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A microearthquake study in the Gulf of Patras region, western Greece, and its seismotectonic interpretation

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SUMMARY

The seismicity of the Gulf of Patras region was monitored from 1983 May to 1984 April by a six-station microearthquake network with 30 km aperture. Over 650 events were located using the revised version of the HYPO71 computer program for determining hypocentral locations. Of these, a subset of 73 well-located events was identified for detailed study.

Analysis of these 73 events reveals a local pattern of seismicity with two main features: firstly, a gap in seismicity in the central and western parts of the Gulf, despite the local presence of active faulting affecting the seabed; secondly, a zone of high seismicity in the northeast part of the Gulf where offshore seismic surveys reveal a zone of intense structural complexity. A northeast dipping, diffuse zone of seismicity extending to a depth of about 25 km has been identified in the northeastern corner of the Gulf, centred on the Rio–Antirrio Channel between the Gulfs of Patras and Corinth.

It is postulated that the active faults in the central part of the Gulf, which are apparently aseismic, are linked via a shallow decollement surface under the Gulf to a deep, WNW–ESE trending fault zone defining the southern margin of the Corinth graben, and extending westwards to Trichonis Lake. The dipping zone of seismicity is related to the latter fault zone, and aseismic faulting in the Gulf of Patras is attributed to the shallow surface of decollement into which the faults sole. A transfer fault zone trending approximately NE–SW is postulated to account for the offset of the Corinth and Patras grabens.

Key words: Microearthquake, Gulf of Patras, seismotectonics

1 INTRODUCTION

The Gulf of Patras contains a WNW–ESE trending neotectonic graben structure (Ferentinos, Brooks & Doutsos 1985). It lies at the western end of a regional system of three grabens (Patras–Corinth–Saronicos; Fig. 1), extending from the vicinity of the western Hellenic trench to the Cyclades volcanic arc of the southern Aegean. Both of these major geological features are associated with active subduction of the eastern Mediterranean lithosphere beneath the Aegean domain. The graben system is a localized manifestation of crustal extension in the Aegean marginal basin behind the active subduction zone of the outer Hellenic arc. Available fault plane solutions for the last 30 yr show the prevailing pattern of crustal stress to involve approximately N–S extension over much of the Aegean region, including the present study area (Ritsema 1974; McKenzie 1978; Drakopoulos & Delibasis 1982; Papazachos *et al.* 1984; Papadopoulos *et al.* 1986; Fig. 2).

North and south of the Gulf, a thin and discontinuous cover of Plio-Quaternary sediments unconformably overlies tectonized Mesozoic and Cenozoic rocks of the Gavrovo and Ionian isopic zones of the outer Hellenides. In particular, these are Cretaceous through Eocene limestones with intervening outcrops of Oligo-Miocene flysch, and they are underlain by evaporites of presumed Permo-Triassic age (British Petroleum Co. Ltd 1971).

Seismic profiling at 3.5 kHz in the Gulf (Ferentinos *et al.* 1985, figs 2 and 3) has shown that offshore Plio-Quaternary sediments are affected by widespread active faulting with a WNW–ESE trend. In the northeast part of the Gulf, profiling records show that concealed faulting affects Pleistocene sequences beneath an undeformed Holocene surface layer. In the immediate vicinity of the Rio–Antirrio Channel (Fig. 3), there is a complicated system of active faults with several trends (Perisoratis, Mitropoulos & Aggelopoulos 1986; Chronis *et al.* 1987).

Onshore neotectonic data from around the Gulf show two groups of normal faults trending WNW–ESE and ENE–WSW (Ferentinos *et al.* 1984, 1985; Doutsos, Kontopoulos & Ferentinos 1985; Doutsos, Kontopoulos & Frydas 1987). To the east of the Gulf NNW–SSE and NNE–SSW normal faulting is observed (Kontopoulos & Doutsos 1985; Doutsos *et al.* 1987). In addition, near-vertical faults have been observed in the same area that cannot be characterized as either dip–slip or strike–slip for lack of fault-plane lineations.

In marked contrast to this N–S extensional regime, large-scale compressional tectonics on NW–SE trending lines characterize the area to the west, in the Kefallinia–

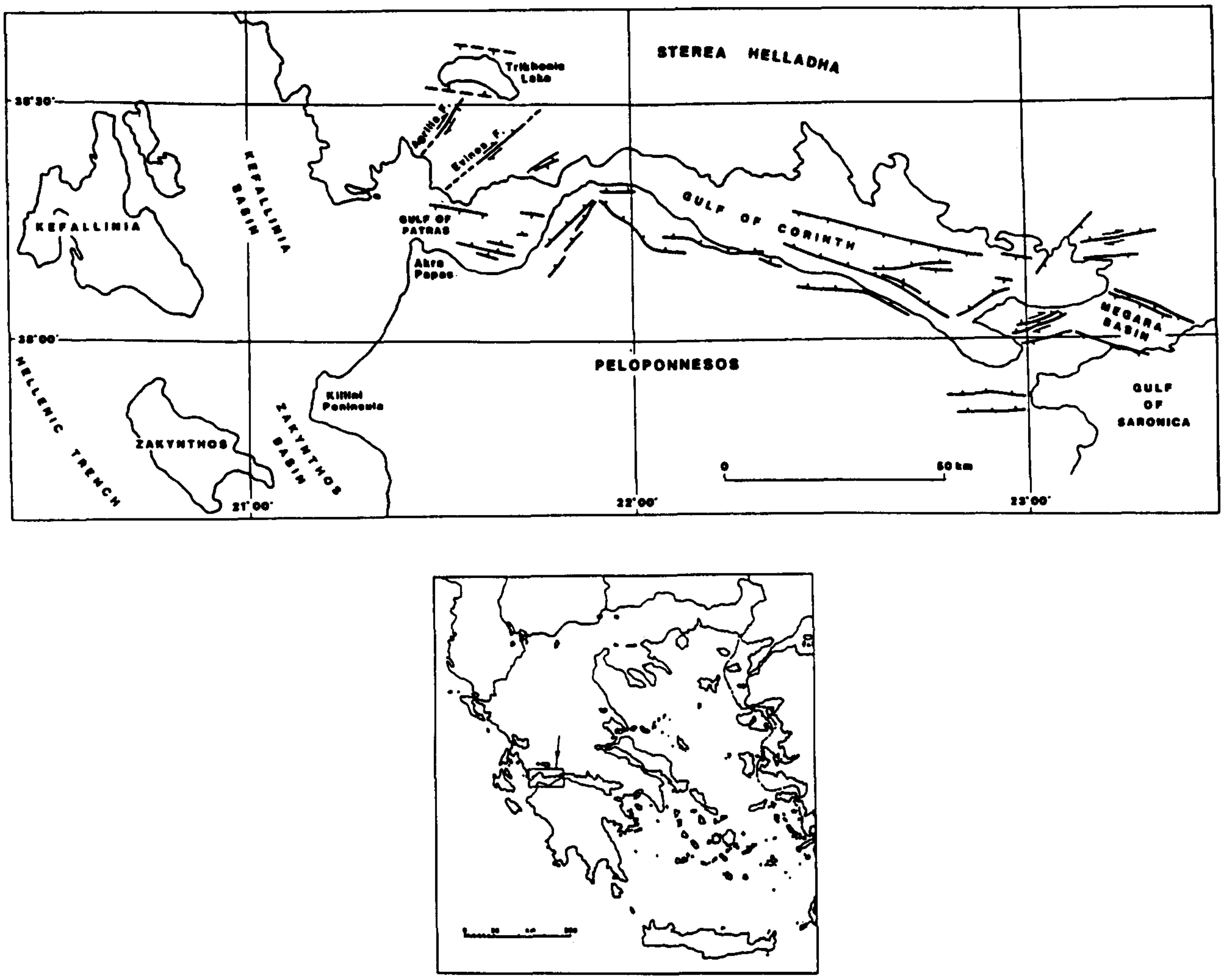

Figure 1. Location map and regional system of neotectonics showing associated grabens (after Ferentinos *et al.* 1985).

Zakynthos Channels, as indicated by seismic profiling surveys (Monopolis & Bruneton 1982; Brooks & Ferentinos 1984). This abrupt change of tectonic regime coincides in position with a change in the fault plane solutions of shallow earthquakes, which indicate E–W compression in this western region (Ritsema 1974; McKenzie 1978; Papazachos *et al.* 1984; Papadoupoulos *et al.* 1986).

The boundary between the two zones of different tectonic style and prevailing stress regime, as deduced from fault plane solutions, passes through the outer part of the Gulf of Patras. This key area represents a marked seismic gap for earthquakes with magnitude larger than $M_s = 4.5$ for the past century (Makropoulos & Burton 1981; Comninakis & Papazachos 1982). Hence, the Gulf of Patras represents an area in which more information on local seismicity might be expected to lead to a deeper understanding of Aegean neotectonics.

2 PATRAS SEISMIC NETWORK

From 1983 May to 1984 April a telemetric microearthquake network (Patras Seismic Network—PATNET) was operated around the Gulf of Patras. PATNET was intended as a preliminary network to establish the local microearthquake distribution in sufficient detail to facilitate the design of a longer term and denser network (Pearce 1984; Melis, Brooks & Pearce 1986). Hence, it consisted of only six stations (PT1 to PT6) numbered clockwise around the Gulf (Fig. 3). Table 1 gives the station locations; the same locations were maintained throughout the recording period except where sites were moved locally as shown to improve recording quality.

PT2 to PT6 were outstations linked by radio through line-of-sight telemetry to receivers at the base station PT1. All outstations consisted of one vertical component, short period Willmore MkIIIA seismometer. Two additional MkIIIA seismometers were operated as horizontals (E–W, N–S) at PT1. At the base station the three component set, the telemetry receivers and a time signal receiver (for RWM Moscow) were linked by cable to a Racal Geostore tape recorder, recording at 15/320 ips. Thus all data were recorded on a single analogue magnetic tape with a common time base provided by the internal clock. The RWM time signal was used as a reference to Universal Time.

The station distribution was inadequate for the determination of very accurate hypocentral locations, especially in view of the absence of sea-bottom seismometers within the Gulf. However, it was found that epicentral parameters

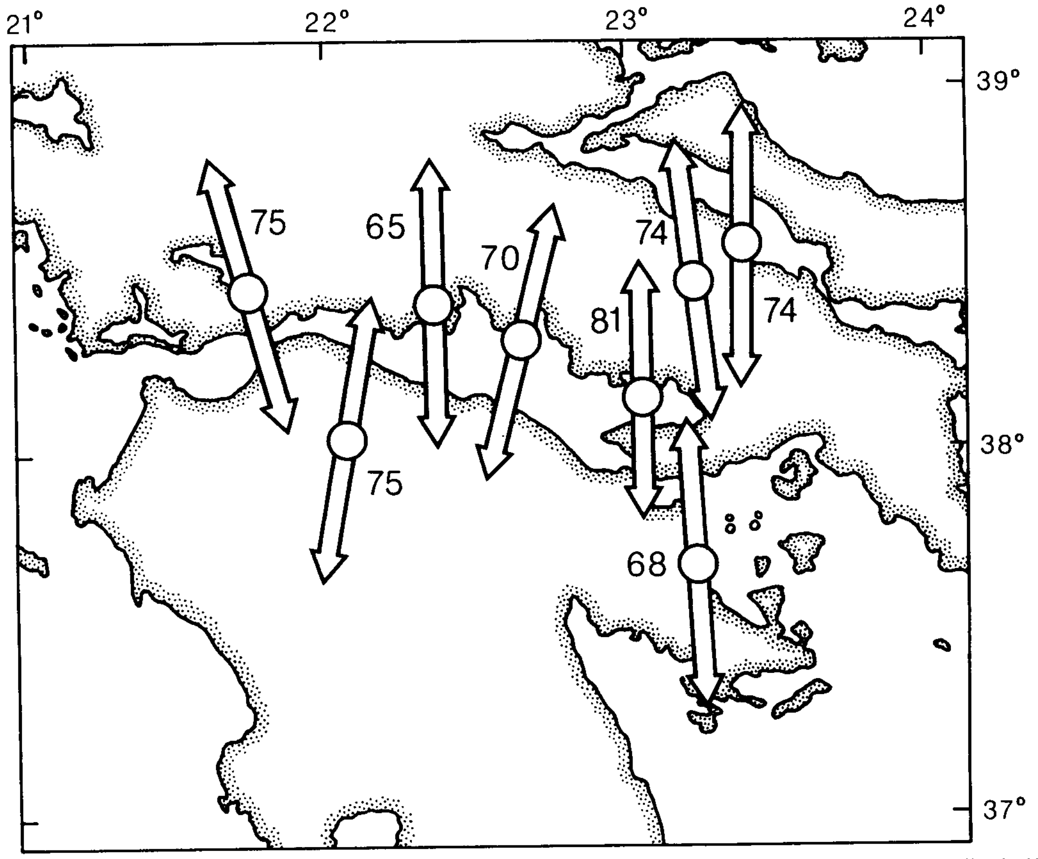

Figure 2. Surface projection of fault plane solutions in central Greece (after Papazachos 1976; Drakopoulos & Delibasis 1982; Papazachos *et al.* 1984; Papadopoulos *et al.* 1986). Arrows indicate the direction of extension and numbers the year of occurrence.

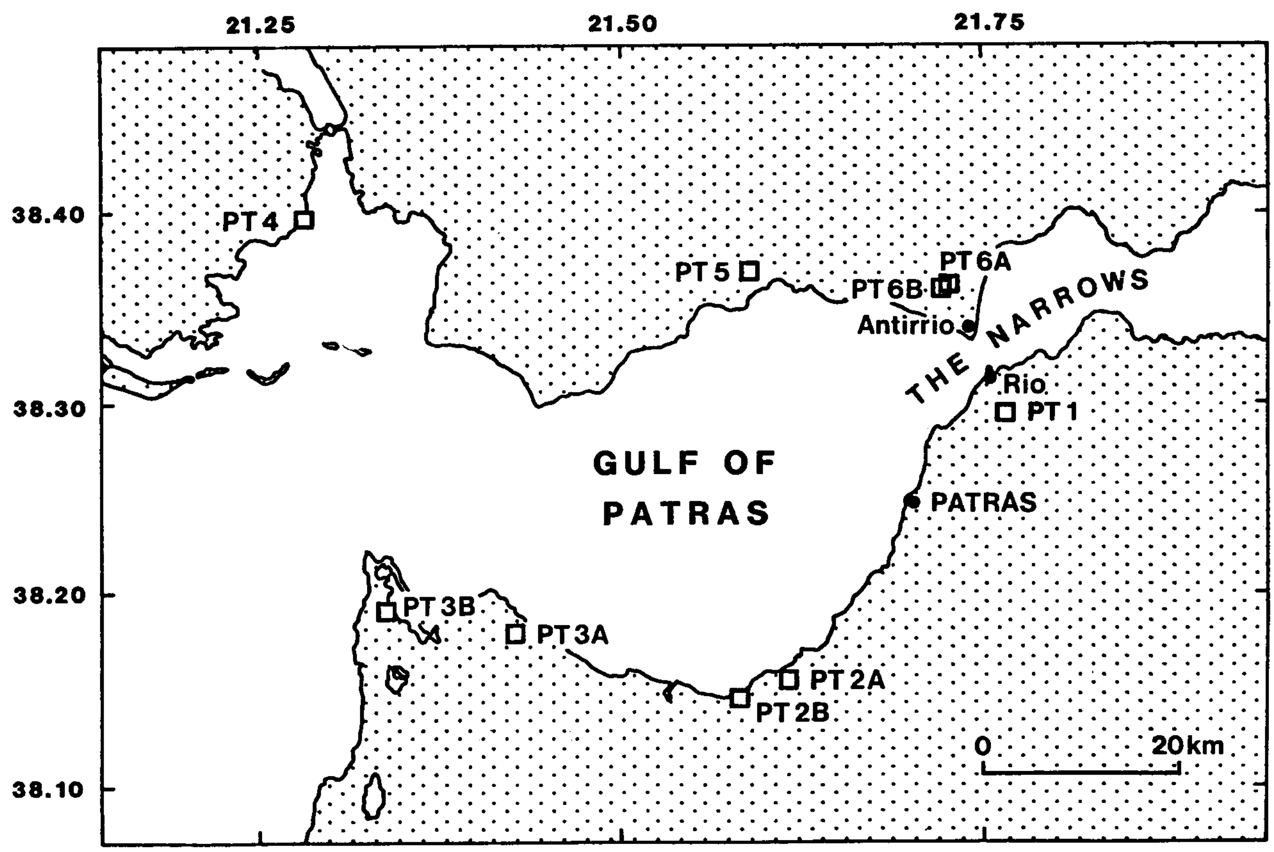

Figure 3. PATNET station locations.

Table 1. PATRAS network station details

No.	Station code	Station site	Component[a]	Latitude N	Longitude E	H (m)	Geological foundation
1	PT1	University of Patras	V E–N N–S	38 17.35	21 47.32	70	Pliocene conglomerate
2	PT2a	Tsoukaleika	V	38 09.17	21 38.73	45	Pliocene conglomerate
	PT2b	Kaminia	V	38 08.58	21 36.80	25	Quaternary marine deposit
3	PT3a	Lakkopetra	V	38 10.59	21 27.99	85	Eocene flysch
	PT3b	Araxos	V	38 11.29	21 22.95	45	Cretaceous limestone
4	PT4	Agia Trias	V	38 23.43	21 19.91	05	Pliocene lagoonal deposit
5	PT5	Kato Vasiliki	V	38 21.72	21 37.21	50	Eocene flysch
6	PT6a	Molykreion	V	38 21.30	21 45.09	80	Pliocene lacustrine deposit
	PT6b	Molykreion	V	38 21.18	21 44.84	55	Pliocene lacustrine deposit

Centre of PATNET PTcp at 38 16.005N 21 33.615E

[a] Willmore MKIIIA seismometers with free period set to 1.0 s and damping to 0.7.

of events within the network were adequately determined when detected by four or more network stations. Focal depths were generally less well determined, both because of the small number of stations and because of extreme lateral variations in velocity structure.

3 DATA PROCESSING

A visual method for event detection was used (Melis 1986). All analogue magnetic tapes were replayed on a Racal Store-14 tape deck and seismograms with time signals were played out at slow speed using a jet pen recorder. This system produced records with a compressed time scale suitable for event identification and time code reading. All detected events were then registered on a scan list, which consisted of event identification (tape number/event number) and a reference time 5 s before the earliest P-arrival at the network. A total of 665 events were detected by at least four stations during a 6 month recording period representing 60 per cent of the overall operating time. Of these events, 301 were recorded by at least five stations.

All events were digitized using a semiautomated ADC system driven by a BBC microcomputer (Melis 1986). Screen display was used for the identification of P- and S-arrivals, length of coda, and maximum amplitude and its period for all six stations. All measurements for each event then were copied on to floppy disc and transferred to a Honeywell main-frame computer for analysis.

4 DATA ANALYSIS

The revised version of the event location computer program HYPO71 (Lee & Lahr 1975; Lee & Stewart 1981) was used to calculate location parameters for all events. In view of the small number of stations and their distribution, careful attention was given to the calculation of epicentral location and the estimation of errors. In some cases, the focal depth could not be estimated accurately. The criteria used to accept the epicentral and depth computations of HYPO71 were:

(i) a minimum of four P-arrivals for epicentral, and five for depth, computation,
(ii) the use of S-arrivals where available and when they improved the location error,
(iii) a maximum of 0.15 s in computed rms error in travel-time residuals,
(iv) a maximum of 5 km in computed location errors,
(v) the epicentral distance to the nearest station (which should approximately equal the computed focal depth; Lee & Stewart 1981), and
(vi) a calculated focal depth not fixed at its initial value.

Another important factor is the assumed crustal model. For the PATNET data an initial model was based on crustal data obtained from the area of Greece and the Ionian Sea by Makris (1977, 1978a,b), together with surface geological data at station sites (IGME 1:50 000 geological map series, British Petroleum Co. Ltd 1971; Ferentinos *et al.* 1985; Kontopoulos & Doutsos 1985). Details of the crustal model are given in Table 2. Using this model, HYPO71 was used to compute initial locations for all events. On the basis of the above criteria 30 well-located events within the network, recorded by at least five stations, were chosen for further study. These events were used to investigate the validity of a half-space P-velocity model in place of the layered crustal model that had been selected initially. Various models with V_P values from 4.0 to 6.4 km s^{-1} were tested. Plots of computed rms errors of travel-time residuals versus half-space V_P values were produced for each event (Fig. 4). Then the value of the half-space V_P at the minimum rms

Table 2. Makris crustal model used for initial hypocentral locations

Depth (km)	P-Velocity (km s^{-1})
0–5	5.0
5–30	6.0
30–47	6.6
47	7.9

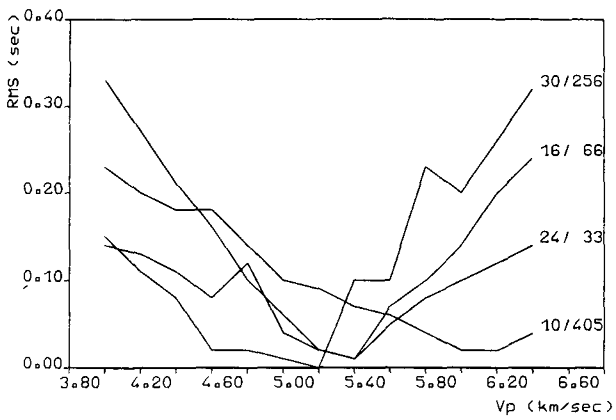

Figure 4. The rms travel-time residuals as a function of half-space V_P values for some of the 30 selected events.

time residual was picked and a histogram was plotted showing the distribution of half-space V_P values yielding a minimum rms time residual (Fig. 5). A weighted mean velocity was calculated and found to be $5.47\,\mathrm{km\,s^{-1}}$. Hence, a half-space velocity model with $V_P = 5.5\,\mathrm{km\,s^{-1}}$ was selected, close to the velocity of $5.7\,\mathrm{km\,s^{-1}}$ used by King *et al.* (1985) for hypocentral determinations in the nearby Gulf of Corinth area.

In the absence of good velocity/depth information, the $5.5\,\mathrm{km\,s^{-1}}$ half-space *P*-velocity crustal model was used in HYPO71 in place of the layered crustal model of Makris to derive locations for all events, without any adverse effect on the computed hypocentres (Melis 1986).

In most cases, only *P*-arrival data were used; only in a few cases was an *S*-arrival at PT1 incorporated into the analysis to improve the final location. Of the 665 events located (Fig. 6), only 73 proved to satisfy all the quality criteria mentioned above, including at least five *P*-arrivals, and these are shown in Fig. 7. Most of the 665 events provided good epicentral locations but only for 73 events was a satisfactory focal depth computed. For these 73 events the computed focal depth location error was less than 5 km. Only these 73 well-located events were used for analysis of the regional seismicity.

PATNET was designed to detect seismicity of local magnitude M_L less than 4. Local magnitudes were computed

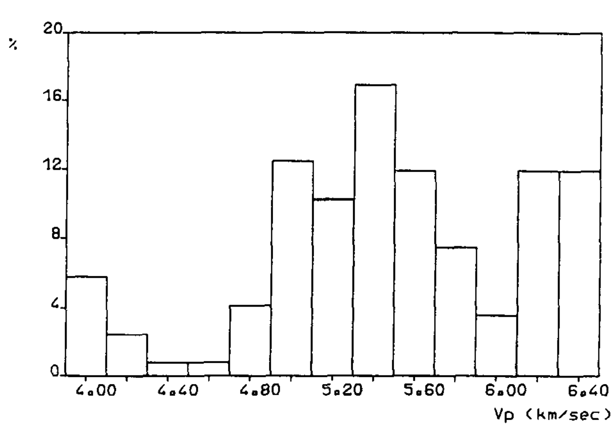

Figure 5. Distribution of V_P (half-space) values at the minimum rms time residual for each of the 30 test events.

using the signal duration of each event at each station. The signal duration was measured for PATNET data as the time in seconds from the *P*-onset to the point where the signal fell to the background noise-level. An empirical formula defined by Kiratzi & Papazachos (1985) for earthquakes recorded by short period seismographs in Greece was used

$$M_L = 2.31 \log(T) + 0.0012D + C,$$

where M_L is the local magnitude, T is the signal duration, D is the epicentral distance (km) and C is a station constant. The constant C was calculated for each station in a least squares sense (Table 3), using seven events which had been assigned a local magnitude M_L by the National Observatory of Athens (Monthly Bulletins 1983).

Hence, defining the constant C for each station, the magnitudes were estimated using HYPO71, replacing the constants of Lee, Bennett & Meagher (1972) by the newly calculated values. Individual calculations of local magnitude at each station were averaged to produce an estimate of local magnitude for each event.

5 SEISMICITY PATTERN

The seismicity pattern (Fig. 7) determined from the PATNET microearthquake data reveals two well defined zones of different seismicity. The area of the narrow strait linking the Gulfs of Patras and Corinth represents a localized region of high seismicity. The central–western part of the Gulf of Patras represents a seismicity gap within which no event was located. This gap exists over a wide range of magnitudes: it is observed here for magnitudes less than $M_L = 4.5$ and in Greek earthquake catalogues (Makropoulos & Burton 1981; Comninakis & Papazachos 1982) for events with magnitudes greater than $M_S = 4.5$ on the Richter scale. Even when all the events in Fig. 6 are included the seismicity pattern persists. This overall pattern can be seen also in the results from two other recent networks established in mainland Greece, namely, the Volos Network (VOLNET) operating from 1983 to 1985 (VOLNET Monthly Bulletins; Melis & Burton 1988) and a network in the Peloponnesos operating for a 6-week period during 1986 (Pedotti 1988).

The part of the Gulf lacking seismicity contains normal growth (synsedimentary) faults extending up to the seafloor (Ferentinos et al. 1985). In the northeastern part of the Gulf, concealed inactive faults under a cover of undeformed Holocene sediments have been observed by Ferentinos *et al.* (1985). However, in the immediate vicinity of the Rio–Antirrio Channel, characterized by a high concentration of seismicity, a complicated system of active faults with several trends has been investigated by Perisoratis *et al.* (1986) and by Chronis *et al.* (1987). It may be noted that ENE–WSW, NNW–SSE and NNE–SSW fault trends are observed locally onshore in Plio-Quaternary sediments (Kontopoulos & Doutsos 1985; Doutsos *et al.* 1985, 1987; Mettos, Rondoyianni & Papadakis 1987). The concentration of seismicity is evidence of high tectonic activity in this zone of structural complexity. The concealed faults in the northeastern part of the Gulf seem to lie at the boundary between this zone and the western aseismic zone.

Generally, events within or close to the network are shallow, lying within a depth range of 2–25 km. A NE–SW

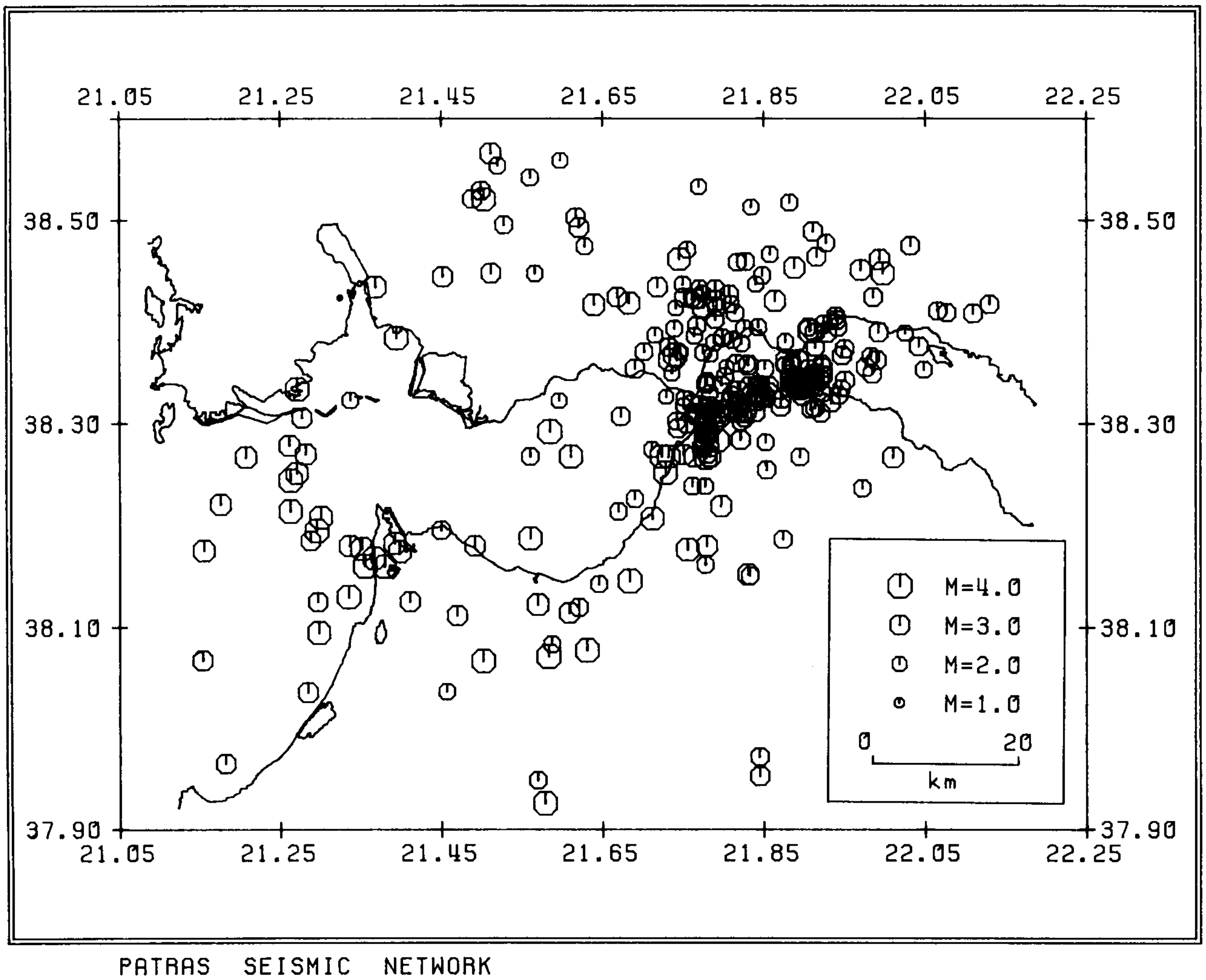

Figure 6. Distribution of all 665 events recorded by PATNET with M_L magnitudes.

section across the high concentration of seismicity in the northeastern part of the Gulf shows the events to lie in a poorly defined dipping zone with focal depths increasing to the northeast (Fig. 8). Further examination of HYPO71 locations computed for events plotted in this cross-section was made, in order to test the reality of the dipping zone, as it was considered to be possibly due, in part, to the station geometry. Two tests were conducted. First, random error in the range of ± 0.3 s was assigned on observed arrival times of four chosen events and new locations were determined using HYPO71. This was repeated twenty times for each event, to give a representation of the shape of the confidence surface appropriate to that event. The purpose of this test was to establish whether the observed feature was merely an artifact of such confidence surfaces. For all events, the observed features of the confidence surface after conducting this test were dissimilar in shape to the observed dipping zone. In the second test, the focal depth for all 73 events was fixed at a range of values from 5 to 40 km at 2.5 km intervals, and a best location (lowest possible rms time residual) was computed for each depth. This was to test whether a true depth convergence at the final focal depth had occurred during location by HYPO71. For all events it was shown that the convergence existed and the location was computed at the depth with the minimum rms residuals. Both tests thus support the reality of the dipping zone. It may be noted that this same zone has been observed in the results from the 1986 Peloponnesos network (Pedotti 1988, fig. 4.6).

6 INTERPRETATION

The dipping zone of seismicity (Fig. 8) can be interpreted as a major fault plane dipping to the northeast, and it may therefore represent a westward continuation of the major basin-bounding fault at the southern margin of the Gulf of Corinth graben (Brooks & Ferentinos 1984). In addition, one of the fault plane solutions given by Papazachos (1976), for earthquakes (38.4°N, 21.7°E, 1975 June 30, December 31) immediately northeast of the Gulf of Patras, determined almost N–S extension (see Fig. 2). Thus, there is evidence for a westward prolongation of the zone of crustal extension occupied by the Gulf of Corinth graben.

It needs to be considered how this postulated westward extension of the deep fault defining the Corinth graben might be related to the pattern of neotectonic faults in the

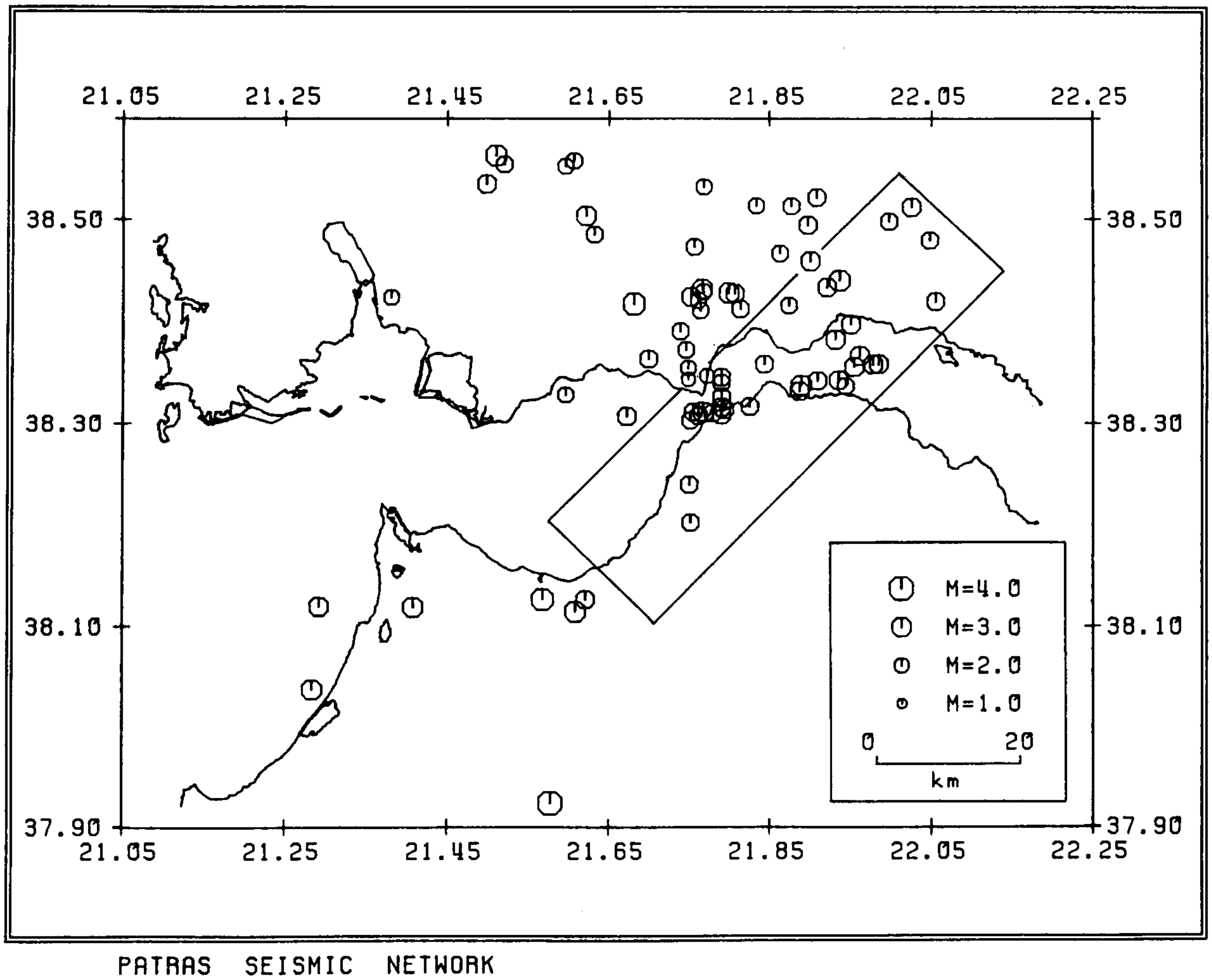

Figure 7. Distribution of the 73 well-located events recorded by PATNET with M_L magnitudes. Box indicates area from which data have been selected to construct the cross-section of seismicity illustrated in Fig. 8.

Table 3. Value of constant *C* for each station, used in determining local magnitude M_L from signal duration

Solution	Code	constant *C*
University of Patras	PT1	−0.73
Tsoukaleika	PT2a	−1.10
Kaminia	PT2b	−1.00
Lakkopetra	PT3a	−0.96
Araxos	PT3b	−1.00
Agia Trias	PT4	−0.68
Kato Vasiliki	PT5	−1.04
Molykreion	PT6b*	−0.78

* PT6a records were not used for event location.

Gulf of Patras area. There, surface faults are observed within the seismic gap and, to the northeast, a complicated system of neotectonic faulting occurs in the zone of high seismicity. A possible explanation is provided by the model illustrated in Fig. 9. The Gulf of Corinth structure is an asymmetrical graben (Brooks & Ferentinos 1984), and its southern bounding fault is postulated to be a listric fault flattening off into a mid-crustal decollement. Fault plane solutions indicate almost N–S crustal extension in this area (Papazachos 1976; King *et al.* 1985; Papadopoulos *et al.* 1986; Fig. 2). Thus, the Gulf of Corinth graben can be schematically defined as shown in Fig. 9.

The southern part of the Gulf of Patras is characterized by active surface faulting which, if dip slip, again indicates almost N–S extension. However, no seismic events are located in the area. The association of surface faulting with aseismicity can be explained if the faults in the Gulf of Patras pass down into a shallow zone of decollement, most probably at the level of the Permo-Triassic evaporites that are known, from offshore drilling, to underlie the Neogene and Quaternary sediments of the Gulf (Ferentinos *et al.* 1985). It is reasonable to postulate that such a decollement zone links northwards with the deep fault associated with high seismicity extending westwards from the Gulf of Corinth.

On this model, the offset of the Corinth and Patras grabens is achieved by means of a transfer fault zone trending aproximately NE–SW near to the southeastern coastline of the Gulf of Patras and through the Rio–Antirrio Channel (Fig. 9). ENE–WSW faulting in adjacent onshore regions has been characterized as normal by Doutsos *et al.*

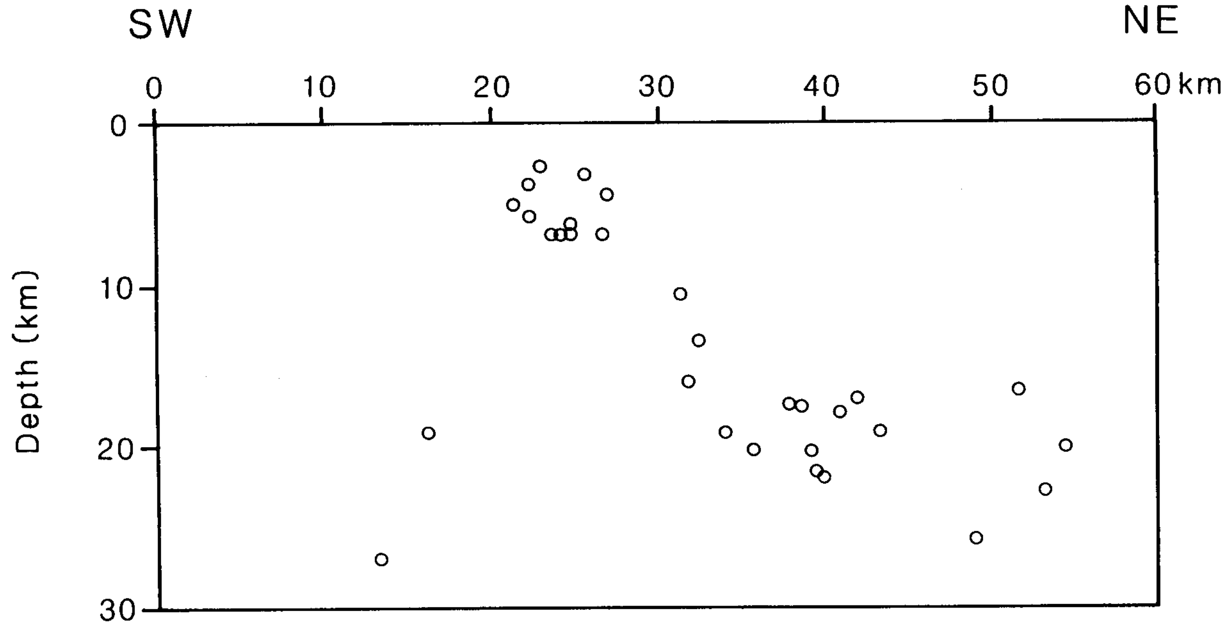

Figure 8. Cross-secton centred on latitude 38.3°N, longitude 21.9°E, showing the dipping zone of seismicity within a vertical crustal slice with a width of 20 km.

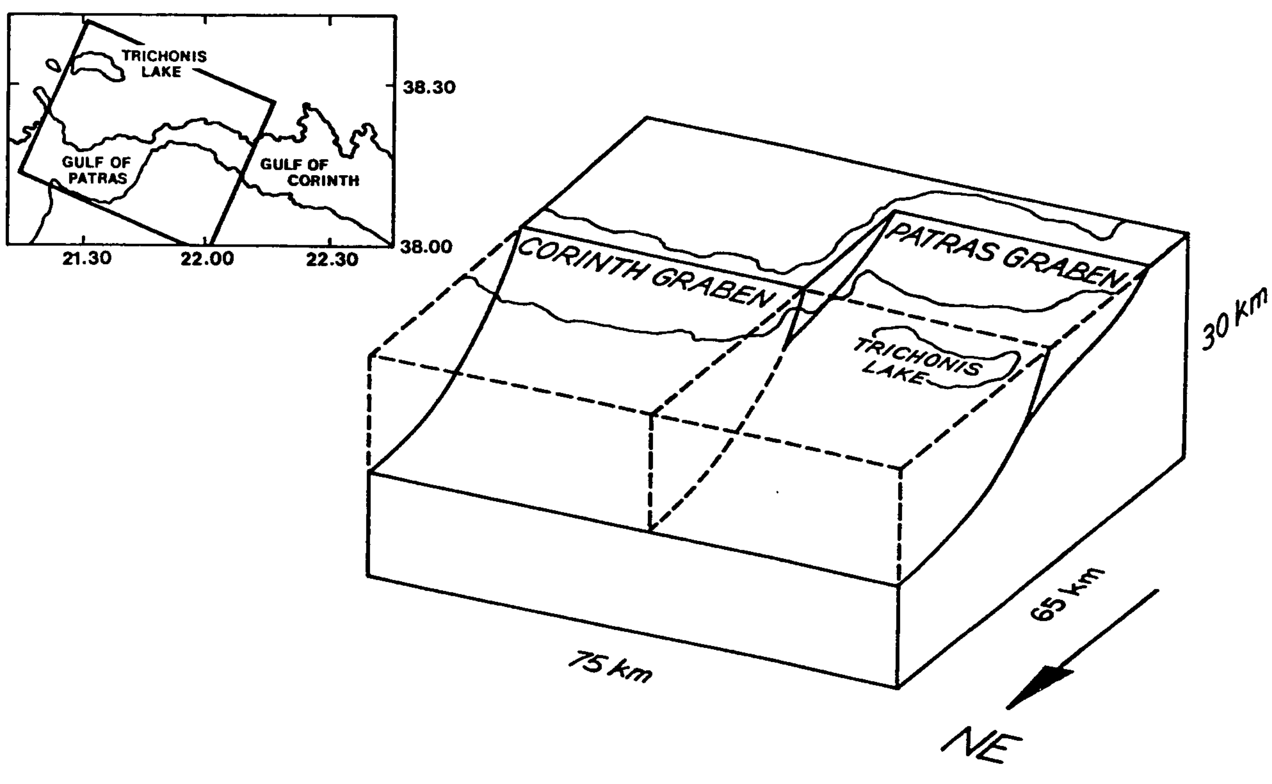

Figure 9. Possible footwall geometry of an extensional fault system in the Corinth–Patras–Trichonis region.

(1987) but marine seismic profiling surveys (Perisoratis *et al.* 1986; Chronis *et al.* 1987) show the offshore zone to be structurally complex, and more work is required to test the hypothesis of dextral strike-slip displacements in this offshore zone. It may be noted, however, that the recent study of seismicity in the Peloponnesos (Pedotti 1988) provides focal mechanism solutions indicating dextral strike-slip faulting along the postulated transfer zone.

Ideas advanced in this paper could be tested more rigorously by means of a denser seismic network in the same area recording over a longer period, as originally planned. This would facilitate the determination of focal mechanisms for local earthquakes, together with more accurate hypocentral and structural parameters.

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