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# RUPTURE ZONES AND ACTIVE CRUSTAL DEFORMATION IN SOUTHERN THESSALIA, CENTRAL GREECE

**Abstract.** The southern Thessalia fracture belt has been selected as an area for pilot multinational studies on short term earthquake prediction. Thus, geophysical information has been used (surface breakages, macroseismic fields, aftershock distribution, hot springs, gravimetric data) to accurately define the rupture zones of the three strongest earthquakes in this belt during the present century (1954 M=7.0, 1957 M=6.8, 1980 M=6.5). It is found that the rupture zones of these three earthquakes have an approximately east-west trend, tend to abut without significant overlap, cover the whole southernmost part of the Thessalia plain, and are due to normal type faulting. Furthermore, reliable fault plane solutions and seismic moment release rates have been used to show that the upper crust in this belt is seismically extending at a rate of 7 mm/yr in a N 6°W direction, and is vertically thinning at a rate of 3 mm/yr, which is in full agreement with normal faulting striking in a N 80°E direction.

## INTRODUCTION

Thessalia is the largest plain in central Greece and covers part of the Pelagonian and Subpelagonian geological zones (Mountrakis, 1985). A recent study of its tectonic regime (Caputo, 1990) suggests that the post-Alpine evolution of Thessalia is the result of two different phases.

The first phase, which took place during the Pliocene, was characterized by a N51°E extensional field, and resulted in the formation of two basins (graben structures) trending in a NW-SE direction, that is, the Karditsa and Larissa basins. The second phase, which has been taking place since the Pleistocene, is characterized by an almost N-S extensional field resulting in a nearly EW fault system, which is imposed over the structures of the first phase. Fault plane solutions for four strong earthquakes of the 1980 and 1985 seismic sequences are in agreement with such an extensional field and such a direction of seismic faulting in southeastern Thessalia at present (Papazachos et al., 1983; Taymaz et al., 1991).

The twentieth century seismic activity in Thessalia is very high (Papazachos, 1991), but historical data, as shown in the seismicity maps (Papazachos and Papazachou, 1989), reveal a peculiar pattern of strong earthquake occurrence in this region: a relatively quite period of about one hundred and twenty years (1787-1905) was preceded by an active period of about one and a half centuries (1621-1787), and has been followed by an active period from 1905 to the present. The latest strong ($M_s$=5.8) earthquake in this region occurred in 1985.

Recent work on long term earthquake prediction (Papazachos, 1991) assigns a high probability to the generation of a strong earthquake ($M \geq 5.7$) in Thessalia during the next decade. This is one of the reasons justifying the selection of Thessalia as an area for a joint multidisciplinary experiment by European countries for the identification of short term earthquake precursors.

It is, however, of great importance for such an experiment to define as accurately as possible



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that part of Thessalia which is subject to the major deformation rate, where the probability of identifying earthquake precursors is highest, and to obtain an estimation of the size of this deformation. This is the main goal of the present paper, which was first released in April 1992 (publication No. 7 of the Geophysical Laboratory) and circulated to the participants of the first EPOC meeting in Athens on June 1, 1992. Work on these rupture zones was recently presented by Papadopoulos (1992).

## SEISMICITY OF THESSALIA

There are several known historical earthquakes with magnitudes between 6 and 7 in the Thessalia plain (Papazachos and Papazachou, 1989). This historical information is useful, but the data available are not complete or accurate enough for quantitative investigation of the seismicity in this region. For this reason, only instrumental data from the present century are used for the investigation in this paper.

Eight seismic sequences with mainshock magnitudes equal to or larger than 6.0 occurred in this region during the present century (1905, 1911, 1930, 1941, 1954, 1955, 1957, 1980), and these sequences include 12 shallow earthquakes with $M_s \geq 6.0$. Table I gives information on these twelve earthquakes.

In addition to this sample of strong earthquakes, three additional complete samples of smaller earthquakes have also been used in the present study. The four samples are thus as follows:

| Time | $M_{min}$ |
|---|---|
| 1901-1990 | 6.0 |
| 1911-1990 | 5.6 |
| 1970-1990 | 4.5 |
| 1981-1990 | 3.8 |

That is, the complete data include: all earthquakes with $M_s \geq 6.0$ from the first time period (1901-1990), all earthquakes with $M_s \geq 5.6$ from the second time period, etc. The completeness is based on information given in the corresponding sources used for these data and on the corresponding diagrams of the cumulative distributions of magnitudes. The data source for the first two samples is the catalogue of Comninakis and Papazachos (1986), and for the last two samples, a catalogue published by Karacostas (1988) and the bulletins of the Seismological Institute of the National Observatory of Athens and of the Geophysical Laboratory of the University of Thessaloniki. Fig. 1 shows a geomorphological map (contour lines of 300 m and 900 m are drawn) of the area where the epicenters of the earthquakes of these four data samples are plotted.

It is observed that most of the seismic activity in Thessalia occurs along a seismic belt which follows the southern boundary of the plain. All four earthquakes of the present century with $M_s \geq 6.5$ occurred in this very active seismic belt. The broken line in Fig. 1 is a least squares fit to the epicenters of the complete data set. Its direction is N79°E.

Fig. 1 also shows another seismic belt along the eastern boundary of Thessalia. Five earthquakes with magnitudes between 6.0 and 6.3 occurred in this belt between 1905 and 1955. It is interesting to note that the activity in this belt (during the period 1905-1955) was followed by the activity in the main southern belt (during the period 1954-1980), where stronger earthquakes (up to $M_s = 7.0$) occurred.

No strong seismic activity has occurred during the present century in northwestern Thessalia, but small shocks occur on this boundary, as Fig. 1 indicates.

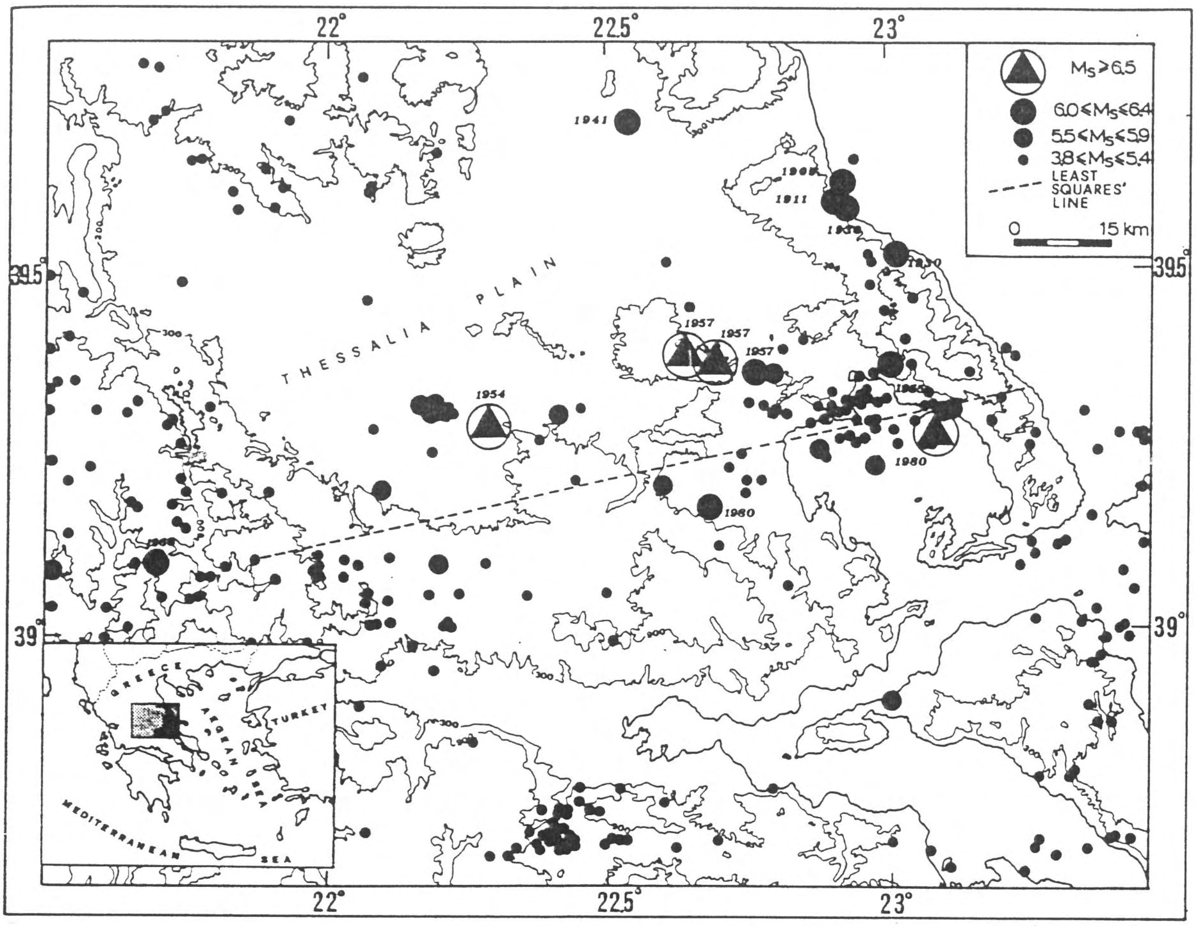


Fig. 1 — Distribution of epicenters of earthquakes in Thessalia and surrounding area during the present century. The seismic belt in southern Thessalia is well defined.

Fig. 1 shows that there are enough data to accurately determine seismicity parameters for the big southern seismic belt, such as the most probable magnitude for a certain time period, etc. These parameters depend on the well known constants a and b of the Gutenberg and Richter (1944) recurrence relation. To determine these constants accurately, a method called by some seismologists (Milne and Davenport, 1969) the "mean value method" has been applied. This method, which is described in detail by Papazachos (1990), can use all complete samples of available data and is suitable in the present case because we have four such data samples.

Fig. 2 shows a plot of the logarithm of the number N of earthquakes with magnitude equal to or larger than M, as a function of the magnitude for a period of 90 years.

The relation

$$logN=5.71-0.80\cdot M \quad (1)$$

fits the data in the least squares sense; that is, the constants a and b for one year period are

$$a=3.76 \qquad b=0.80. \quad (2)$$

The magnitude M of the most probable maximum earthquake with this magnitude or larger in a time period t is given by the relation

$$M=\frac{a}{b}+\frac{logt}{b} \quad (3)$$

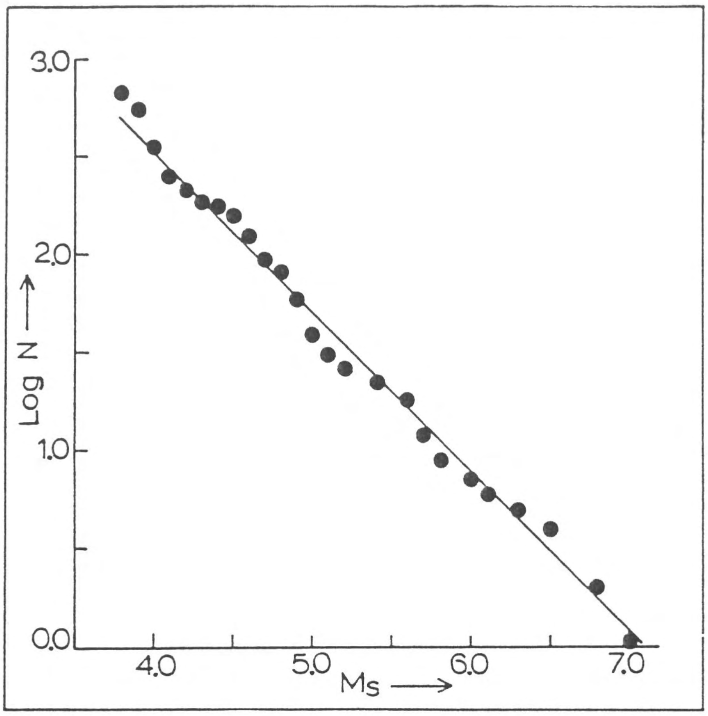


Fig. 2 — Logarithm of the cumulative frequency N of earthquakes in the southern Thessalia seismic belt as a function of the surface wave magnitude M for a period of 90 years.

For t=5 years and t=10 years, we get M=5.6 and M=5.9, respectively. This means that there is a high chance of a strong (M≥5.5) earthquake during the next few years in southern Thessalia.

## RUPTURE ZONES IN SOUTHERN THESSALIA

Table 1 shows that all strong ($M_s \geq 6.0$) earthquakes which have occurred in Thessalia during the last 38 years had their foci in the very active fracture belt which follows the southern boundary of the Thessalia plain. This provides the possibility for accurate determinations of the rupture zones along this fracture belt, because there is much seismological and other geophysical information for these relatively recent earthquakes.

Surface breakage is the most direct information for estimating the dimensions of the rupture zone of a past earthquake. Precise location of aftershocks, however, is the most simple and effective method for estimating the extent of primary faulting in earthquakes (Sykes, 1971;

**Table 1 — Date, epicenter ($\phi$, $\lambda$), focal depth h (n stands for normal), and surface wave magnitude $M_s$ for all shocks with $M_s \geq 6.0$ in Thessalia during the present century.**

| Date | Origin time | $\phi$°N | $\lambda$°E | h (km) | $M_s$ |
|---|---|---|---|---|---|
| 1905, Jan. 20 | 02:32:30 | 39.62 | 22.91 | n | 6.3 |
| 1911, Oct. 22 | 22:31:45 | 39.59 | 22.90 | n | 6.0 |
| 1930, Feb. 23 | 18:19:12 | 39.58 | 22.92 | n | 6.0 |
| 1930, Mar. 31 | 12:33:48 | 39.52 | 23.02 | n | 6.1 |
| 1941, Mar. 01 | 03:52:47 | 39.70 | 22.54 | n | 6.3 |
| 1954, Apr. 30 | 13:02:36 | 39.28 | 22.29 | 7 | 7.0 |
| 1955, Apr. 19 | 16:47:19 | 39.37 | 23.00 | n | 6.2 |
| 1957, Mar. 08 | 12:14:14 | 39.38 | 22.63 | n | 6.5 |
| 1957, Mar. 08 | 12:21:13 | 39.36 | 22.69 | 6 | 6.8 |
| 1957, Mar. 08 | 23:35:09 | 39.36 | 22.76 | 9 | 6.0 |
| 1980, Jul. 09 | 02:11:56 | 39.27 | 23.09 | 16 | 6.5 |
| 1980, Jul. 09 | 02:35:52 | 39.16 | 22.68 | 18 | 6.1 |

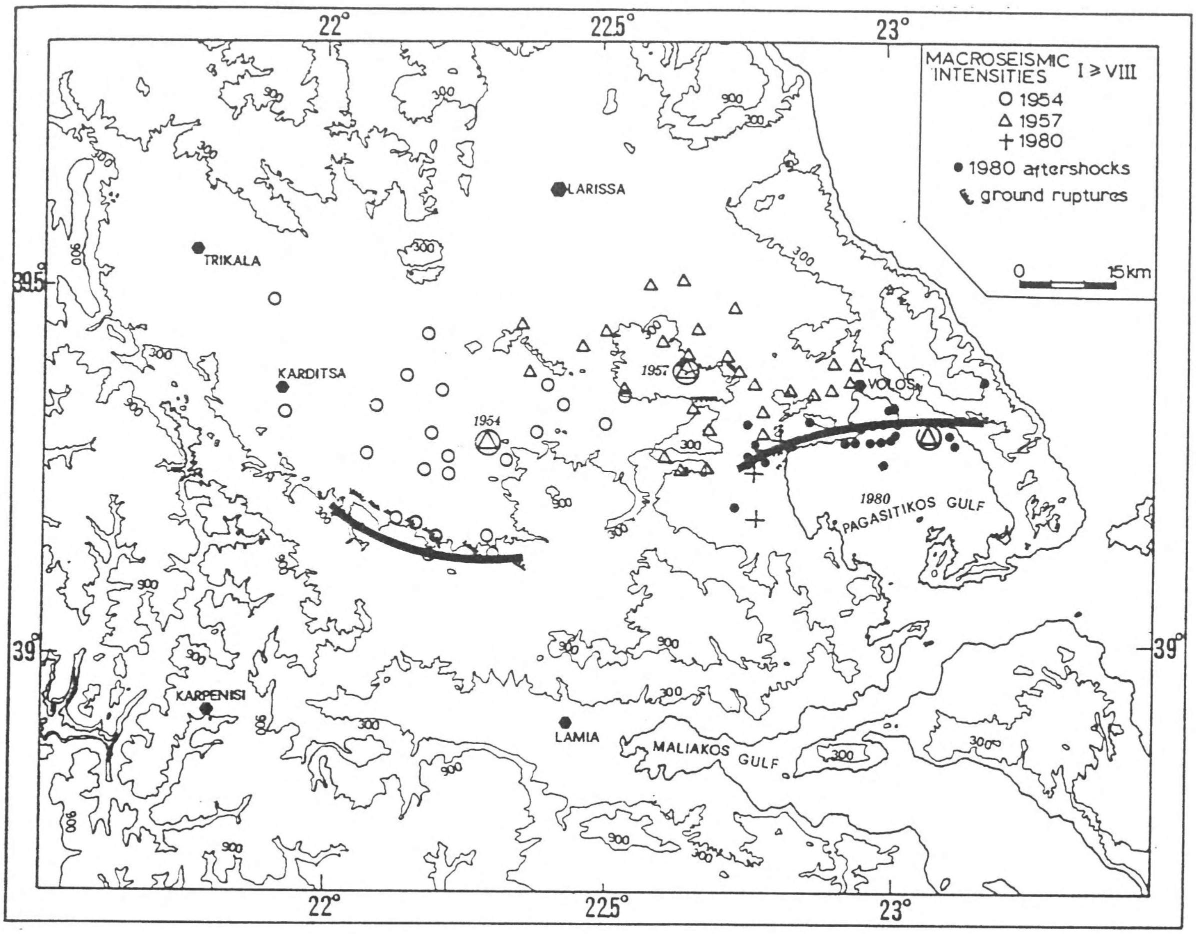


Fig. 3 — Maximum macroseismic intensities and ground ruptures for the 1954, 1957 and 1980 earthquakes in southern Thessalia, and the spatial distribution of aftershocks of the 1980 earthquake. The thick lines indicate the southern boundary of the rupture zone of the 1954 earthquake, and the northern boundary of the rupture zone of the 1980 earthquake, respectively.

Karakaisis et al., 1985). Data on high values of macroseismic intensities, which define the meizoseismal area of an earthquake, have also been used for mapping rupture zones of past earthquakes (Kelleher, 1972). All these three methods are here apaplied as an initial step in the determination of the rupture zones of past earthquakes in southern Thessalia.

These three methods (surface breakage, aftershock distribution, meizoseismal area) are effectively applicable for the determination of rupture zones only of large past earthquakes. For this reason, rupture zones for the mainshocks with $M_s \geq 6.5$ were determined; that is, for the following main shocks: 1) 1954, $M_s=7.0$, 2) 1957, $M_s=6.8$. 3) 1980, $M_s=6.5$ (see Table 1 and Fig. 1).

Fig. 3 shows the epicenters of these three mainshocks and a plot of the data available for the three methods.

The observed tensional ground breakages in the western part of the belt (Papastamatiou and Mouyaris, 1986; Ambraseys and Jackson, 1990) dip almost to the north and define fairly well part of the southern boundary of the rupture zone of the 1954 ($M_s=7.0$) earthquake, while the meizoseismal area of this earthquake, defined by the large macroseismic intensities (open circles) covers an area north of the surface breakage. The dimensions of this area fit well the size of the rupture zone for an earthquake of this magnitude, according to relations between the size of the rupture zones (fault length, surface) and magnitudes of earthquakes in Greece (Papazachos and Papazachou, 1989). All this information indicates that the earthquake was produced by a normal fault striking in an approximately east-west direction and diping to the north. It is probably a listric fault, that is, the dip angle is large (almost vertical) near the surface in the southern boundary of the rupture zone and this angle decreases with depth to the north. No precise locations of aftershocks are available for this seismic sequence.

Neither precise locations of aftershocks nor the extent of ground rupture are known for the earthquake of 1957 ($M_s$=6.8). The meizoseismal area for this earthquake, however, as this area is indicated by the spatial distribution of the large macroseismic intensities (triangles in Fig. 3), defines fairly well its rupture zone, which is also elongated in an approximately east-west direction. The size of the area is approximately equal to the size of the rupture zone expected for an earthquake of this magnitude. This earthquake was also produced by a normal fault, which strikes roughly east-west, but there is no evidence for its dip direction. Seven minutes before the mainshock, a very strong foreshock ($M_s$=6.5) occurred (see Fig. 1). It is probable that one of these twin earthquakes occurred on a fault dipping to the north and the other on an antithetic fault dipping to the south.

Precise locations of aftershocks are available for the 1980 ($M_s$=6.5) earthquake (Papazachos et al., 1983), and the epicenters of the strongest of these aftershocks ($M_s \geq 4.5$) are shown (black circles) in Fig. 3. Some surface tensional breakage and the location of three sites (crosses) with high macroseismic intensities are also shown in this figure. It is clear that the rupture zone of the 1980 earthquake also has an almost east-west trend, and reliable fault plane solutions for four earthquakes of this zone also support this conclusion (Papazachos et al., 1983; Taymaz et al., 1991). The surface breakage and the geomorphology of the area indicate that this is a normal listric fault dipping to the south. According to Burton et al. (1991), it consists of two parts separated by a barrier: the western part is associated with the Almiros graben and the eastern part with the Pagasitikos gulf.

The thick solid lines in Fig. 3 show the southern boundary of the rupture zone of the 1954 earthquake and of the northern boundary of the rupture zone of the 1980 earthquake, respectively. These lines also indicate where the faults of these earthquakes are expected to reach the earth's surface.

In addition to the previously described information, macroseismic intensities observed outside the meizoseismal zone but at distances from the epicenter smaller than 150 km have been also used to determine the rupture zones of these three earthquakes. Papazachos (1992) has shown that the macroseismic intensities within these distances ($\Delta < 150$ km) depend mainly on the radiation pattern at the focus of the earthquake, and has developed a model of anisotropic radiation to interprete the spatial distribution of the intensities. This model was used to draw theoretical isoseismals. That isoseismal of which the maximum dimension is equal to the expected fault length of the earthquake, on the basis of a relation between this length L and the magnitude M of the earthquake, is assumed to define the rupture zone of the earthquake. Such a relation for Greece has been derived by Papazachos and Papazachou (1989) and is given by

$$logL = 0.51 \cdot M_s - 1.85, \tag{4}$$

where L is in kilometers and $M_s$ is the surface wave magnitude. Figure 4 shows the rupture zones of the three strong earthquakes in southern Thessalia (1954, 1957, 1980). Comparison of Figs. 3 and 4 shows that this technique gives very consistent results with the results of the direct methods for the three rupture zones of southern Thessalia. The thick lines are the same as those in Fig. 3. The three stars in this figure show three hot springs (1 Smokovo, 2 Kaitsa, 3 Agrapidia), and it is very interesting to see that the sites of the springs follow the trend of the 1954 earthquake fault, indicating that these hot springs are genetically related to this fault.

It is clear that the rupture zones of these three large ($M_s \geq 6.5$) successive (1954-1957-1980) earthquakes in southern Thessalia trend in an approximately east-west direction, tend to abut without significant overlap, and cover the whole southernmost part of the plain of Thessalia.

## GRAVIMETRIC EVIDENCE

The role of gravimetry in delineating tectonic features was recognised a long time ago. Its

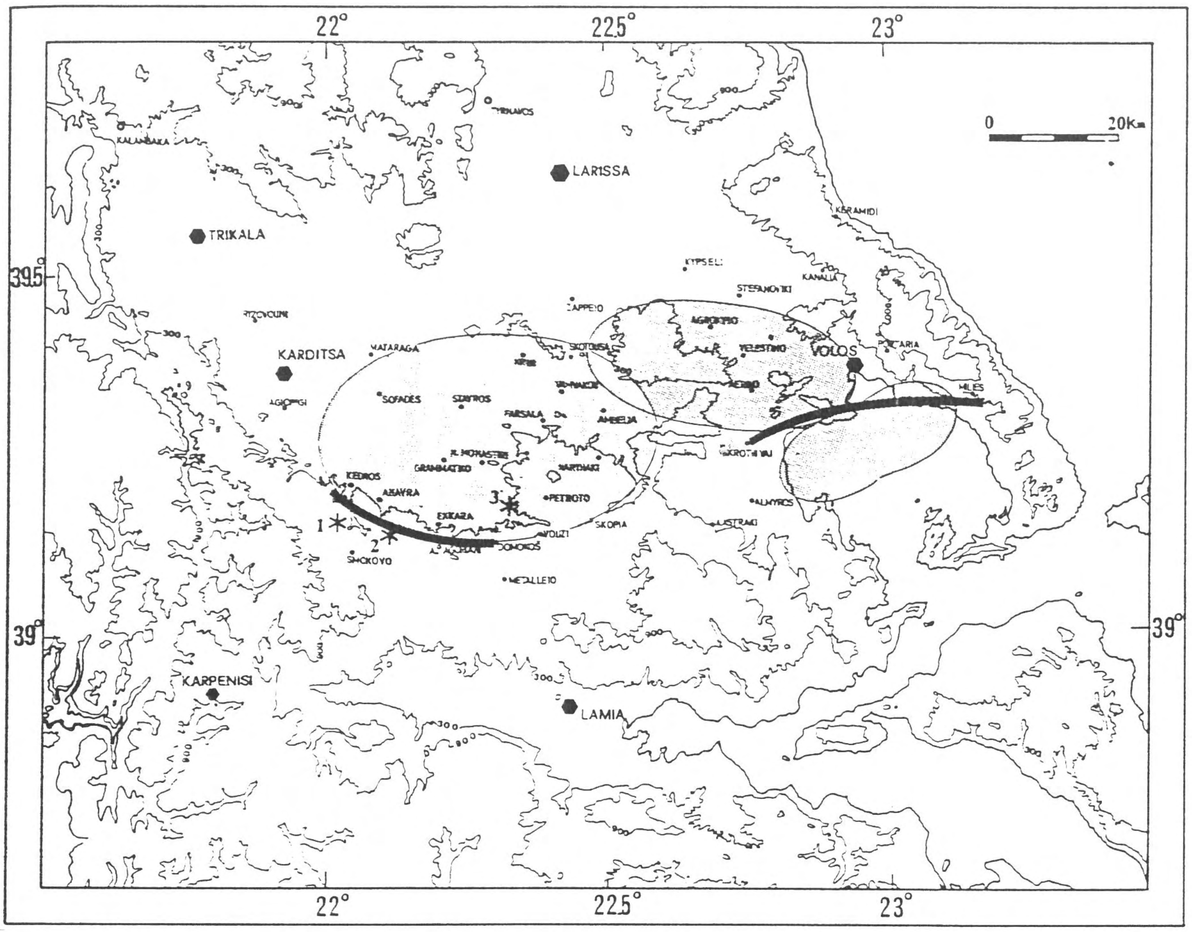


Fig. 4 — Rupture zones of the 1954, 1957, 1980 earthquakes in southern Thessalia determined on the basis of an anisotropic radiation modelling of macroseismic intensities. Stars show the sites of thermal springs (1 Smokovo, 2 Kaitsa, 3 Agrapidia).

practical application and limitation due to inherent ambiguity in interpretation are very well established nowadays. As an example, we can quote a recent work by Papazachos and Panagiotopoulos (1992), where seismological and gravimetric data were combined to define fracture zones in the southern Aegean. The present study aims at assisting the location of fracture zones in Thessalia, in combination with the seismological techniques presented in the preceding paragraphs.

The data used comprise part of the regional Bouguer anomaly map of Greece which was compiled by Lagios et al. (1988). They used data collected by Makris and Stavrou (1984), offshore data by Morelli et al. (1975) and several other data sets acquired earlier for oil and mineral exploration. The mean density of measurements is one station in every 4 km.

Data were digitized and gridded in order to produce a new matrix with Bouguer anomaly values at points spaced 5 km apart. A second order trend was extracted, which was attributed to deep causes, perhaps to an undulation of the Moho depth beneath the Aegean area.

Data were then transformed into the wavenumber domain employing the Cooley-Tukey FFT algorithm (Cooley and Tukey, 1965). Next, they were multiplied by the corresponding operators. Inverse transformation of the product yielded the first and second derivatives and the upward continuation of the original data.

Fig. 5 shows the map of the field continued upwards by 5 km, i.e. by 1 grid unit. This map was chosen as representative of the relatively long wavelength anomalies which are assumed to reflect major tectonic features. The upward continuation reveals broad anomalies due to deep or wide bodies and suppresses short wavelength anomalies of shallow origin (Agarwal and Lal, 1972).

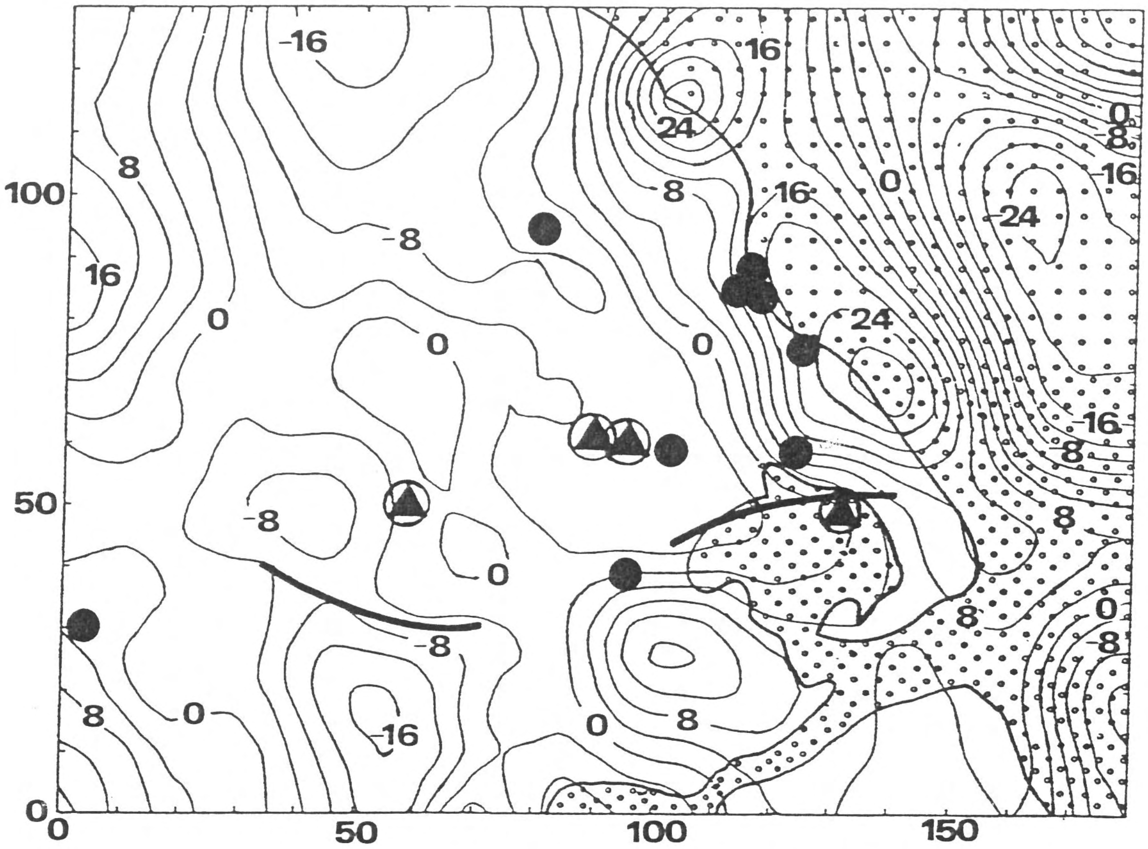


Fig. 5 — Upward continuation of the gravimetric field by 5 km. The dotted area is off-shore. The distribution of epicenters of earthquakes of $M_s \geq 6.0$ has also been plotted.

The lines which define the boundaries of the two rupture zones (1954, 1980) mentioned previously have been also drawn in the same figure. A portion of the easternmost one corresponds to an eastward deflection of the contours defining the main feature of the map, which is described in the next section. Furthermore, the other portion of the boundary coincides with the region where intense E-W faulting is observed at the surface (Caputo and Pavlides, 1991).

The boundary of the westernmost rupture zone shows an alignment with the contours of the relevant gravity anomalies. It also corresponds well with the topographic features and the fractures seen at the ground surface.

The same conclusions can be drawn from the maps of the first and second vertical derivatives which are not shown in this study.

An elongated pattern dominates the eastern part of the map ranging north of the city of Volos up to the uppermost margin of the Thessalia plain. It strikes approximately NW-SE and reflects the eastern faulted boundary of the plain where several earthquakes with magnitudes up to 6.3 occurred between 1905 and 1955.

## ACTIVE DEFORMATION IN THE FRACTURE BELT OF SOUTHERN THESSALIA

Work on active deformation in Thessalia as a whole has been performed by Papazachos and Kiratzi (1992), and in the broader area of central Greece by Ambraseys and Jackson (1990). Here, we attempt an estimation of the active deformation in the seismic belt of southern Thessalia, since this belt is now accurately defined.

The method of analysis applied in the present paper for the estimation of active crustal deformation in this belt is the one proposed by Papazachos and Kiratzi (1992), which is mainly based on Kostrov's (1974) and Jackson and McKenzie's (1988) formulations. The velocity of deformation is calculated in two steps. Initially, the "size" of the deformation, represented by the annual scalar moment rate $\dot{M}_o$ is calculated using the following relations defined by Molnar (1979):

$$\dot{M}_o = \frac{A}{1-B} \cdot M_{o,\,max}^{(1-B)} \cdot \tag{5}$$

where $M_{o,\,max}$ is the moment of the largest ever observed earthquake of the zone, and

$$A = 10^{\left(a+\frac{bd}{c}\right)} \text{ and } B = \frac{b}{c}, \tag{6}$$

where a and b are the constants of the Gutenberg and Richter relation, and c and d are the constants of the moment-magnitude relation:

$$\log M_o = c\, M_s + d. \tag{7}$$

The values of the constants a and b for this belt are those given by the relations (2), $M_{max}=7.0$ (see Table 1) and $c=1.5$ and $d=15.89$ (Papazachos and Kiratzi, 1992). In this way, the moment rate value was found to be $\dot{M}_0 = 7.6*10^{24}$ dyne cm/yr.

Then, the "shape" of the deformation is defined for this seismic belt as is described in the following.

A "representative focal mechanism tensor" F is calculated for each seismic belt using the following relation:

$$F = \frac{\dot{M}}{\dot{M}_o} = \frac{\sum_{n=1}^{N} M^n/\tau}{\sum_{n=1}^{N} M_o^n/\tau} = \frac{\sum_{n=1}^{N} M^n}{\sum_{n=1}^{N} M_o^n}, \tag{8}$$

where N is the number of all focal mechanisms available in each belt, $M^n$ is the moment tensor, and $M_o^n$ the corresponding scalar moment for the nth focal mechanism. The moment tensor $M^n$ is calculated from

$$M^n = M_o^n\, F^n, \tag{9}$$

where $F^n$ is a function of the strike, dip and rake of the focal mechanism (Aki and Richards, 1980).

The equations initially defined by Kostrov (1974) and Jackson and McKenzie (1988) can now be transformed and the strain rate and velocity tensors calculated using the following relations:

$$\epsilon_{ij} = \frac{1}{2\,\mu\, V} \dot{M}_o\, F_{ij} \quad i,\, j=1,\, 2,\, 3, \tag{10}$$

$$U_{ii} = \frac{1}{2\,\mu\, I_k\, I_j} \dot{M}_o\, F_{ii} \quad i\neq k,\; k\neq j,\; j\neq i, \quad i=1,\, 2,\, 3,$$

$$U_{12} = \frac{1}{\mu\, I_1\, I_2} \dot{M}_o\, F_{12}\,, \tag{11}$$

$$U_{i3} = \frac{1}{\mu\, I_1\, I_3} \dot{M}_o\, F_{i3} \quad i=1,\, 2,$$

**Table 2 — Fault plane parameters (strike $\alpha$, dip $\delta$, rake $\lambda$) for four earthquakes in the southern Thessalia fault belt.**

| Date | Origin time | $\phi^{o}$N | $\lambda^{o}$E | $M_s$ | $\alpha^{o}$ | $\delta^{o}$ | $\lambda^{o}$ | Ref |
|---|---|---|---|---|---|---|---|---|
| 1980, July 09 | 02:10:20 | 39.3 | 22.9 | 5.6 | 82 | 42 | — 79 | 1 |
| 1980, July 09 | 02:11:57 | 39.3 | 22.9 | 6.5 | 81 | 40 | — 90 | 1 |
| 1980, July 09 | 02:35:52 | 39.2 | 22.7 | 6.1 | 81 | 40 | — 90 | 1 |
| 1985, Apr. 30 | 18:14:13 | 39.3 | 22.8 | 5.8 | 77 | 50 | —105 | 2 |

1. Papazachos et al. (1983), 2. Taymaz et al. (1991)

where $\mu$ is the bulk modulus ($=3\cdot10^{11}$ dyne/cm$^2$), V is the volume of the deforming zone, $l_1$ and $l_2$ denote the length and the width of the seismic zone, respectively, while $l_3$ denotes the depth of the seismogenic layer. The reference system, $0x_1x_2x_3$ , is the zone's $0l_1l_2l_3$ system. It is obvious that since F is calculated in the North-East-Down system (using eqn. (8)), a rotation of F in the reference system of the zone is necessary before it is incorporated in eqns. (10) and (11).

To determine the tensor F, we need fault plane solutions (strike, dip, and rake) and the corresponding magnitudes. Table 2 gives the most reliable fault plane solutions which are available for this belt. The number in the last column of the table corresponds to the references written at the bottom of the table.

The components of the tensor F for this belt are

$$\begin{matrix} 0.96 & -0.15 & 0.15 \\ -0.15 & 0.20 & -0.14 \\ 0.15 & -0.14 & -0.98 \end{matrix} \qquad (12)$$

From this we can get the eigenvalues of F. It is clear from eqn. (8) that the deviation of these eigenvalues from 1,0 and —1 is a measure of the consistency of the fault plane solutions (specifically, of the P, T and null axis). The eigenvalues corresponding to eqn. (12) are practically equal to 1, 0, —1, showing that these fault plane solutions are very similar.

Eqns. (10) and (11) show that in addition to $\dot{M}_o$ and F we need also $l_1$ , $l_2$ , $l_3$ . The value of $l_3$ was taken as 15 km, which is the thickness of the seismogenic layer in this belt, as we have already mentioned. $l_1$ and $l_2$ were calculated in the following way suggested by Papazachos and Kiratzi (1992).

Since we are interested in the active part of the belt and because we need to have a belt of rectangular shape, we assumed that the epicentral area of this belt has such a shape. Thus, the length, the width and the azimuth of this belt were calculated as follows: using the complete data set, we calculated the coordinates of the center of the belt which we chose as the center of a coordinate system, with the x and y axes parallel to the NS and EW direction, respectively. Assuming that the earthquakes in the belt follow a linear pattern, we calculated a least squares best fit line for the zone. The azimuth $\xi$ is taken to be the azimuth of the line with north. The projections of the most distant epicenters (from the center of the zone) onto this line define the length $l_1$ of the zone. The width $l_2$ of the zone is taken to be $4\sigma$, where $\sigma$ is the standard deviation of the earthquake epicenters from the line.

In this way the following values were found for the length and the width of the belt:

$$l_1=121 \text{ km}, \quad l_2=35 \text{ km and } l_3=15 \text{ km}.$$

The values of the parameters given by eqns. (4), (7) and (12) were used to calculate the values, v, the azimuth, and the plunge of the maximum and the minimum deformation velocity, together with the deformation velocity in the vertical direction. The results are as follows:

| *vel.* | *Az.* | *plunge* |
|---|---|---|
| *0.70* | *174* | *—5* |
| *—0.00* | *84* | *0* |
| *—0.30* | | |

It is seen that this belt is extending at a rate of 7 mm/yr in a N 6°W direction, and is thinning at a rate of —3mm/yr in the vertical direction. These values, however, express only the seismic part of the deformation. If we make the reasonable assumption that the whole deformation (seismic and aseismic) is about double the seismic one (Papazachos and Kiratzi, 1992), we conclude that the total horizontal extension in southern Thessalia is 1.4 cm/yr, and its total vertical thinning is 0.6 cm/yr.

## DISCUSSION

The results of the present paper justify the selection of southern Thessalia as an area for pilot studies on short term earthquake prediction for two main reasons.

First, the rupture zones and the focal mechanisms of the past strong earthquakes are well known. It is reasonable to expect in the future earthquakes of a similar mechanism which will break part of or the whole rupture zone of each of these three shocks. This is a great help in the selection of the sites for the several kinds of observations (installation of instruments, etc).

Second, the active deformation rate in this belt is very high at present, and it is very probable that this deformation will continue to be high for some time. Therefore, the probability of the generation of moderate magnitude ($M_s \geq 5.0$) or even of large magnitude ($M_s \geq 6.0$) earthquakes during the next decade is high. In addition to that, such a strong total deformation rate (1.4 cm/yr) can be easily detected by geodetic or gravimetric methods, and such work would contribute to a better understanding of some basic geotectonic properties of the area.

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