Evidence for Stability in Coda Q Associated with the Egion (Central Greece) Earthquake of 15 June 1995
by G-Akis Tselentis

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^-1 before and after large earthquakes. Table 1 depicts some of these observations.

Recently, Tselentis (1993) reported a characteristic decrease in Q^-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

<table>
<thead>
<tr>
<th>Author</th>
<th>Region</th>
<th>Variation of Q^-1</th>
</tr>
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<tbody>
<tr>
<td>Wyss (1985)</td>
<td>Hawaii</td>
<td>30% increase before Songpan 7.2 earthquake</td>
</tr>
<tr>
<td>Gusev and Lemzikov (1985)</td>
<td>Kuril-Kamchatka</td>
<td>20% increase before three 8 earthquakes</td>
</tr>
<tr>
<td>Novelo et al (1985)</td>
<td>Mexico</td>
<td>30% increase before one 7.6 earthquake</td>
</tr>
<tr>
<td>Tsukuda (1985)</td>
<td>Japan</td>
<td>15% increase before one 6.2 earthquake</td>
</tr>
<tr>
<td>Sato (1986)</td>
<td>Central Japan</td>
<td>High Q^-1 before one 6.0 earthquake</td>
</tr>
<tr>
<td>Jin and Aki (1986)</td>
<td>China</td>
<td>300% increase before Tangshan 7.8 earthquake</td>
</tr>
<tr>
<td>Jin and Aki (1986)</td>
<td>China</td>
<td>300% increase before Haicheng 7.3 earthquake</td>
</tr>
<tr>
<td>Wang et al. (1989)</td>
<td>E. Taiwan</td>
<td>Increase after one 6.1 earthquake</td>
</tr>
<tr>
<td>Lee et al. (1986)</td>
<td>California and Nevada</td>
<td>Variation in Q_c before large earthquakes</td>
</tr>
<tr>
<td>Scherbaum and Kisslinger (1985)</td>
<td>Aleutian</td>
<td>No variation</td>
</tr>
<tr>
<td>Robinson (1987)</td>
<td>New Zealand</td>
<td>Q_c decreased when the number of all events relatively to the large increased</td>
</tr>
<tr>
<td>Peng et al. (1987)</td>
<td>California</td>
<td>Increase after one 5.7 earthquake</td>
</tr>
<tr>
<td>Jin and Aki (1989)</td>
<td>California</td>
<td>Q_c increased when the number of small events relatively to the large increased</td>
</tr>
<tr>
<td>Tselentis (1993)</td>
<td>Central Greece</td>
<td>Decrease after two earthquakes</td>
</tr>
<tr>
<td>Aster et al. (1996)</td>
<td>California</td>
<td>No variation</td>
</tr>
<tr>
<td>Antolik et al. (1996)</td>
<td>California</td>
<td>No variation</td>
</tr>
</tbody>
</table>
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 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-SEE cross section. The big star denotes the main event (Tselentis et al., 1996). Shadow area corresponds to the outer circle of Figure 2.
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.
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 $38^\circ18'54''$ N, $22^\circ08'46''$ 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

$$A_s(t) = \sqrt{A^2(t) + H^2(t)},$$

(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

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

(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

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

(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

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

(4)

Finally, $Q_c$ is calculated as

$$Q_c = \log_{10} (\pi f/\nu),$$

(5)

where $\nu$ 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 $(Q_f=4Hz = 183, Q_f=8Hz = 266)$ and after $(Q_f=4Hz = 184, Q_f=8Hz = 273)$ 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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