Relationship between the 13 May 1995 Kozani–Grevena (NW Greece) earthquake and the Polyphyto artificial lake

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Received 3 December 1997; accepted 13 July 1998

Abstract

On 13 May 1995 a strong earthquake of $M_s = 6.6$ struck the cities of Kozani and Grevena in northwestern Greece. This region is characterized by low seismicity. In the same area, three hydroelectric dams have been operating for the last 30 years. One of them, the Polyphyto dam, is located only 40 km from the epicenter of the 13 May earthquake.

In the present work all available seismological aspects of the main event, such as focal properties and source parameters together with foreshock and aftershock characteristics of the earthquake sequence are considered in order to examine whether the water level changes behind the Polyphyto dam induced the unusual seismic activity in this area. Thus, a detailed examination of the seismic activity is made and this is compared to the seismotectonic regime of the region and the reservoir loading from 1976 to 1995. The results show that there is no obvious correlation between seasonal or sharp fluctuations in the water level and the seismicity of the region (except once during 1989). Moreover, comparison with other previous cases of induced seismicity in Greece and in other countries shows no similarities to the Kozani–Grevena earthquake sequence.

It is concluded that the Kozani–Grevena earthquake is therefore an event in the framework of the regional seismicity rather than an event triggered by the impounding of the Polyphyto artificial Lake. © 1998 Elsevier Science B.V. All rights reserved.

Keywords: Greece; Induced seismicity

1. Introduction

The energy release during an earthquake is the result of complex geophysical processes in the Earth’s interior. Sometimes, external processes trigger earthquakes. Such a triggering effect is the impounding of artificial lakes. Over the last 30 years much work has been done on the reservoir-induced seismicity and several examples of the association of earthquakes with artificial lakes are now available (Gupta et al., 1972a,b; Papazachos, 1973; Drakopoulos, 1974; Simpson, 1976; Zoback and Hickman, 1982; Simpson et al., 1985; Ohtake, 1986; Ibenbrahim et al., 1989; Kebeasy and Gharib, 1992; Rajendran, 1992; Knoll et al., 1996; Sobolev et al., 1996; Yungo et al., 1996). In most cases the main shock was shallow and the fore- shock activity initiated or increased considerably after the impounding of the reservoirs. The epicen-
ters were mostly located within a distance of 25 km from the lakes. Among the factors affecting the seismic activity are the rate of the water level increase, the duration of the loading, the maximum levels reached and the time period for which the high levels are retained.

Two processes of stress modification have been suggested as the dominant mechanisms that are possibly responsible for triggering of earthquakes by large reservoirs (Simpson et al., 1988): (1) the direct effect of loading, through increased elastic shear stress; (2) the effect of increased pore pressure, through decreased effective normal stress.

On 13 May 1995, a strong earthquake, $M_s=6.6$, occurred in the northwestern part of Greece (Fig. 1) and caused serious damage in the Kozani-Grevena region. The maximum observed macroseismic intensity was IX+ on the MM scale. The main shock was preceded by several foreshocks and followed by intense aftershock activity. The epicenter of the main shock is located ca. 40 km from the Polyphuto artificial lake (FD) where a dam was constructed by the Public Power Corporation (PPC) in the early 1970s (Fig. 1).

In this study we investigate the foreshock and aftershock activity of this earthquake as well as its focal mechanism and source parameters. An attempt is also made to examine whether any relationship exists between the occurrence of the main shock and the variations of the water level in the reservoir.

2. Seismotectonic regime and seismicity of the area

Greece is part of the convergence zone between the Eurasia and the African lithospheric plates. In the Greek mainland earthquake result mainly from north-south crustal extension and associated east-west normal faulting (Mckenzie, 1978; Angelier, et al., 1982; Papadopoulos et al., 1986; Papazachos et al., 1986; Pavlides and Moutrakis, 1987; Armi), et al., 1992).

The basement of the Kozani-Grevena area belongs to the Pelagonian geotectonic zone and consists of Pre-Alpine and Alpine rocks like gneisses, and schists, covered by limestones and flysch. Over these units the Vourinos massif exists formed mainly by ophiolites NW-SE in direction. On both sides of Vourinos two large basins predominate NW-SE in direction. The Meso-Hellenic basin is located at the west and the Kozani basin at the east, both being filled with molasic sediments conglomerates, sandstones, marls and silts. Over them lacustrine, fluvial and torrential deposits of Neogene age and Quarternary deposits of talus cones and extended fluvial terraces are present.

From a morphotectonic point of view the area is dominated by a NE-SW striking zone of normal faulting. It consists of three segments (Figs. 1 and 2) with a total length of ca 80 km bounding the southern shore of Lake Kozani as well as the southern shore of the Polyphuto artificial lake. The fault zone cuts almost perpendicularly the Vourinos massif structure as well as the two basins. The Servia fault, with a trend of N 60° and a dip of 60° toward the NW (Figs. 1 and 2), constitutes the northern segment of the zone and the most prominent feature of the area. At the base of the fault a clear scarp is present, which suggests that the fault was active at least during the last 10 000 years (Armijo et al., 1992). After the large earthquake of 13 May 1995, a surface rupture was observed along a preexisting normal fault ca 10 km to the northwest of the major Servia fault. It had a length of 8 km with a normal slip of 2.4 cm (Pavlides et al., 1995; Papanastassiou et al., 1998).

According to archives of monasteries located in the area, during the last three centuries three earthquakes occurred at the area of Servia, in 1695, 1766 and 1852. They caused extensive damage in the city of Servia and the surrounding villages (Papanastassiou et al., 1998). During the present century three strong earthquakes were recorded close to the epicenter of the 1995 main shock (Fig. 1). They occurred on 7 December 1922 (40.01°N-21.31°E, $M_s=5.5$), 25 March 1943 (40.41°N-21.89°E, $M_s=5.3$) and 25 October 1984 (40.11°N-21.62°E, $M_s=5.6$) (Makropoulos et al., 1989; Papazachos and Papazachou, 1997). So, all information suggests that the area under investigation could not be considered as of low seismicity but as an area of sparse seismological information.
3. The 13 May 1995 earthquake

The main shock occurred at 08:47 GMT. It was preceded by some significant foreshocks of $3.5 < M_c < 4.5$, the last of them taking place ca 15 s before the main event. The relocated position for the main shock (Fig. 1) was 40.12°N–21.67°E with a focal depth of 15 km. The relocated positions of the foreshocks were placed to the south of the main shock with depths < 16 km. The CMT scalar moment given by Harvard was $7.6 \times 10^{18}$ Nm.
Immediately after the main shock, the Institute of Geodynamics deployed local network in the region using ten stations. The aftershock activity was continuously monitored for 50 days from 14 May to 4 July. The activity was intense during the first week when >1000 events of \( M_L \geq 1.5 \) were recorded per day (Papanastassiou et al., 1998). The majority of the events were located to the south of the main shock and north–northwest of the observed surface rupture (Fig. 2). On both ends of the observed surface rupture two clusters of activity were observed. The cluster located at the western end was larger and well defined. The other one had a direction towards the northeast while its spatial distribution was diffused. At the eastern end, just south of the village of Paleochori, there was a smaller group of aftershocks following a linear trend NW–SE in direction (Fig. 2).

On a cross section (Fig. 3) perpendicular to the fault trace, the main cluster appears as a clear zone located north of a fault plane dipping to the northwest which reaches the surface at the mapped location. The activity was intense during the first week when >1000 events of \( M_L \geq 1.5 \) were recorded per day (Papanastassiou et al., 1998). The majority of the events were located to the south of the main shock and north–northwest of the observed surface rupture (Fig. 2). On both ends of the observed surface rupture two clusters of activity were observed. The cluster located at the western end was larger and well defined. The other one had a direction towards the northeast while its spatial distribution was diffused. At the eastern end, just south of the village of Paleochori, there was a smaller group of aftershocks following a linear trend NW–SE in direction (Fig. 2).

4. Focal mechanism and source parameters

By using P-wave polarities from the permanent network of Institute of Geodynamics and others...
provided by international agencies, the fault plane solution for the main shock was determined. The solution (Fig. 1) indicates pure normal faulting on planes N240° and N72° in direction, dipping 35° to the NW and 56° to the SE, respectively. The first plane coincides with the geometry of the observed faulting. The CMT solution determined by Harvard also suggests a pure normal faulting. The trends and dips of the first and second planes are N240°/31° NW and N70°/59° SE, respectively, which are almost identical with the mechanism determined in this study.

From the spectral analysis of digital short-period data Chouliaras and Stavrakakis (1997) obtained the values in Table 1 for the source parameters. Meyer et al. (1996) using Synthetic Aperture Radar (SAR) interferometry modeling to characterize the coseismic displacement field concluded that the large displacement did not reach the surface and was restricted to a depth of between 4 and 15 km.

Table 1

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seismic moment</td>
<td>$M_0 = 2.53 \times 10^{26}$ dyne cm$^{-1}$</td>
</tr>
<tr>
<td>Corner frequency</td>
<td>$f_c = 0.06$ Hz</td>
</tr>
<tr>
<td>Fault radius</td>
<td>$L = 26.1$ km</td>
</tr>
<tr>
<td>Stress drop</td>
<td>$\Delta = 6.26$ bar</td>
</tr>
<tr>
<td>Mean displacement</td>
<td>$u = 39.4$ cm</td>
</tr>
</tbody>
</table>

5. Examination of the seismic activity in comparison to the reservoir loading

In order to investigate if there is any correlation between the seismic activity of the region and the reservoir loading, we examined:
(1) the details of the seismicity before the reservoir loading; and
(2) the temporal and spatial distribution of the seismic activity compared with other characteristic cases of induced seismicity.

The Polyphyto artificial lake was constructed by the PPC in the early 1970s. The maximum water depth is 95 m just behind the dam, while at the other end of the lake the maximum water depth is only 10 m. The maximum volume of the water in the reservoir is almost $1.3 \times 10^9$ km$^3$. In 1976, the PPC installed in the region a local seismological network to monitor the microseismic activity. Therefore, in the frame of our analysis the data from the PPC network were included. Fig. 4 illustrates the epicenter distribution of all earthquakes recorded by the PPC network from 1976 to May 1995. The spatial distribution shows that there are two regions of intense seismicity at the southwestern part of the Polyphyto artificial lake.

In order to investigate if there is any relation between the water depth fluctuations in the Polyphyto artificial lake and the local seismicity...
before the occurrence of the main shock, the variation of the water depth was examined against the variation in the seismicity from 1976 to May 1995.

A visual inspection of the diagrams in Fig. 5 leads to the following results:

1. The seismic activity was low in the interval 1976–1983. The average number of earthquakes was about ten events per month, while some fluctuations in activity occurred.

2. During 1984 an increase of the number of earthquakes took place, while at the same time the water depth in the reservoir varied from 80 to 92 m.

3. In the same year, the largest earthquake occurred in the region. Its magnitude was $M_s = 5.4$ and the intensity felt was of VI (MM).

4. From 1985 to 1986 the water level did not change significantly but at the same time the number of microearthquakes decreased.

5. By the beginning of 1989 the water depth decreased to its lowest level of 78 m and 7 months later the number of microearthquakes increased to its highest value at ca 19 earthquakes per month.

6. Since then, the water depth fluctuates 'regularly' between 82 and 93 m, while at the same time the occurrence of microearthquakes does not seem to be affected by fluctuations in the water level.

7. Before and after the large earthquake of 13
microearthquakes is rather low in the area of Polyphyto Dam.

6. Comparison to previous cases of induced earthquakes

Microseismic data are not available for the period before the construction of Polyphyto Dam. Therefore it is difficult to judge whether or not the activity from 1976 onwards has been affected by the impounding of the water in the reservoir.

According to Simpson et al. (1988) the seismic activity at the reservoirs suggests that two primary types of induced seismicity are observed:

(1) a rapid response, in which the seismicity follows immediately on first loading of the reservoir, consisting primarily of low magnitude swarm-like activity, is confined to the immediate reservoir area, and is closely correlated with changes in water level in the reservoir; and

(2) a delayed response, in which the seismicity follows with a significant delay after first filling, is often associated with large magnitude earthquakes, may extend significantly beyond the confines of the reservoir, and may not show an immediate correlation with major changes in the reservoir level.

Often, the reservoir water level may go through a number of apparently similar annual cycles in the water level between first impounding and the onset of significant seismicity.

We compare the 1995 Kozani–Grevena case to the Kremasta (Greece) earthquake, which has been documented. In 1966 an earthquake of $M_s = 6.3$ occurred in the region of the Kremasta artificial lake. It was preceded by about 740 foreshocks of magnitude $2.4 \leq M_s \leq 4.6$ and was followed by $> 2600$ aftershocks (Comninakis et al., 1968). In Fig. 6 the foreshock–aftershock pattern of the Kremasta earthquake is shown (Gupta et al., 1972a). They classified this pattern to the type II model of foreshock–aftershock sequences according to Mogi (1963). In the case of the 1995 Kozani–Grevena earthquake, only a few foreshocks with a magnitude $3.5 < M_s < 4.5$...
In the case of the 1967 Koyna and 1975 Oroville the main event occurred 5 and 7 years after filling of the reservoir, respectively (Simpson et al., 1988). In both cases, however, the reservoir water level goes through a number of apparently similar annual cycles. In the case of Koyna reservoir the seasonal variation in water level, from 15 to 50 m, were rather sharp. Gupta (1983) identified a rate of loading exceeding 12 m per week as a necessary but not a sufficient condition for earthquakes to occur in the vicinity of the Koyna Dam. In the case of Oroville the seasonal variations were <25 m. But during a drought in 1974–1975, the water level dropped almost 50 m and the possible connection of the 1975 earthquake with this rapid change in water level has been discussed by Toppozada and Morrison (1982). In the case of the 1995 Kozani–Grevena earthquake the seasonal variation in the water level was not so sharp.

In the case of Aswan, the main earthquake (1981, $M = 5.3$) occurred 17 years after the first impounding of the reservoir and after a number of annual cycles. Also in this case the fluctuations of the water level were very sharp. From 1964, the year of the first impounding, till 1981 when the largest earthquake occurred the water level fluctuations were >50 m.

In the case of the 1995 Kozani–Grevena earthquake the seasonal variations of the water level were not so sharp. Therefore, even the case of delayed response induced seismicity could not be considered as in the case of the Kozani–Grevena earthquake.

As a follow-up we examined the cases of delayed response. In the cases of 1967 in Koyna and 1975 in Oroville the main event occurred 5 and 7 years after filling of the reservoir, respectively (Simpson et al., 1988). In both cases, however, the reservoir water level goes through a number of apparently similar annual cycles. In the case of Koyna reservoir the seasonal variation in water level, from 15 to 50 m, were rather sharp. Gupta (1983) identified a rate of loading exceeding 12 m per week as a necessary but not a sufficient condition for earthquakes to occur in the vicinity of the Koyna Dam. In the case of Oroville the seasonal variations were <25 m. But during a drought in 1974–1975, the water level dropped almost 50 m and the possible connection of the 1975 earthquake with this rapid change in water level has been discussed by Toppozada and Morrison (1982). In the case of the 1995 Kozani–Grevena earthquake the seasonal variation in the water level was not so sharp.

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7. Discussion and conclusions

In seeking induced earthquakes caused by the impounding of reservoirs, two different questions should be answered.

1. Is there any increase in the number of earthquakes in the region after the filling of the reservoir?
2. Is there any correlation between large earthquakes and seasonal or sharp fluctuations in the water level?

In the case of the 13 May 1995 Kozani–Grevena earthquake, it is not possible to answer with confidence the first question due to the lack of seismological data in the region before the construction of the dams, although comparison with other cases of induced seismicity shows that no similarities exist in such situations. In order to make clear that the seismic activity related to the 13 May 1995 earthquake shows no similarities to previous cases of induced earthquakes, in Fig. 7 the seismic activity is plotted several months before and after the main earthquake. In the same figure, the fluctuations of the water depth are plotted for the same time period. Comparison with the seismic activity of the Kremasta induced earthquake (Fig 6) shows that in the case of the Kozani–Grevena earthquake the foreshock activity was low. Moreover, the rate of decrease of aftershock activity is also low. The activity lasted 2 years after the occurrence of the main shock and includes aftershocks with fairly large magnitudes. In the same period, there was no significant fluctu-
A change of this order in the water load could affect the local microseismicity, but may not trigger a large earthquake. Only once, in 1989, 7 months after the sharpest decrease of the water depth (ca. 15 m) did the number of microearthquakes in the region increase to its highest value. Hence, there is no clear correlation between the regional seismicity and the water fluctuations.

Comparison with other cases of induced earthquakes and the complex seismotectonic regime of the region suggest that the 13 May 1995 Kozani-Grevena earthquake is an event in the framework of the regional seismicity rather than an event triggered by the load of Polyphyto artificial Lake.

Acknowledgment

The authors are grateful to Professors H.K. Gupta, D.W. Simpson and S. Yunga for their critical comments and suggestions. The authors also thank the Public Power Corporation of Greece for providing data from their network as well as the water elevations at Polyphyto artificial Lake.

References


