretical curves of the energy density at a given lapse time and hypocentral distance can be calculated by using the equation described by Zeng et al. (1991) and Zeng (1991):

$$E(r, t) \approx E_0 e^{-t / T} \left[ \frac{\delta_r}{4\pi V_s^2 t} + \frac{H(t - \frac{t}{T})}{4\pi V_s^2 t} \ln \left( 1 + \frac{t}{T} \right) - \frac{1}{4\pi V_s^2 t} \right]$$

$$+ cH \left( \frac{r}{V_s} \right)^2 \left[ \frac{3\eta_s}{4\pi V_s^2 t} - \frac{\eta_s t}{V_s^2} \right] e^{-t / T},$$

where $E_0$ is the energy released by the earthquake, $t$ is the time from the origin of the event, $H(t)$ is the Heaviside function, $\delta_r$ is the Kronecker delta, $V_s$ is the shear wave velocity, $\ln$ is the natural logarithm, $\eta_s$ is the attenuation coefficient, $\eta_c$ is the coda attenuation coefficient, $c$ is the speed of sound, $V_s$ is the shear wave velocity, $T$ is the surface wave time, $t$ is the time from the origin of the event, $r$ is the hypocentral distance, $H(t)$ is the Heaviside function, $\delta_r$ is the Kronecker delta, and $\ln$ is the natural logarithm.

**The CODA-Normalization Method**

This method (Aki 1980) is based on the assumption that, for weak scattering media, the distribution of coda wave energy at a fixed reference time ($t_c$) is nearly uniform and the seismic energy is uniformly distributed in a volume surrounding the source. It consists in dividing the spectral amplitude of the S waves by that of coda waves measured at the lapse time $t_c$. In this way, it is possible to eliminate source and site effects. Briefly, $Q_s$ is calculated from the slope of the function given in the following equation:

$$\ln \left[ \frac{A_S(f, r)}{A_C(f, t_c)} \right]_{r + \Delta r} = - \frac{\pi f}{Q_s(f)} V_s r + c_s(f).$$

In eq. (1), $A_S$ and $A_C$ are the spectral amplitude of direct S waves and coda waves, respectively, $f$ is the frequency, $r$ is the hypocentral distance, $\gamma$ is the exponent of geometrical spreading and $V_s$ means the S waves velocity. The constant $c_s$ includes the scattering characteristics of the medium and the site amplification effect. Finally, the symbol $\langle \rangle_{r + \Delta r}$ denotes the average within a small hypocentral distance range of $r \pm \Delta r$. By using many S waves from events at different azimuths, the effects of source radiation pattern are considered negligible (Aki 1980). $Q_s^{-1}(f)$ can be estimated from a linear regression of $\ln \left[ \frac{A_S(f, r)}{A_C(f, t_c)} \right]_{r + \Delta r}$ versus $r$, by means of the least-squares method.
Figure 3. Coda-normalized $S$-wave amplitudes versus hypocentral distance for different bands centred at frequency $f$. The value of $Q_S$ obtained by least-squares linear regression and its relative standard deviation is shown in each frame.

with:

$$c = E_0 \left[ 1 - (1 + \eta_S V_S t) e^{-\eta_S V_S t} \right]$$

where $E_0$ is the total incident wave energy at $t = 0$, $H$ is the Heavy-side function, $t$ is the traveltime, $r$ is the hypocentral distance and $\alpha = V_S t / r$. The intrinsic attenuation factor $\eta_I = 2 \pi f V_S Q_I^{-1}$ and the scattering attenuation factor $\eta_S = 2 \pi f V_S Q_S^{-1}$ may be, therefore, expressed in terms of total attenuation $L^{-1} = \eta_I + \eta_S$ and seismic albedo $B_0 = \eta_S / \eta_I + \eta_S$.

**DATA ANALYSIS AND RESULTS**

A total of about 700 waveforms, selected on the basis of a good signal-to-noise ratio at the end of the coda, was considered suitable for the application of the coda-normalization method (Aki 1980) and for the MLTW A analysis (Hoshiba et al. 1991). A time window of 6 s for the $S$ waves and for the coda, at a lapse time of 40 s from the origin time, were used for the estimation of $A_S (f, r)$ and $A_C (f, t_C)$ of eq. (1). The average $S$- to coda amplitude ratios were calculated over six frequency bands centred at 1.5, 3, 6, 9, 12 and 15 Hz, with bandwidths of 0.5 for 1.5 Hz, 1 for 3 Hz and 2 for the remaining frequencies. The calculated $S$ to coda wave amplitude ratios were multiplied for hypocentral distance $r$, considering the exponent of geometrical spreading $\gamma$ as one, since we used body waves. Finally, eq. (1) was applied to our data and $\ln \left( A_S (f, r) / r / A_C (f, t_C) \right)$ versus hypocentral distance $r$ was plotted for each frequency band. $Q_S^{-1}$ was estimated by means of least-squares linear regression, taking into account an average $V_S = 3.3 \text{ km s}^{-1}$ according to the velocity model for the area by Musumeci et al. (2003). Fig. 3 shows the results obtained for each frequency. It is worth stressing that data are narrowly scattered in points around the regression line for all frequency bands. The number of data used for the $Q_S$ computation increases with increasing central frequencies. A lower number of data was suitable for the analysis at low frequencies (1.5 and 3 Hz), probably due to the presence of a
large effect of noise induced by sea waves (e.g. Chung & Sato H. 2001).

\(Q_S\) increases as a function of frequency according to the power law \(Q_S = Q_S f^n\) where \(Q_S\) is the value of \(Q_S\) at 1 Hz and \(n\) is the frequency-dependence coefficient. The average attenuation law for the whole area is \(Q_S = 49 f^{0.88}\) (Fig. 4).

What we found is in good agreement with the results obtained in a previous study carried out in the area by Giampiccolo et al. (2003), who applied the frequency decay method (e.g. Bianco et al. 1999) to a slightly different data set. In particular, from the S-wave spectra, they obtained \(Q_S = 190\) in the low-frequency range (<5 Hz) and \(Q_S = 700\) in the high-frequency range (>16 Hz). As it can be observed in Fig. 4, these values (empty circles) follow the same distribution around the best-fit line of the \(Q_S\) values estimated in the present study.

In order to investigate the possible variation of seismic attenuation with the source-to-receiver path, we estimated \(Q_S\) at each single station. The attenuation laws obtained are reported in Fig. 5 together with the average attenuation law found for the area.

In general, we observe a similar trend among the stations, except at station SR2, for which a stronger frequency dependence \((n = 1.16)\) was found.

A second step of our analysis regarded the application of the MLTWA technique to obtain a detailed separation of scattering and intrinsic attenuation in the area. The analysis was performed for the hypocentral distance range 20–80 km. Each seismogram was band-pass filtered in the frequency bands stated above (1.5, 3, 6, 9, 12 and 15 Hz). Then, the integrated observed energies as a function of distance and frequency were calculated by measuring the \(r_{ms}\) amplitudes for three selected windows, defined at 0–12 s, 12–24 s and 24–36 s, starting from the \(S\)-wave onset. The observed energy was corrected for earthquake sources and station site effects by performing a normalization, as described in eq. (2), for the energy contained in the fixed reference time interval from the origin time \((t_{ref} = 40 \pm 10\ s)\), selected to avoid contamination with direct S waves. The normalized energies were corrected for geometrical spreading by multiplying by \(4\pi r^2\) and compared with theoretical seismic energies. The comparison between observed and theoretical seismic energies at each frequency band and time window is reported in Fig. 6. The first time window (0–12 s) shows a larger data scatter than the later two time windows. This effect has been observed in different areas (e.g. Fehler et al. 1992; Mayeda et al. 1992; Jin et al. 1994; Akinci et al. 1995; Pujades et al. 1997; Akinci & Eydoğan 2000; Bianco et al. 2002) and can be attributed to the simplicity of the model underlying the method. In particular, differences in source radiation pattern among the events, not taken into account by the coda normalization, and/or local structure under the seismic stations, may affect the first time window more strongly than the later ones; in fact, later coda waves sample grater volumes and average over a greater number of scatterers with a higher degree of azimuthal distribution around the source and the receiver (Pujades et al. 1997).

The best-fit \(L^{-1}_e\) and \(B_0\) estimates are also reported at the top of the pictures. The estimation of the best-fit \(L^{-1}_e\) and \(B_0\) was performed by using the misfit function between observed data and model by Hoshiba (1991):

\[
M(B_0, L^{-1}_e) = \sum_{k=1}^{N} \sum_{r} \left( E_t(r) - E_i(r) \right)^2 ,
\]

where \(E_t(r)\) are the energies measured from the experimental data and \(E_i(r)\) are the theoretical ones. The best-fit model parameters are obtained by calculating the minimum residual between the predicted and observed data as a function of \(L^{-1}_e\) and \(B_0\). Optimal solutions for \(L^{-1}_e\) and \(B_0\) are found using a grid search. To estimate the error intervals in \(L^{-1}_e\) and \(B_0\), we used the F-distribution for 40 degrees of freedom at the 60 per cent confidence level (e.g. Bianco et al. 2002, 2005). By plotting the misfit function (eq. 3) normalized by its minimum, we accepted as possible solutions those with value lower than 1.2 (for details on this procedure, see Mayeda et al. 1992).

Fig. 7 shows the residuals of the 6489 pairs for all the analysed frequency bands.

From the best \(B_0\) and \(L^{-1}_e\) values we estimated the quality factors for total \((Q^{-1}_i)\), scattering \((Q^{-1}_s)\) and intrinsic \((Q^{-1}_i)\) attenuation by using the following relations:

\[
Q^{-1}_i = L^{-1}_e \frac{V_S}{2\pi f} , \quad Q^{-1}_s = B_0 Q^{-1}_i , \quad Q^{-1}_i = (1 - B_0) Q^{-1}_s .
\]

Total, scattering and intrinsic \(Q\) were compared to the observed coda-\(Q\) \((Q_{\text{coda}})\) calculated in a previous work (Giampiccolo et al. 2004), and to the expected coda-\(Q\) \((Q_{\text{exp}})\) calculated in this study by using the following expression (e.g. Hoshiba 1991; Mayeda et al. 1992):

\[
Q^{-1}_{\text{exp}} = Q^{-1}_s \left[ \frac{1 - C_2 + 2C_3 g v t + 3C_4 (g v t)^2 + ...}{1 + C_2 g v t + C_3 (g v t)^2} \right] + Q^{-1}_i ,
\]

where \(C_2\) is the coefficient for the 0th scattering order. We cut-off at the 10th order for our analysis.

In Table 1, the values of \(B_0, L^{-1}_e, Q^{-1}_i, Q^{-1}_s, Q^{-1}_i, Q^{-1}_{\text{coda}}\) and \(Q^{-1}_{\text{exp}}\) for each frequency are listed.

Fitting the values of \(Q^{-1}_i, Q^{-1}_s, Q^{-1}_i\) and \(Q^{-1}_{\text{coda}}\) by means of the relation \(Q = Q_0 f^n\) we obtained the following attenuation laws: \(Q_T = 38 f^{1.44}\), \(Q_s = 44 f^{1.8}\) and \(Q_i = 192 f^{0.86}\).

In Fig. 8 the scattering, intrinsic, total and coda-\(Q\) expected are shown together with the coda-\(Q\) observed, calculated at a lapse time equal to 40 s and under the hypothesis of a single scattering process (Giampiccolo et al. 2004).

**DISCUSSION AND CONCLUSIONS**

The present work is an extension of previous studies on attenuation of seismic waves carried out in southeastern Sicily (Giampiccolo et al. 2002, 2003, 2004; de Lorenzo et al. 2004).
First, we applied the coda-normalization method (Aki 1980) to a data set composed of about 180 selected seismic events to investigate in detail on $Q_S$ frequency dependence and possible variations of seismic attenuation with the source-station path. We obtained an average attenuation law $Q_S = 49f^{-0.38}$. It is worth stressing that our $Q_S$ estimates correlate well, both in absolute value and frequency dependence, with those obtained by Giampiccolo et al. (2003) who applied the frequency decay method (e.g. Bianco et al. 1999) to a slightly different data set recorded in the same area. In fact, these values follow the same distribution around the best-fit line of the $Q_S$ values estimated in the present study. This indicates the reliability and efficiency of both methods for attenuation measurements. Moreover, our results are in good agreement with the frequency dependence exponents obtained for other seismically active areas in different parts of the world (e.g. Japan, Yoshimoto et al. 1993; northern Greece, Hatzidimitriou 1995; Turkey, Akinci & Eydo˘gan 1996; and Horasan & Boztepe-Guney 2004), in which processes as faulting are likely to introduce strong heterogeneities.

When searching for a $Q_S$ dependence on source-to-receiver path, no significant differences in the attenuation law calculated at each single station were found. The $Q_S$ values at 1 Hz are, within the errors, close among the stations and the frequency dependence coefficient $n$ does not seem to vary among them. Only station SR2 shows a stronger frequency dependence, being $n = 1.16$. Very local site effects could be invoked to explain this clear difference with respect to the average $n$ ($\sim0.88$).

Finally, if we compare the average attenuation law of $S$ waves $Q_S = 49f^{-0.38}$ with that of coda waves ($Q_C = 75 \pm 10f^{0.9\pm0.06}$), calculated at a lapse time of the same order of the $S$-wave travel path (25 s) by Giampiccolo et al. (2004), we observe that the frequency dependence of the coda-$Q$ is roughly the same as that of the S wave $Q$. Also $Q_S$ and $Q_C$ values are, within the associated errors, comparable, except at low frequencies (around 1.5 Hz) were $Q_C$ is slightly larger than $Q_S$.

According to the energy flux model (Frankel 1991) the similarity between the coda and the $S$-wave decay at one frequency implies that intrinsic attenuation is the dominant cause of attenuation at that frequency. The results of our study show that below 1.5 Hz the coda and the amplitude decay of $S$ waves with distance are not comparable. This indicates that scattering attenuation at low frequencies becomes important.

In order to accurately quantify the separate amount of scattering and intrinsic absorption in the frequency range 1.5–15 Hz, we applied the MLTWA technique (Hoshiba et al. 1991) to the same data set. We then compared the new results with those obtained by Giampiccolo et al. (2004) who attempted an estimate of $Q_S^{-1}$ and $Q_C^{-1}$ in southeastern Sicily by applying the Wennerberg (1993) method.

We found that the seismic albedo increases with decreasing frequency, being higher than 0.5 below 3 Hz; this indicates that scattering attenuation dominates over intrinsic absorption. At frequencies above 3 Hz the intrinsic absorption dominates over scattering attenuation. This result is in agreement with what was found by Giampiccolo et al. (2004) and it is probably due to the fact that the seismic wavelengths are lower than the average size of the heterogeneities in the region. All the attenuation parameters are frequency dependent and the coefficient $n$ ranges between 0.86 and 1.8. The highest value corresponds to $Q_C$. A decrease of $Q_C^{-1}$ faster than $f^{-1}$ with increasing frequency implies that the medium is characterized by a Gaussian autocorrelation function (Wu 1985). As suggested by several other authors (e.g. Mayeda et al. 1992; Akinci et al. 1995; Canas et al. 1998; Akinci & Eydo˘gan 2000) this strong frequency dependence could be related to the size of heterogeneities. In such a hypothesis, a strong frequency dependence of $Q_C^{-1}$ occurs.

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**Figure 5.** Attenuation laws calculated at each seismic station (black thin lines) and average attenuation law calculated for the study area (grey thick line).
Figure 6. Observed integrated seismic energies for the 0–12 s (crosses), 12–24 s (diamonds) and 24–36 s (squares) lapse time windows calculated for the hypocentral distance range 20–80 km. The continuous lines represent the theoretical curves. Frequency and corresponding best fit $B_0$ and $L_0^{-1}$ are also reported.
Figure 7. Map of the residuals normalized to their minimum at all frequency bands for the hypocentral distance range 20–80 km. Grey cross indicates the pair $B_0$ and $L_e^{-1}$ of minimum residual.
Table 1. Inverse extinction length, seismic albedo, total attenuation, scattering and intrinsic attenuation, observed and expected coda attenuation are shown. Six frequency bands have been studied.

| Frequency (Hz) | $L_{-1}^{-1}$ (km$^{-1}$) | $B_0$ | $Q_T^{-1}$ | $Q_c^{-1}$ | $Q_i^{-1}$ | $Q_{Cobs}^{-1}$ | $Q_{Cexp}^{-1}$ |
|---------------|-------------------------|------|-----------|-----------|-----------|---------------|---------------|
| 1.5           | 0.054                   | 0.82 | 0.0189    | 0.0156    | 0.0034    | 0.0064        | 0.0037        |
| 3             | 0.028                   | 0.52 | 0.0049    | 0.0026    | 0.0024    | 0.0004        | 0.0047        |
| 6             | 0.019                   | 0.38 | 0.0017    | 0.0006    | 0.0010    | 0.0024        | 0.0009        |
| 9             | 0.019                   | 0.34 | 0.0011    | 0.0004    | 0.0007    | 0.0016        | 0.0007        |
| 12            | 0.019                   | 0.34 | 0.0008    | 0.0003    | 0.0005    | 0.0012        | 0.0005        |
| 15            | 0.023                   | 0.32 | 0.0008    | 0.0003    | 0.0005    | 0.0011        | 0.0005        |

Figure 8. Total, scattering, intrinsic, observed coda-$Q$ and expected coda-$Q$ versus frequency.

when the heterogeneities responsible for the scattering are, at least, comparable with the wavelength for the lowest frequencies analysed (around 2 km). A comparison between $Q$-coda observed ($Q_{Cobs}$) and $Q$-coda expected ($Q_{Cexp}$) was also performed. It is worth stressing that poor fit is observed between the two parameters at all frequencies. We found that $Q_{Cexp}$ is similar to $Q_i^{-1}$, in agreement with finite difference simulation (Frankel & Wennerberg 1987) and laboratory experiments (Matsunami 1991). However, $Q_{Cobs}$ is closer to but greater than $Q_i^{-1}$, at least at frequencies above 3 Hz, contradicting both theoretical predictions and the empirical results of Giampiccolo et al. (2004). This discrepancy between predictions and observations has been observed also by other authors who applied the MLTW A technique under the assumption of uniform distribution of scatterers (e.g. Mayeda et al. 1992; Pujades et al. 1997; Akinci & Eydoğan 2000; Bianco et al. 2005). A possible interpretation is that the model fitting for the observed energy–distance relation at multiple lapse time windows did not work well at all distances. The assumption of a uniform distribution of scatterers may be unrealistic because it is widely accepted that heterogeneity decreases with increasing depth. Any depth-dependent attenuation mechanism in the lithosphere will cause the departure from the idealized uniform and homogeneous case. Therefore, models incorporating non-uniform distribution of scatterers (e.g. Hoshiba 1997; Hoshiba et al. 2001) and the possibility that intrinsic dissipation may decrease with depth could be more appropriate than those assuming a uniform earth model.

Interestingly, Giampiccolo et al. (2004) found increasing values of coda-$Q$ with lapse time which can be reasonably interpreted in terms of a non-uniform medium and depth-decreasing intrinsic attenuation in the lithosphere of southeastern Sicily. On this basis, the quality factors estimated in this study might be apparent, since the simplified model of the medium assumed may provide some ambiguity (e.g. Margerin et al. 1999; Hoshiba et al. 2001). In spite of this, the application of different methods (Wennerberg 1993; Hoshiba et al. 1991) to very similar data sets led to the result that intrinsic attenuation dominates over scattering at frequencies higher than 3 Hz; this means that we are dealing with the main attenuative properties of the lithosphere in the study area. However, the simplicity of the theoretical models used to measure the attenuation parameters and the strong assumptions underlying them hinder the resolution of fine features of the attenuation processes.

Several authors have tried to separately estimate $Q_i^{-1}$ and $Q_T^{-1}$ in different regions worldwide by applying the MLTW A method. Our results are in agreement with those obtained in many other tectonic regions which show dominant intrinsic attenuation, at least at frequencies higher than 3 Hz (e.g. Japan, Hoshiba 1993; western
Figure 9. Trend of total (a), intrinsic absorption (b) and scattering attenuation (c) with frequency for different regions: Southern Appenine, Italy (Bianco et al. 2002); southeastern Sicily (this study); southern California (Jin et al. 1994); central California (Mayeda et al. 1992); eastern Turkey (Akinci & Eydoğan 2000); Kanto-Tokay, Japan (Hoshiba 1993); northern Chile (Hoshiba et al. 2001); southern Spain and western Anatolia (Akinci et al. 1995); southeastern Iberian peninsula (Pujades et al. 1997); northeastern Venezuela (Ugalde et al. 1998); central Colombian Andes (Ugalde et al. 2002); southern Netherlands (Goutbeek et al. 2004); northeastern Italy (Bianco et al. 2005); northeastern Sicily (Tuvé et al. 2004) and Mt Etna volcano, Sicily (Del Pezzo et al. 1996).
Anatolia, Akinci et al. 1995; southeastern Iberian peninsula, Pujades et al. 1997; northeastern Venezuela, Ugalde et al. 1998; southern Appenine, Bianco et al. 2002; central Colombian Andes, Ugalde et al. 2002; southern Netherlands, Goutbeek et al. 2004; northeastern Italy, Bianco et al. 2005). In Fig. 9 we compare the attenuation parameters obtained in southeastern Sicily with the pattern of $Q_{s}^{-1}$, $Q_{c}^{-1}$ and $Q_{v}^{-1}$ with frequency observed in several areas of the world. It can be seen that our $Q_{s}^{-1}$ estimate is generally lower than that observed in most of other tectonic regions here quoted. In particular, southeastern Sicily is less attenuating than southern and central California, eastern Turkey, southern Spain and southern Netherlands, at least at frequencies higher than 3 Hz. Conversely, the total attenuation seems to be comparable with that observed in central Colombian Andes, southern Appenine (Italy) and northeastern Italy. Following the interpretation of the results obtained by Bianco et al. (2005) in northeastern Italy, the low attenuation values found in southeastern Sicily could be related to the elastic properties of the thick sequence of carbonate rocks which characterizes the study region. The trend of the intrinsic absorption with frequency in the study area is comparable with that observed in northeastern Venezuela, central Colombion Andes and southeastern Italy, whereas the scattering process seems to be similar to that observed in western Anatolia (at least above 6 Hz). On the other hand, in general, southeastern Sicily appears more heterogeneous than southern Appenine (Italy), northeastern Venezuela, northern Chile, northeastern Italy and southern India, at the scale length corresponding to the analysed frequencies.

An interesting comparison can be carried out with scattering and intrinsic $Q_{s}^{-1}$ obtained at the volcanic area of Mt. Etna, located nearby the northern margin of the Hyblean Foreland. At Mt. Etna both scattering and intrinsic absorption are higher in southeastern Sicily. Moreover, $Q_{c}^{-1}$ is close to $Q_{v}^{-1}$ in the frequency range 1–18 Hz. This means that heterogeneity, which generates the scattering phenomena (explained with the presence of different direction of faults, medium with strong layering, magma intrusions, source of heat flows, etc.) plays a key role in determining the higher seismic attenuation at Mt. Etna with respect to the tectonic environment here investigated.

Finally, the here obtained $Q_{c}^{-1}$ and $Q_{v}^{-1}$ values are also lower than those obtained in northeastern Sicily (Tuvé et al. 2004). This is not surprising because, even though both areas are tectonically active, they have significantly different geological and tectonic settings.

Our results evidence a weakly heterogeneous crust in southeastern Sicily. This well correlates with the seismic pattern and the tectonic characteristics of the area, which is basically characterized by large scale heterogeneities (e.g. Bianca et al. 1999; Azzaro & Barbano 2000; Musumeci et al. 2005), as also supported by a recent 3-D velocity tomography study (Scarfi et al. 2005).

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