Pure appl. geophys. 150 (1997) 75–89
0033–4553/97/010075–15 \$ 1.50 + 0.20/0

Pure and Applied Geophysics

The Winter 1991–1992 Earthquake Sequence at Cephalonia Island, Western Greece

G.-A. Tselentis,[1] N. S. Melis,[1] E. Sokos[1] and P. Beltas[1]

Abstract—Properties of the earthquake sequence of the 23 January, 1992 (5.8 M_S) earthquake in Cephalonia Island, Western Greece, are investigated. The parameter b in the frequency-magnitude relation is found equal to 1.084 while the decay parameter p of the time distribution of the aftershock sequence is found equal to 0.991. The principal parameters method is applied to the aftershock sequence and the average strikes of N156°E (±20°) and N61°E (±10°) were obtained denoting also very small dips. The major strike which resulted from the aftershock epicentre distribution was NW–SE, similar to Miocene-Neogene basins on Cephalonia Island.

Key words: Earthquakes, seismotectonics, Greece, Ionian Islands.

1. Introduction

On January 23 1992 (04hr 24min 16sec) an earthquake of magnitude 5.8 M_S occurred off the NW coast of Cephalonia Island, Western Greece. The hypocenter of the earthquake (38.35°N, 20.32°E, depth = 17 km) was located in the Gulf of Mirtu (Figure 1). This earthquake was preceded by a foreshock of magnitude 4.5 M_L, which occurred 3 hours prior to the main shock and was followed by numerous aftershocks.

In the present paper, the time, magnitude and space distribution of the foreshock and aftershock activity are investigated. A primary objective of the paper is to infer major directions of rupture from the spatial and temporal evolution of the aftershock sequence.

2. Tectonic Setting

Cephalonia is located at the northwestern end of the Hellenic arc, in an area where the Adriatic collision takes place (Sorel, 1976; Sorel *et al.*, 1976) as Appulia converges with the Aegean microplate (Figure 1a). This is one of the most seismically active areas in Greece and the surrounding regions. The area is

[1] Seismology Laboratory, University of Patras, Rio 261 10, Greece.

characterized by a compressional stress regime producing low-angle thrusting (McKENZIE, 1972, 1978; MERCIER *et al.*, 1972, 1976). The collision is characterized by horizontal compression almost perpendicular to the Hellenic arc (LE PICHON and ANGELIER, 1979, 1981; MERCIER *et al.*, 1987; HATZFELD *et al.*, 1990) and it is thought to have been initiated during the Miocene (McKENZIE, 1972; LE PICHON and ANGELIER, 1979, 1981).

An escarpment in the bathymetry NW of the island indicates the existence of a strike-slip fault trending NE–SW (FINETTI, 1976; UNDERHILL, 1989; Figure 1). This is also indicated as an offset of the regional seismicity north of the island (FINNETI, 1976, 1982; ANDERSON and JACKSON, 1987). UNDERHILL (1988, 1989) postulated that dextral strike-slip faulting is due to strain accommodation within the main system of contraction. MERCIER (1981) proposed the existence of reverse as well as strike-slip faults in the area and this is confirmed by focal mechanism solutions (RITSEMA, 1974; McKENZIE, 1978; PAPAZACHOS *et al.*, 1984; ANDERSON, 1987; ANDERSON and JACKSON, 1987; HATZFELD *et al.*, 1990; PAPAZACHOS *et al.*, 1991; Figure 1b).

A simplified tectonic map of the island (after UNDERHILL, 1989; Figure 1b) shows the NE–SW trending strike-slip system offshore. Miocene and Neogene basins such as those of Argostoli, Gulf of Mirtu, Gulf of Agia Kiriaki, Lixouri and Markopoulo are of the two main characteristic trends, NE–SW and NW–SE. These basins and the E–W trending piggy-back basins (UNDERHILL, 1989) which were formed in the eastern part of Cephalonia during the thrust migration of the Hellenide orogeny towards Western Greece, are areas of recent shallow seismic activity (strike-slip and extensional mechanisms are proposed). In addition, there is deeper seismicity (mostly compressional mechanisms), which is related to the subduction. At this point, it is interesting to mention that PAPAZACHOS *et al.* (1992) have observed two dominant modes of crustal deformation (N–S extension and E–W compression) to coexist in the major area of Cephalonia and the adjacent central Ionian Islands.

3. Station Coverage and Data Collection

The main seismic event on Cephalonia occurred towards the NW part of a local microearthquake network installed by the University of Patras Seismology Labora-

Figure 1

(a) Major tectonic features in Western Greece. The box indicates the study area. (b) Map of Cephalonia and Ithaki (after UNDERHILL, 1989) showing the Late Miocene-Neogene basins (shaded). Normal faults are shown with square teeth on the downthrown side, thrusts are shown with triangles on the uplifted side and strike-slip faults with the two arrows marking the movement on both sides. Fault plane solutions are shown as lower hemisphere equal area projections with shaded quadrants of compression and unshaded quadrants of tension (after RITSEMA, 1974; McKENZIE, 1978; PAPAZACHOS *et al.*, 1984; ANDERSON, 1987; ANDERSON and JACKSON, 1987; HATZFELD *et al.*, 1990; PAPAZACHOS *et al.*, 1991).

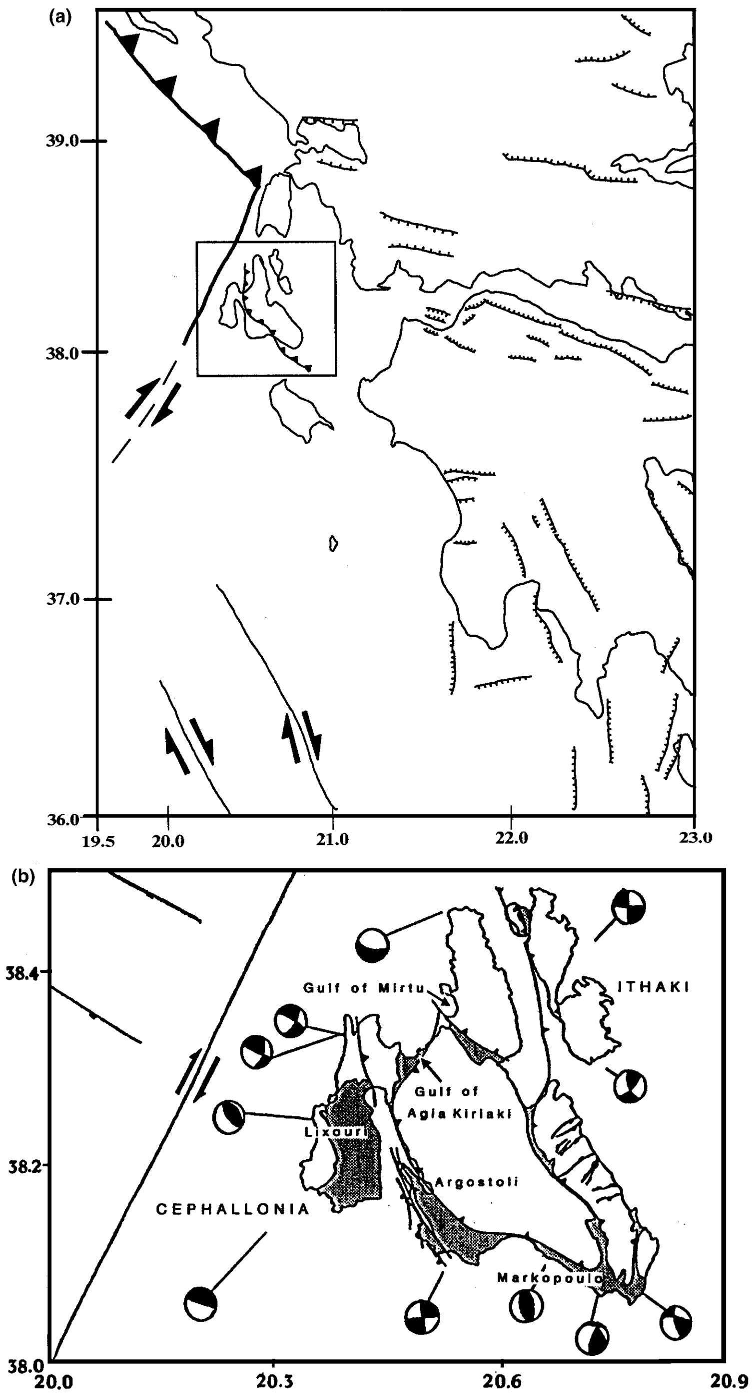
(a)
39.0
38.0
37.0
36.0
19.5
20.0
21.0
22.0
23.0
(b)
38.4
38.2
38.0
20.0
20.3
20.6
20.9
Gulf of Mirtu
ITHAKI
Gulf of
Agia Kiriaki
Lixouri
Argostoli
CEPHALLONIA
Markopoulo

Table 1

Final V_P crustal velocity model used for earthquake location

Velocity (km/sec)	Depth (km)
5.7	0.0
6.0	14.0
6.4	20.0
7.9	39.0

tory in order to study the seismic activity and the tectonic evolution of the region defined by the Gulf of Patras graben system and the Cephalonia Island.

The network started operation in September 1991 and consisted of six short period stations, each with one vertical component (1 Hz) seismometer operating at 60 dB dynamic range and in a low noise environment. The signals were radio-telemetered via FM subcarriers to the central recording site at Patras Seismological Centre in real time. There, each channel signal was antialias filtered with a 30 Hz Butterworth low-pass filter, sampled at 100 Hz and converted to digital form with a resolution of 16 bits.

The STA/LTA technique was employed for event triggering while the HYPO71 program (LEE and LAHR, 1975; LEE and VALDES, 1985) was used for hypocentral and magnitude determinations, employing both *P* and *S* phases recorded at all six stations and the coda duration.

Special care was taken for the correct location of the events. After a conventional hypocentral determination using the standard procedure, events with more than five *P* arrivals and three *S* arrivals were selected and relocated, employing various velocity models for the Aegean area and Western Greece (e.g. MAKRIS, 1977; PANAGIOTOPOULOS and PAPAZACHOS, 1985; MELIS, 1986; PEDOTTI, 1988). The model (Table 1) which resulted in the smallest RMS time residuals was selected and all the events were relocated using this model. Finally, all the epicenters were again determined, using the recommended station delays obtained from the previous run of HYPO71. A value of 1.78 for V_P/V_S ratio was used following MELIS *et al.* (1989). The events with an RMS time residual less than 0.1 and error on the epicentral determination of less than 5 km were accepted as well-located events.

The magnitudes (M_L) of the recorded events were determined using a relation of the form (LEE *et al.*, 1972):

$$M_L = a + b \log T + cD \tag{1}$$

where T is the signal duration in seconds, D is the epicentral distance in kilometers and a, b, c are constants ($a = -0.9$, $b = 2$, $c = 0.0035$) determined by the usual statistical procedure, multiple regression analysis, for earthquakes for which M_L

was known, taken as determined by the National Observatory of Athens for the 2-year period of operation of the network.

The above formula was used to determine earthquake magnitude at each station. The average of the determinations was taken as the local magnitude for each earthquake.

In order to reduce the magnitude threshold to include small events recorded only by the closest station (PT06) to the epicentral region of the main shock, the following procedure was adopted. First, the linear dependence of M_L on duration was verified for all the well-located events by constructing the diagram $M_L - 2\log T$ versus $t_S - t_P$, where the arrival times were read on the digital recordings of station PT06. This was done for M_L values within the range 1.8–4. A least-squares fit to the data resulted in the following equation (Figure 2):

$$M_L = -1.866 + 2\log T + 0.376(t_S - t_P). \qquad (2)$$

Next the arrival times of P and S phases of all the recorded earthquakes at station PT06 were read from the analog recordings, with an estimated accuracy of better than 0.033 sec, and the corresponding magnitudes were assessed using eq. (2).

4. Regional Seismicity before the Main Shock

We first examine the temporal and spatial distribution of seismicity in the Western Greece region prior to the main shock. A time and distance range of 65 days and 100 km respectively was selected, fulfilling the limits proposed by JONES and MOLNAR (1979) as necessary for the detection of foreshock activity.

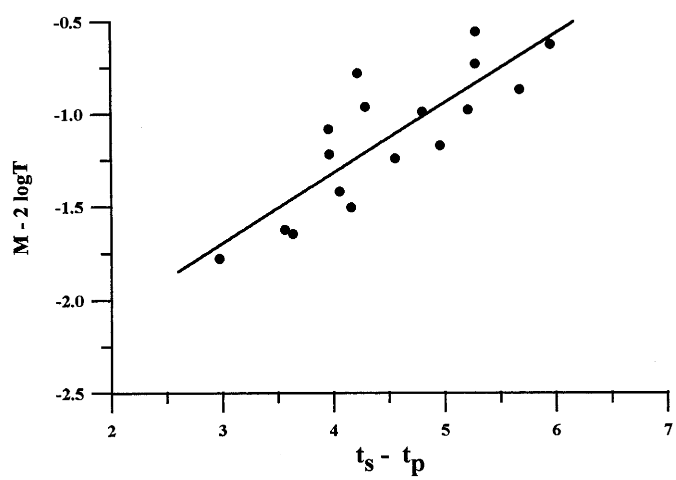

Figure 2
$M_L - 2\log T$ versus $t_S - t_P$ diagram with least-squares regression line for station PT06.

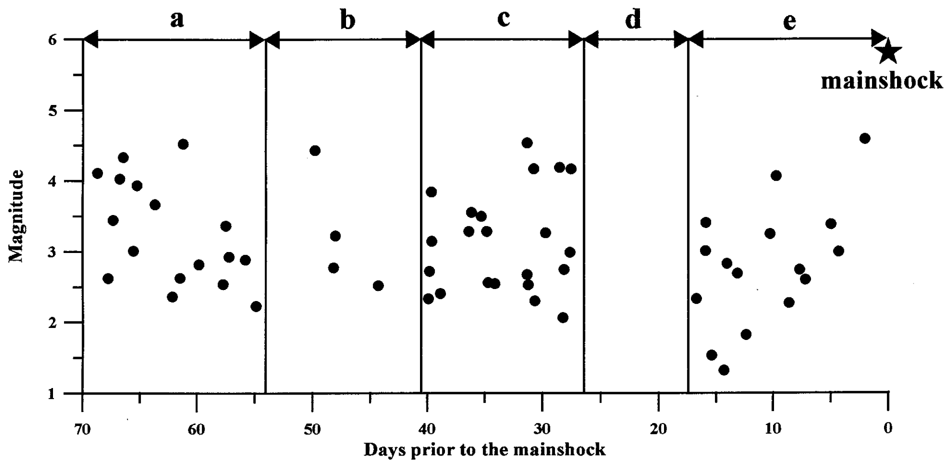

Figure 3
Diagram showing the distribution of local magnitudes of events occurring prior to the main event. (a) to (e) are the time subperiods referred to in the text.

The magnitude-time diagram for the above period is shown in Figure 3. The entire period can be divided into three subperiods (*a*, *c*, *e*) characterized as seismically active, separated by two seismically inactive subperiods (*b*, *d*).

Figure 4 depicts the spatial distribution of the epicenters for each of the above four seismically active subperiods. During the first subperiod a (15/11/91–29/11/91)

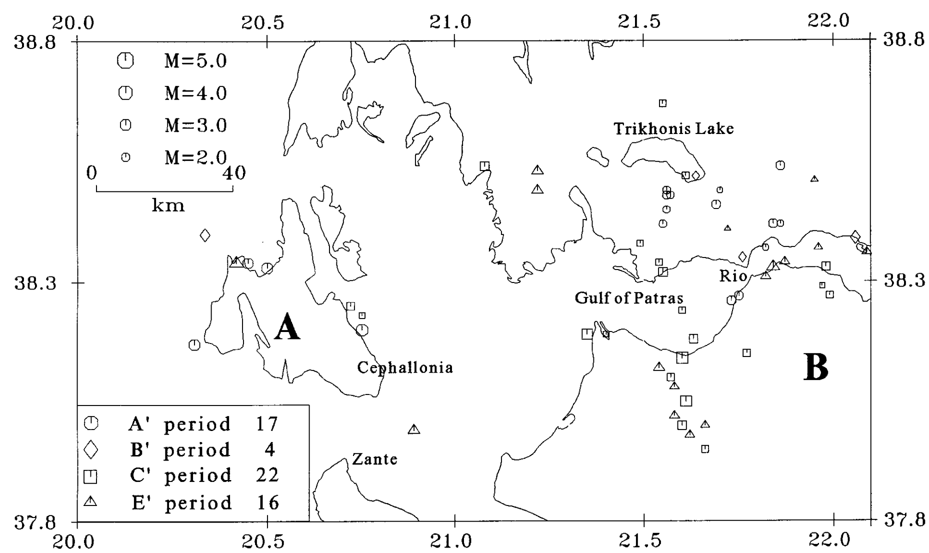

Figure 4
Diagram showing the distribution of epicenters during the seismically active periods (a) to (e) referred to in the text prior to the main event. *M* is local magnitude.

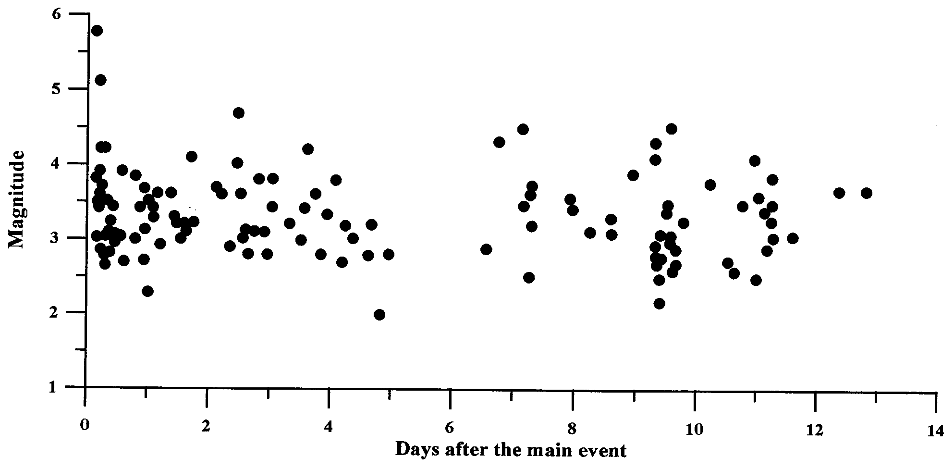

Figure 5
Distribution of local magnitudes of aftershocks in days after the main shock.

disperse seismicity occurred mainly in the region between Lake Trikhonis and Rio Strait. On November 29 1991 an event (3.4 M_L) occurred close to the epicentral region of the main shock (hereafter noted as area A) in the Gulf of Mirtu NW of the island of Cephalonia (Figure 4).

During the second subperiod *b* (30/11/91–13/12/91) four events were recorded, and one of them originated in area A (Figure 4). The third subperiod *c* (14/12/91–26/12/91) is characterized by an earthquake swarm in the Gulf of Patras (hereafter noted as area B) and at a distance of about 120 km from the main event (Figure 4). This swarm is followed by a subperiod of complete quiescence of (27/12/91–6/1/92) and then by a subperiod *e* involving a continuous increase in seismic activity (Figure 3) up to the occurrence of a foreshock (4.5 M_L) on January 21 1992 in area A (Figure 4).

5. The Aftershock Sequence

Following the main event, there were 247 aftershocks during the period 23/1/92 to 4/2/92. Of those events, 142 were determined using the digital recordings of the University of Patras Seismology Laboratory while the magnitudes of the rest were assessed from the analog recordings of station PT06, following the procedure described above.

Figure 5 shows the variation of aftershock magnitudes with time, while Figure 6 exhibits the logarithm of the frequency (n) of aftershocks (events/12 hours) as a function of the logarithm of time. This diagram clearly evidences that the time distribution of aftershocks follows the power law proposed by UTSU (1962):

$$\log n = A - p \log t. \tag{3}$$

The straight line which corresponds to this relation is shown in Figure 6 and the parameter p has the value 0.991. Values between 0.7 and 1.9 have been found for aftershock sequences elsewhere in the area of Greece (PAPAZACHOS, 1974).

The cumulative frequency distribution for the aftershocks is shown in Figure 7 and results in a b value of 1.084, similar to the value of 1.030 proposed by HATZIDIMITRIOU (1984) for the area of Western Greece.

Figure 8 shows the spatial distribution of epicenters for all the well-located events for the 20-day period after the main shock. The aftershock sequence indicates that a secondary fracture zone trending NW–SE was activated after the main event. This direction is consistent with the system of basins existing on the island (Figure 1). It is also consistent with the NW–SE trend shown from the focal mechanism solution given for the main shock (USGS preliminary results 1992; see Figure 8). Although the mechanism is considered as clearly compressional, these basins (e.g., Argostoli and Lixouri basins) are considered to be mainly strike-slip systems (''transpressional'') in order to justify their existence in the Cephalonia tectonic model. In addition, a strike-slip component has been observed on the faults bounding these basins (LEKKAS, pers. comm.). However, the main shock could also be considered as a subduction quake which was linked through the aftershock sequence with the major neotectonic features at the surface. The shortage of focal mechanisms from the aftershock sequence that occurred along these basins limits further interpretation.

In a recent paper based on an analysis of the 1983 7.0 M_S Cephalonia earthquake, SCORDILIS *et al.* (1985) proposed the existence of a transform fault between the islands of Cephalonia and Zante. MAKROPOULOS *et al.* (1989) applied the principal parameters method to the aftershock sequence of the same earthquake

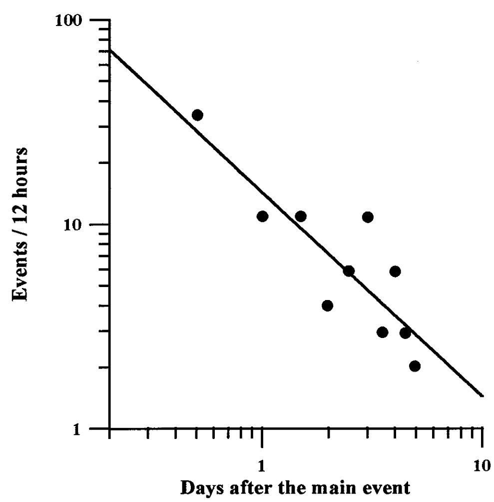

Figure 6
Number of events every 12 hours versus time in days after the main shock.

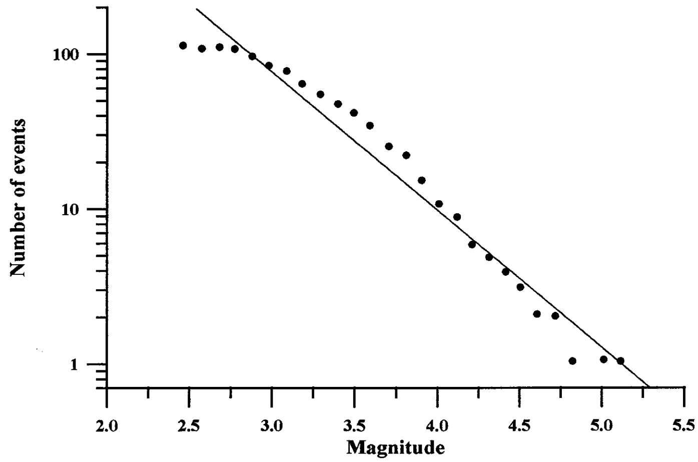

Figure 7
Diagram showing the frequency–magnitude relation and the least-squares regression line.

and reached similar conclusions. Figure 9 depicts the aftershock activity and focal mechanism solution corresponding to the 1983 earthquake and illustrates clearly that it is distributed perpendicularly to the trend of the aftershock sequence presented in Figure 8. The two trends are consistent with the accepted neotectonic regime, which involves both NE–SW and SW–NE trending basins (Figure 1).

6. Principal Parameter Cluster Analysis of the Aftershock Data

In order to understand the rupture propagation of the aftershock sequence, the principal parameters method of EBBLING and MICHELINI (1986) has been applied to the data. The same technique has been applied successfully in other aftershock sequences in Greece (TSELENTIS, 1989; MAKROPOULOS *et al.*, 1989).

An earthquake sequence is considered to take place on numerous faults within an active seismic volume embedded in a regional stress field. Each rupture causes a redistribution of the stress field and the accumulation of stress towards the boundaries of the existing faults. These high stress regions have a higher probability of being the location of the next earthquake. This progressive rupture model is supported by the observation that aftershocks are clustered together in both space and time, suggesting that they are interdependent. It is thus expected that the analysis of the spatial distribution of successive events could provide useful information regarding the evolution in space and time of the rupture process.

Based on the above assumptions, the method consists of studying the spread or the variance-covariance matrix of successive sets of temporally arranged events within the aftershock sequence. This matrix may be regarded as defining the spatial ellipsoid, fitted through the foci, whose axes are the eigenvectors and have lengths equal to the square roots of the corresponding eigenvalues.

The principal parameters method was employed to investigate the 1992 earthquake sequence and to find any correlation to the 1983 sequence. For the applicaton of the method, a circle of 15 km radius located at 15.2 km depth, thus centered at the epicenter defined area, was chosen. The optimal width of the sliding window was chosen equal to 25 hypocenters for both earthquake sequences (MAKROPOULOS *et al.*, 1989), after examining the average ratio between the intermediate and smallest eigenvalues (MICHELINI and BOLT, 1986) of the scatter matrices for different window widths.

Figures 10a,b display the obtained eigenvectors with the largest and smaller eigenvalues, plotted on a lower hemisphere equal area projection for the 1983 and the 1992 events, respectively. The clustering of the smallest eigenvectors around the center of the projection circle indicates that all obtained solutions should be expected to have very small dips.

The average strikes obtained for the 1983 earthquake sequence can be separated into two groups: one depicting an average strike of N135°E ($\pm 20°$) and one with

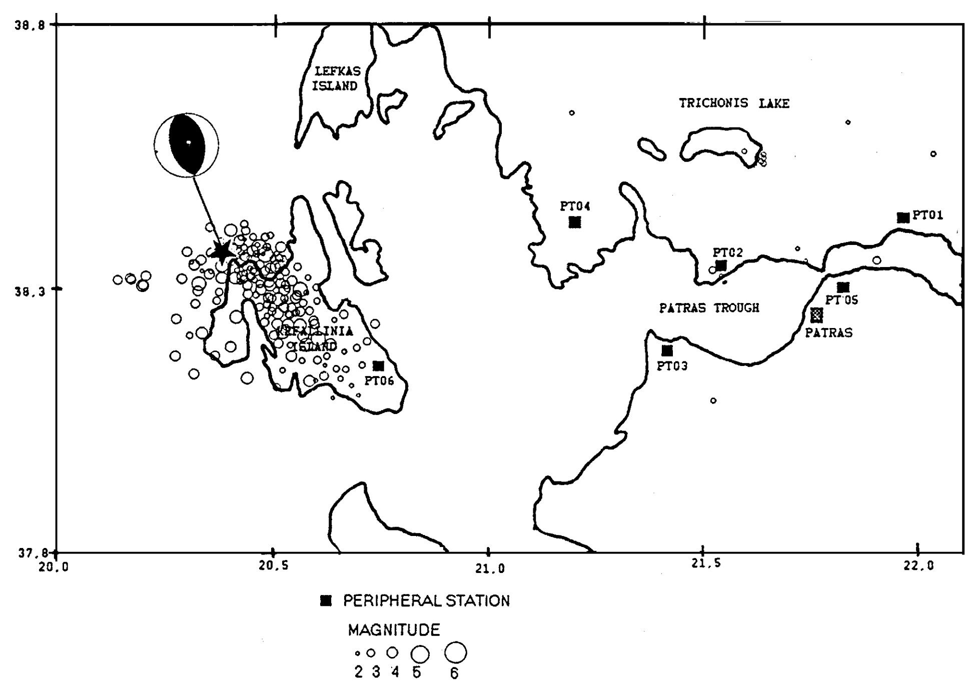

Figure 8

Map of the aftershock sequence. The epicenter of the main shock is marked with a star. The focal mechanism solution shown is taken from the USGS preliminary results (1992). The Patras Seismic Network is also shown with each station indicated by a filled square.

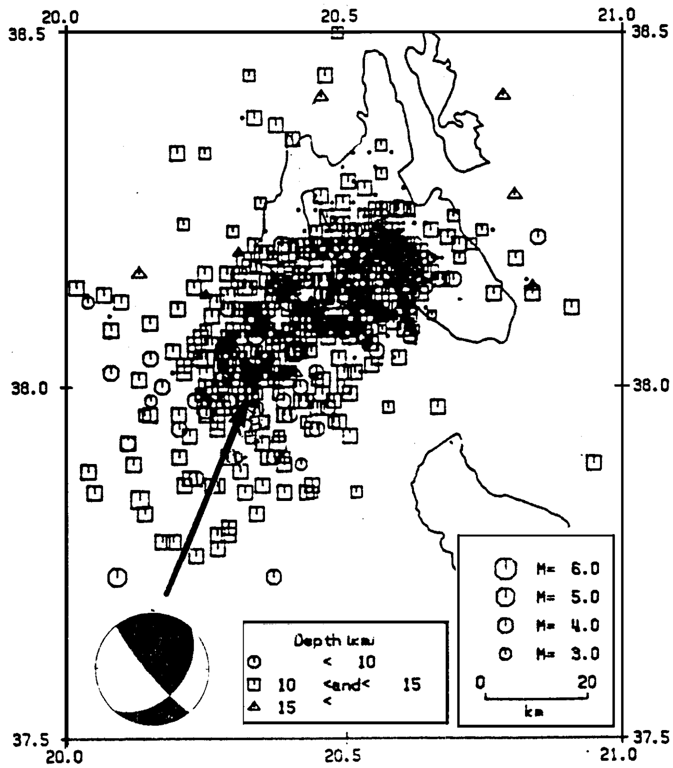

Figure 9
Map of the aftershock sequence in 1983. M is local magnitude. The focal mechanism presented for the main shock is taken from SCORDILIS *et al.* (1985).

N74°E ($\pm$10°). Similar results were obtained for the second earthquake sequence with average strikes of N156°E ($\pm$20°) and N61°E ($\pm$10°). It is worth noting that these strikes are in agreement with the dominant trends of deformation given by PAPAZACHOS *et al.* (1992) for the area. These are N174°E and N83°E, and show almost horizontal deformation, which explains the strike-slip component that exists in the area. Figure 11 depicts the above obtained average strikes and compares them with the trends of the aftershock sequences of the two events.

7. Conclusions

The analysis of the aftershock sequence of the Cephalonia Island January 23 1992 earthquake of magnitude 5.8 M_S, showed a NW–SE trending tectonic feature corresponding to the strike direction of the Argostoli and Lixouri, two of the local Neogene basins. However, a focal mechanism solution corresponding to the main

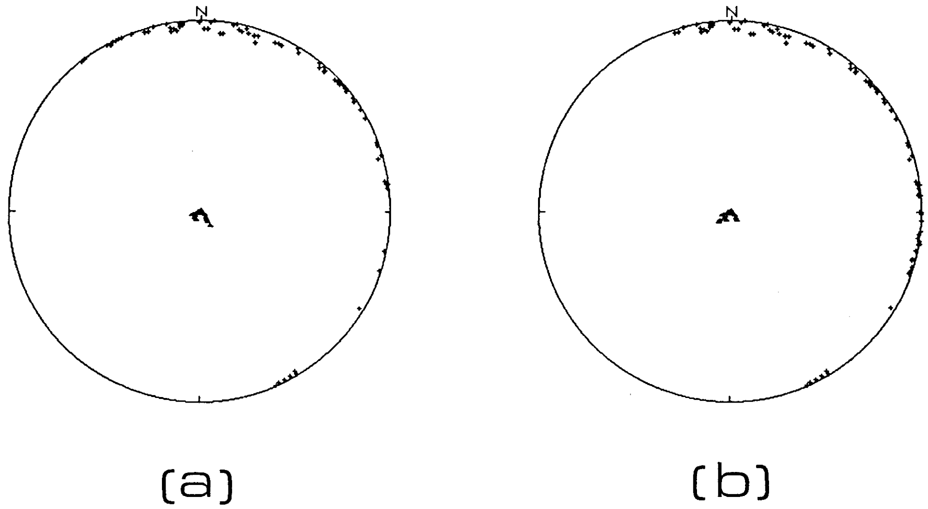

Figure 10
Lower hemisphere equal area projections of the eigenvectors obtained from the application of the principal parameters method to the aftershock sequence of the 1983 event (a) and the 1992 event (b).

shock showed a compressional regime active on faulting of similar direction, with the tectonic feature denoted by the aftershock sequence.

Strike-slip component observed on the faults bounding the basins (LEKKAS, pers. comm.) results from the transpressional or transtensional character of the basins on the island. However, the main shock being a subduction event and linked with these basins is also possible, as the uncertainty in depth determination of the aftershock sequence events does not exclude this.

Further investigation of the local seismicity is recommended by means of a denser network operating on Cephalonia. This would be expected to reveal the links in seismicity between subduction events and faulting along the boundaries of the major Miocene-Neogene basins of the island and, in particular, the possibility of reactivation of the basin-bounding faults by major subduction events.

Acknowledgments

We are grateful to Professor M. Brooks, who suggested improvements to the original manuscript and Dr. Th. Doutsos for his constructive criticism. We also thank Dr. K. Kadinsky-Cade and Prof. H. Berckhemer for their fruitful comments and recommendations.

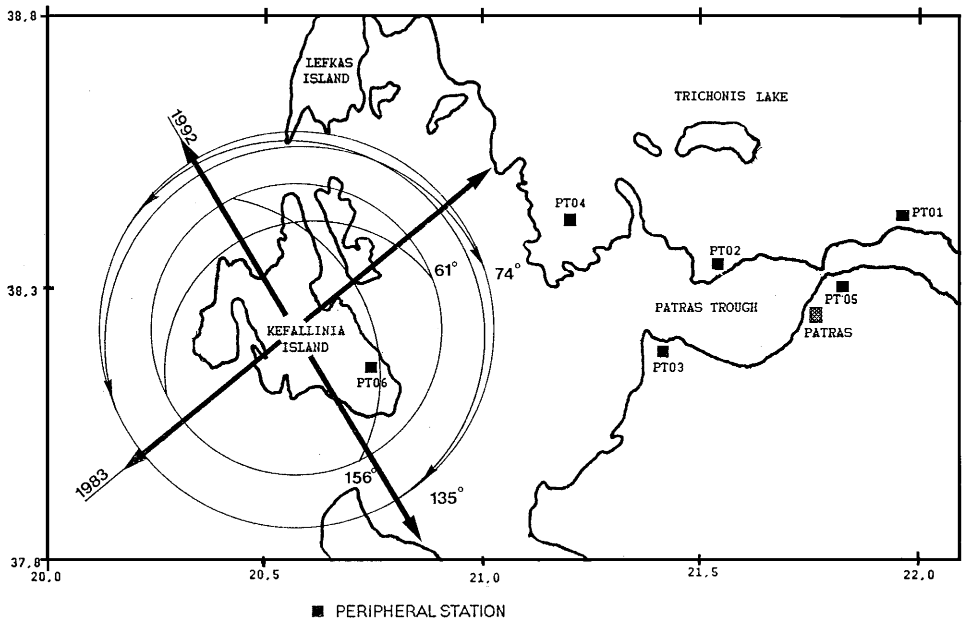

Figure 11
Diagram showing the 1983 and 1992 trends of aftershocks, marked with arrows, and the results of the principal parameters method applied to both sequences.

REFERENCES

ANDERSON, H. (1987), *Is the Adriatic an African Promontory?* Geology *15*, 212–215.

ANDERSON, H., and JACKSON, J. (1987), *The Deep Seismicity of the Tyrrhenian Sea*, Geophys. J. R. Astr. Soc. *91*, 613–637.

EBBLING, C., and MICHELINI, A. (1986), *A Principal Parameter Analysis of Aftershock Sequences: Applied to the 1977 Friuli, Italy, Sequence*, Ann. Geophys. *4*, 473–480.

FERENTINOS, G., BROOKS, M., and DOUTSOS, TH. (1985), *Quaternary Tectonics in the Gulf of Patras, Western Greece*, J. Struct. Geol. *7*, 713–717.

FINNETI, I. (1976), *Mediterranean Ridge. A Young Submerged Chain Associated with the Hellenic Arc*, Boll. Geofis. teor. applic. *15*, 263–341.

FINNETI, I. (1982), *Structure, Stratigraphy and Evolution of the Central Mediterranean Sea*, Boll. Geofis. teor. applic. *24*, 247–312.

HATZFELD, D., PEDOTTI, G., HATZIDIMITRIOU, P., and MAKROPOULOS, K. (1990), *The Strain Pattern in the Western Hellenic Arc Deduced from a Microearthquake Survey*, Geophys. J. Int. *101*, 181–202.

HATZIDIMITRIOU, P. M. (1984), *Seismogenic Volumes and Seismic Sources of the Aegean and Surrounding Areas*, Ph.D. Thesis, University of Thessaloniki.

JONES, L., and MOLNAR, P. (1979), *Some Characteristics of Foreshocks and their Possible Relationship to Earthquake Prediction and Premonitory Slip of Faults*, J. Geophys. Res. *84*, 3596–3608.

LEE, W. H. K., and LAHR, J. C. (1975), *HYPO71 (Revised): A Computer Program for Determining Hypocentre, Magnitude, and First Motion Pattern of Local Earthquakes*, U.S. Geological Survey, OPEN FILE REPORT 75–311.

LEE, W. H. K., and VALDES, C. M. (1985), *HYPO71PC: A Personal Computer Version of the HYPO71 Earthquake Location Program*, U.S. Geological Survey, OPEN FILE REPORT 85–749.

LEE, W. H. K., BENNET, R. E., and MEAGHER, K. L. (1972), *A Method of Estimating Magnitude of Local Earthquakes from Signal Duration*, U.S. Geological Survey, OPEN FILE REPORT 1–28.

LE PICHON, X., and ANGELIER, J. (1979), *The Hellenic Arc and Trench System: A Key to the Neotectonic Evolution of the Eastern Mediterranean Area*, Tectonophysics *60*, 1–42.

LE PICHON, S., and ANGELIER, J. (1981), *The Aegean Sea*, R. Soc. Lond. Phil. Trans. *A300*, 357–372.

MAKRIS, J. (1977), *Geophysical Investigations of the Hellenides*, Hamburger Geophys. Einzelschriften *34*, Wittenborn; Hamburg, Reihe A, 124pp.

MAKROPOULOS, K. C., DRAKOPOULOS, J. K., and LATOUSAKIS, J. B. (1989a), *A Revised and Extended Earthquake Catalogue for Greece since 1900*, Geophys., J. Int. *98*, 391–394.

MAKROPOULOS, K. C., VOULGARIS, N. S., DRAKOPOULOS, J. K., and TSELENTIS, A.-G. (1989b), *Analysis of Earthquake Sequences Using the Principal Parameters Method. Case Histories in Greece*. 4th Intern. Symposium on the Analysis of Seismicity and Seismic Risk. Sept. 4–9, Bechyne, Czechoslovakia, Proceed. *1*, 57–64.

MCKENZIE, D. P. (1972), *Active Tectonics of the Mediterranean Region*, Geophys. J. R. Astr. Soc. *30*, 109–185.

MCKENZIE, D. P. (1978), *Active Tectonics of the Alpine-Himalayan Belt: The Aegean Sea and Surrounding Regions*, Geophys. J. R. Astr. Soc. *55*, 217–254.

MELIS, N. S. (1986), *Geophysical Studies of the Gulf of Patras Region Using the Patras Seismic Network, 1983–84*, M.Sc. Thesis, University of Wales.

MELIS, N. S., BROOKS, M., and PEARCE, R. G. (1989), *A Microearthquake Network in the Gulf of Patras Region, Western Greece, and its Seismotectonic Interpretation*, Geophys. J. R. Astr. Soc. *98*, 515–524.

MERCIER, J.-L. (1981), *Extensional-compressional Tectonics Association with the Aegean Arc: Comparison with the Andean Cordillera of South Peru-North Bolivia*, R. Soc. Lond. Phil. Trans. *A300*, 337–355.

MERCIER, J.-L., SOREL, D., and SIMEAKIS, K. (1987), *Changes in the State of Stress in the Overriding Plate of a Subduction Zone: The Aegean Arc from the Pliocene to the Present*, Annal. Tecton. *1*, 20–39.

MERCIER, J.-L., CAREY, E., PHILIP, H., and SOREL, D. (1976), *La Neotectonique Plio-Quaternaire de l'Arc Egeen Externe et de la Mer Egee et ses Relations avec Seismicite*, Bull. Soc. Geol. Fr. *18*, 159–176.

MERCIER, J.-L., BOUSQUET, B., DELIBASIS, N., DRAKOPOULOS, I., KERAUDREN, B., LEMEILLE, F., and SOREL, D. (1972), *Deformations en Compression dans le Quaternaire des Rivages Ioniennes (Cephalonie, Grece)*, Donnees Neotectoniques et Seismiques, Academie de Sciences Comptes Rendus *275*, 2307–2310.

MICHELINI, A., and BOLT, B. A. (1986), *Application of the Principal Parameters Method to the 1983 Coalinga, California, Aftershock Sequence*, Bull. Seismol. Soc. Am. *76*, 409–420.

NATIONAL OBSERVATORY OF ATHENS (1987–1993), *Greek Network Earthquake Monthly Bulletins*. Athens.

PANAGIOTOPOULOS, D. G., and PAPAZACHOS, B. C. (1985), *Travel Times of P_n Waves in the Aegean and Surrounding Area*, Geophys. J. R. Astr. Soc. *80*, 165–176.

PAPAZACHOS, B. C. (1974), *On Certain Aftershock and Foreshock Parameters in the Area of Greece*, Annal. di Geof. *27*, 497–515.

PAPAZACHOS, B. C., KIRATZI, A. A., and PAPADIMITRIOU, E. (1991), *Regional Focal Mechanisms for Earthquakes in the Aegean Area*, Pure appl. geophys. *136*, 405–420.

PAPAZACHOS, B. C., KIRATZI, A. A., HATZIDIMITRIOU, P. M., and ROCCA, A. C. (1984), *Seismic Faults in the Aegean Area*, Tectonophysics *106*, 71–85.

PAPAZACHOS, C. B., KIRATZI, A. A., and PAPAZACHOS, B. C. (1992), *Rates of Active Crustal Deformation in the Aegean and the Surrounding Area*, J. Geodyn. *16*, 147–179.

PEDOTTI, G. (1988), *Etude Sismotectonique du Peloponnese et Reponse Sismique d'une Vallée Sedimentaire en Grece du Nord*, These, Universite Joseph Fourier, Grenoble.

RITSEMA, A. R. (1974), *The Earthquake Mechanism of the Balkan Region*, Netherl. Meteorol. Inst. Sci. Rep. *74-4*, 1–36.

SCORDILIS, E. M., KARAKAISIS, G. F., KARAKOSTAS, B. G., PANAGIOTOPOULOS, D. G., COMNINAKIS, P. E., and PAPAZACHOS, B. C. (1985), *Evidence for Transform Faulting in the Ionian Sea: The Cephalonia Island Earthquake Sequence of 1983*, Pure appl. geophys. *123*, 388–397.

SOREL, D. (1976), *Etude néotectonique des isles ioniennes de Cephalonie et Zanthe et de L'Elide occidentale (Grece)*, These du 3e cycle, Universite de Paris-Sud, Centre d'Orsay.

SOREL, D., NESTEROFF, W. D., LIMOND, J., LEMEILLE, F., and SEBRIER, M. (1976), *Mise en Evidence de Structures Compressives Sous-Marine Plio-Pleistocene dans l'Arc Egeen Externe au Large de Levkas (Isles Ioniennes, Grece)*, Paris, Academie des Sciences Comptes Rendus *282*, 2045–2048.

TSELENTIS, A.-G. (1989), *Stress Tensor Determination at the Aftershock Area of Kalamata, Sept. 13, 1986 (S. Greece) Earthquake, Using Akaike's Bayesian Information Criterion (ABIC), and Coulombs Failure Criterion*, 25th General Assembly, IASPEI, Istanbul, Turkey.

UNDERHILL, J. R. (1988), *Triassic Evaporites and Plio-Quaternary Diapirism in Western Greece*, J. Geol. Soc. London *145*, 269–282.

UNDERHILL, J. R. (1989), *Late Cenozoic Deformation of the Hellenide Foreland, Western Greece*, Bull. Geol. Soc. Am. *101*, 613–634.

U.S. DEPT. OF INTERIOR/GEOL. SURVEY (1992), *Preliminary Determination of Epicentres*, National Earthquake Information Center, Monthly listing, January 1992.

UTSU, T. (1961), *A Statistical Study on the Occurrence of Aftershocks*, Geophys. Mag. *30*, 521–605.

(Received May 10, 1996, accepted April 5, 1997)