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DEPTH-DEPENDENT SEISMIC ATTENUATION IN THE GRANADA ZONE (SOUTHERN SPAIN)

BY J. M. IBÁÑEZ, E. DEL PEZZO, F. DE MIGUEL, M. HERRAIZ, G. ALGUACIL, AND J. MORALES

ABSTRACT

Coda-Q estimates for the Granada Basin (Southern Spain) are obtained by analyzing earthquakes occurring within or close to the borders of this area. The data set is composed of 54 earthquakes with local magnitudes ranging between 2.2 and 4.0 and with epicentral distances to the nearest station smaller than 10 km. A single-scattering process is assumed. Coda-Q values obtained show a clear dependence on frequency according to the relationship $Q = Q_o f^n$, where n ranges from 0.81 to 0.89. After removing the main site effects and discussing the possible multiple-scattering influence, the lapse-time dependence is interpreted as due to attenuation variations with depth. This result agrees with the variation of coda Q obtained by using different sets of events with increasing hypocentral depth ranges.

INTRODUCTION

Studies of attenuation and velocity structure of the seismic media in the high-frequency range (1 to 30 Hz) are fundamental to understanding the seismo-tectonic features of a given area. Recently, it has been possible to regionalize the attenuation properties of the seismic regions using the "coda method" to estimate the quality factor Q. This method is based on the properties of the coda waves, which are scattered waves from heterogeneities of the medium (Aki, 1969). As coda waves are the result of random processes that take place in the medium, they reflect averaged path effects instead of source or direct path-dependent characteristics. Therefore, problems concerning radiation pattern or directivity of the source, which are decisive when the so called "direct methods" for finding Q are used, can be avoided. In addition, coda methods are less sensitive to precision in source location and site effects than those based on direct waves. Therefore, coda wave analysis provides us with a single-station method which gives a stable estimate of a sort of space-averaged Q. At present, the physical meaning of this coda Q remains quite controversial. Aki (1980a,b) stated that this parameter accounts for both intrinsic and scattering losses, while more recently Frankel and Wennerberg (1987), by using different theoretical assumptions (energy flux model), have suggested that coda Q is most due to intrinsic attenuation. Moreover, the scattering process is far from being well understood. Analysis based on the single-scattering model (Aki and Chouet, 1975) assumes that the scattering is a weak process and implicitly accept a violation of the energy conservation law. At the far end, the diffusion model satisfies this law by accepting that all the seismic energy is scattered through a strong scattering mechanism. Intermediate models assume multiple scattering (see, for example, the two dimensional model established by Shang and Gao, 1988, which satisfies the energy conservation law). This phenomenon seems to occur in most seismic media for long lapse times after the origin time of the earthquake (Gao *et al.*, 1983; Frankel and Wennerberg, 1987; Wu and Aki, 1988).

In spite of these shortcomings, single-scattering coda Q is widely used in seismotectonic studies, and it has been successfully applied for comparison among different tectonic areas (Phillips and Aki, 1986; Herraiz and Espinosa, 1987; Havskov *et al.*, 1989; Steck *et al.*, 1989). The comparison takes place through the coda-Q frequency dependence given by the expression

$$Q_c = Q_o * f^n \tag{1}$$

where Q_o is the Q value of $T = 1$ sec and n ranges from 0.4 to 1.1 (e.g., Gagnepain-Beyneix, 1987). This frequency dependence shows regional variations, both in the Qo and n parameters.

Recently a coda-Q dependence on lapse-times has been clearly stated by several workers (Roecker *et al.*, 1982; Pulli, 1984; Gagnepain-Beyneix, 1987). This dependence may be due to a Q variation with depth and/or to some wrong assumptions of the single-scattering model.

The aim of this paper is to measure the lapse-time dependence of the coda Q in the Granada Basin, southern Spain, and to study the possibility of a real depth dependence of attenuation in this area.

Geological Setting

The Granada basin (Fig. 1) is located in the central sector of the Betic Cordillera (south-east Spain). It is one of the largest intra-mountain Neogene basins of the Betic ranges. The Granada Basin is younger than the last alpine event which structured the Betic cordillera, ranging from the upper Miocene to the present. All the seismic stations used, of the Andalusian Seismic Network (RSA) and the Portable Seismic Network (RSP), are situated on hard rock, with the exception of the ACHM station which is located on a 1.4 km thick sedimentary layer (Morales *et al.*, 1990).

Local seismicity has been monitored since 1983 in the region using a local network, allowing to correlate the main tectonic features with the pattern of the hypocentres (Vidal, 1986; Vidal *et al.*, 1987).

Strong variations in the crustal thickness were found by Deep Seismic Soundings (Udías and Suriñach, 1980). Crust reaches 42 km towards the north-east border of the zone under study and becomes as thin as 14 km towards the south beneath the Alboran Sea. The maximum thickness gradient is just beneath the Granada basin, where preliminary and yet unpublished studies, based on gravimetric data of a hydrocarbon survey, reveal a crustal thickness of about 25 km (Morales, unpublished data). The P-wave average velocity in the Betic Cordillera varies locally from 6.0 km/sec to 6.3 km/sec. The P_n velocity lies close to 8.1 km/sec with strong lateral variations as we enter the Alboran Sea, where the velocity is reduced to 7.5–7.9 km/sec (Banda, 1988).

Data

Data used in this study were recorded by two short-period digital seismic networks which operate in the Granada Basin (Alguacil, 1986) (Fig. 2). The network consists of seven permanent stations (RSA) and six portable stations (RSP), all of them with vertical component seismometers operating at 65 dB dynamical range and in a low-noise environment. All of them use 1 Hz vertical seismometers Mark L4-C or Kinemetrics Ranger SS-1. The signals are radio-telemetered via FM subcarriers to the central recording site at the Cartuja Observatory in real time (Alguacil, 1986). There, each channel signal is filtered for aliasing with a 30-Hz seven-pole Butter-

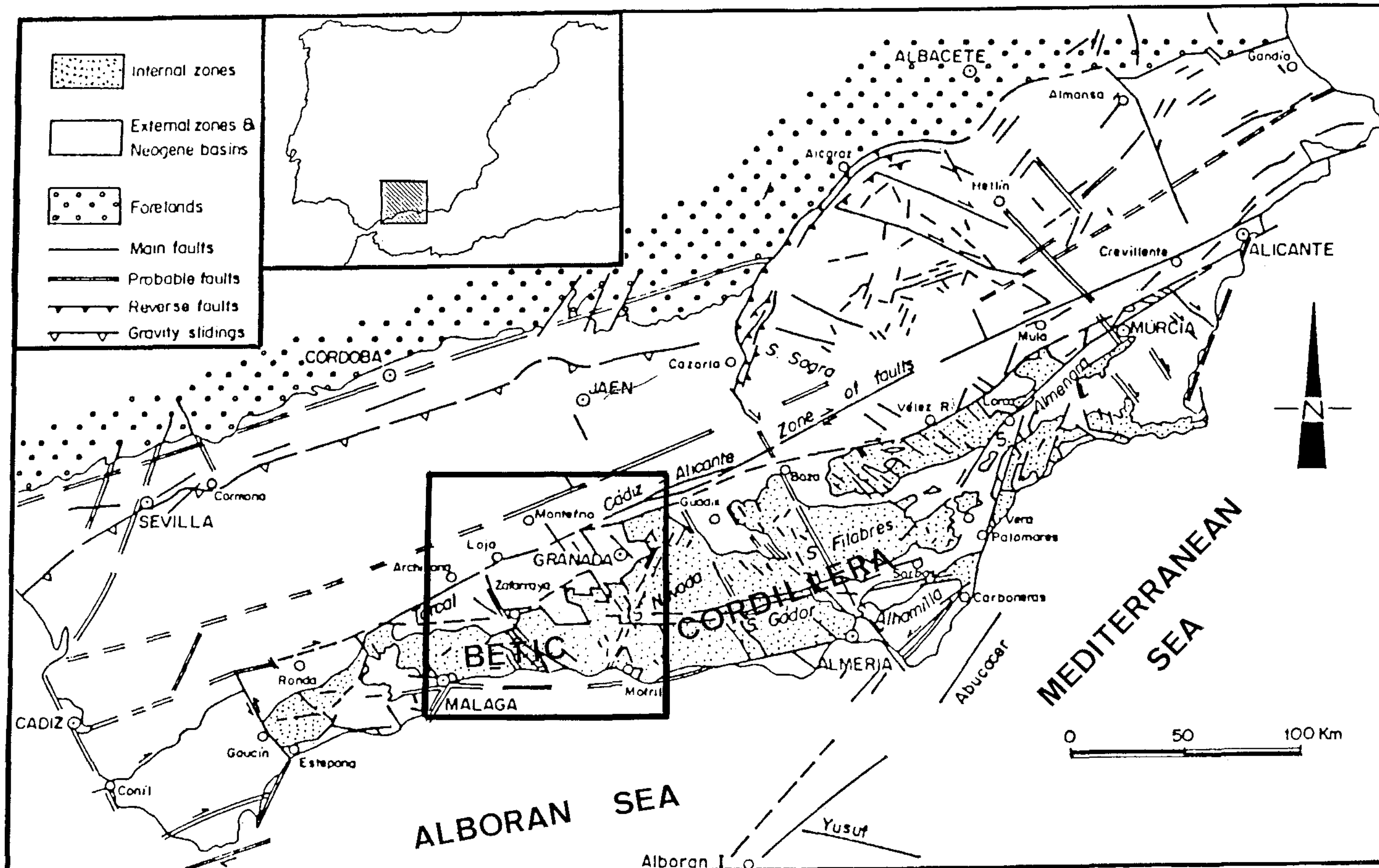

FIG. 1. Tectonic sketch and main fault systems of the Betics Region, Southern Spain, (from Sanz de Galdeano, 1988). The analyzed area is shown in a rectangle.

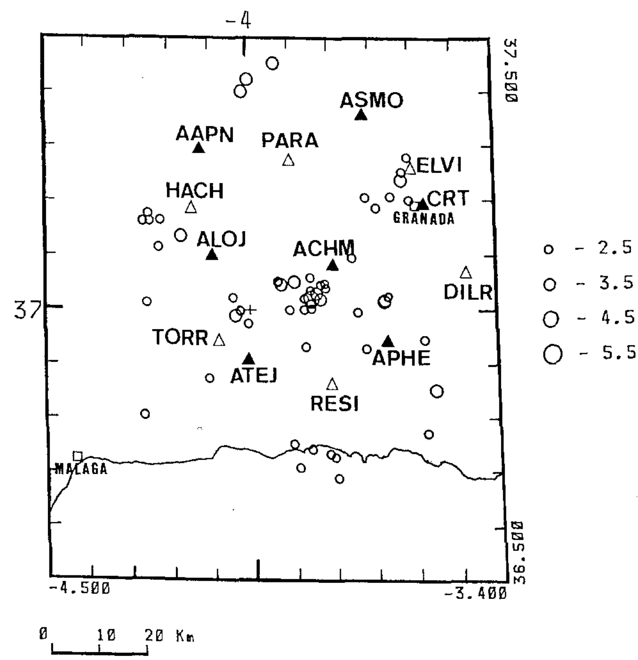

FIG. 2. Distribution map of the earthquakes used in this work. The sizes of the circles are proportional to the event magnitudes. Solid triangles represent the stations of the RSA network, and open triangles represent the stations of the RSP network used in this study. The data used were selected with a magnitude ranging from 2.2 to 4.0 and an epicentral distance from the nearest station smaller than about 10 km.

worth low-pass filter, samples at 100 sps, and converted to digital form with a resolution of 12 bits.

The hardware of the data acquisition system consists of a PC with several plug-in boards: a 16-channel A/D converter, a real-time clock modified to use an external high stability oscillator, and an input-output card used for clock synchronization with time signals. Each network has its own data acquisition system, although all stations could be recorded by one single system. The overal response is flat (−3dB) to ground velocity in the band 1 to 30 Hz, the lower limit being posed by the seismometer and the higher one by the cut-off frequency of the antialias filter. The electronics design, the magnification, and the sites (hard bedrock, far from cultural noise sources) have been carefully chosen so as to achieve a practical dynamic range of 66 to 72 dB.

The acquisition software (Guirao, 1989) is interrupt-driven, so the recording on disk can be performed simultaneously with the data acquisition and the algorithms of event-detection and cut-off. As a result, there is no memory-size limitation for the length of the event, as there is in other systems (Lee *et al.*, 1988).

Two hundred earthquakes were selected on the basis of a first rough visual inspection. This first selection was carried out by selecting any earthquake that was recorded in at least two of the stations with a magnitude greater than 1.5. A further discrimination shortened the sample to events with magnitude, defined by De Miguel *et al.* (1988), formulae obtained from the duration of the recordings, ranging from 2.2 to 4.0, and an epicentral distance from the nearest station smaller than about 10 km. This selection was completed by analysing the signal-to-noise ratio of the end of the coda, taking those seismograms which

TABLE 1

HYPOCENTRAL PARAMETERS OF EVENTS USED

#	Date (YY/MM/DD)	Time (GMT) HH:MM:SS.SS	Localization Longitude (N)	Localization Latitude (W)	Magnitude	Depth	N^*
1	88/09/15	00:27:33.89	36.990	3.739	2.7	9.5	6
2	88/09/28	17:23:16.14	36.679	3.792	2.5	6.9	6
3	88/10/07	14:56:00.98	36.745	3.861	2.6	1.3	6
4	88/10/12	10:49:29.92	37.023	4.047	2.2	25.7	7
5	88/10/13	13:24:37.06	36.813	4.178	2.2	7.4	2
6	88/10/14	19:25:47.55	36.996	4.284	2.5	16.5	9
7	88/10/21	12:05:24.31	37.026	3.892	2.7	15.1	12
8	88/10/22	16:13:29.30	37.024	3.846	2.2	14.4	11
9	88/10/23	20:39:52.25	37.188	3.689	2.4	9.4	11
10	88/10/26	13:55:55.69	36.715	3.559	2.6	2.4	9
11	88/10/31	03:18:17.17	37.141	4.181	3.4	8.4	12
12	88/11/14	11:20:32.15	37.043	3.830	2.6	7.4	5
13	88/11/17	21:33:28.63	37.376	4.227	3.2	41.7	12
14	88/11/21	11:02:12.66	37.169	4.227	2.3	13.7	3
15	88/11/25	19:01:56.28	36.689	3.491	2.9	5.7	4
16	88/12/01	16:13:52.66	36.941	3.570	2.7	10.8	10
17	88/12/03	02:33:50.22	37.170	4.257	2.3	9.5	6
18	88/12/03	02:39:02.06	37.132	4.279	2.4	8.6	5
19	88/12/03	03:16:24.91	37.167	4.258	2.3	9.4	6
20	88/12/05	20:12:26.05	37.028	3.837	4.0	14.0	10
21	88/12/06	06:09:15.26	37.021	3.846	3.2	13.6	10
22	88/12/06	20:15:35.28	37.022	3.844	3.8	13.6	10
23	88/12/07	18:50:25.31	37.025	3.852	3.6	13.8	11
24	88/12/07	20:38:01.40	37.017	3.854	2.9	11.4	6
25	88/12/10	00:26:20.32	37.024	3.719	2.4	19.0	10
26	88/12/12	16:52:00.58	36.755	3.906	2.7	10.8	4
27	88/12/14	17:00:07.19	37.050	3.930	3.8	15.6	11
28	88/12/18	09:39:54.11	36.807	4.269	2.4	2.8	3
29	88/12/22	00:14:56.86	37.026	3.662	2.5	14.2	8
30	88/12/23	16:37:38.47	36.732	3.663	2.7	5.1	8
31	88/12/25	13:56:00.36	37.020	3.660	3.6	11.0	10
32	89/01/03	05:32:08.46	37.206	3.629	2.6	9.4	6
33	89/01/05	20:18:35.11	37.018	3.864	2.4	23.1	7
34	89/01/05	20:24:12.72	37.010	3.865	2.9	24.0	11
35	89/01/06	06:33:50.23	37.016	3.856	2.3	20.0	6
36	89/01/06	10:53:38.75	37.441	3.978	3.0	32.6	11
37	89/01/12	18:32:54.94	37.453	3.934	3.0	43.1	11
38	89/01/21	13:35:52.64	37.280	3.610	2.5	14.8	9
39	89/02/08	12:53:11.20	37.252	3.625	2.8	0.1	11
40	89/02/13	12:23:32.33	37.001	3.903	2.4	16.3	4
41	89/02/15	00:05:03.28	36.992	4.034	3.7	17.2	11
42	89/02/15	01:55:05.27	36.995	4.032	3.4	17.3	11
43	89/02/15	13:19:16.92	36.979	4.007	2.6	0.1	3
44	89/02/21	16:46:42.53	36.935	3.871	2.9	13.2	8
45	89/02/23	16:35:41.86	37.095	3.750	2.3	20.6	4
46	89/02/24	05:54:20.64	37.204	3.606	2.6	17.6	8
47	89/02/28	04:43:50.36	36.926	3.719	2.5	14.4	5
48	89/03/02	13:31:36.67	37.236	3.629	3.0	20.7	11
49	89/03/08	21:40:20.44	37.056	3.930	2.5	14.6	10
50	89/03/16	16:28:28.42	36.735	3.813	2.5	8.1	4
51	89/03/17	16:29:49.27	36.710	3.890	2.5	6.3	6
52	89/03/22	16:08:21.80	37.055	3.895	3.1	16.4	11
53	89/04/05	09:17:52.73	37.116	4.116	2.5	5.8	7
54	89/04/10	23:42:12.25	37.061	3.857	2.5	11.7	11

In all cases the RMS is smaller than 0.32 sec, the epicentral error is smaller than 3.2 km and the depth error is smaller then 6.7 km.

* Total number of seismograms, for each event, used to calculate Q_c.

present a signal-to-noise ratio greater than 5 as the end of the coda. In this way, only events occurring inside, or at least near, the borders of the networks were chosen, and 54 earthquakes were finally selected and processed (Fig. 2 and Table 1), corresponding to a total of 426 analyzed seismograms.

Method and Data Processing

In the present analysis the single-scattering hypothesis is assumed. Coda is interpreted as generated by randomly but uniformly distributed heterogeneities (Aki and Chouet, 1975). Assuming weak and single back-scattering we can express the amplitude of the coda envelope, A_c, as a function of the lapse time from the origin, t, as:

$$A_c(f, t) = A_o(f)t^{-1}\exp(-\pi ft/Q_c) \tag{2}$$

where A_o is a source factor, the power of the lapse time accounts for the geometrical spreading for body waves, and Q_c is the quality factor of coda waves. Expression (2) strictly holds when station and source are coincident.

Sato (1977) developed a single isotropic scattering model that overcomes this last restriction. His model assumes the weak scattering hypothesis. In addition, Sato's model implies a physical meaning different from that stated by Aki and Chouet (1975), because it requires the wave-lengths to be comparable with the size of heterogeneities instead of being longer, as the back scattering model assumes.

Sato's model yields the expression

$$A_c(f, t) = A_oK(r, a)\exp(-\pi ft/Q_c) \tag{3}$$

where $K(r, a)$ is a function of the station-source distance, r, and is defined as

$$K(r, a) = \frac{1}{r}\frac{1}{a}\ln\left(\frac{a+1}{a-1}\right) \tag{4}$$

In its turn, $a = (t/t_s)$, t_s being the travel time of S waves. Both expressions (2) and (3) are used in the present paper. For a coda start time greater than 2^*ts, Rautian and Khalturin (1978) showed that the coda parameters A_o and Q_c obtained by using the expression (2) are not dependent on source-station distance. The present data show a coda start time almost always greater than $2 \cdot t_s$. Checks were made using both formulas in fitting the data, and no variations between the estimates were found. Therefore, expression (2) has been used. Taking the algorithm of both sides in this formula we obtain

$$\ln(A_c(f, t)t) = \ln A_o(f) - \pi ft/Q_c \tag{5}$$

This equation can be used to least-square fit the data and to invert for Q_c, and A_o for each frequency. $A_c(f, t)$ can be estimated by pass band filtering the signal in a frequency band whose center is f, and the coda envelope can be obtained taking the rms of the signal in a sliding time window (Aki and Chouet, 1975); alternatively $A(f, t)$ can be estimated using a moving fast Fourier transform window (Lee *et al.*, 1986). In the present study, seismograms have been numerically filtered using a five-pole Butterworth filter centered successively at central frequencies, fc, of 1.5, 3, 6, 12, and 18 Hz, and with a bandwidth equal to 0.7 f_c. Although the envelope is

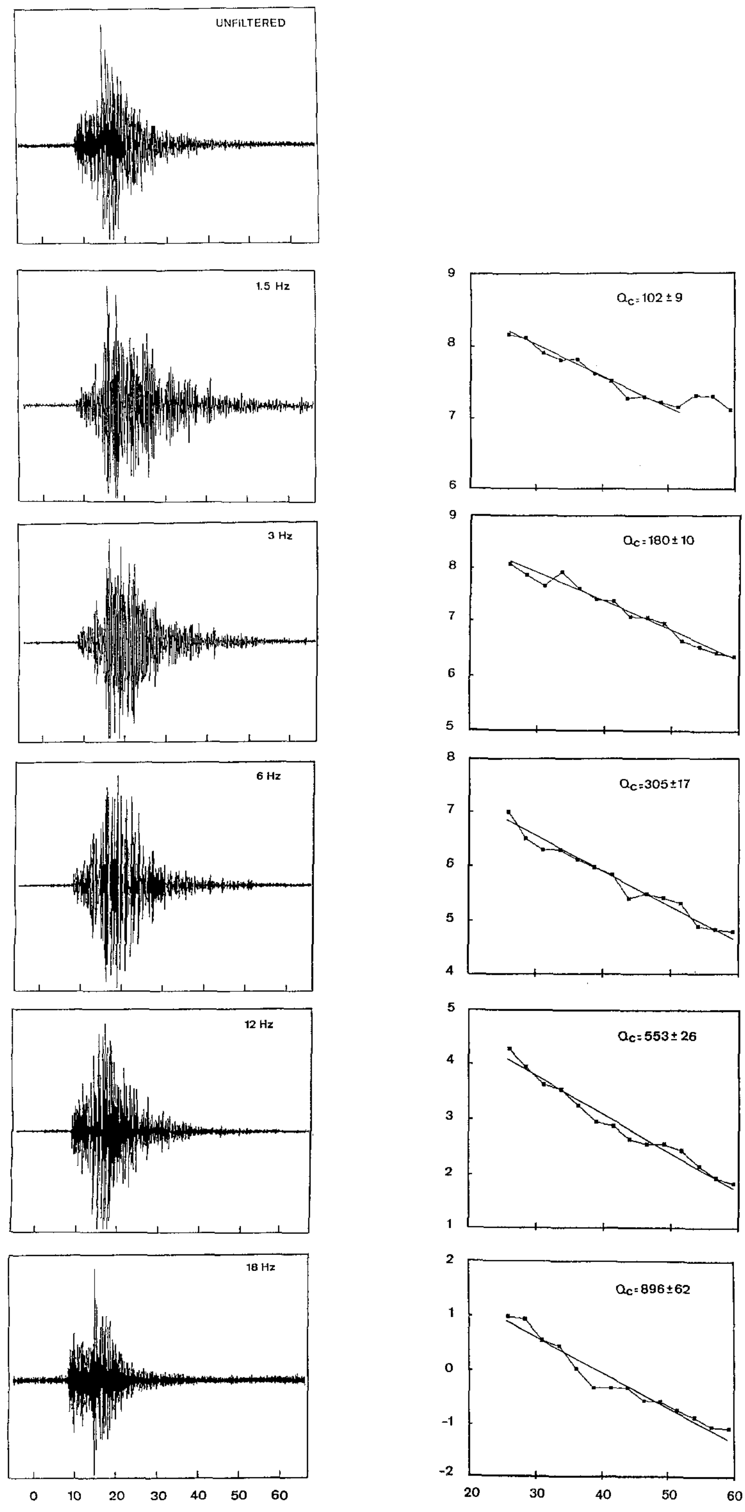

FIG. 3. An example of adjustment of Q_c. This record is the event 37 recorded in AAPN. To the right of each filtered seismogram the quantity $\ln(A(w, t) \cdot t)$ versus the lapse time is presented. The continual line above each curve presents the adjustment obtained.

usually calculated by taking the rms, a different method has been used in this paper. This approach calculates the Fourier spectrum of the filtered signal in a time window of 5.12 sec which slides along the coda in stetps of 2.56 sec and evaluates the integral of the spectrum in the same band of the filter at each step. The values of this integral constitute a smoother envelope of the coda than that obtained using the rms approach.

The analyzed time interval corresponds to the section of the seismograms in which the coda decay is regular. The beginning and the end of this interval have been empirically found by using the approach defined in Ibáñez *et al.* (1990). This approximation consists of defining the coda at the point in the seismogram, after the direct S phase in which the numerical function $\ln(A(w, t) \cdot t)$ begins to decay in a regular manner. In all the cases analyzed, this time coincided with more than $2 \cdot t_s$. A further selection of data has been performed after the least-square fit. Data with a correlation coefficient greater or equal to 0.80 were selected. This later selection, which satisfies the correlation coefficient criteria, permits that the effects on the coda envelope that make it not follow expression in a regular manner, such as the site and noise effect (Ibáñez, 1990), be eliminated. Figure 3 presents an example of how the Q_c and A_0 values are fitted to the coda envelopes for different frequency bands, with one example of the filtered seismograms for the same frequency bands.

Results

An interesting effect was noticed at the ACHM station, located approximately in the middle of the sedimentary basin. Coda waves in the 1.5 and 3 Hz band exhibited remarkable reverberations for magnitudes greater than approximately 3.0. This effect is suspected to be generated by the sedimentary layer of the basin which reaches its greatest thickness beneath ACHM station, and it will be the subject of future studies (Ibáñez, 1990). In the present paper, coda envelopes showing this effect have been eliminated to minimize possible site dependence in the estimates of coda parameters.

In order to investigate on a possible lapse time dependence of the estimated coda Q, we have carried out the analysis of each seismogram using four coda lengths, measured from the beginning of the coda, from 0 to 20, 0 to 40, 0 to 60, and 0 to 80 sec, if the duration of the coda permitted it. These coda lengths correspond to four different lapse times, 30, 50, 70, and 90 sec, measured from the origin time of the earthquake. Results are reported in Table 2. Figure 4 shows a Q_c dependence

TABLE 2

MEASURED QUALITY FACTOR Q_c FOR DIFFERENT LAPSE TIME

f (Hz)	30 sec (1)		50 sec (2)		70 sec (3)		90 sec (4)	
	Q_c	N^*	Q_c	N^*	Q_c	N^*	Q_c	N^*
1.5	109.2 ± 24.2	169	142.3 ± 25.0	208	172.9 ± 24.0	101	223.6 ± 21.9	17
3	190.6 + 38.3	253	242.6 ± 34.7	303	298.0 ± 37.8	142	402.8 ± 34.4	16
6	358.9 ± 57.4	312	439.4 ± 50.3	327	542.3 ± 49.7	145	771.1 ± 48.9	16
12	656.6 ± 82.7	332	809.5 ± 77.8	320	1009.2 ± 83.6	143	1427.7 ± 85.4	17
18	944.2 ± 102.6	316	1203.2 ± 118.6	302	1477.4 ± 115.6	132	2024.5 ± 103.0	16

* N represents the number of fit with a correlation coefficient equal or greater than 0.8 (1) $Q_c = (75.1 \pm 9.4)\, f^{(0.87\pm0.04)}$ correlation coefficient = 0.9998; (2) $Q_c = (97.0 \pm 8.8)\, f^{(0.86\pm0.03)}$ correlation coefficient = 0.9993; (3) $Q_c = (118.1 \pm 8.9)\, f^{(0.86\pm0.02)}$ correlation coefficient = 0.9995; (4) $Q_c = (157.4 \pm 8.1)\, f^{(0.89\pm0.03)}$ correlation coefficient = 0.9999.

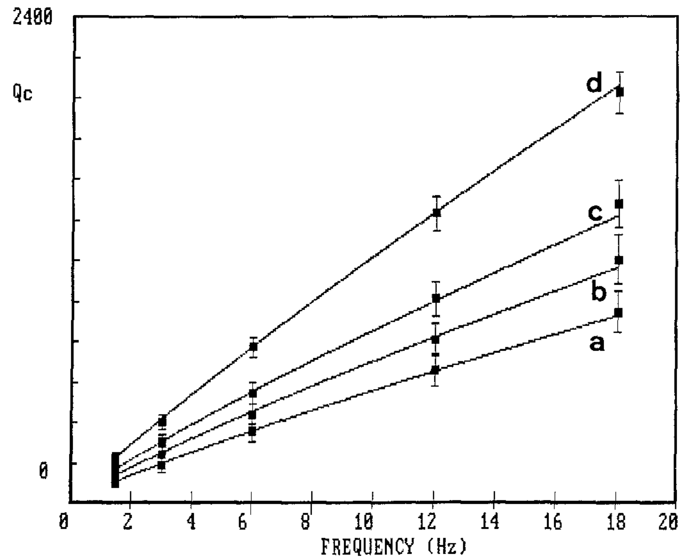

FIG. 4. Plots of Q coda versus frequency, in linear scale, having different coda lengths. Curve (a) is with a lapse time not greater than 30 sec, (b) 50 sec, (c) 70 sec, and finally (d) 90 sec. The values are reported in Table 2. The error bars represent the average errors of Q_c obtained by applying the propagation of error theory.

TABLE 3

ESTIMATED Q_c FOR DIFFERENT FOCAL DEPTH EVENT GROUPS*

f (Hz)	0–10 km (1)		10–20 km (2)		20–30 km (3)		30–45 km (4)	
	Q_c	N†	Q_c	N†	Q_c	N†	Q_c	N†
1.5	148.2 ± 20.6	26	167.6 ± 24.8	63	160.5 ± 24.8	6	196.3 ± 30.7	16
3	251.5 ± 28.0	40	271.5 ± 32.3	81	264.4 ± 38.4	20	355.5 ± 44.5	27
6	428.7 ± 36.8	42	490.6 ± 45.7	80	436.2 ± 75.8	21	653.3 ± 64.1	29
12	789.4 ± 63.2	43	924.8 ± 82.1	81	798.0 ± 75.8	20	1213.6 ± 94.0	30
18	1190.2 ± 91.1	40	1351.1 ± 123.0	69	1242.9 ± 152.4	20	1782.7 ± 114.8	30

* All of events have the same lapse time, 70 sec.

† N represents the number of fit with corelation coefficient equal or greater than 0.8 (1) $Q_c = (102.0 \pm 7.3)\, f^{(0.83\pm0.03)}$ correlation coefficient = 0.9986; (2) $Q_c = (112.5 \pm 8.4)\, f^{(0.85\pm0.03)}$ correlation coefficient = 0.9985; (3) $Q_c = (109.8 \pm 8.3)\, f^{(0.81\pm0.04)}$ correlation coefficient = 0.9970; (4) $Q_c = (135.3 \pm 6.4)\, f^{(0.89\pm0.03)}$ correlation coefficient = 0.9999.

on frequency for each of the four subsets. A Q_c dependence on lapse time can also be inferred by the plots. The data fit the empirical relationship (1) well for the four subsets showing an increase of Q_o with increasing lapse-time, while n remains substantially unchanged.

Roecker *et al.* (1982) and Gagnepain-Beyneix (1987) interpreted similar results, obtained respectively in Hindu–Kush and France, as due to variations of attenuation with depth. The same results can be partly interpreted as due to multiple-scattering effects affecting the late coda (Gao *et al.*, 1983). A check for the present interpretation of the lapse-time dependence as a depth effect has been performed grouping the selected events according to their focal depth. Four classes of events with hypocentral depth ranging respectively from 0 to 10, 10 to 20, 20 to 30, and 30 to 45 km have been chosen. Because of the above mentioned dependence of Q_c on lapse time, all coda envelopes were chosen in the lapse time of 70 sec, because in

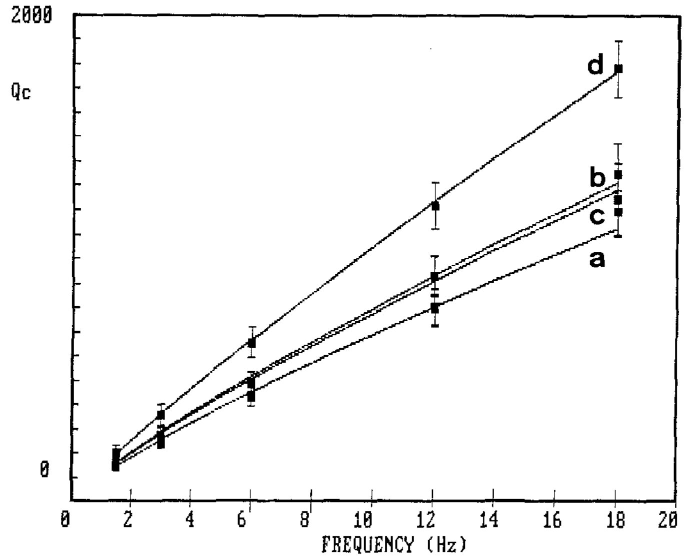

FIG. 5. Representations of Q_c versus frequency for four different sets of earthquakes. Curve (a) represents events with focal depth ranging from 0 to 10 km, (b) from 10 to 20 km, (c) from 20 to 30 km, and (d) from 30 to 45 km. All of the events have the same lapse time, 70 sec.

this interval the maximum number of data was available. Results are listed in Table 3 and plotted in Figure 5. Data fit the empirical relationship for each of the four sets of events in a clear manner. No variations of n parameter are evident, while substantial changes in the parameter Q_o can be observed. Q_o increases with depth. This result shows that there is an increase of Q_c with the depths of the volumes over which Q_c is averaged.

We remark that the lapse time was taken to be approximately constant for each of the four subsets of events, in such a way that, if present, multiple scattering would affect the data at each depth range in the same manner. We point out once more that the estimates of coda Q are insensitive to hypocentral distance correction. Checks were made using Sato's expression (3) obtaining the same results.

Discussion and Conclusions

The physical meaning of the coda Q parameter in the short-period range has been extensively revised during the last 10 years. At present, two main points seem to be important in the development of experimental studies on coda waves: (a) the lapse-time dependence of coda Q, and (b) the site of dependence of coda Q.

Point (a) is inconsistent with the random but homogeneous distribution of scatterers in a medium with a consistent intrinsic attenuation. In such a case, the lapse-time dependence could be generated if the distribution of scatterers is not homogeneous (for example, fewer scatterers at deeper depths), if the intrinsic Q of the medium is not constant, or if multiple scattering occurs in the late coda.

Point (b) has been recently shown (Phillips and Aki, 1986; Steck *et al.*, 1989). One of the possible explanations is that this site effect, which is most present in the early coda, is due to a combined effect of near site attenuation and multiple scattering.

In the present paper, data show a visible dependence of coda Q on the depth of events. As coda of deeper events give information of a deeper portion of the lithosphere than the shallower ones, we infer from this result that attenuation is actually depth dependent. We exclude multiple-scattering effects because, having taken the same lapse-time values for events having increasing depth, the multiple scattering should affect all the coda envelopes in the same way. The values of Q_c obtained by means of coda-wave analysis contain averaged information of a volume of earth sampled by the scattered waves. If we accept the single-scattering model, this volume is defined by a revolution ellipsoid with focus in the hypocenter and in the recording station. The length of the larger semiaxis is defined by $v \cdot t/2$ (Lee *et al.*, 1986), where v is the velocity of the waves which form the coda, between 3 and 3.5 km/sec, and t the lapse time of the analyzed coda. The width of this ellipsoid will be defined by the smaller semiaxis.

The variation of Q_c with the lapse time can thus be explained by comparing the different volumes sampled for different lapse times. If an average velocity of 3.2 km/sec is used for the S waves (Ibañez *et al.*, 1988) for a lapse time of 30 sec, the length of the larger semiaxis is approximately 50 km, and 145 km for the lapse time of 90 sec, clearly indicating the different sampling volumes for each one. Comparing the volume sampled by the scattered waves for a fixed lapse time (70 sec) for seismic sources located at different depths, we observe that, for the set of events at greater depth, the larger semiaxis is about 150 km long, decreasing to about 110 for superficial earthquakes. Clearly, the effects of multiple scattering will shorten these volumes.

We remember that the site effects were minimized, eliminating from the analysis the coda envelopes showing reverberations, which are presumed to be generated by the presence of a sedimentary basin. This minimization of the site effects has been carried out mainly in the selection of information according to the correlation coefficient obtained when applying the expression (5).

Our conclusions are that attenuation is depth dependent in the Granada Basin zone and that the lapse-time dependence of coda Q can be attributed to a variation of attenuation with depth. Canas *et al.* (1988) found evidence of variation of Q_β with the depth in a wider region (Iberian Peninsula) which includes the same, using Rayleigh waves, which is in accordance with the results obtained in this study for Q_c. Steck *et al.* (1989) have showed that the single scattering is inappropriate for early coda when site effects are important, as in the Mono Craters area, California. These authors interpret the coda Q inferred by the early coda as a near site Q. As we have eliminated the main site effects in the used data and, in addition, we averaged the coda Q over the stations, we reasonably can feel that the Q calculated using the early portion of the coda is a typical value of attenuation in the upper crust. As previously mentioned, geological and DSS data indicate an upper mantle as shallow as about 25 km beneath the area under study. This evidence supports our conclusion of a depth-dependent attenuation in the Granada Basin.

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INSTITUTO ANDALUZ DE GEOFÍSICA Y PDS
OBSERVATORIO DE CARTUJA
UNIVERSIDAD DE GRANADA
18071 GRANADA SPAIN
(J.M.I., F.D.M., G.A., J.M.)

ISTITUTO DI SCIENZE DELLA TERRA
UNIVERSITÁ DI CATANIA
CORSO ITALIA 55
95129 CATANIA ITALY
(E.D.P.)

CÁTEDRA DE GEOFÍSICA
FACULTAD DE CIENCIAS FÍSICAS
UNIVERESIDAD COMPLUTENSE
28040 MADRID, SPAIN
(M.H.)

INSTITUTO DE CIENCIAS DE CONSTRUCCIÓN
"EDUARDO TORROJA" (CSIC)
SERRANO GALVACHE S/N
28030 MADRID SPAIN
(M.H.)

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