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Evidence for Stability in Coda Q Associated with the Egion (Central Greece) Earthquake of 15 June 1995

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Abstract One hundred seventy-two seismograms, recorded at one seismological station located close to the source region of the Egion ($Ms = 6.2$) 15 June 1995 earthquake, were used to measure values of Q from the decay of the earthquake coda. The collected data were compared between events that occurred before and after the main event. The analysis showed no significant temporal variation in the value of coda Q for the region.

Introduction

Various observations in different regions of the world have indicated a striking difference in the rate of amplitude decay between tectonically active and static regions. The decay rate of coda amplitude is in general a very stable parameter common to all local earthquakes located near a given station and is insensitive to source and receiver locations within the region, reflecting the average properties of the crust.

Many researchers have reported temporal changes of the coda attenuation parameter Q_c^{-1} before and after large earthquakes. Table 1 depicts some of these observations.

Recently, Tselentis (1993) reported a characteristic decrease in Q_c^{-1} after two medium earthquakes in central Greece. This stimulated our interest for a more detailed investigation of the relation between seismic attenuation and seismicity in the tectonically active region of central Greece.

In this region, the seismological laboratory of Patras University has established a permanent microearthquake network, and in addition to the seismic activity, it also monitors continuously a range of physical parameters, well known as earthquake precursors (Tselentis and Ifantis, 1996), such as the Earth's electric field, electromagnetic anomalies, groundwater temperature, and chemistry and acoustic emission.

A large earthquake of magnitude 6.2 M_L occurred on 15 June 1995 at 00:15 GMT in the western end of the Gulf of Corinth (Fig. 1) within the eastern region covered by the network. This earthquake was followed 15 minutes later by a 5.4 M_L aftershock and a rich aftershock sequence.

This large body of data provides a unique opportunity to examine the temporal variation of seismic attenuation before and after the mainshock.

Table 1
Observed Temporal Changes in Seismic Attenuation (Q^{-1}) Associated with Earthquake Activity

Author	Region	Variation of Q^{-1}
Wyss (1985)	Hawaii	30% increase before Songpan 7.2 earthquake
Gusev and Lemzikov (1985)	Kuril-Kamchatka	20% increase before three 8 earthquakes
Novelo *et al* (1985)	Mexico	30% increase before one 7.6 earthquake
Tsukuda (1985)	Japan	15% increase before one 6.2 earthquake
Sato (1986)	Central Japan	High Q^{-1} before one 6.0 earthquake
Jin and Aki (1986)	China	300% increase before Tangshan 7.8 earthquake
Jin and Aki (1986)	China	300% increase before Haicheng 7.3 earthquake
Wang *et al.* (1989)	E. Taiwan	Increase after one 6.1 earthquake
Lee *et al.* (1986)	California and Nevada	Variation in Q_c before large earthquakes
Scherbaum and Kisslinger (1985)	Aleutian	No variation
Robinson (1987)	New Zealand	Q_c decreased when the number of all events relatively to the large increased
Peng *et al.* (1987)	California	Increase after one 5.7 earthquake
Jin and Aki (1989)	California	Q_c increased when the number of small events relatively to the large increased
Tselentis (1993)	Central Greece	Decrease after two earthquakes
Aster *et al.* (1996)	California	No variation
Antolik *et al.* (1996)	California	No variation

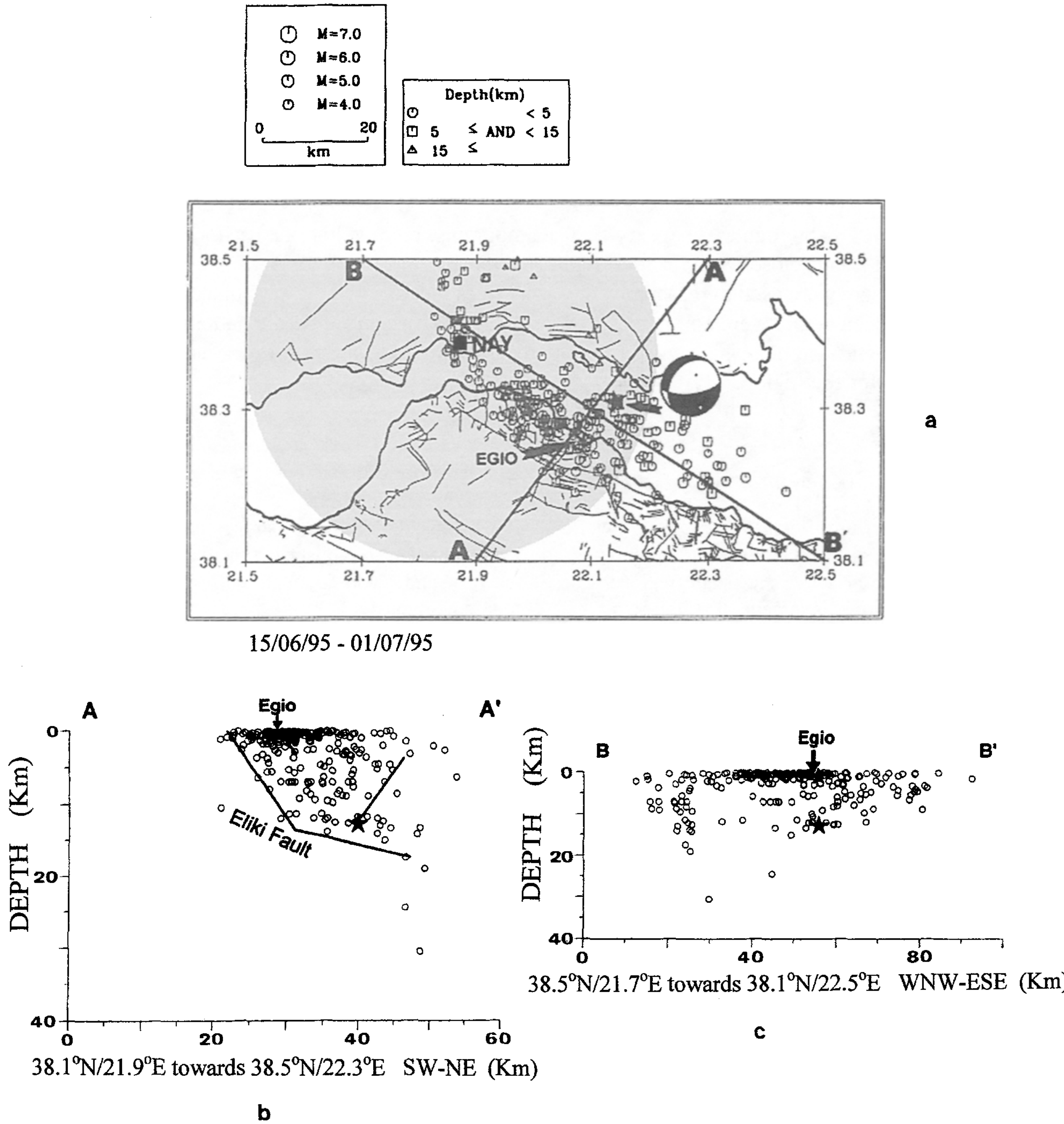

Figure 1. (a) Spatial distribution of the best located aftershocks during the first 17 days of the aftershock sequence and focal mechanism of the main event, (b) AA′ SW–NE and (c) BB′ WNW–ESE cross section. The big star denotes the main event (Tselentis *et al.*, 1996). Shadow area corresponds to the outer circle of Figure 2.

Data

The present investigation is based on data recorded by station NAY (Fig. 1) of the Patras University seismological network (PATNET). This station is located within the western part of the aftershock region and was operating at the time of the earthquake. It is equipped with a vertical-component 1-Hz seismometer operating at 60-dB dynamic range in a low-noise environment. The recorded signals are radio-linked using FM subcarriers to the central recording site at Patras University, where they are antialias filtered with a 30-Hz Butterworth low-pass filter, sampled at 100 Hz and converted to digital form with a resolution of 32 bits.

To minimize the effects of the source function and path propagation on the Q parameter, we have confined the events used in this study to be in the magnitude range of $2.5 < M_L < 3.0$ and within a cubic volume approximately 30 km on each side. The relative location of the recording station and

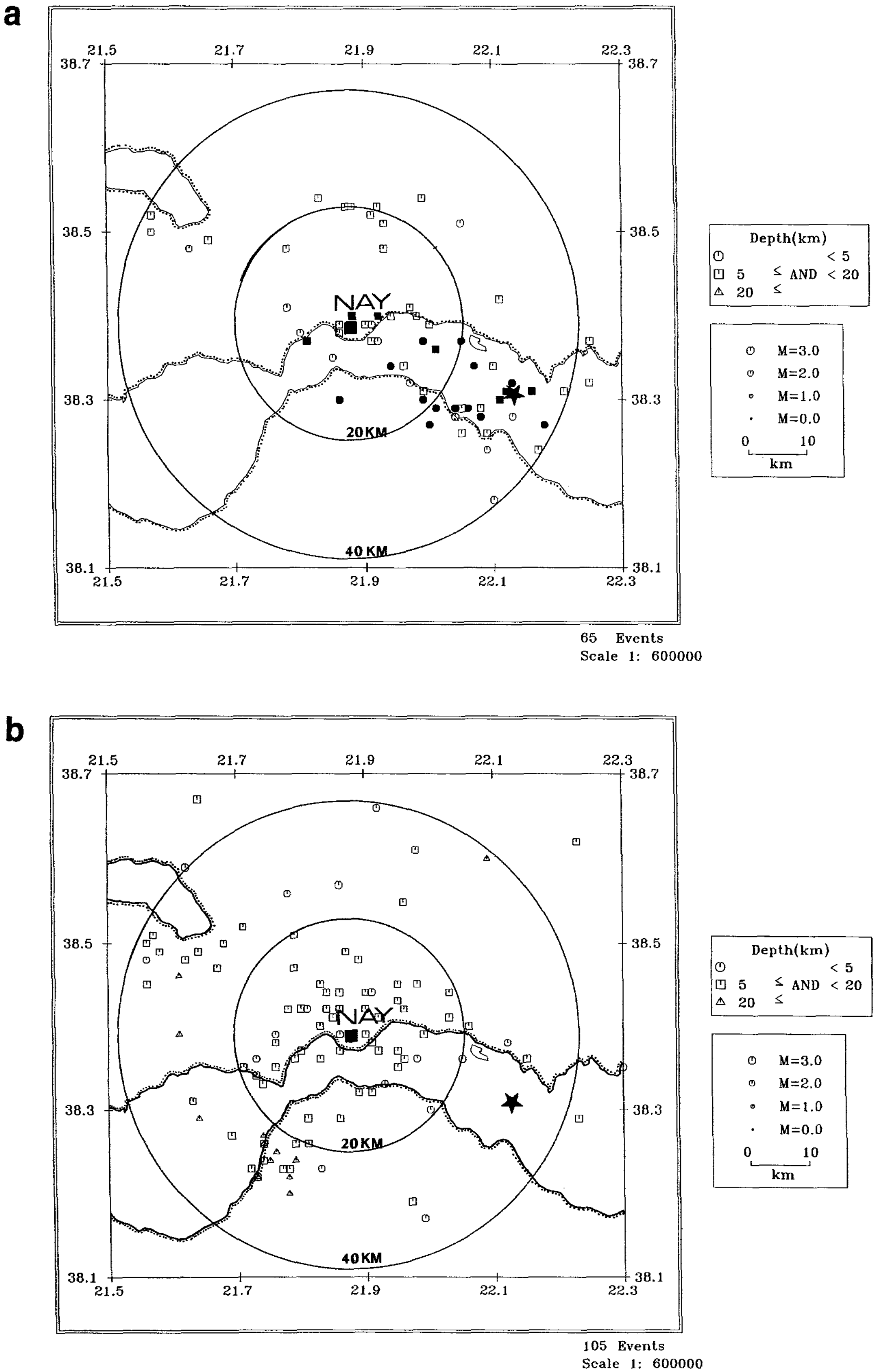

Figure 2. Epicentral distribution of the (a) aftershocks, (b) prior to the main event earthquakes used. Black symbols in (a) correspond to the events shown in the insert of Figure 3. Star denotes the main event.

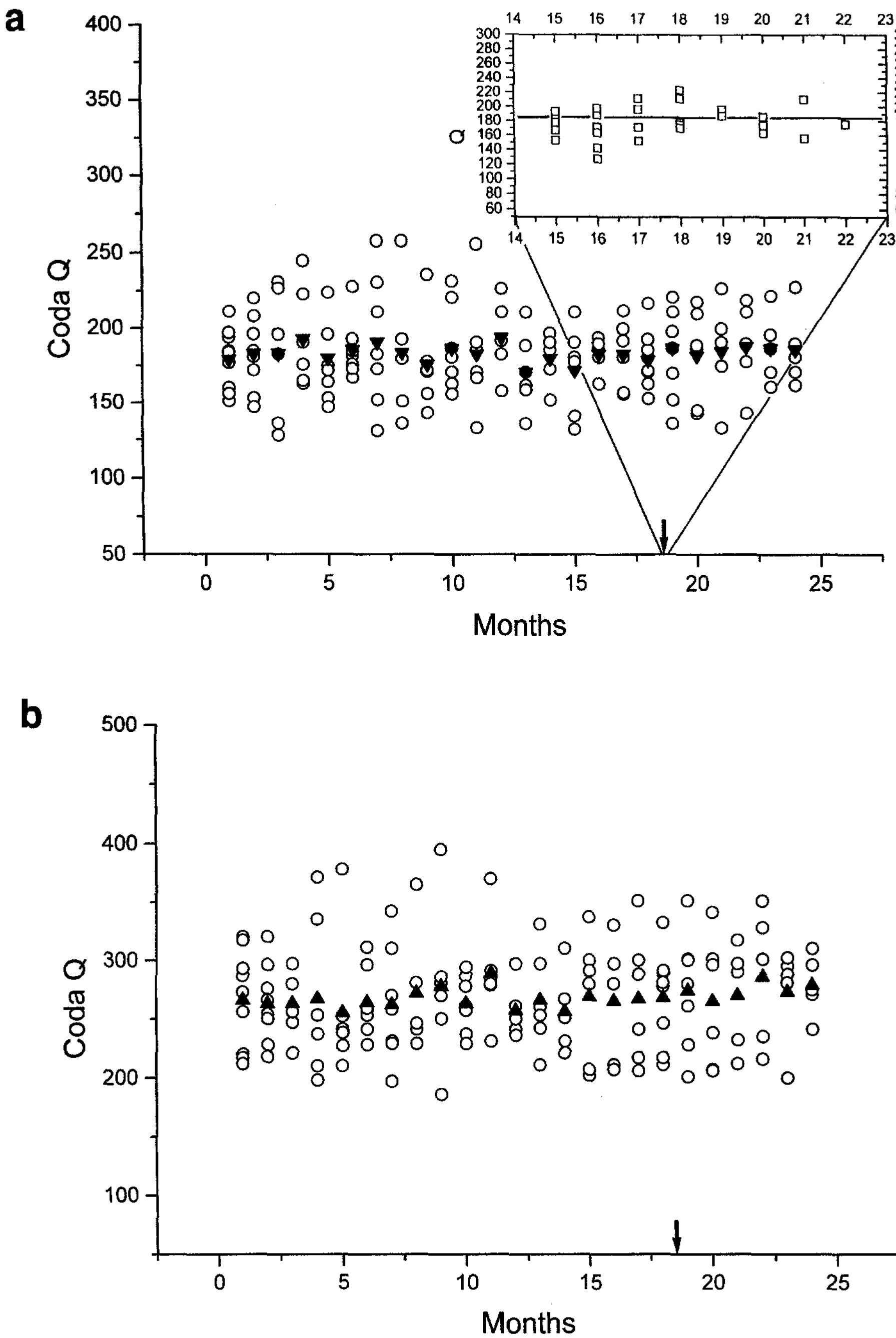

Figure 3. Temporal variation of coda Q before and after the mainshock (a) for 4 Hz and (b) 8 Hz central frequencies. Insert in (a) depicts the corresponding coda Q variation within the first week after the mainshock. Black triangles denote monthly averages.

all source points make the corresponding ellipsoidal sampling volumes (Pulli, 1984) largely overlapping, and thus the obtained Q^{-1} values should be closely related to attenuation of the medium close to the source region.

The mainshock, as located by PATNET, is placed at 38°18′54″ N, 22°08′46″ E and at a depth of 12.8 km. A large number of aftershocks followed, and the principal spatial characteristics of the best located aftershocks (Tselentis *et al.*, 1996) are illustrated in Figure 1. The distribution of hypocenters in cross section does not immediately suggest a planar distribution but rather defines a volume about 15 km (depth) by 35 km (NW–SE) and by 20 km (NE–SW).

The above region defines the sampling volume from which coda Q information is sought; the narrow magnitude range confines the source excitation to an approximately uniform frequency band (Wang *et al.*, 1989).

Sixty-five earthquakes in the magnitude range $2.5 < M_L < 3.0$ (Fig. 2a) that occurred after the main event have been selected for deriving the coda Q parameter. In addition to the above data set, we have also collected 106 events in the same region for a period of 17 months preceding the mainshock (Fig. 2b). Normally, waves in the interval 20 to 30 sec after the S arrival have been used to determine the decay rate.

Method

The coda Q values were obtained as follows (Woodgold, 1994). Three-pole Butterworth filters are applied with center frequencies at 2, 4, 8, and 12 Hz. The envelopes of the filtered traces are averaged over 5-sec sliding windows with the beginning of the windows 2 sec apart. The envelope is calculated from

$$A_e(t) = \sqrt{A^2(t) + H^2(t)}, \tag{1}$$

where A is the seismic amplitude at time t and $H(t)$ the Hilbert transform of $A(t)$. We also averaged envelopes in a noise window, and the signal amplitude is estimated as

$$A_s = \sqrt{A_T^2 - A_N^2}, \tag{2}$$

where A_T is the averaged envelope for the window and A_N is the averaged envelope for the noise. Following Woodgold (1994), the function

$$F(t) = \log_{10}\left[\sqrt{\frac{t}{K''(a)}}\,A_s\right] \tag{3}$$

is calculated, and for each event-station-frequency combination, a regression line is fit to $F(t)$, where t is the lapse time at the center of the 5-sec window, $\alpha = t/t_s$, t_s is the lapse time of s-wave arrival, and

$$K''(a) = \sqrt{1 - (1/a^2)}. \tag{4}$$

Finally, Q_c is calculated as

$$Q_c = \log_{10} e\pi f/s, \tag{5}$$

where s is the slope from the regression and f is the frequency in Hertz.

Results

Temporal variations of coda Q values for earthquakes occurring between January 1994 and December 1995 and for frequencies 4 and 8 Hz are depicted in Figures 3a and 3b, respectively. In the same diagram, we present also the corresponding monthly averages.

Judging from these diagrams, we cannot assess any significant change in Q_c. The overall means of the measured coda Q are similar before ($Q_{f=4\text{Hz}} = 183$, $Q_{f=8\text{Hz}} = 266$) and after ($Q_{f=4\text{Hz}} = 184$, $Q_{f=8\text{Hz}} = 273$) the main event.

In a similar research, Wang *et al.* (1989) observed a significant drop of coda Q immediately after the Hualien 1986 earthquake in E. Taiwan. This drop lasted approximately two days before returning to ambient level. To investigate for a similar phenomenon in our case, we calculated the coda Q corresponding to 30 aftershocks occurring within 1 week of the main event. The results are also presented as an insert in Figure 3a. No drop in the value of Q_c can be seen.

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