Seismicity and associated strain of central Greece between 1890 and 1988

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SUMMARY

We examined the seismicity of central Greece between 1890 and 1988, using macroseismic and instrumental data, to ask two questions: (1) does the seismicity of this period reveal all the major tectonic structures that are known to be active?; and (2) what are the likely strains associated with the seismicity over this period? Many known active structures have been effectively aseismic for the last hundred years, and even the inclusion of all known large events earlier than 1890 reveals no activity associated with the NE coast of Evia, Gulf of Argos, or graben NE of Mt Parnassos. It is clear that even 100 years' data are inadequate for either a reasonable assessment of seismic risk or for a confident estimation of maximum magnitude. However, we are aware of no earthquakes in central Greece during the last 200 yr that were larger than $M_s 7.0$. It is probable that the maximum magnitude is restricted by the maximum length of fault segments, which appears to be around 15–20 km. The earthquakes of $M_s \geqslant 5.8$ during 1890–1988 can account for a N–S displacement of around 45–70 cm (with maximum and minimum estimates a factor of two greater and smaller than this) across part of a 1890–1900 triangulation network in central Greece that was resurveyed in 1988. The contribution of smaller events may increase this displacement by about 50 per cent. This cumulative seismic displacement is similar to that estimated from the geodetic work (about 100 cm), but a detailed comparison of the two sets of observations will be reported elsewhere. A re-evaluation of all the important earthquakes of 1890–1988 in central Greece is presented in the Appendix, which summarizes information of use to both earth scientists and engineers.

Key words: earthquake damage, earthquakes, Greece, hazard, magnitude, strain.

INTRODUCTION

In this paper we examine the seismicity of central Greece (36.3°–39.5°N, 21.0°–25.5°E) over a hundred year period, from 1890–1988. We attempt to assess the location, magnitude, and likely style of faulting for all earthquakes with $M_s \geqslant 5.8$, and then to estimate the total strain across central Greece in this period. There were two principal reasons for carrying out this study. The first was to see whether the seismicity over one hundred years correlates better with the active structures revealed in the geomorphology than does the seismicity over the last 25 yr for which good instrumental data are available. The second was to estimate the overall strain accounted for by the seismicity in the period 1890–1988, in order to compare it with the strain measured geodetically over the same time interval.

The geodetic results, and their detailed comparison with the strain estimated from the seismicity, will be reported elsewhere.

The Aegean Sea, and its surrounding coastal regions, is one of the most rapidly extending areas on the continents today, with a N–S rate of extension across the whole province, i.e. between Crete and Bulgaria, of around 40–60 mm yr^{-1} (McKenzie 1978; Le Pichon & Angelier 1979; Jackson & McKenzie 1988b). Jackson & McKenzie (1988a, b) examined the seismicity of the Aegean extensional province during the period 1908–1981 and estimated that earthquakes with $M_s > 5.9$ accounted for a N–S extensional rate, averaged over the whole province, of 20–60 mm yr^{-1}. Thus the seismicity apparently can account for most of the expected regional strain, unlike for example, in the Zagros mountains of Iran or in the Hellenic Trench,

where the most of the strain must occur aseismically (Jackson & McKenzie 1988a, b). Jackson & McKenzie's studies estimate seismic moments from surface wave magnitudes using empirical $M_s:M_o$ relations. The M_s values they used had been recalculated from amplitude and period data and are presented here in detail for the first time.

The tectonic activity in central Greece is dramatically expressed in the topography, geomorphology and movement of the coastline relative to sea level. The structure of the region is dominated by large graben that have been active since at least the lower Pliocene, are bounded by faults that affect Quaternary or Recent formations, and strongly influence the present-day pattern of drainage and sedimentation (e.g. Mercier *et al.* 1976, 1979, 1987, 1989; Roberts & Jackson 1990). On several of these faults slip is known to have occurred at the surface during recent or historic earthquakes (see Appendix). However, the structures that are known to have been active in the Quaternary are not all delineated by the seismicity of the last 25–30 yr, for which high-quality instrumental locations are available (Fig. 1). In Fig. 1 we have plotted epicentres for earthquakes only with $M_s \geqslant 5.0$, as only these may reasonably be expected to correlate with structures visible at the surface: smaller earthquakes, which have source dimensions less than 1–2 km, are known to occur throughout the region, and often represent internal deformation of the blocks bounded by larger faults (e.g. King *et al.* 1983, 1985; Hatzfeld *et al.*

1987, 1989, 1990). The epicentres in Fig. 1, which are reported by the US Geological Survey, are known to be in error by up to about 20 km , and are probably systematically shifted northwards (Soufleris 1980; Soufleris *et al.* 1982; Jackson *et al.* 1982a). Nevertheless, it is clear that although there is seismicity associated with some large structures, such as the Gulf of Corinth, others, such as the North and South Evia, Saros and Argos Gulfs, the Sperchios (Lamia) and Evrotas (Sparta) graben and the basin NE of Evia show no significant seismic activity since 1961. An assessment of the seismic activity over a longer period is clearly desirable, and it was with this in mind that we addressed the historical record. In contrast with other studies, where both macroseismic and instrumental data have been used without specifying the background information, we have compiled a summary description of the largest earthquakes in the Appendix and its accompanying figures.

DATA

We used information from published and unpublished macroseismic and instrumental sources as well as from our own field studies. The area included in our study is shown in Fig. 2. In includes the part of the 1890–1900 triangulation survey that was resurveyed by Global Positioning System (GPS) instruments in 1988 (Billiris *et al.* 1989), but also
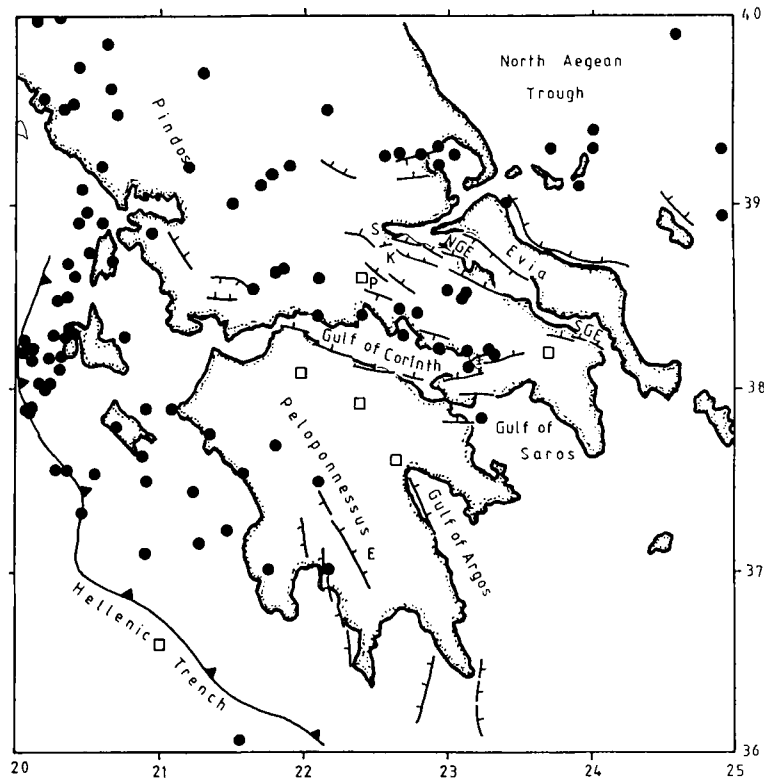
Figure 1. Seismicity of central Greece during the period 1962–1988. Locations are from the USGS Preliminary Determination of Epicenters, and only those of $m_b \geqslant 5.0$ have been included. Those with depths greater than 50 km are indicated as squares. Faulting associated with major tectonic structures on land and in the coastal regions of central Greece is summarized from Lyon-Caen *et al.* (1988), Roberts & Jackson (1990) and our unpublished work. This map shows only major active faults and is not complete: it is certain that many have been omitted, particularly from offshore regions and in remote areas north of the western Gulf of Corinth and in the Pindos. Abbreviations are: North Gulf of Evia (NGE); South Gulf of Evia (SGE); Sperchios (Lamia) graben (S) and Evrotas (Sparta) graben (E).
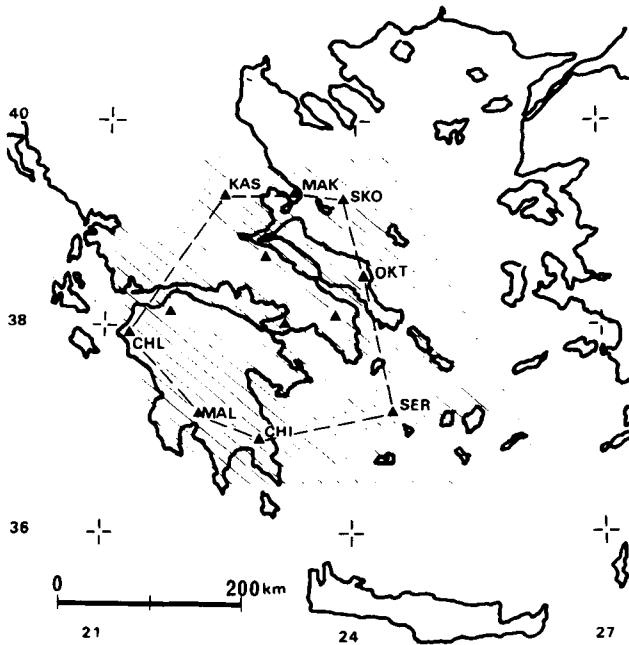
Figure 2. Map showing the area in which we examined seismicity during the period 1890–1988 (dashed), and the area enclosed by the resurveyed part of the 1890–1900 triangulation network (broken line). Reoccupied triangulation stations are shown by triangles.

extends this area to include that of a new GPS network installed in 1989.

For the period after 1900 we think we have identified all events of $M_s \geqslant 5.8$. Surface wave magnitudes were recalculated using amplitudes from Milne instruments for the early period (Ambraseys & Melville 1982), or using the Prague formula with amplitudes and periods (Vanek *et al.* 1962) for events after 1905. The purpose of recalculating M_s was not only to derive a homogeneous body of data, but also to assess the uncertainties in assessing this parameter from single station estimates. The recalculated average magnitudes have mean standard deviations of 0.20 ± 0.05 and 0.23 ± 0.06, with and without station corrections respectively. They differ little (± 0.2) from values calculated by Karnik (1968) for events before 1956, and confirm the reliability of his magnitude estimates. For earthquakes up to 1955 they also agree with values in Papazachos & Comninakis (1982), who seem to have adopted Karnik's estimates, but for later events, M_s values given by these authors are on average 0.3 ± 0.2 greater than our estimates. These differences are larger if averaged over shorter periods of time. For example, if we compare M_s values we have recalculated for Greece down to magnitude 4.8 for the period 1909–1922 with those published by Makropoulos & Burton (1981), who have used basically only two stations (Uppsala and Kiruna), we find that their values are systematically smaller by 0.25 units. For the period 1923–1952 the situation is reversed; their M_s estimates are consistently larger by the same amount. Between 1952 and 1978 there is again a reversal and their average M_s values are larger by 0.1 units but with some differences of up to ± 0.50. For earthquakes before 1900 we assessed magnitudes from macroseismic data, using the empirical formula derived for the region (Ambraseys 1988), which for a particular

earthquake relates M_s to the average radii of isoseismals of different intensity. Table 1 lists all the earthquakes with $M_s \geqslant 5.8$ and their largest foreshocks and aftershocks ($M_s \geqslant 5.0$).

The macroseismic effects, and when possible the teleseismic data from each of the events, including the aftershocks, in Table 1 were then re-examined. In the first instance, in addition to published information in technical papers, unpublished data, in particular from press reports and special studies, were retrieved and used together with field studies to re-evaluate the far- and near-field effects of each earthquake. Where possible, teleseismic data were used to assess the source mechanism, magnitude and seismic moment. The results are summarized in the Appendix, and in Table 1. Summary descriptions of the epicentral effects of each earthquake, contained in the Appendix, contain information that we regard as important to earth scientists and engineers. For earthquakes where there is published information the annotation is kept as brief as possible, and we have added all pertinent references. Information in the local press is too extensive to be quoted in detail, and it is collectively cited as 'Press'. Place names are those at the time of the earthquake, or in the original sources of information, and no attempt has been made to give corresponding modern names.

Almost all entries in the Appendix for shallow earthquakes of $M_s \geqslant 6.0$ are accompanied by an isoseismal and epicentral map. These maps show a selection of the places affected. Intensity maps up to 1965 have been uniformly re-evaluated using the MSK scale (MSK 1981), and in most cases are broadly consistent with similar maps produced by Shebalin, Karnik & Hadzievski (1974) and Papazachos *et al.* (1983). Differences do exist between our intensity maps and those prepared by other authors, particularly for the larger events and for lower isoseismals. The reason for this is that these authors rarely use observations from outside their own country. Epicentral maps give additional information and supplement the summary description of each event. They also permit relatively accurate location of the epicentral region and help in associating the earthquake with local tectonic structures.

A few earthquakes within the re-surveyed triangulation network seem to have had sub-crustal focal depths. Those of 1925 July 6 and 1938 September 18 were assigned depths of 120 and 100 km by Gutenberg & Richter (1948) on the basis of relative amplitudes of *P*-, *S*- and surface waves. The earthquake of 1962 August 28 was assigned an intermediate depth by the ISS, while those of 1964 July 17, 1965 March 31 and 1972 September 13 were located at 155, 45 and 75 km depth by the ISC. These depths are based on arrival times alone and lack precision. They are most probably sub-crustal and they all exhibit the macroseismic peculiarities of this class of event (e.g. Ambraseys 1967), i.e.: large epicentral area of disproportionately low intensity for their magnitude; large felt area, often extending from S. Italy and Sicily to Egypt; minimal aftershock activity; long duration of shaking, and noticeable strong, long-period effects at large epicentral distances.

However, there are also other events in the region that show some of these macroseismic characteristics of sub-crustal focal depths, but which are genuinely shallow events. The earthquakes of 1909 May 30, 1927 July 1 and

Table 1. List of earthquakes ($M_s \geqslant 5.8$) in central Greece and larger aftershocks during the period 1890–1988.

No.	Date	Time	Coord	I_O	h	M_S (dMs)	N	m_b	m_p	m_B	M_L	Str	Dip	Rake	C	Affected Region	log(Mo) (dyn.cm)	Fig
1	1893 Mar 26	2300	–	–	–	5.1 (0.3)		–	–	–	–	–	–	–	0	Thiva	–	–
2	Mar 27	1655	–	–	–	5.2 (0.3)		–	–	–	–	–	–	–	0	"	–	–
3	May 22	1931	–	–	–	5.5 (0.3)		–	–	–	–	–	–	–	0	"	–	–
4	May 23	2029	38.30-23.25m	VIII+	10*	6.0		–	–	–	–	270	45	-090	0	"	–	1;2
5	May 24	1028	–	–	–	5.1 (0.3)		–	–	–	–	–	–	–	0	"	–	–
6	1894 Apr 20	1716	38.65-23.16m	IX+	6*	6.4 (0.3)		–	–	–	–	290	45	-070	0	Martino	–	3;4
7	Apr 27	1945	38.66-23.03m	IX+	11*	6.9 (0.3)		–	–	–	–	290	45	-070	0	Atalanti	–	5;6
8	1898 Jun 2	2159i	37.00-22.00m	(VII)	80*	–		–	–	–	–	(155	45	-090)	3	Megalopolis	–	7
9	1899 Jan 22	0817	37.20-21.69m	VIII	9*	6.11(0.16)	2*	–	–	–	–	180	45	-090	2	Kiparisia	–	8;9
10	1909 May 30	0614	38.44-22.14m	VIII+	10*	6.27(0.20)	12	–	–	6.9	–	090	74	-115	2	Fokis	–	10;11
11	Jun 15	2331	39.05-22.36m	(VII)	–	5.79(0.21)	9	–	–	6.1	–	305	45	-090	0	S.Domokos	–	–
12	1911 Oct 22	2231	39.55-22.78m	(VII)	–	5.87(0.32)	9	–	–	–	–	284	56	-090	2	N.Volos	–	–
13	1914 Oct 17	0622	38.34-23.46m	IX	10*	6.24(0.18)	11	–	–	–	–	100	45	-090	0	Thiva	–	12;13
14	Oct 17	1042	38.29-23.20m	(VII)	–	5.48(0.28)	6	–	–	–	–	–	–	–	0	"	–	–
15	1915 Jun 4	1722	39.10-21.40m	(VIII)	9*	6.10(0.27)	11	–	–	–	–	200	56	-040	2	Agrafa	–	14;15
16	1916 Feb 6	1440	39.00-23.50m	(VII)	–	5.81(0.21)	8	–	–	6.5	–	300	45	-070	0	Skiathos	–	–
17	Sep 27	1502	38.85-22.85m	(VI+)	–	5.85(0.18)	9	–	–	–	–	290	45	-070	0	N.Evia	–	–
18	1917 Dec 24	0914	38.45-21.75m	VII	–	5.84(0.17)	10	–	–	–	–	090	74	-115	0	Nafpaktos	–	–
19	1923 Dec 5	2056	39.85-23.60	(VII+)	6*	6.43(0.29)	10	–	–	–	–	135	45	-090	2	Kasandra	–	16
20	1925 Jul 6	1216i	37.79-21.94	(VI+)	70	5.77(0.27)	15	–	–	6.8	–	(295	45	-070)	1	Achaia	–	–
21	1926 Aug 30	1138i	36.76-23.16	–	75	6.63(0.30)	18	–	–	7.4	–	(340	45	-090)	3	S.Peloponnesus	–	–
22	1927 Jul 1	0819	36.78-22.25	(VIII)	20*	6.43(0.29)	31	7.0	–	7.3	–	335	45	-090	2	Laconia	–	17;18
23	1928 Apr 22	1901	(37.94-23.00)	–	–	5.00( )	1	–	–	–	–	–	–	–	0	Corinth	–	–
24	Apr 22	1959	37.94-23.00m	VII	–	5.21(0.19)	2	–	–	–	–	–	–	–	0	"	–	–
25	Apr 22	2013	38.03-22.83m	(IX)	9*	6.32(0.29)	26	–	–	6.4	–	285	40	-070	0	"	–	19;20
26	Apr 25	0031	(38.03-22.83)	(VI)	–	5.18(0.22)	8	–	–	–	–	–	–	–	0	"	–	–
27	Apr 29	0949	(38.03-22.85)	(VI)	–	5.22(0.18)	11	–	–	–	–	–	–	–	0	"	–	–
28	Jun 7	1253	38.01-22.95m	VII	–	5.00( )	1	–	–	–	–	–	–	–	0	"	–	–
29	1930 Feb 23	1819	(39.60-23.17)	(VII+)	–	5.91(0.19)	30	–	–	6.2	–	310	45	-090	2	Magnesia	–	21;22
30	Mar 31	1233	39.60-23.17m	(VIII)	10*	6.02(0.22)	27	–	–	–	–	310	45	-090	2	"	–	–
31	Apr 17	2006	37.82-23.06m	IX	6*	5.88(0.17)	30	–	–	–	–	270	45	-090	0	SE Corinth	–	23;24
32	1938 Jul 20	0023	38.30-23.79m	VIII+	6*	6.07(0.11)	27	–	–	–	–	290	45	-070	0	Oropos	–	25;26
33	Sep 18	C751	38.27-22.47	(V+)	50	5.83(0.20)	11	–	–	6.2	–	(295	45	-070)	1	Galaxidi S.	–	–
34	1941 Mar 1	0353	39.70-22.54m	VIII+	10*	6.10(0.22)	16	–	–	6.1	–	150	45	-090	2	Larisa	–	27;28
35		0751	(39.70-22.54)	–	–	5.05(0.30)	5	–	–	–	–	–	–	–	2	"	–	–
36	1947 Oct 6	1956	36.80-22.00	(VIII)	–	6.92(0.24)	24	–	–	6.5	–	200	45	-080	2	Mesinia	–	29;30
37	1954 Apr 30	1302	39.22-22.28m	IX	9*	6.72(0.26)	19	6.2	–	–	–	300	45	-090	0	Sofades	–	31;32
38	Apr 30	1933	39.20-22.20	–	–	5.06(0.30)	8	–	–	–	–	–	–	–	0	"	–	–
39	May 4	1643	39.25-22.15	VII	–	5.65(0.22)	16	–	–	–	–	–	–	–	0	"	–	–
40	May 4	1645	39.20-22.35	–	–	5.45(0.25)	2	–	–	–	–	–	–	–	0	"	–	–
41	May 25	2203	39.20-22.30	VI	8*	5.58(0.18)	11	–	–	–	–	–	–	–	0	"	–	–
42	1955 Jan 3	0107	39.15-22.30	VII	–	5.49(0.21)	14	–	–	–	–	–	–	–	0	"	–	–
43	Jan 8	0753	39.25-22.10	VI	–	5.00(0.20)	6	–	–	–	–	–	–	–	0	"	–	–
44	Apr 19	1647	39.30-23.03m	(VIII)	6*	6.24(0.18)	21	6.1	–	6.3	–	090	45	-090	2	Volos	–	33;34
45	Apr 21	0718	(39.30-23.03)	(VII+)	–	5.91(0.22)	11	–	–	–	–	090	45	-090	2	"	–	–
46	1956 Jul 9	0311	36.70-25.80	(VIII)	–	7.16(0.33)	19	–	–	7.7	–	060	45	-090		Amorgos Isl.	–	35
47	Jul 9	0324	36.60-25.70	–	–	6.62(0.30)	3	–	–	7.2	–	–	–	–		"	–	–
48	Jul 9	0622	36.80-25.90	–	–	5.32(0.18)	3	5.6	–	–	–	–	–	–		"	–	–
49	Jul 10	0301	36.70-26.20	–	–	5.64(0.30)	7	5.6	–	–	–	–	–	–		"	–	–
50	Jul 30	0541	35.70-26.10	–	–	5.26(0.30)	7	5.6	–	–	–	–	–	–		"	–	–
51	Jul 30	0914	35.90-26.00	–	–	5.81(0.34)	13	6.3	–	–	–	–	–	–		"	–	–
52	Jul 30	1039	35.80-25.90	–	–	5.24(0.20)	7	5.4	–	–	–	–	–	–		"	–	–
53	Sep 6	1146	36.00-25.60	–	–	5.34(0.40)	12	5.5	–	–	–	–	–	–		"	–	–
54	1957 Mar 8	1214	39.34-22.68	–	–	6.49(0.11)	11	–	–	6.8	–	–	–	–	2	Velestino	–	–
55	Mar 8	1221	39.36-22.63m	IX+	11*	6.63(0.20)	14	–	–	6.9	–	110	45	-090	2	"	–	36;37
56	Mar 8	2335	39.35-22.75m	VIII	–	5.94(0.11)	9	–	–	–	–	–	–	–	2	"	–	–
57	Mar 11	0931	39.35-22.80m	VI	–	5.05(0.30)	4	–	–	–	–	–	–	–	2	"	–	–
58	Mar 28	2226	39.34-22.68	VI+	–	5.14(0.30)	3	–	–	–	–	–	–	–	2	"	–	–
59	May 21	1324	39.42-22.81	VIII	–	5.45(0.28)	5	–	–	–	–	–	–	–	2	"	–	–
60	Nov 26	0815	39.32-22.80m	VI	–	5.15(0.21)	4	–	–	–	–	–	–	–	2	"	–	–
61	Nov 26	1150	39.37-22.64	VI	–	5.11(0.25)	2	–	–	–	–	–	–	–	2	"	–	–
62	Nov 27	0308	39.37-22.65	VI+	–	5.45(0.22)	4	–	–	–	–	–	–	–	2	"	–	–
63	1962 Aug 28	1059i	37.80-22.88	(VII)	95	5.95(0.40)	6	–	–	6.9	–	(295	45	-070)	1	N.Peloponnesus	–	–
64	1964 Jul 17	0234i	38.05-23.61	(V)	155	–		5.8	–	–	–	(270	45	-090)	1	Mesinia	–	–
65	1965 Mar 9	1757	39.34-23.80	(VIII)		6.46(0.31)	13	5.7	–	–	–	040	90	180	2	Alonisos	–	38
66	Mar 9	1837	39.28-23.93	–	–	5.25(0.18)	12	5.2	–	–	–	–	–	–	2	"	–	–
67	Mar 9	1946	39.12-23.86	–	–	5.10(0.20)	4	5.1	–	–	–	–	–	–	2	"	–	–
68	Mar 10	0136	39.08-23.77	–	–	5.45(0.22)	4	5.0	–	–	–	–	–	–	2	"	–	–
69	Mar 13	0408	39.11-23.97	–	–	5.05(0.30)	4	4.9	–	–	–	–	–	–	2	"	–	–
70	Mar 13	0409	39.03-23.68	(V+)	–	5.15(0.24)	12	5.1	–	–	–	–	–	–	2	"	–	–
71	1965 Mar 31	0947i	38.38-22.26	(VII+)	45	6.65(0.19)	8	6.3	–	–	–	(136	76	080)	1	Aetolia	–	39
72	Apr 5	0312	37.48-21.97m	VIII	10*	5.93(0.31)	12	5.5	–	–	–	125	74	-032	0	Megalopolis	–	40;41
73	Jul 6	0318	38.27-22.37	(VII+)	–	6.38(0.27)	11	5.8	6.0	–	–	090	74	-115	0	Eratini	–	42
74	1966 Feb 5	0202	39.07-21.79m	VIII	15*	6.20(0.15)	13	5.6	6.0	–	6.0	252	66	-100	2	Kremasta	–	43
75	Feb 5	0258	39.00-21.60m	VII	–	5.05(0.25)	4	5.0	–	–	–	–	–	–	2	"	–	–
76	Oct 29	0239	38.80-21.18m	VII+	–	5.85(0.23)	12	5.8	–	–	5.6	330	45	-090	2	Amfilohia	–	–
77	1967 Mar 4	1758	39.25-24.60	(V+)	–	6.75(0.30)	11	6.0	6.5	–	–	098	60	-115	2	Skiros Isl.	–	44
78	May 1	0709	39.45-21.20m	VIII	10*	6.19(0.24)	13	5.5	–	–	6.2	200	56	-040	2	Tzumerka	–	–
79	1970 Apr 8	1350	38.23-22.60	(VII)	6*	6.18(0.29)	43	5.8	–	–	5.4	090	74	-115	0	Antikyra	–	45
80	Apr 20	1539	38.25-22.70	(VI)	–	5.45(0.21)	12	–	–	–	–	–	–	–	0	"	–	–
81	1972 Sep 13	0413i	38.00-22.40	(VII)	75	5.39(0.16)	10	5.8	6.2	–	5.4	295	45	-070	1	Corinth	–	–
82	1975 Dec 31	0945	38.50-21.65	(VII+)	–	5.72(0.41)	24	5.2	5.8	–	5.1	090	45	-090	0	N.Nafpaktos	–	–
83	1980 Jul 9	0210	39.26-22.93	(V+)	–	5.45(0.20)	3	5.2	5.7	–	5.2	–	–	–	0	Almyros	–	–
84	Jul 9	0211	39.23-22.76m	VIII+	10*	6.44(0.21)	13	–	–	6.0	6.0	270	50	-090	0	"	25.94	46
85	Jul 9	0236	39.23-22.59	–	–	6.01(0.25)	10	5.7	6.1	–	5.6	(270	50	-090)	0	"	–	–
86	Jul 9	0601	39.31-22.87	–	–	5.03(0.28)	14	5.1	5.1	–	4.8	–	–	–	0	"	–	–
87	Jul 10	1939	39.32-22.93	VII	–	5.45(0.21)	5	5.3	5.6	–	5.0	–	–	–	0	"	24.49	–

Table 1. (*continued*)

No.	Year	Date	Time	Epicentre	I_O	h	M_S	dMs N	m_b	m_p	m_B	M_L	Str	Dip	Rake	C		log(Mo) (dyn.cm)	fig
88		Aug 11	0915	39.26–22.72	VI	–	4.85(0.24)21		5.2	5.4	–	4.7	–	–	–	0	"	23.88	–
89	1981	Feb 24	2053	38.10–22.84	(VIII)	15*	6.69(0.21)44		6.1	6.5	–	6.8	285	40	–070	0	Alkionides	25.95	47;48
90		Feb 25	0235	38.13–23.05	–	–	6.41(0.23)43		5.7	6.4	–	6.1	250	42	–080	0	"	25.58	–
91		Mar 4	2158	38.18–23.17	VIII	–	6.24(0.40)54		5.8	6.3	–	6.3	068	47	–082	0	"	25.45	–
92		Mar 5	0659	38.20–23.13	–	–	5.22(0.18)36		5.3	5.7	–	5.0	–	–	–	0	"	24.26	–
93		Mar 7	1134	38.16–23.24	–	–	5.05(0.22)23		5.4	5.7	–	5.1	–	–	–	0	"	24.18	–
94		Dec 27	1739	38.90–24.92	(VI+)	–	6.39(0.36)44		5.4	5.9	–	6.0	045	90	180	2	Skiros Isl	25.52	–
95	1986	Sep 13	1724	37.02–22.16	VIII+	8	5.75(0.22)105		6.0	6.1	–	5.5	–	–	–	2	Kalamata	24.85	–

Notes on Table 1

Time is GMT; (i) indicates event deeper than normal.
Epicentre: (m) = macroseismic. A location in brackets is assumed to be the same as for the main shock.
Io: Intensity is epicentral unless shown in brackets which implies offshore epicentre or maximum intensity observed, not necessarily epicentral, in MSK scale.
h: • focal depth in km, assessed from macroseismic data using the method in Ambraseys (1985). h=Focal depths from teleseismic waveform modeling for five events are: no.78, 11km; no.89, 12km; no.90, 7km, no.91, 4km, no.95, 5km.
Ms: Recalculated surface wave magnitude; M* is from Milne instruments.
dM: standard deviation of magnitude estimate.
N: number of single–magnitude determinations used.
m_b: short–period body wave magnitude determined from interval of 5sec after first onset .
m_p: body wave magnitude from intermediate–period instruments in the interval of about 20sec after first onset.
m_B: body–wave magnitudes from P, PP and S waves recorded by medium–period instruments.
M_L: Local magnitude.
Strike, Dip and Rake of adopted source mechanism, assuming all events to be crustal.
C: code number; 0:shallow event within triangulation network, 1: subcrustal event within network; 2: shallow outside network, 3: subcrustal outside network.

1947 October 6, for instance, like some other shallow events in the area, show strongly elongated felt areas extending in the NW–SE direction (see Figs A10, A17, A29 in the Appendix), with their lower isoseismals extending along the Adriatic coast to the NW and towards Egypt in the SE. This isoseismal pattern may be related to high absorption in the upper mantle beneath the Aegean (Papazachos & Comninakis 1971), or may be reflecting the geological structure of the thrust sheets that comprise the Pindos mountains, which strike NW–SE and impose a strong structural 'grain' on the country.

We have attempted to estimate focal depth using macroseismic data (Ambraseys 1985), and the values so obtained are shown in Table 1, marked with an asterisk. These values are of an index nature, but they agree with the few well-determined depths based on teleseismic waveform analysis to within 5 km. The true focus may not be at the macroseismically determined depth, but at least these depths are not liable to the gross errors possible with early teleseismic focal depths based on arrival times alone or with those based on crude empirical relationships.

The question now arises: how complete are our data in Table 1? For small ($M_s < 5.0$) and earlier (pre-1900) events, particularly those in remote and relatively uninhabited parts of the Pindos, our information cannot be complete; although, from the macroseismic point of view, it is homogeneous. However, the combined use of local press reports with bulletins from regional seismograph stations guarantees that all events of $M_s \geq 5.0$ have been detected. Earthquakes of $M_s \geq 5.0$ would be felt with intensities of V (MSK) or more within distances of 45 km. It is very unlikely that such events have escaped the notice of both the local press and station bulletins. In the present case we cannot make an *a priori* assumption of stationarity of the seismic

activity in the region and consequently we cannot use a sample completeness test (Stepp 1973) to answer formally the question of how complete is our data set. However, it is reasonable to suppose that for $M_s \geq 5.0$ the data set in Fig. 3 is complete for the whole period 1890–1988. Note that Fig. 3 contains many events of $5.0 \leq M_s < 5.8$ that are not in Table 1, which contains only those in this range that we consider foreshocks or aftershocks of larger earthquakes.

ESTIMATION OF STRAIN FROM SEISMICITY

One of the purposes of this study was to estimate the strain accounted for by the seismicity during the period 1890–1988 within the part of 1890–1900 triangulation survey that was resurveyed by Billiris *et al.* (1989) using GPS instruments in 1988. The area that was resurveyed is shown in Fig. 2.

Method

The technique used was that of Kostrov (1974), and we followed the procedure described by Jackson & McKenzie (1988a, b). In Kostrov's technique the strain tensor is estimated by summing the seismic moment tensors in a volume of known dimensions. The seismic moment tensors are constructed from the fault plane solution (strike, dip, rake) and scalar seismic moment, M_o, (proportional to the fault area and seismic slip) for each event.

Where possible, the fault plane solution we used was that obtained from teleseismic first motion or waveform observations. Otherwise we estimated the strike, dip and rake from either surface faulting associated with the earthquake, or from the orientation of nearby active faulting (even though there was no conclusive evidence for surface
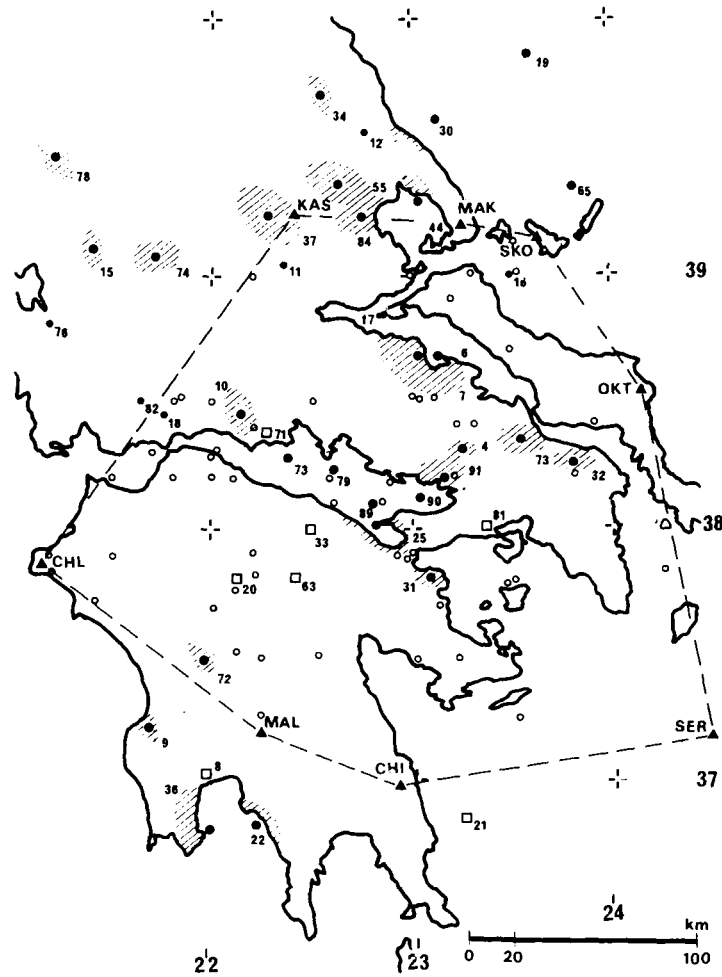
Figure 3. Earthquakes identified during the period 1890–1988. Numbers identify the events in Table 1. Filled circles are shallow events of $M_s \geq 5.8$. Macroseismic epicentral areas are shown dashed. Numbered squares are probable sub-crustal events. Small open circles are those of $5.0 \leq M_s < 5.8$, and include many events not in Table 1. The broken line encloses the area of the resurveyed triangulation network.

faulting in the earthquake), or from nearby events whose fault plane solutions were known. Our estimated fault plane solutions, and our reasons for choosing them, are given in Table 1 and the Appendix.

We estimated all the scalar seismic moments from an empirical relationship between moment and surface wave magnitude (within the geodetic network only five events of $M_s \geq 5.8$, those of the 1980 Almyros and 1981 Gulf of Corinth sequences, had moments calculated directly from teleseismic observations). With the abundance of accurate M_o determinations from centroid-moment tensor inversions using the Global Digital Seismic Network, it is clear that there are regional variations from the global average M_s–M_o relation (Ekström & Dziewonski 1988). In particular it appears that the Aegean region may yield somewhat higher M_s values than predicted from a global M_s–M_o relation (Jackson & McKenzie 1988b; Ekström & England 1989). Fig. 5 shows Ekström & Dziewonski's (1988) global average M_s–M_o relation (solid line) on which we have superimposed the data for 63 earthquakes in the Aegean region, most of which plot above this relation. We therefore calculated moments using both the global relation and a regional relation based on M_o values from centroid-moment tensor inversions over a 10 yr period 1977–87: $\log (M_o) = 19.24 +$

M_s, if M_s is ≤ 7.2 (provided by G. Ekström, personal communication). The global relation is likely, if anything, to overestimate the moments, and hence strain (M_o in dyne cm).

The main source of uncertainty in the seismic strain estimate is likely to come from the uncertainty in M_o, with the uncertainty caused by errors in strike, dip and rake being relatively small (Molnar & Deng 1984; Jackson & McKenzie 1988a, b). We therefore estimate the uncertainty in strain from the uncertainty in observed surface wave magnitude (see Table 1). Because of the possibility of systematic bias, we do not calculate standard errors, but simply extreme maximum and minimum possible values, allowing for all the magnitudes to be under- or overestimated. We thus obtain a range of maximum to mimimum values of strain for both the global and regional M_s–M_o relations. The maximum and minimum values of strain typically differ by a factor of two from that obtained from the tabulated M_s values, reflecting a typical uncertainty of ±0.25 units in surface wave magnitude.

It is also necessary to specify the dimensions of the zone. We took the length (E–W) to be 250 km, the width (N–S) to be 230 km, and the thickness to be 10 km. The most significant error is associated with the thickness of the layer
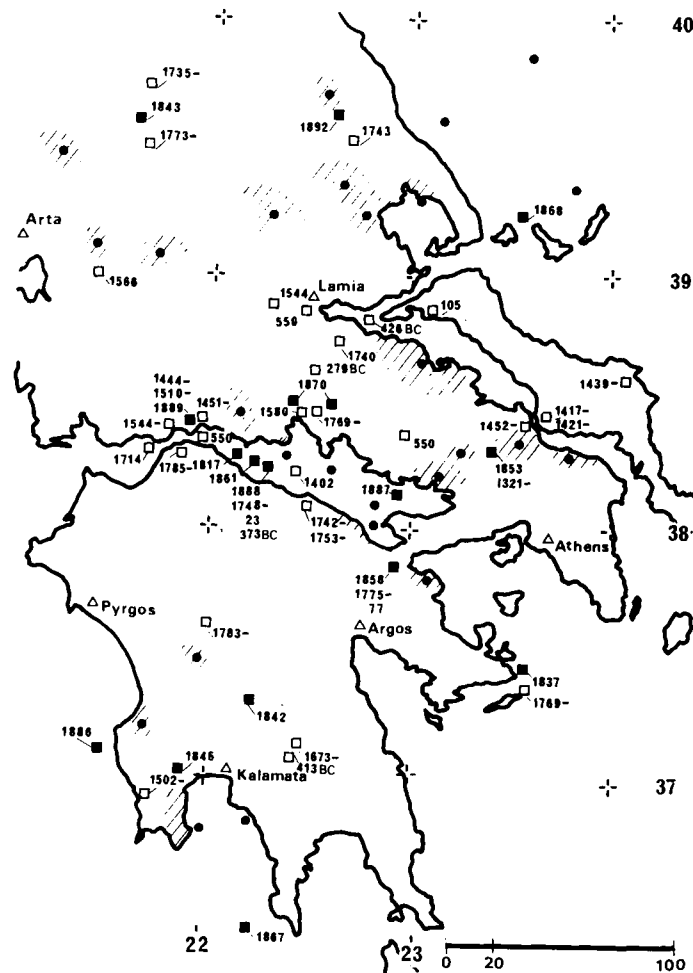
Figure 4. The distribution of all earthquakes known to us of $M_s \geqslant 6.0$. Solid circles are the events listed in Table 1 (1890–1988), and their epicentral regions are shaded. Solid squares are the approximate epicentres of 19th century earthquakes. Open squares are approximate epicentres of events earlier than 1800. The pre-1890 events are labelled with the year of their occurrence: those that may have been smaller than M_s 6.0 are labelled with a bar.
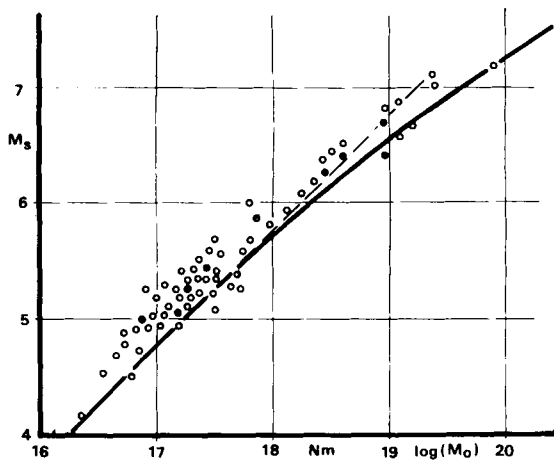
Figure 5. M_s–M_o relationships for shallow events in the Aegean region since 1967 (34–42°N; 20–29°E). Filled circles are events in the study area. M_o values, in N m, are taken from Harvard moment tensor solutions and other studies. M_s values were recalculated. The solid curve is the global relation of Ekström & Dziewonski (1988). The dashed line is Ekström's regional relation, discussed in the text. (1 dyne cm = 10^{-7} N m.)

within which seismic energy is released. Like Jackson & McKenzie (1988a, b) we chose 10 km, based on the focal depth of large earthquakes in the Aegean region, but this could be in error by a few km: we suspect that the errors associated with M_o estimation are much more important. A final uncertainty is how to treat earthquakes that we think probably have sub-crustal focal depths. They do not contribute to the deformation of the upper seismogenic layer and should not be included in the strain analysis. However, it is not always easy to distinguish these deeper events using macroseismic data, and we may have misidentified some of the early events as deep. We have therefore estimated the strain both including and excluding the apparently deep events within the geodetic network. As will be seen, the number and size of possible deep events is small, and their inclusion or exclusion makes no significant difference.

Results

Our results are listed in Table 2. The axes of the strain tensor are oriented approximately N–S (180°), E–W (89°)

Table 2. Displacements (cm) calculated from earthquakes of $M_s \geqslant 5.8$ in the period 1890–1988.

Direction	minimum	probable	maximum	Ms-Mo relation
E-W	0.9	2.0	4.8	global
N-S	33.1	71.5	168.0	(shallow events only)
E-W	0.7	1.2	2.2	regional
N-S	25.4	45.1	82.5	(shallow events only)
E-W	-1.2	-2.0	-2.7	global
N-S	34.9	74.4	174.0	(shallow and deep events)
E-W	-0.8	-1.1	-1.5	regional
N-S	27.2	48.0	87.7	(shallow and deep events)

Strain rates (x $10{-}16s{-}1$)

Direction	minimum	probable	maximum	Ms-Mo relation
E-W	0.1	0.3	0.6	global
N-S	4.6	9.8	23.1	(shallow events only)
vertical	-4.7	-10.1	-23.8	
E-W	0.1	0.2	0.3	regional
N-S	3.5	6.2	11.4	(shallow events only)
vertical	-3.6	-6.4	-11.6	

(Extension is positive)

and vertically, and the overall style of deformation across the network is N–S extension, balanced by crustal thinning, with no significant movement of material in the E–W direction. This is no surprise, as it could have been predicted by inspection of the geology and fault plane solutions in the region (McKenzie & Jackson 1983, 1986). The earthquakes of $M_s \geqslant 5.8$ in the period 1890–1988 account for N–S displacements across the geodetic network of about 45 cm (regional M_s–M_o relation) to 72 cm (global M_s–M_o relation), with maximum and minimum values being about factors of two greater and less than these values, and with the possibly deep events making no significant difference. These displacements correspond to N–S velocities of $4–7\,mm\,yr^{-1}$ across the geodetic network, compared with $20–60\,mm\,yr^{-1}$ across the Aegean as a whole
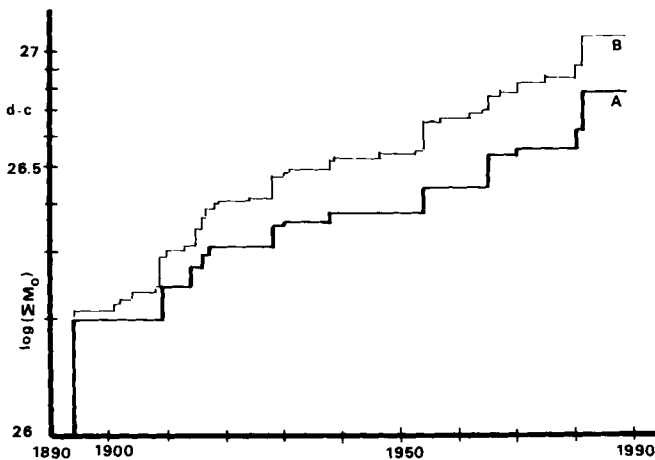
Figure 6. Cumulative moment versus time for shallow earthquakes within the resurveyed geodetic network, based on Ekström's regional M_s–M_o relation. Curve A is for all events of $M_s \geqslant 5.8$. Curve B includes all events of $M_s \geqslant 5.0$.

(Jackson & McKenzie 1988b).

In summing the moments we only included earthquakes of $M_s \geqslant 5.8$. The contribution of smaller events to the total strain is small compared with the errors that anyway exist from uncertainties in M_o. This can be shown by extending the lower magnitude to 5.0. Fig. 6 shows the cumulative moment versus time for shallow shocks within the geodetic network, using Ekström's regional M_s–M_o relation. Curve A is for shocks of $M_s \geqslant 5.8$, and shows a total moment of 7.8×10^{19} Nm. Curve B includes all the smaller events down to $M_s 5.0$, showing a total moment of 9.6×10^{19} N m: an increase of 23 per cent. The contribution of even smaller magnitudes to the total may be assessed from the bilinear frequency distribution that fits the magnitude data in Table 1. Using Ekström's regional M_s–M_o relation, and assuming that the frequency distribution is valid for $M_s < 5.0$, we may assess the contribution of smaller events to the total moment calculated for the range $M_s 5.8–6.9$ (Båth 1978; Molnar 1979). Thus, by including earthquakes greater than $M_s 5.0$ or 4.0, we increase the total moment by 23 or 60 per cent respectively. These increments derived from Eström's regional M_s–M_o relation ($M_s = -19.24 + \log M_o$) are greater than those obtained from earlier global M_s–M_o relations, such as that of Kanamori & Anderson (1975), ($M_s = -10.7 + 2/3 \log M_o$) ($M_o$ in dyne cm). The reason for this is that the slope of Ekström's relation is steeper, which accounts for the larger contribution to M_o from smaller magnitude events. We note, in passing, that Fig. 6 confirms the suggestion of Jackson & McKenzie (1988a) that, in Greece, the moment rate averaged over a 30 yr period is representative of longer periods (in this case 100 yr).

These displacements estimated from the seismicity are similar in orientation and magnitude to those calculated from the resurveying of the 1890–1900 geodetic network, which found about 100 cm of N–S extension (Billiris *et al.* 1989), but a detailed comparison of the two will be reported elsewhere.

DISCUSSION

Seismicity distribution

The distribution of the larger earthquakes in the interval 1890–1988 (Fig. 3) is rather similar to that of the period 1961–1987 (Fig. 1). With the exception of the North Gulf of Evia (1894) and the environs of the plain of Thiva (1893, 1914, 1938), little activity is seen associated with the major topographic structures that are aseismic in Fig. 1. We therefore undertook a preliminary examination of the seismicity of the area prior to 1890.

The historical seismicity of central Greece is imperfectly known. The period from 1810 onwards is relatively well documented. However, further back in time it becomes increasingly difficult to find data, and the minimum size of the earthquake for which there is guaranteed detection increases. For the period before the 14th century only a few earthquakes are known and most of these are for the period between the 5th century BC and 6th century AD. For the following seven centuries, not a single event can be assessed with certainty.

Figure 4 shows the distribution of all the earthquakes known to us with estimated magnitudes of 6.0 or greater. 19th century and earlier events are still under investigation, and both their location and size are to some extent uncertain. Location errors for epicentres on land are unlikely to be greater than 30 km, but uncertainties in magnitude for some of these events are considerable. This figure is revealing. Earthquakes are associated with most of the major active morphological features in central Greece, including several that look quiet in Figs 1 and 3. The main structures that stand out as aseismic in Fig. 4 are the NE coast of Evia, the graben bounding the NE flanks of Parnassos and Kallidromon and the Gulf of Argos. It is difficult to know how to interpret the apparent quiescence of these features: they all exhibit morphology suggesting that they were active in the Quaternary, and yet several of them are in relatively remote regions, where earthquakes could have been missed by contemporary chroniclers, or have so far escaped our notice. Some, such as the fault escarpment along the NE flank of Kallidromon, may no longer be active (e.g. Philip 1974; Mercier *et al.* 1979). Rapid recent uplift along the NE coast of Evia suggests that the fault bounding the basin offshore is still active (Roberts & Jackson 1990), and seismic reflection records suggest recent activity on faults within and bounding the Gulf of Argos (T. van Andel & C. Perissoratis, personal communication).

The other clear feature of Fig. 4 is the intense activity in the Gulfs of Evia and Corinth: but these are the only regions in central Greece for which the historical record is better known after the 14th century. Thus Fig. 4 shows a concentration of epicentres in the Gulf of Corinth that is to some extent artificially distorted by the distribution of reporting centres; though no less interesting from our point of view because of that.

Maximum magnitude

There is no evidence, so far, that any of the earthquakes shown in Fig. 4 have exceeded M_s 7.0 in size. This is not, in our view, surprising. Earthquakes larger than M_s 7.0 require faults of 30–40 km or longer, and as pointed out by Roberts & Jackson (1990), we are unaware of such faults in central Greece. Although some of the major topographic features, such as the north Gulf of Evia and Gulf of Corinth, are longer than 100 km, the normal faults that bound them are segmented, with individual fault segments being no longer than about 15 km: a length sufficient only for earthquakes of M_s 6.0–6.5. Jackson & White (1989) argue that a maximum segment length of 15–20 km is common for continental normal faults worldwide: although normal faulting earthquakes larger than M_s 7.0 are known elsewhere, they are generally multiple events, with individual sub-events apparently not larger than equivalent to M_s 6.5. A maximum likely magnitude for central Greece has obvious implications for seismic risk. Note, however, that strike–slip events larger than M_s 7.0 are known offshore in and around the North Aegean Trough.

Surface faulting

Earthquakes with M_s as small as 5.7 are known to be associated with surface faulting in Greece (e.g. the Kalamata earthquake of 1986 September 13). In the period 1980–1988 surface faulting was observed following the earthquakes of 1894 (Martino), 1954 (Sofades), 1980 (Anhialos), 1981 (Alkionides) and 1986 (Kalamata) in central and southern Greece. There are several others for which reports are equivocal, or where known faults that outcrop in the epicentral region were not examined or discussed in contemporary accounts: such as those of 1861 December 26 near Aigio, 1870 August 1 in Fokis (Ambraseys & Panteloupos 1989), 1966 September 1 in Megalopolis, and 1966 October 29 near Amfilohia. We think it very likely that faulting associated with even the larger events in the past has been missed.

Vulnerability and damage

Changes in the predominant type of construction in the study area have been rapid after about 1930 and for larger urban areas there has been a considerable improvement after 1950, making uniform assessment of damage for the whole period difficult.

In general, the vulnerability of the average dwelling in central Greece is lower than in other parts of the Eastern Mediterranean region (Ambraseys & Jackson 1981). This, together with the fact that many of the larger shocks occur as clusters of two or three events, closely spaced in time (Table 1), and are often preceded by foreshocks or have offshore epicentres, accounts for the relatively small number of casualties (Ambraseys & Jackson 1981).

CONCLUSIONS

We examined the seismicity of central Greece during the period 1890–1988 with two questions in mind: does the seismicity over 100 yr have a closer association with the known active tectonic structures than does the better-known seismicity since the early 1960s?; and what are the

likely strains associated with the seismicity over this time interval?

The distribution of the larger earthquakes in the period 1890–1988 resembles quite closely that of the larger events in the shorter interval 1962–1988. A number of large tectonic structures have been effectively aseismic since 1890. The seismicity record for the period before 1890, though imperfect, reveals activity associated with more of these tectonic structures: the most striking feature of the seismicity of this longer period being the intense activity in the Gulf of Corinth and between Evia and the mainland. However, we have no record of major earthquakes associated with the NE coast of Evia, the Gulf of Argos, or the graben bounding the NE flanks of Parnassos and Kallidromon except perhaps in the early events of 426 BC, 279 BC, 105 AD and 550 AD (Fig. 4). The implication is clear: that even 100 years' seismicity is insufficient to reveal all the major active tectonic structures, and assessments of seismic risk should take a longer perspective. Also important for seismic risk is the lack of evidence (so far) for earthquakes significantly larger than M_s 7.0 in central Greece: an observation that is consistent with Roberts & Jackson's (1990) remark that the major graben are bounded by faults that are segmented, and are not continuous for more than about 15–20 km.

An assessment of the seismic strain over the period 1890–1988 required an estimate of the moment tensor for each earthquake, constructed from the strike, dip and rake combined with the seismic moment. It is surprisingly easy to estimate the likely strike and dip, even for older events with no teleseismic constraints. Many of the earthquakes have obvious active faults in their epicentral regions, and we suspect that surface ruptures may have gone unnoticed in the past. The greatest uncertainty in the strain estimates comes from the M_s–M_o relation used to obtain seismic moments. We estimate that the earthquakes of $M_s \geqslant 5.8$ can account for about 45–70 cm of N–S extension across the resurveyed part of the 1890–1900 triangulation network, with maximum and minimum values a factor of two greater and smaller than this. The contribution of smaller earthquakes, with $M_s < 5.8$, will increase these displacements by around 50 per cent. The orientation of the strain tensor indicates that N–S extension is balanced almost entirely by crustal thinning, with very little movement of material in the E–W direction. These results will be compared in detail with the geodetic observations, which indicate about 100 cm of N–S extension, elsewhere.

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APPENDIX

The Appendix deals with the felt effects of the larger events that we consider to be important for earth scientists and engineers. The preferred estimates of the focal mechanisms are given in brackets, in the sequence (strike: dip:rake), the convention following that of Aki & Richards (1980). In the references that follow each event, P stands for press reports. U stands for unpublished information from one or more of the following sources: Railways Piraeus–Athens–Peloponnesus (1895–); Hellenic State Railways (1915–); Cable and Wireless Ltd (1912–48): Red Cross Relief Reports (Athens & Geneva, 1926–55); Ministry of Reconstruction, Damage Statistics (1946–51); Prime Minister's Office Reports: Klimakion Ektakton Anangon

(1954–61); Syntonistiko Klimakio Arogis (1953–65); Ypourgion Koinonikis Pronias: Diefthinsis Laikis Stegis (1964–70); Ypourgion Dimosion Ergon: Ypiresia Oikismou (1963–); Ypiresia Apokatastaseos Seismoplikton (1978–); Hydrographic Service of the Navy: Fyla Periorismenis Epiteliks Hriseos (1948–61); Ypiresia Erevnon Ypedafous (1952–62). Starred references contain an isoseismal map. G stands for the Annual Bulletin of the Seismological Institute of the National Observatory in Athens published by A. Galanopoulos between 1951 and 1976 that covers the years 1950–1967.

1893 May 23 Thiva (*Ms* 6.0)

Preceded by two damaging foreshocks, the earthquakes affected a zone between Thespiae and the town of Thiva. The foreshocks of March 27 and May 22 caused some damage to a number of villages on the east facing slopes of Mt Elikon as well as in Thiva where a few houses collapsed. The main shock, which was widely felt in central Greece (Fig. A1) was centred to the east of Mt Elikon and, among other places, ruined Thiva, the only large urban centre in the region (population 3300). About 100 of its 1200 houses collapsed and another 800 were damaged. The effects of the shock were more serious at Pyri and Ag. Theodori where two people were injured. Another six hamlets including Lutufi were ruined, without casualities, the population having already been warned by the foreshocks (Fig. A2).

With the exception of ground cracks reported in the vicinity of Thiva as well as from Mulki, near the swamps of the Kopais Lake which at the time was artificially drained,
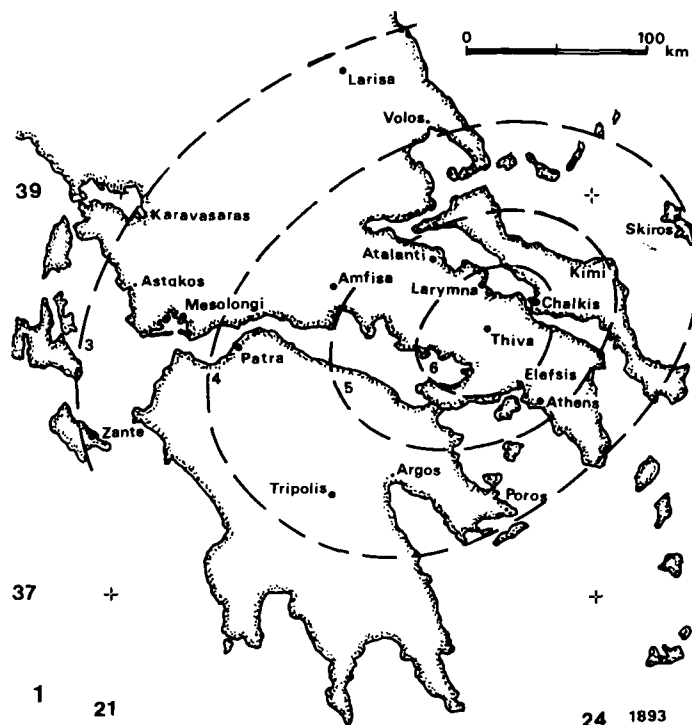
Figure A1. 1893 May 23 Thiva earthquake (*M*ₛ 6.0).
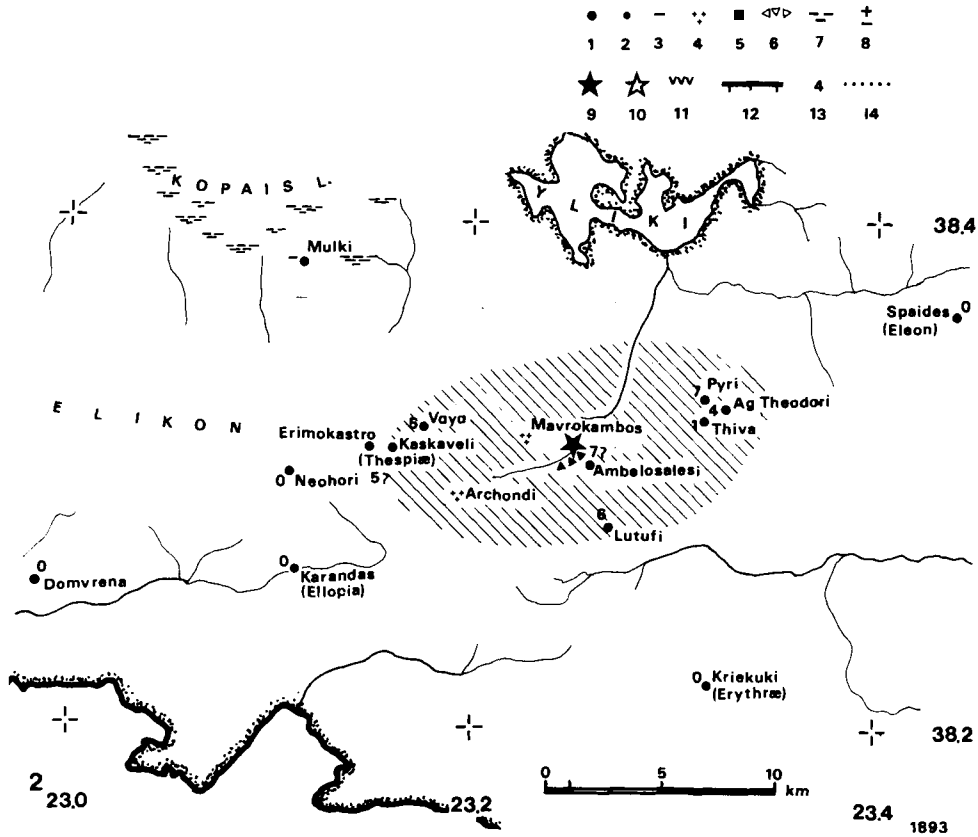
Figure A2. (1) Sites for which information is sufficient to estimate intensity. (2) Sites with insufficient macroseismic data to assess intensity. (3) Sites where the shock was not felt. (4) Small settlements, monasteries, farming communities (mostly abandoned today) affected by the shock. (5) Submarine cable-breaks associated with earthquake. (6) Areas of landslides, rockfalls and unspecified ground deformations associated with event. Full triangles indicate large-scale effects. (7) Sites of reported ground liquefaction, slumping of level ground and flow slides. (8) Coseismic elevation on subsidence of coast. (9) Estimated epicentre of main shock. (10) Adopted macroseismic location of aftershock. (11) Seismic sea-wave associated with event. (12) Ground deformations of tectonic origin. (13) Number indicates proportion of houses (in tenths of total number exposed to the earthquake) which were totally destroyed or damaged beyond repair. O indicates 1–9 per cent of total ruined. A bar implies 0 per cent losses. (14) Railway lines. Shading shows extent of epicentral area deduced from maximum damage effects. Scale is in kilometres.

there is no information about ground deformations. Rockfalls and ground cracks, presumably due to landslides, were reported from near the sites of Archondi, Mavrokambos and Ambelosalesi, where the shock was particularly severe. From the nature of local faulting in the Thiva plain we assume E–W striking normal faulting (270:45: – 090). A young fault with E–W strike, downthrown to the S, is suggested by the geomorphology of the epicentral region between Kaskaveli (mod. Leontari) and Ambelosalesi (mod. Ambelohori)

P; Mitzopoulos (1894); Hanusz (1895); Eginitis (1899a)

1894 April 20; Martino (*Ms* 6.4)

Without warning, an earthquake which was the largest foreshock of the earthquake of April 27 in the Gulf of Atalanti (Fig. A3), almost totally destroyed the villages between Skandaraga and Martino (Fig. A4), killing 223 people. In places along the west coast of the Gulf, from Almyra to Livanates, the coastal plain liquefied and the shore slumped into the sea. Between Proskina and Atalanti the shock caused ground cracks in the valley floor and triggered landslides and rockfalls. There is no contemporary
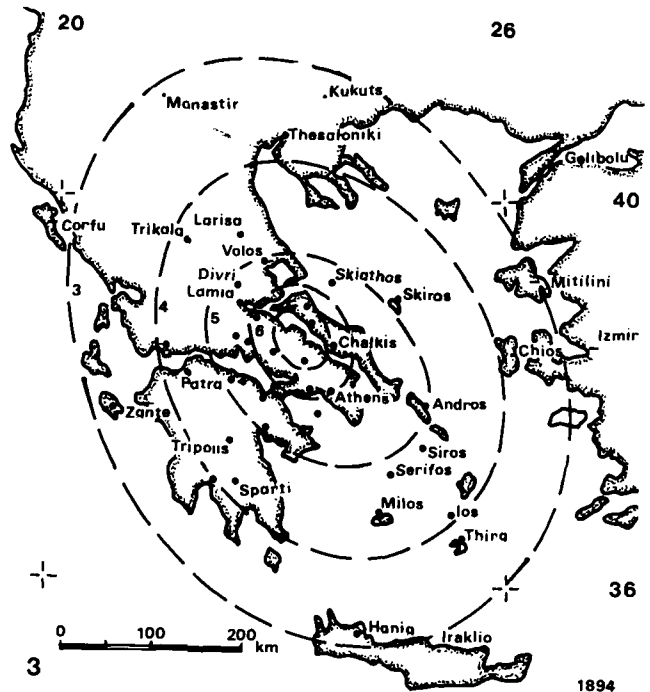
Figure A3. 1894 April 20; Martino earthquake (*M*s 6.4).
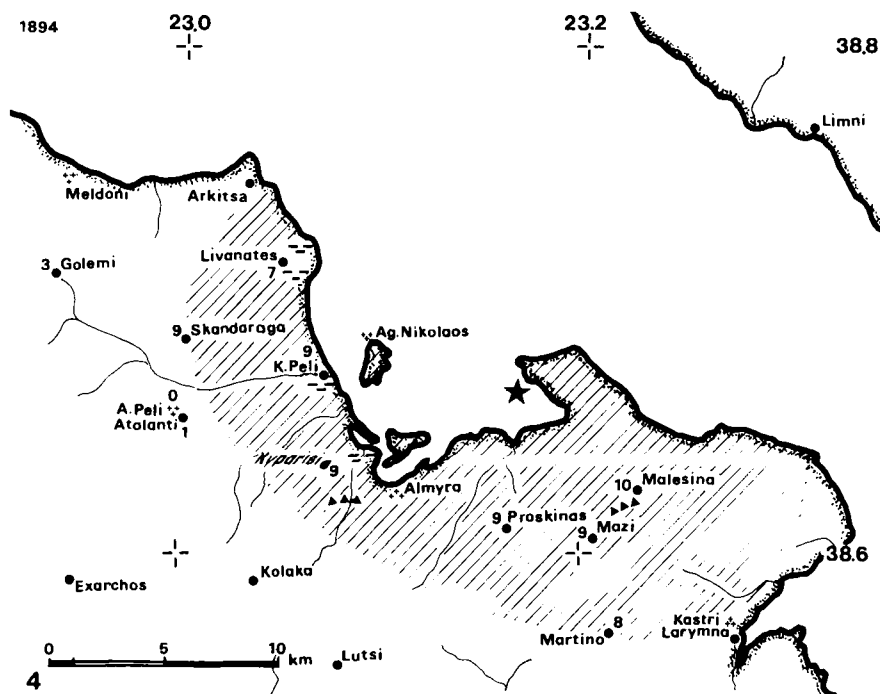
Figure A4. 1894 April 20; Martino earthquake (M_s 6.4).

evidence of faulting and the shock did not trigger a seismic sea-wave. Sporadic damage extended within a radius of about 40 km, and it was particularly severe in the Kopais swamp area in the south. The shock was felt over a relatively large area and was recorded by primitive seismographs and magnetometers up to distances of 18° from the epicentre. the focal mechanism of the earthquake is assumed to be the same as in the event of 1984 April 27 (290:45: − 070).

1894 April 27 Atalanti (*Ms* 6.9)

Preceded by the destructive shock of April 20 and its continuing series of aftershocks, the main shock completed the destruction around the Gulf of Atalanti (Fig. A5). Because of their close spacing in time it is difficult to separate the damage caused by these two shocks. However, the damage distribution of the main shock (Fig. A6) shows a clear northwestward migration of the epicentral area in both the total ruin of the villages northwest of Martino, already affected by the foreshock, as well as the heavy damage of the villages between Skandaraga, Ag. Theodori and Ag. Konstandinos: a region not seriously affected by the first earthquake. In all, the earthquake destroyed 3800 houses and killed 30, injuring about 90 people.

The earthquake was associated with surface faulting that extended from near Martino in the southeast to near Atalanti in the northwest, a distance of about 25 km. The surface breaks mapped shortly after the earthquake and the complicated remnants of these ruptures still visible today show clearly normal faulting with the NE block downthrown by about 100 cm on average, and the reactivation of a pre-existing system of faults striking 300°. Half as large throws are attested by the permenent subsidence of the coast-line between Arkitsa and Almyra, both of unconsolid-

ated alluvium and rock, and also by the submergence of the isthmus of Gaiduronisi. However, uplifted *lithophaga* borings near Livanates attest to long-term uplift of this part of the coast.

Northwest of Atalanti a series of short scarps still visible today, if attributed to this earthquake, perhaps suggest an
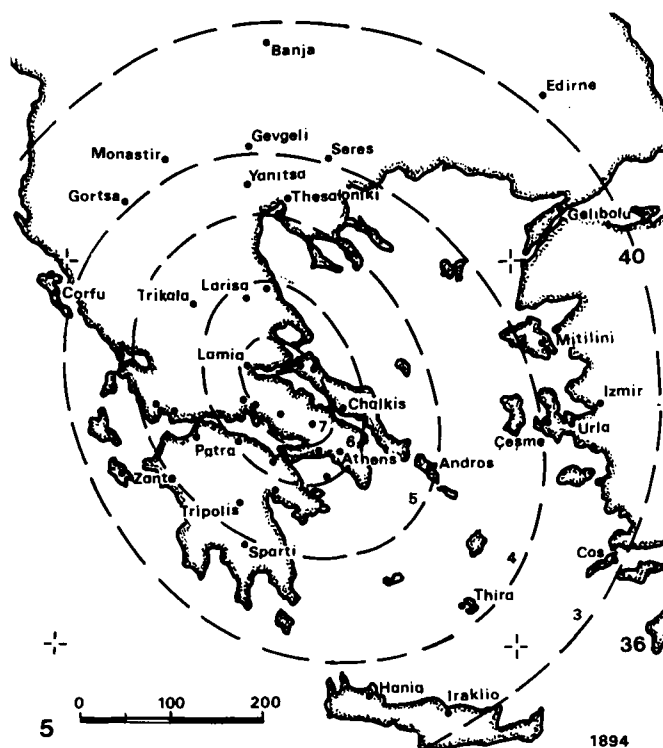
Figure A5. 1894 April 27; Atalanti earthquake (M_s 6.9).
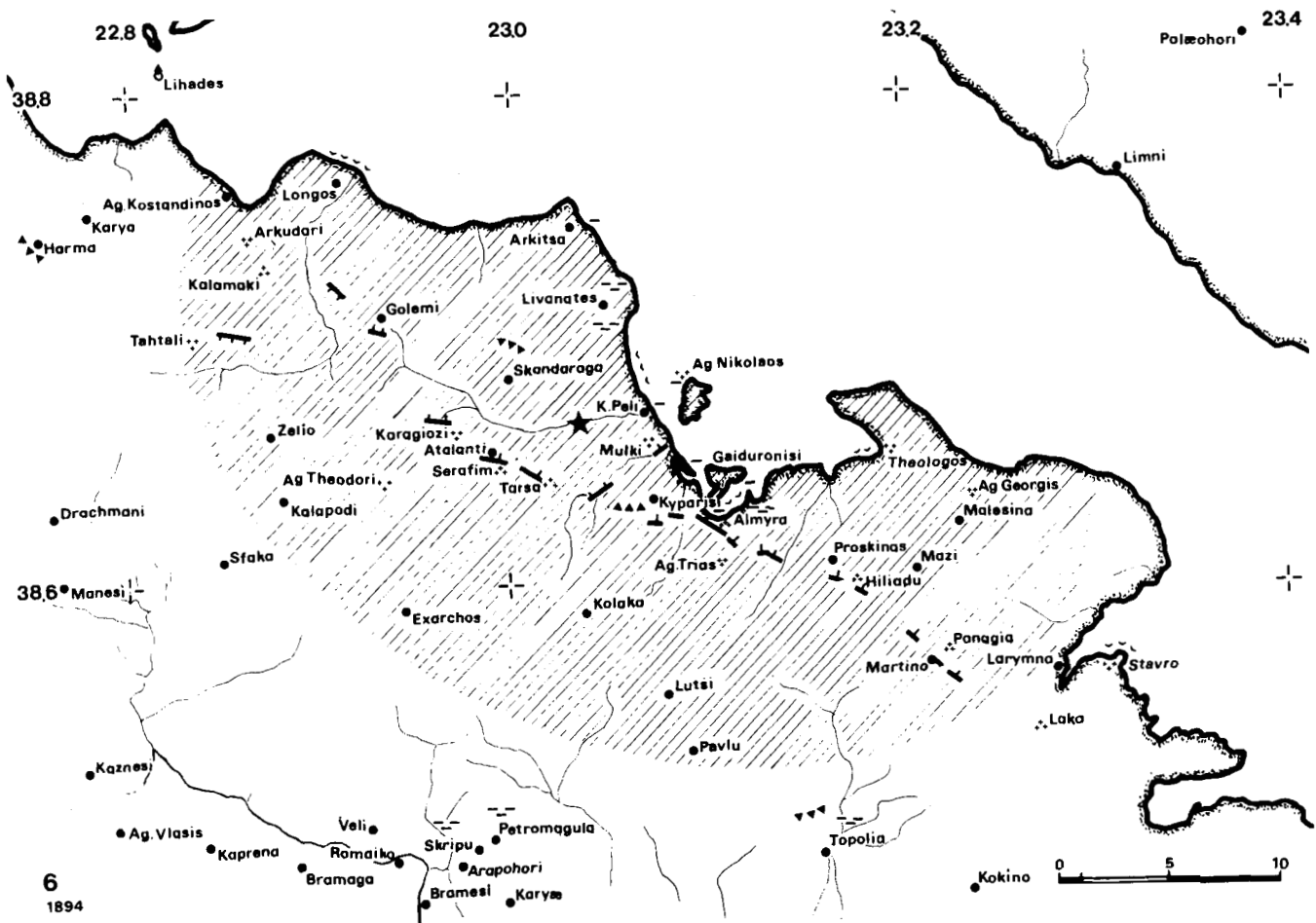
Figure A6. 1894 April 27; Atalanti earthquake (M_s 6.9).

extension of the rupture zone for another 15 km in a direction 290°. All features show normal faulting, but more often with the SW side downthrown by a few tens of centimetres, in the same sense as the topography. None of these features can be associated conclusively with the 1894 event.

The shock triggered landslides in shales and limestones, forming scarps which run for a few hundreds of metres, occasionally following normal faults that probably were not activated at depth. These scarps and other ground fractures, the tectonic origin of which cannot be established, have been interpreted in the past as part of the fault zone that was shown in early reports extending from the southeast of Larymna to northwest of Ag. Kostandinos and from there to Gardiki east of Lamia to the northwest. Thus the focal mechanism was normal faulting (290:45: − 070). The fault strike is taken from the Martino–Atalanti break, with dip and rake from those of nearby exposed fault faces.

Liquefaction of the ground occurred in places up to 40 km from the epicentral area and caused damage not only in the region of Kopais and Topolia but also in Kato Peli where the quay sunk into the sea.

A seismic sea-wave flooded the coast of the Gulf of Atalanti and its effects were reported from many places along the coast from Stavro in the southeast to Longos in the northwest. The wave caused no damage, and because it

occurred at night, details about its height are not known. The earthquake was felt on board ships in the Gulf of Evia and at lighthouses in the region, to which it caused no damage. A few old churches in the region were ruined; the church of St Georgis in the monastery of Malesina was totally destroyed. An inscription dates its construction in 1512 and its history suggests no serious damage between that data and 1894.

The shock was recorded by early seismographs and magnetometers up to distances of 22°. The duration of the earthquake at these stations was on average three times longer than that of the foreshock of April 20.

P; Mitzopoulos (1894; 1895); Papavasiliou (1894a; b; 1895) Philippson (1894); Skouphos (1894); Hanusz (1895); Eginites (1899a, b); *Sieberg (1932); *Shebalin et al. (1974); Roberts & Jackson (1990)

1898 June 2; Peloponnesus (*Ms* ?)

This earthquake was not preceded or followed by strong shocks and it was felt over a large area, from south Italy and eastern Sicily to the Aegean Sea, but not in north Greece (Fig. A7). The shock caused some damage and the collapse of a few old houses at a number of places in the Peloponnesus, spaced many tens of kilometres apart, with
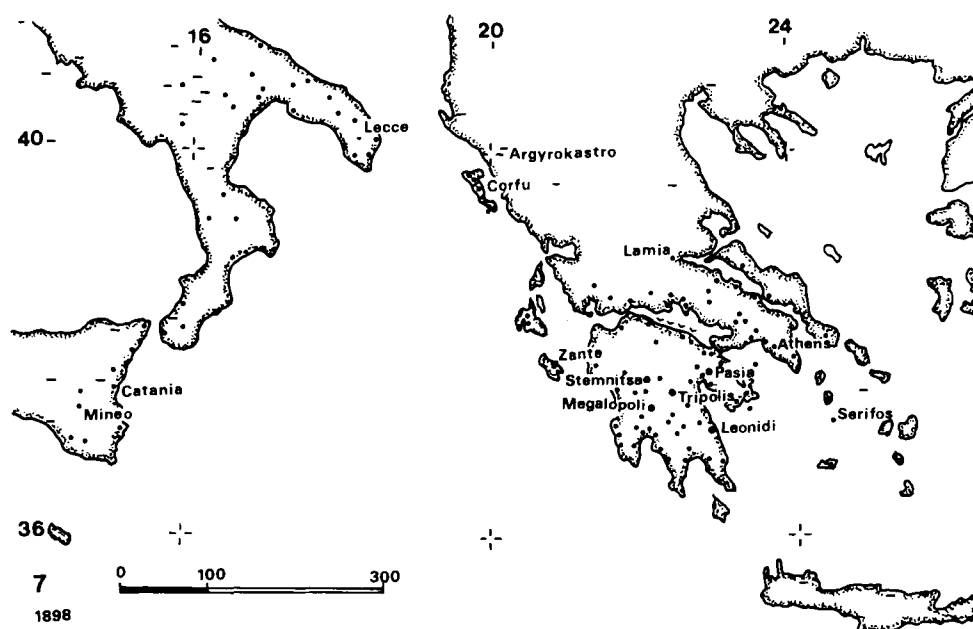
Figure A7. 1898 June 2; Peloponnesus earthquake.

no damage in between them: i.e. at Leonidi, Pasia, Stemnitsa, Megalopolis and in Tripolis, the largest of these towns, where two-story houses were badly cracked and 16 of them were ruined. The earthquake caused no casualties and the total material loss did not exceed one tenth of that incurred by the Thiva earthquake of 1893.

The earthquake caused waves that flooded the coast of the Gulf of Corinth but no breaks of submarine cables. Also, the shock caused water to rise in wells in low lying areas and sporadic liquefaction near Pasia.

The shock was recorded by early undamped seismographs at epicentral distances of 11° and it was of long duration. The macroseismic characteristics of this event suggest a focal depth deeper than normal in the vicinity of the Peloponnesus.

P; Cancani (1899); Eginites (1899b); Mitzopoulos (1900); Baratta (1901); *Shebalin *et al.* (1974); *Sheberg (1932)

1899 January 22; Kiparisia (*Ms* 6.1)

This earthquake in SW Peloponnesus consisted of two shocks spaced 2 min apart, each lasting about 7 s, that caused considerable damage but no fatalities (Fig. A8). The villages most affected were between Blemiani, Raftopulo, Hristiano and Halazoni, an area that had suffered considerable damage from the large, intermediate depth earthquake of 1886 April 27. The damage here, as elsewhere in the Peloponnesus, was cumulative but not serious. Within the epicentral region, Fig. A9, about 250 houses collapsed and 46 people were injured. The shock caused liquefaction and slumping of the coastal material and landslides and rockfalls further inland. Outside this region damage was widespread but minor, the shock causing slides and changes in the yield of spring water as far as Kalamata, Varvitsa and Dimitsana. In the marshes of Mesini, west of Kalamata, there was extensive liquefaction of the ground

and damage to railway embankments and telegraph lines. There is no evidence of ground deformations of tectonic origin. Between north of Kyparisia and Kalamata the coast appears to be sinking. The focal mechanism of this event (180:45 : − 090) is very uncertain. The shock may have been offshore and associated with the Hellenic Trench. There is also a suggestion of E–W faulting onshore, seen in the satellite imagery. A microearthquake survey in 1986 found E–W striking normal faults and thrust faults in the same region (Hatzfeld *et al.* 1990). The 1986 Kalamata earthquake (M_s 5.7) occurred 40 km ESE of the 1899 event, and involved normal faulting on a plane striking 210°. We have assumed normal faulting on a N–S plane for the 1899 event, but have no confidence in this assumption.

The earthquake was followed by a seismic sea-wave, about 1 m high that flooded the coast at Marathos and other places not specified including the coast of Kiparisia. In the island of Zante the height of the wave was 0.4 m. The shock caused no damage to the submarine cables between Zante and the mainland at Tripiti and Katakolo.

The shock was felt throughout the Peloponnesus and it was perceptible along the coast of Epiros and Albania as far as Mineo and Catania in Sicily. However, it was barely perceptible in Athens and it was not felt in the rest of Greece or in the islands of the Aegean, Fig. A8.

The earthquake was recorded by 20 seismographic stations up to a distance of 20° and was followed by relatively few, weak aftershocks that continued till early May.

P; *Galanopoulos (1941); Mitzopoulos (1900); Eginitis (1901); *Sieberg (1932); Flemming, Czartoryska & Hunter (1971); Kraft, Rapp & Aschenbrenner (1975)

1909 May 30; Fokis (*Ms* 6.3)

This earthquake affected the north central part of the Gulf of Corinth and in particular the sparsely inhabited region
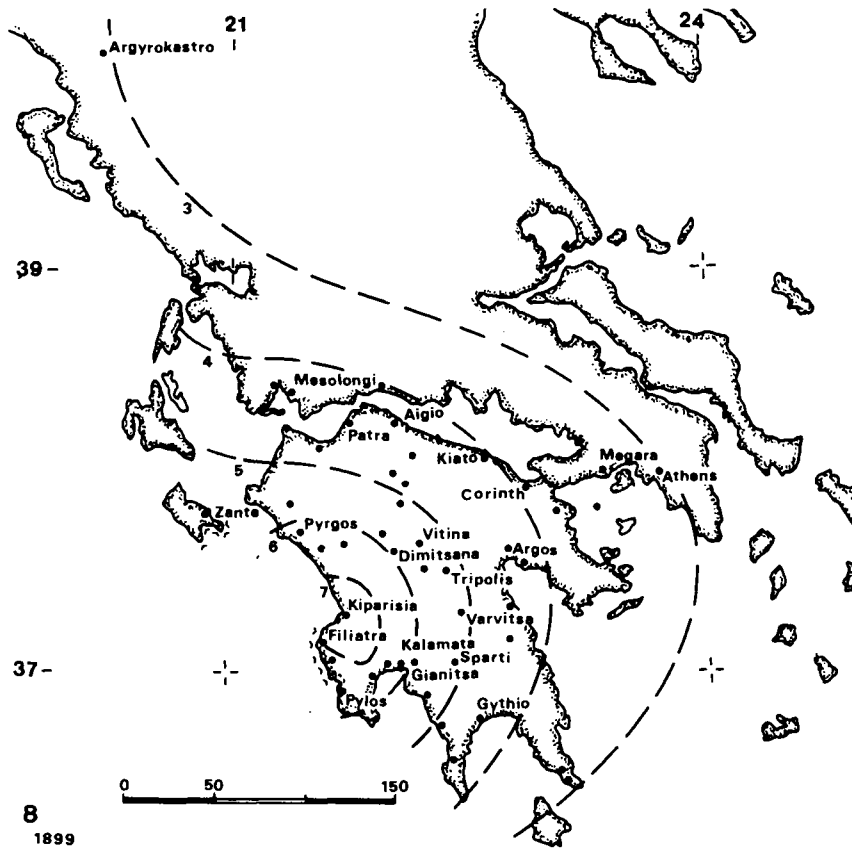
Figure A8. 1899 January 22; Kiparisia earthquake (M_s 6.1).
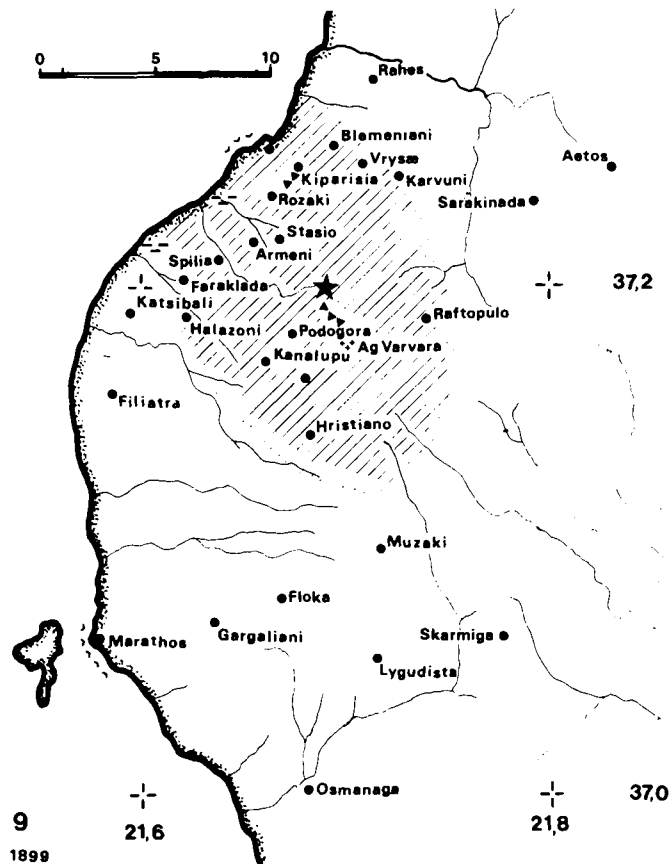
Figure A9. 1899 January 22; Kiparisia earthquake (M_s 6.1).
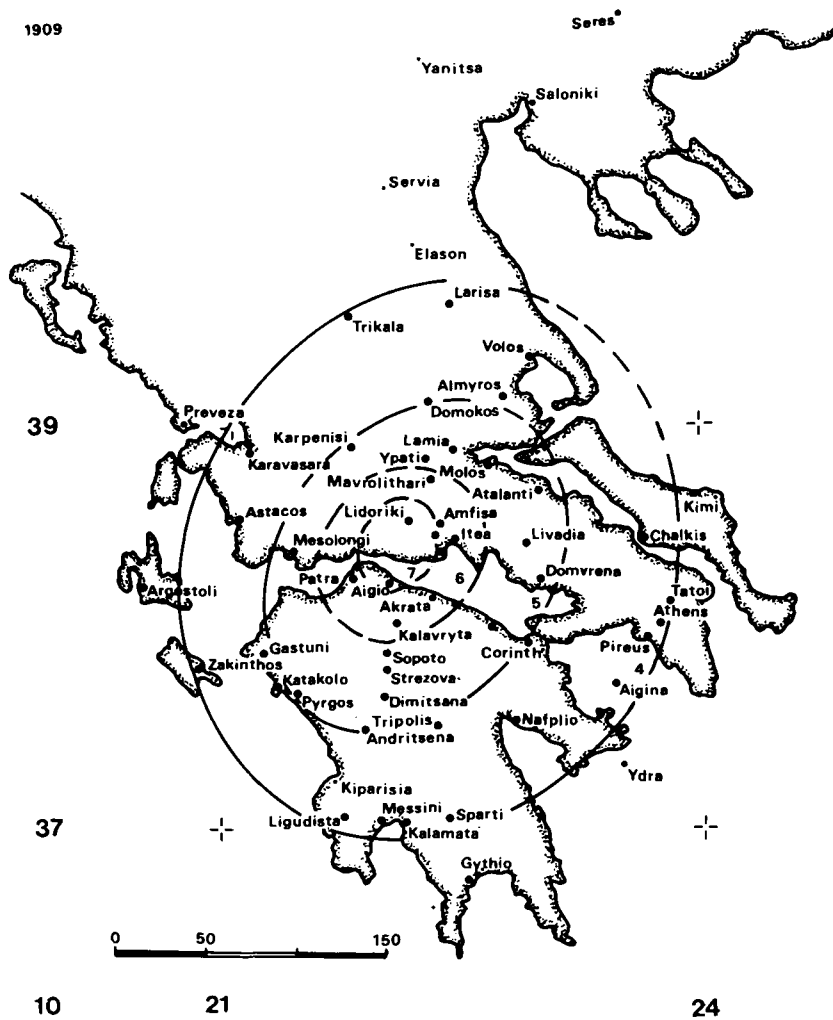
Figure A10. 1909 May 30; Fokis earthquake (M_s 6.3).

between the River Mornos and the coast, Fig. A11. It caused extensive damage, about which we know little detail, in the area between Lidorki, Sergula and Vitrinitsa where almost all houses collapsed with casualties. In this area springs of water dried up and in places, not as yet identified, the ground opened up. In Lidoriki a few houses collapsed and many were ruined; telegraph lines were thrown down and landslides blocked roads. Duvia was totally destroyed mainly due to landsides and slumping of ground, Fig. A11. We have assumed a focal mechanism (090:74: − 115), similar to the earthquake of 1970 April 8, farther east.

Damage extended outside this area, particularly north-west of Aigio where much of the destruction was due to landslides and liquefaction of the ground. The shock caused panic and sporadic damage in the towns of Amfisa, Aigio and Patras where a number of public buildings were badly cracked. The shock was perceptible as far as Saloniki and Seres in the north and as for as Mesina in south Italy, Fig. A10.

As a result of the earthquake the submarine telegraph cable along the Gulf of Corinth broke off Akrata, Fig. A10, presumably due to a submarine slide. However there is no evidence of a seismic sea-wave associated with this event. Two more cable breaks occurred two months later, on July

11 and August 28, which were caused by submarine slides, but not associated with seismic activity.

The earthquake was followed by a short aftershock sequence of small events, which, however, caused some damage in the region between Vitrinitsa Sotena.

P; Eginites (1912a, b); Malladra (1925); Heezen & Johnson (1966); *Shebalin et al. (1974)

1914 October 17; Thiva (Ms 6.2)

The earthquake occurred without a foreshock and caused heavy damage in the region east of Thiva to Dramesi, Fig. A13, a sparsely populated area where a few people were injured. The villages of Vratsi and Mustafa were totally ruined and segments of the railway line between Thiva and Chalkis, shown by inverted triangles in Fig. A13, were damaged. Telegraph lines to Chalkis were thrown down and a number isolated churches in the region were ruined. In Thiva about 15 houses collapsed and many became uninhabitable. Sporadic damage extended as far as Livadia, Chalkis and Megara, the shock causing panic in Athens and Piraeus, Fig. A12.

A second shock 4 hr later added to the damage and
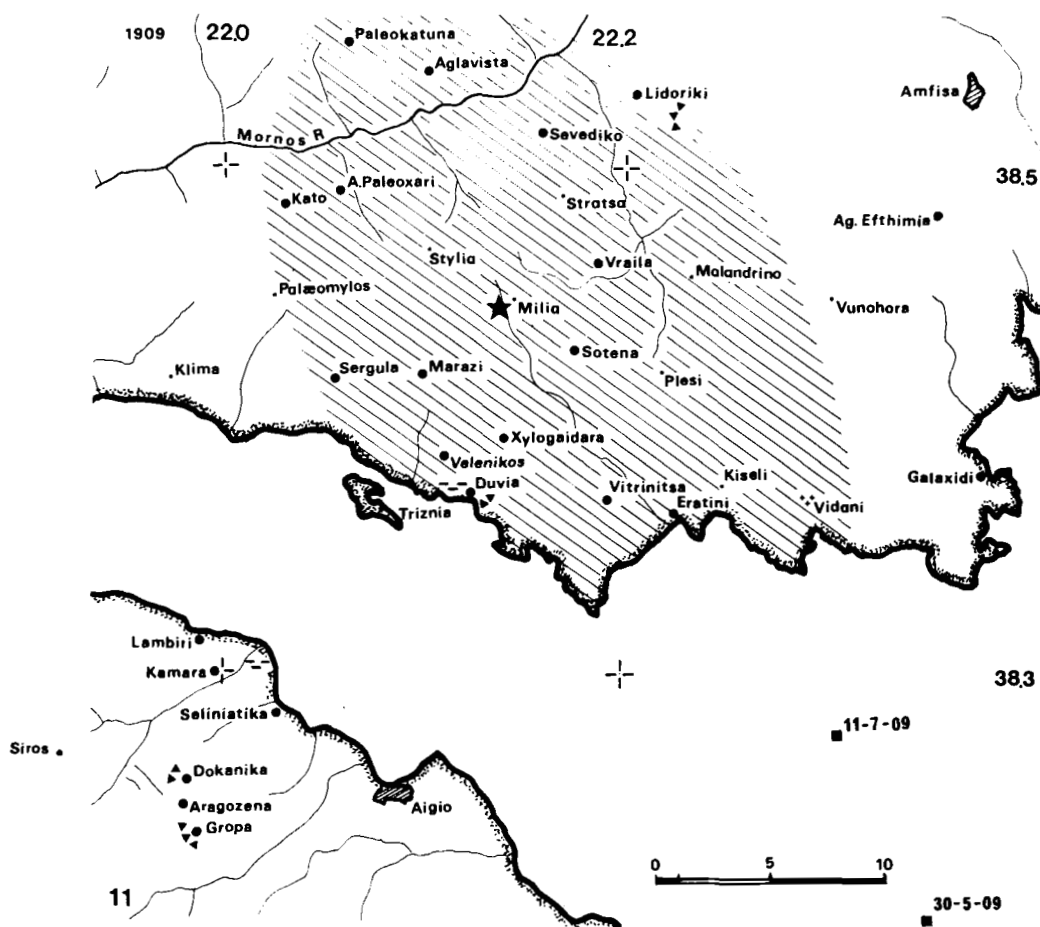
Figure A11. 1909 May 30; Fokis earthquake (M_s 6.3).

enlarged the epicentral area to the southwest of Thiva. This aftershock caused maximum damage in the area between Palaeopanagia and Kriekuki, particularly at Kapareli, but not in Vaya. It is not possible to determine from contemporary reports how much of the damage was attributable to each shock. However, a clear westwards migration of seismicity is evident in the macroseismic reports. There is no information about ground deformations of any kind. From numerous E–W striking escarpments bounding limestone ridges in the epicentral regions we assume normal faulting (100:45: – 090), probably in the region of Mustafades (Kalithea) and Chlebotsari (mod. Asopia)

P; Eginitis (1928a); *Goulandris (1928)

1915 June 4; Agrafa (*Ms* 6.1)

This earthquake has a remote and sparsely populated epicentral region in the upper reaches of the Acheloos river in the mountains of Valtos and Agrafa, Fig. A14. We know that in the region between Dimari, Granitsa, Katavothra and Sukaresti houses were destroyed with casualities, but details are lacking. The hamlets of Avlaki and Vovrano on the Acheloos were totally destroyed and rockfalls added to the damage here and near Yanuti. The shock damaged two old masonry bridges and affected the yield of springs,

causing sporadic damage over a relatively large area. However, information of serious damage is available only from a few places and this is insufficient to delineate the epicentral region, Fig. A15. We have assumed the fault mechanism (200:56: – 040) to be the same as that of the earthquake of 1967 May 1, further north. This is uncertain as the epicentral region lies between the Gulf of Corinth, where normal faults have E–W strikes, and Epirus, where N–S striking normal faults occur (cf. 1967 May 1)

P; De Giorgio (1917); Critikos (1926)

1923 December 5; Kasandra (*Ms* 6.4)

This earthquake had an offshore epicentre south of Cape Kasandra of the Chalkidiki peninsula, where a few coastal villages suffered considerable damage with no casualties. The shock damaged the tower of the lighthouse at Akra Posidion and caused slumping of the ground in the bay of Furka. There is no evidence for a seismic sea-wave and it caused no damage to the submarine cables in the region.

The shock was felt mainly to the north of the epicentral region in Greece and in south Bulgaria and Yugoslavia as far as Sofia, but not in south and west Greece (Fig. A16). It was not followed by any significant aftershocks. We assume this earthquake to be associated with normal faulting (135:45: – 090) on a plane striking NW–SE. Such faulting is
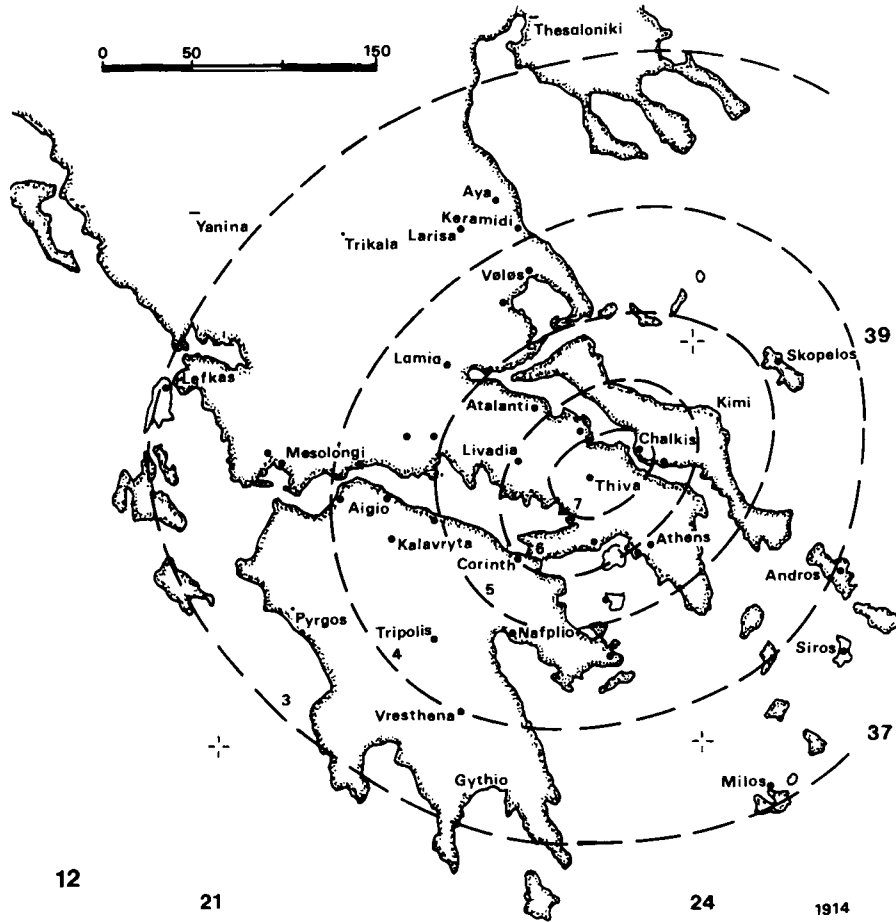
Figure A12. 1914 October 17; Thiva earthquake (M_s 6.2).
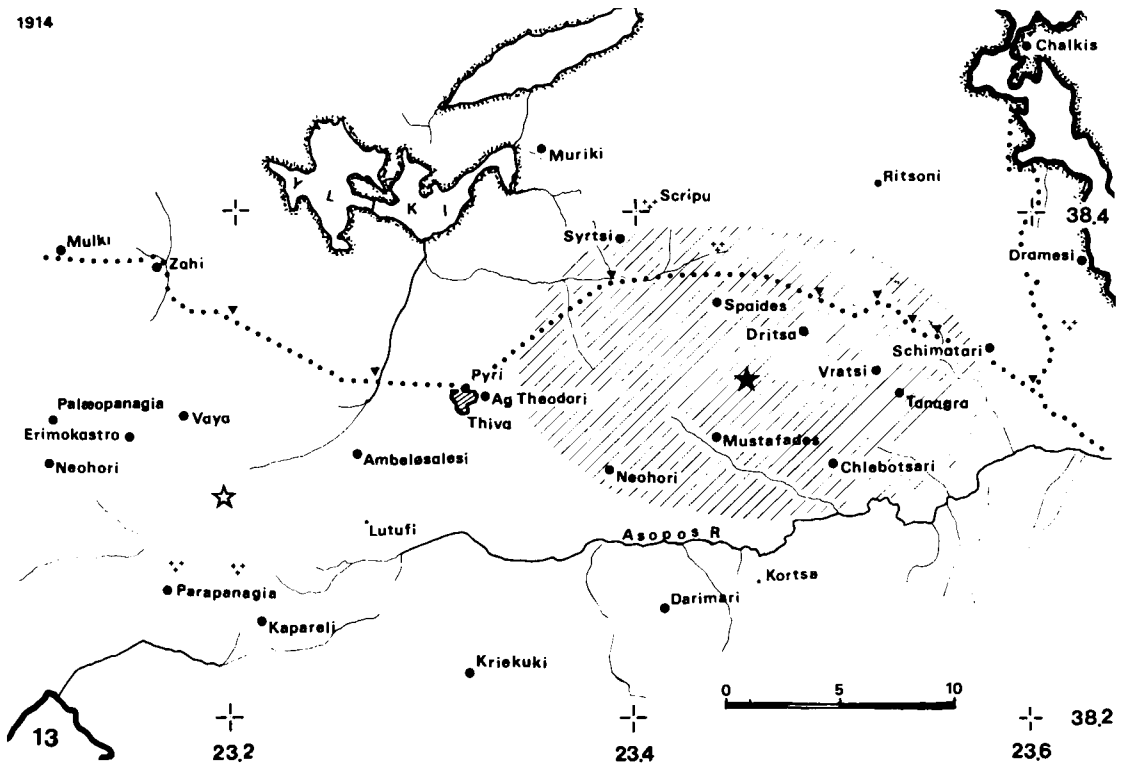
Figure A13. 1914 October 17; Thiva earthquake (M_s 6.2).
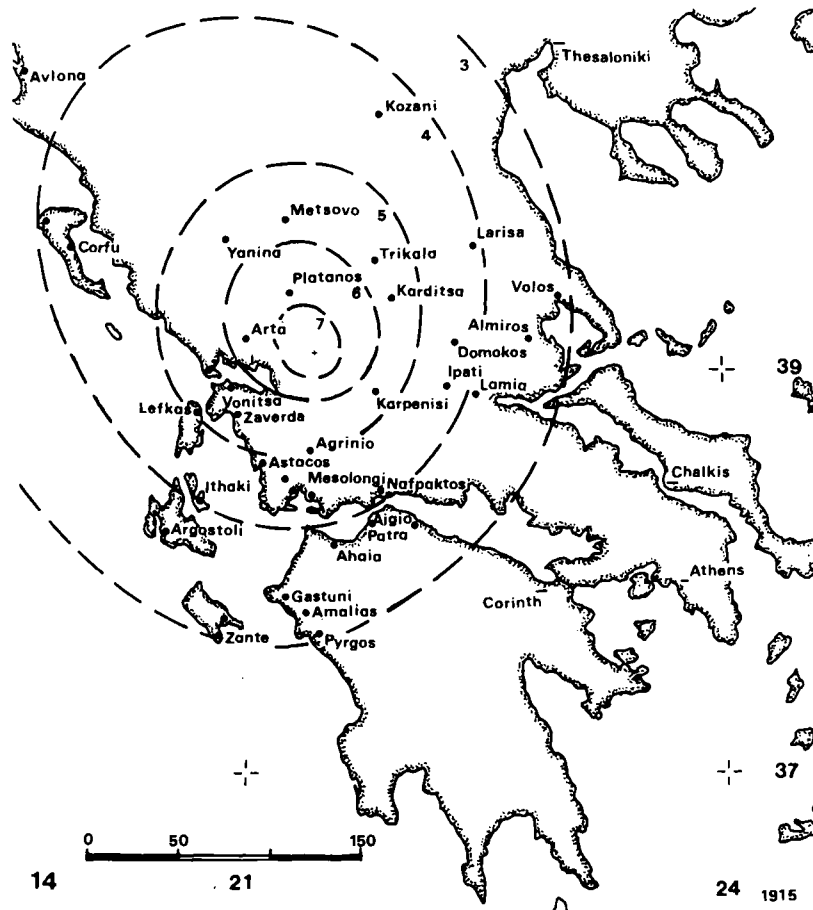
Figure A14. 1915 June 4; Agrafa earthquake (M_s 6.1).
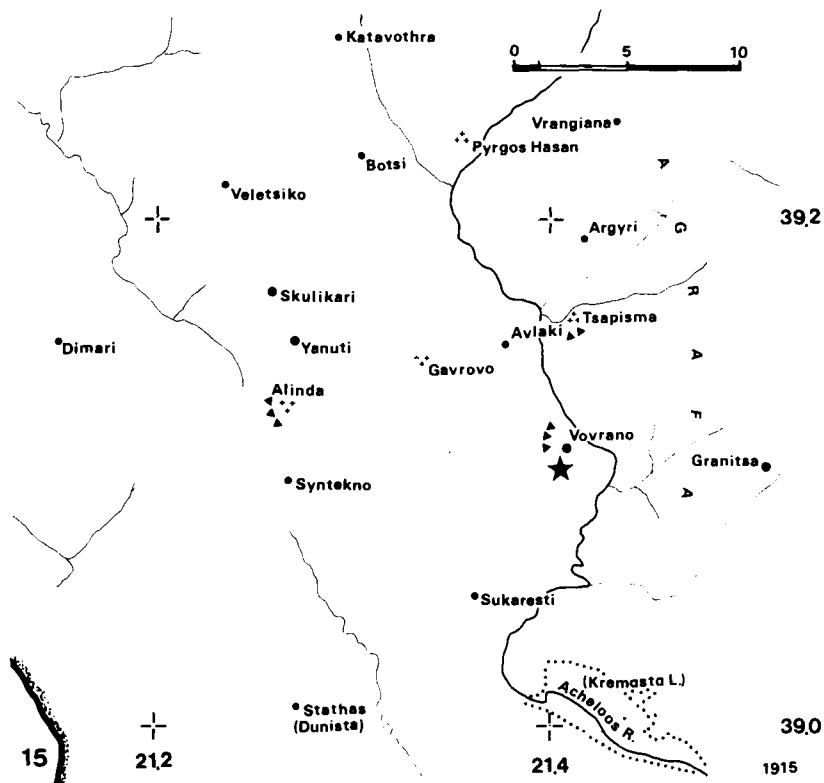
Figure A15. 1915 June 4; Agrafa earthquake (M_s 6.1).
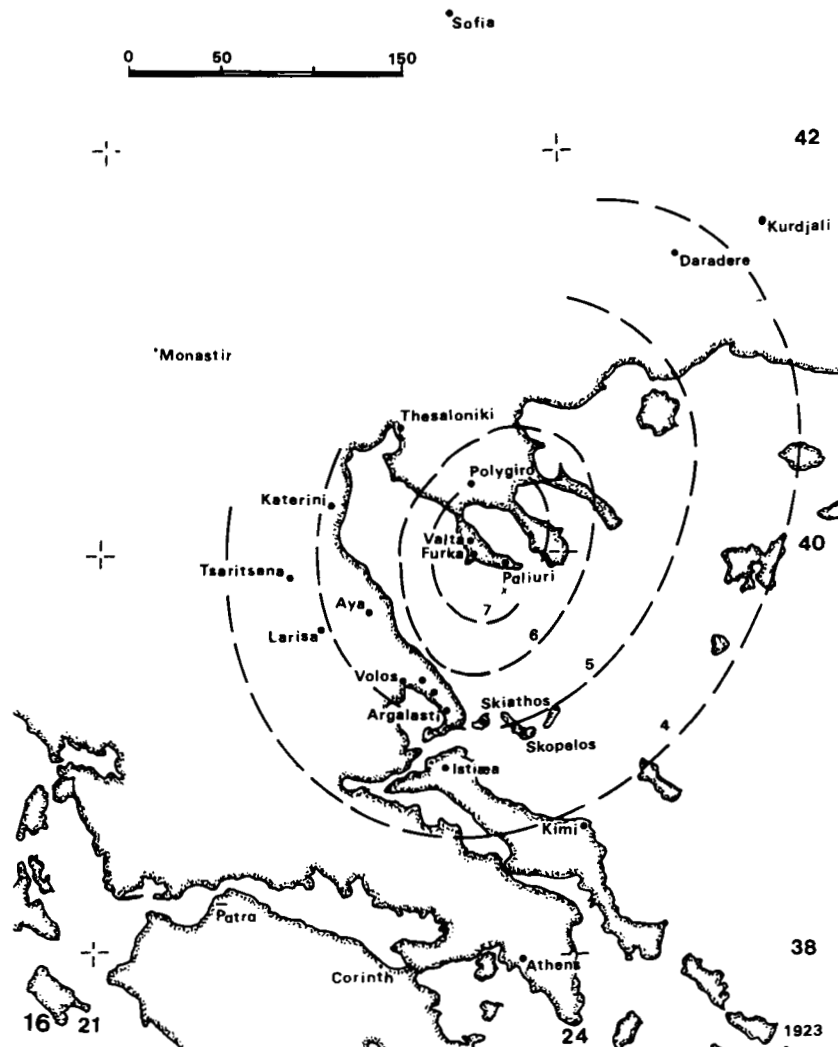
Figure A16. 1923 December 5; Kasandra earthquake (M_s 6.4).

seen in the N. Aegean trough (McKenzie 1978) and is responsible for the three peninsulas of Chalkidiki: but strike slip faulting on NE–SW planes also occurs in this region.

P; Critikos (1928); Kirof (1931); Makropoulos & Burton (1981)

1927 July 1; Laconia (*Ms* 6.4)

The earthquake was felt over a very large area that included Sicily, Greece, Crete, the Nile Delta and Cairo, Fig. A17. Using the amplitudes of *P*-, *S*- and surface waves Gutenberg discriminated this shock from the shallow events of the region (Gutenberg & Richter 1948, and Worksheet no: 86:1010), and estimated a focal depth of about 120 km. Using first-arrival readings only, Makropoulos & Burton (1981) estimated a focal depth of 45 km, and ISS treats this as a shallow event. If we exclude the extreme points at which the shock was felt in Sicily and Egypt, where the mode of the intensity observations did not exceed II (MSK), the intensity distribution shown in Fig. A17 is consistent with a crustal event. We assume a normal fault parallel to the coast, (335:45:−090). The event could be deeper,

associated with the subduction zone, but Hatzfeld *et al.* (1989) found no seismicity deeper than 45–50 km in the Gulf of Messinia—the deeper events occur further east.

The earthquake was felt as two groups of shocks of long period that caused relatively small damage at widely separated localities. However, concentrated damage did occur on the east coast of the. Gulf of Mesinia along the west-facing slopes of Mt Taygetos, Fig. A18. Between Avia, Arna and Arepolos almost all villages were ruined and hundreds of houses collapsed without casualties. Rockfalls blocked country roads, mainly in the region between Avia and Lefktro.

The shock was not associated with a seismic sea wave and no abnormal fluctuation of the sea was reported from the tide station at Kalamata. The earthquake was not preceded or followed by significant activity.

P; Critikos (1932b); Gutenber & Richter (1948); Shebalin *et al.* (1974); Makropoulos & Burton (1981)

1928 April 22; Corinth (*Ms* 6.3)

The Corinth earthquake was preceded by two shocks and followed by three aftershocks of magnitude 5.0–5.2,
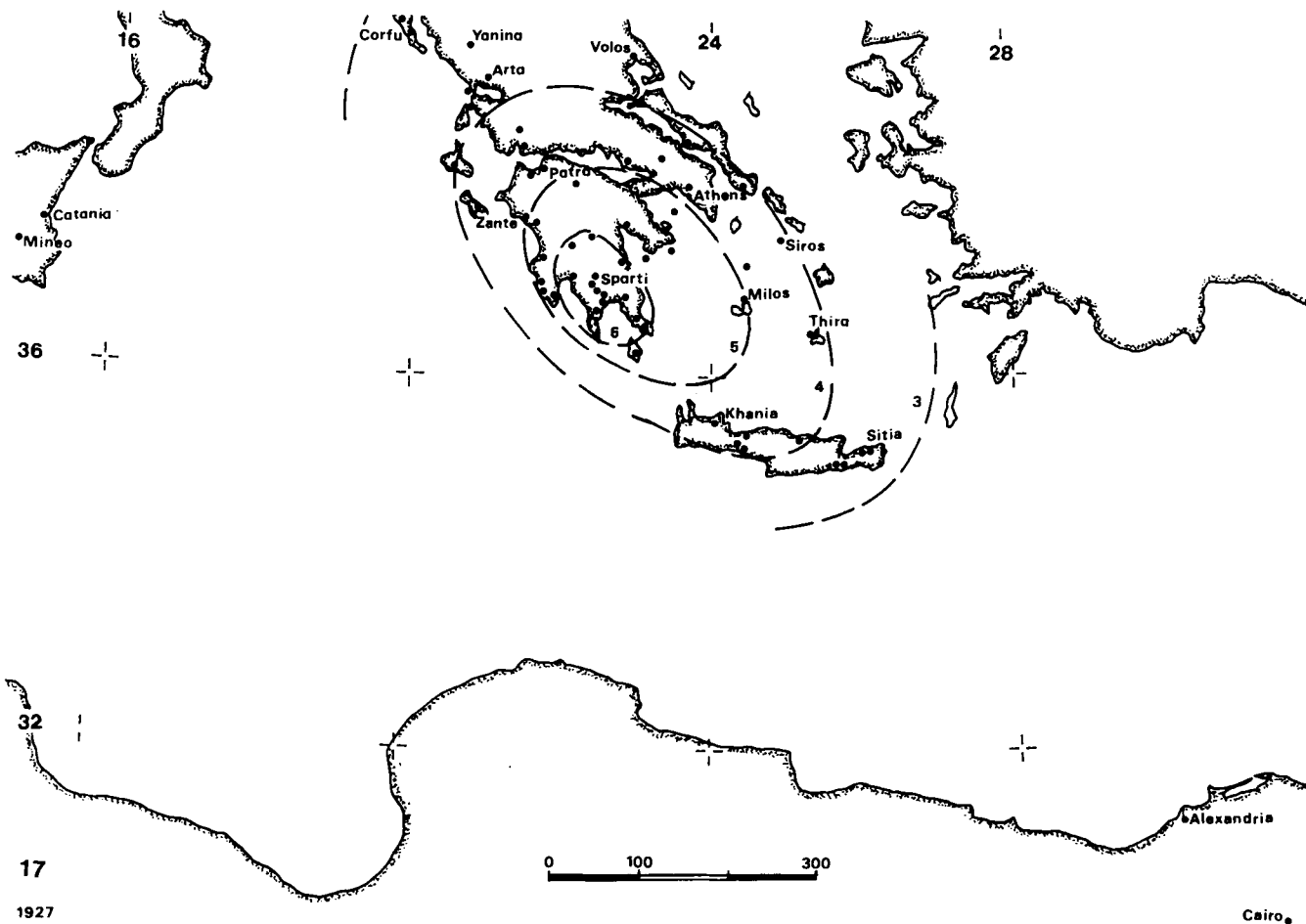
Figure A17. 1927 July 1; Laconia earthquake (M_s 6.4).
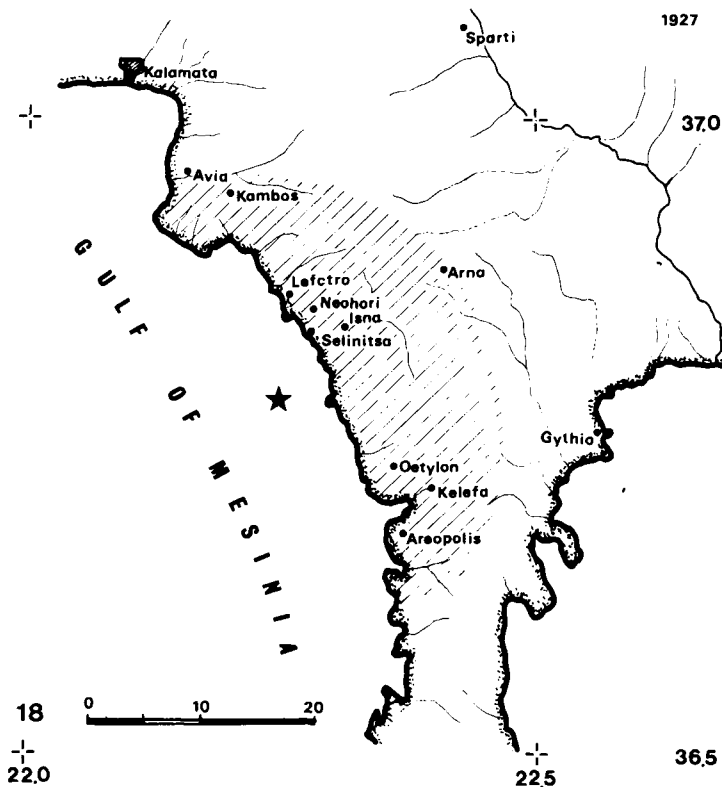
Figure A18. 1927 July 1; Laconia earthquake (M_s 6.4).

Fig. A19. The main shock destroyed more than 3000 houses and killed 22 people in the region between Skinos, Kalamaki, Corinth and Xylokastro, Fig. A20. The largest urban centre, Corinth, was almost totally ruined and the districts of Perahora, Asos, Neratzia, Dimino and Xylokastro suffered equally heavy losses. This heavy damage was mainly due to the low quality of the local type of construction and also because of foundation failures. The observed distribution of damage implies an offshore epicentre, somewhere between Perahora and Kiato, not far from where the submarine telegraph cable off Xylokastro failed, presumably due to a flow-side, 22 km NW of Corinth, Fig. A20. There is no evidence that the earthquake was associated with a seismic sea-wave. The tide gauge at Isthmia does show an abnormal fluctuation of the sea level, but this irregular tidal regime

observed throughout the Aegean Sea during the period April 21 to 25 is known to have been caused by abnormal barometric changes.

Ground deformations, mainly due to slumping, were reported from the port area of Corinth and from the coastal areas near Lutraki, Kalamaki and Xylokastro. No substantial surface faulting was reported from this earthquake, either because it might have been missed, or because it was offshore. Short ground ruptures striking N–70–W were reported from half-way between Lutraki and Posidonia. Other cracks, also a few tens of metres in length, were found following a pre-existing fault about 300 m SE of the railway bridge on the Corinth canal, striking N–70–E. The shocks triggered rockfalls from the Gerania Mt. above Skinos, Susaki and Patapios, and caused the collapse of
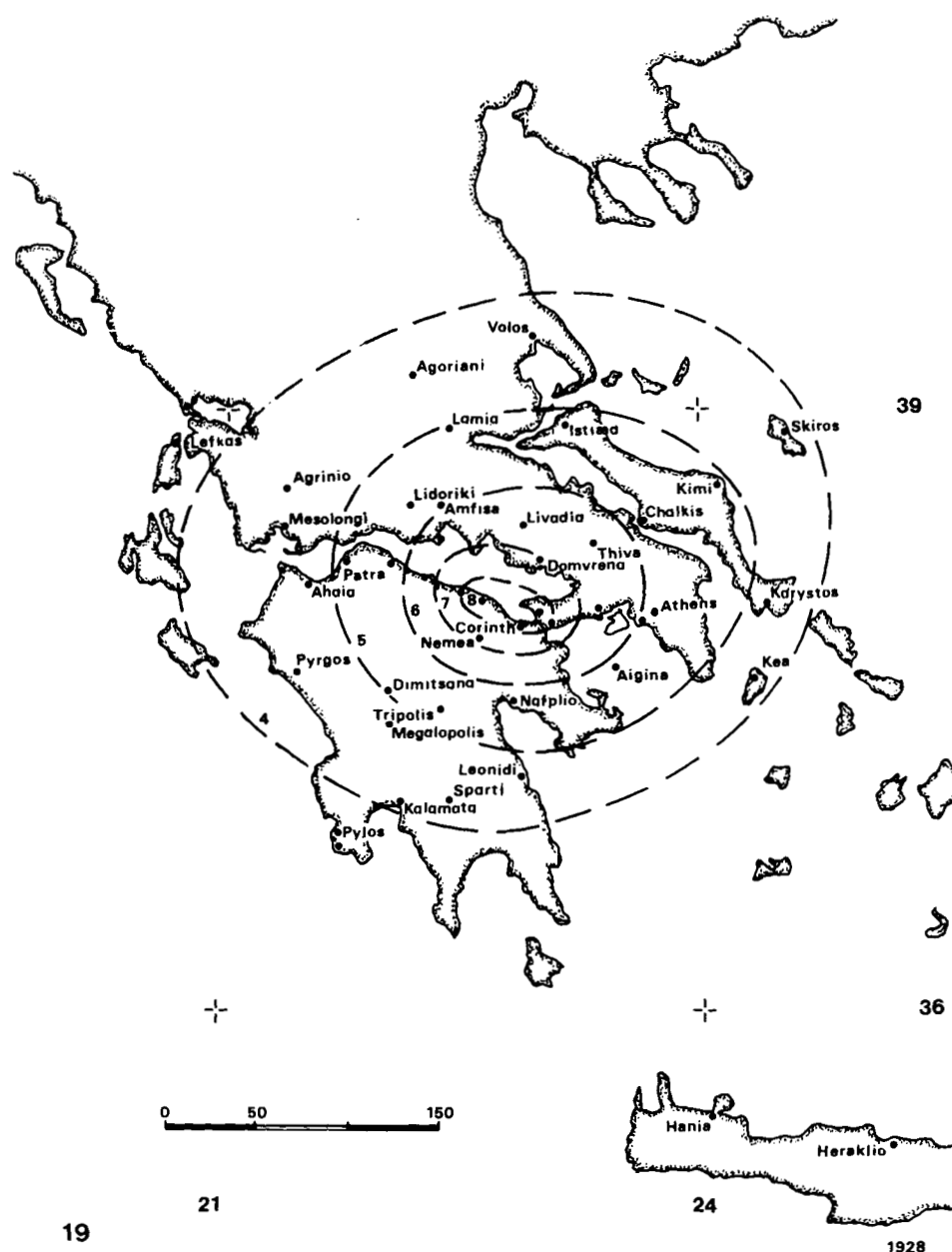
Figure A19. 1928 April 22; Corinth earthquake (M_s 6.3).
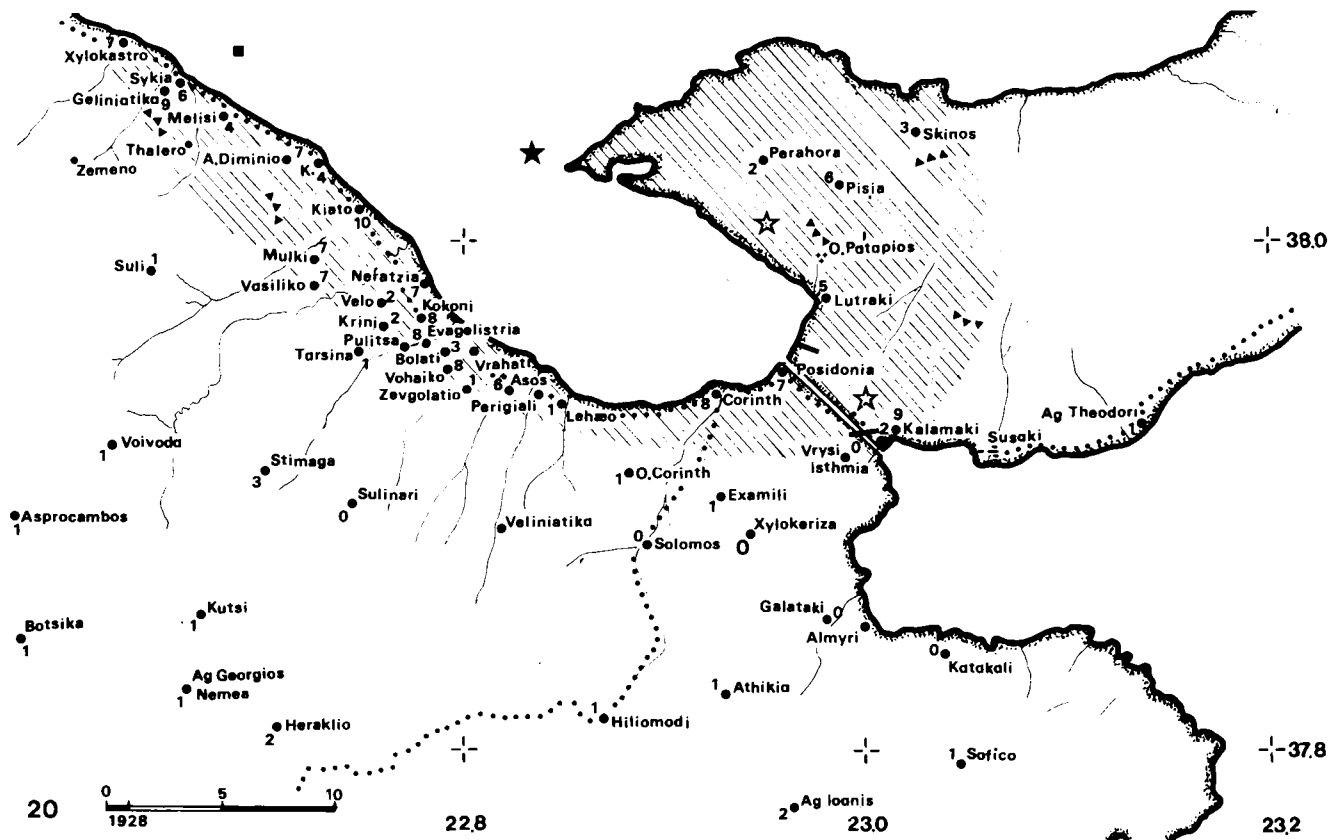
Figure A20. 1928 April 22; Corinth earthquake (M_s 6.3).

steep slopes in marls near Kiato and Melisi. We have assumed normal faulting (285:40: −070), the same as in the earthquake of 1982 February 24. Normal faults striking WNW dominate the structure of this region, particularly offshore (Perissoratis *et al.* 1986).

Macroseismic information suggests that the foreshocks had epicentres near the Corinth isthmus, and that those of the three aftershocks were near the Gerania mountains to the north. The earthquakes caused no damage to the Corinth canal and second-order leveling along the railway line through Corinth showed no changes.

The shock was not felt very far to the north. Fig. A19. Later authors extend the isoseismals into Bulgaria confusing felt reports from the earthquakes in Philipople of April 20 and 25 with those of the Corinth earthquake.

P; Ktenas 1928; Roussopoulos (1928); Tanakadate (1928); Sieberg (1928; 1932); Eginitis (1928b; c; 1931); Critikos (1932b); Heezen *et al.* (1966); Shebalin *et al.* (1974); Drakopoulos, Leventakis & Roussopoulos (1978)

1930 March 31; Magnesia (*Ms* 6.0)

Two shocks of almost equal magnitude with epicentres offshore from Magnesia caused considerable damage to villages in Mt Pilion, Fig. A21. The first shock of February 23 affected the region of Keramidi and Sklithro where a number of houses were ruined. Fig. A22. The second, and larger shock of March 31, affected the same area as well as

villages as far as Makrirahi, an area in which almost all houses became uninhabitable. The shocks caused no causalties. In places they triggered slides from the steep slopes on Mt Pilion, that added to the damage and blocked roads. In Volos, the largest urban centre in the region (population 42 000) a few old houses and public buildings were in part destroyed and the quay walls in the harbour settled excessively. We have assumed the mechanism to be normal faulting (310:45: −090) with a strike parallel to the coast.

P; Critikos (1932a, b)

1930 April 17; SE Corinth (*Ms* 5.9)

This earthquake was not preceded or followed by shocks of any magnitude. It affected primarily the west coast of the Saronikos Gulf, Fig. A23, where a few villages and hamlets were totally ruined and a small number of people were injured. In the bay of Almyri the beach slumped into the sea and the landing place of Kehriae was badly cracked. Rockfalls were triggered from the mountain south of Katsikuli and above Porto Frango, Fig. A24. As a result of the earthquake the depth of water over the bank west of the islet of Evraeos increased between the bathymetric surveys of 1927 and summer 1930. We assume the mechanism to be normal faulting with E–W strike, similar to the known fault along the shore between Vlasi and Frango, Collier (1988). the tide meter at Isthmia showed no abnormal fluctuation of the sea-level in the Saronic Gulf. Outside the epicentral area

Figure A21. 1930 March 31; Magnesia earthquake (M_s 6.0).
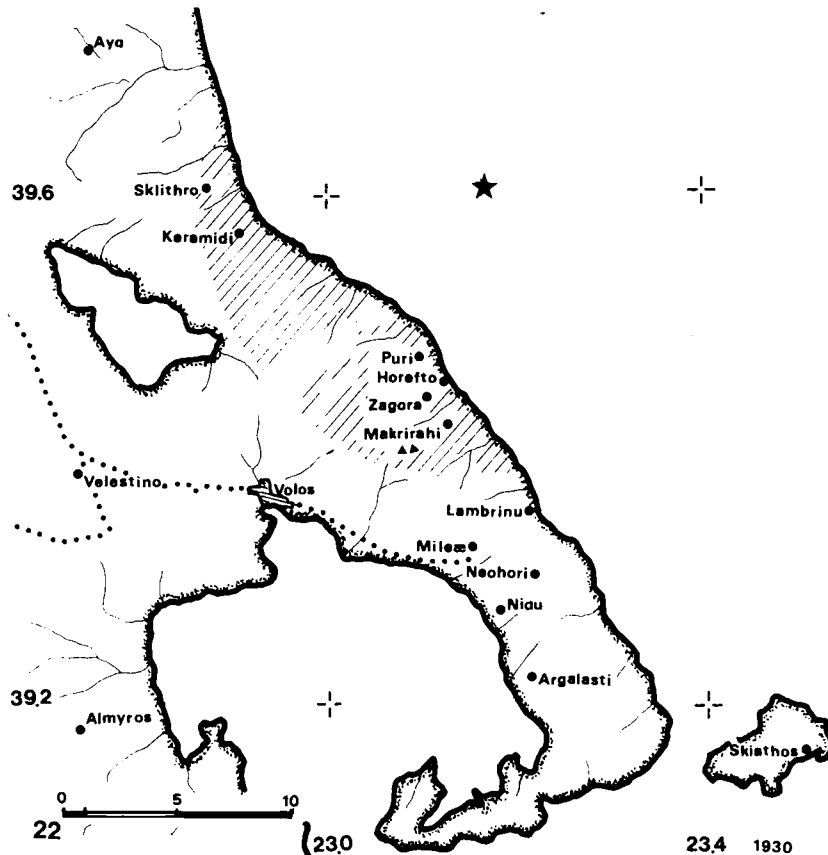
Figure A22. 1930 March 31; Magnesia earthquake (M_s 6.0).
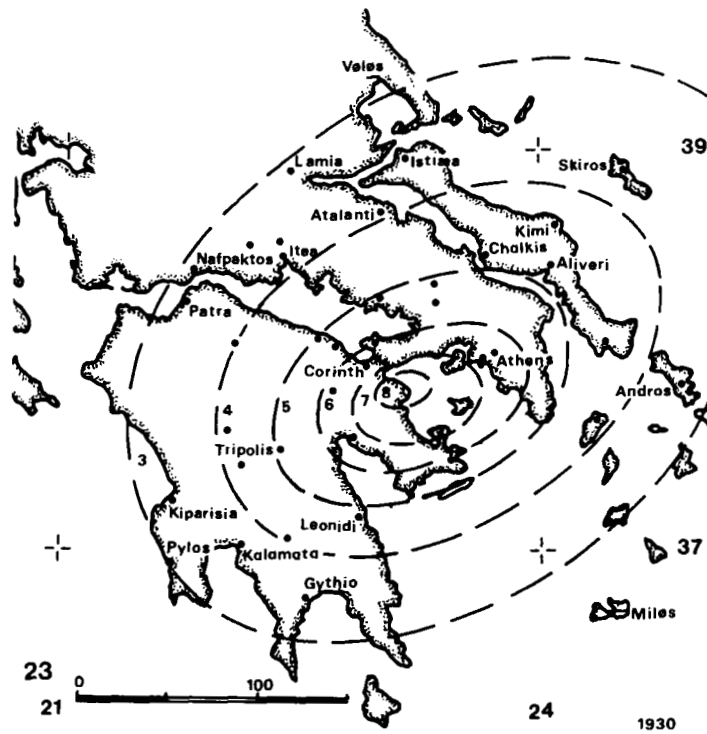
Figure A23. 1930 April 17; SE Corinth earthquake (M_s 5.9).
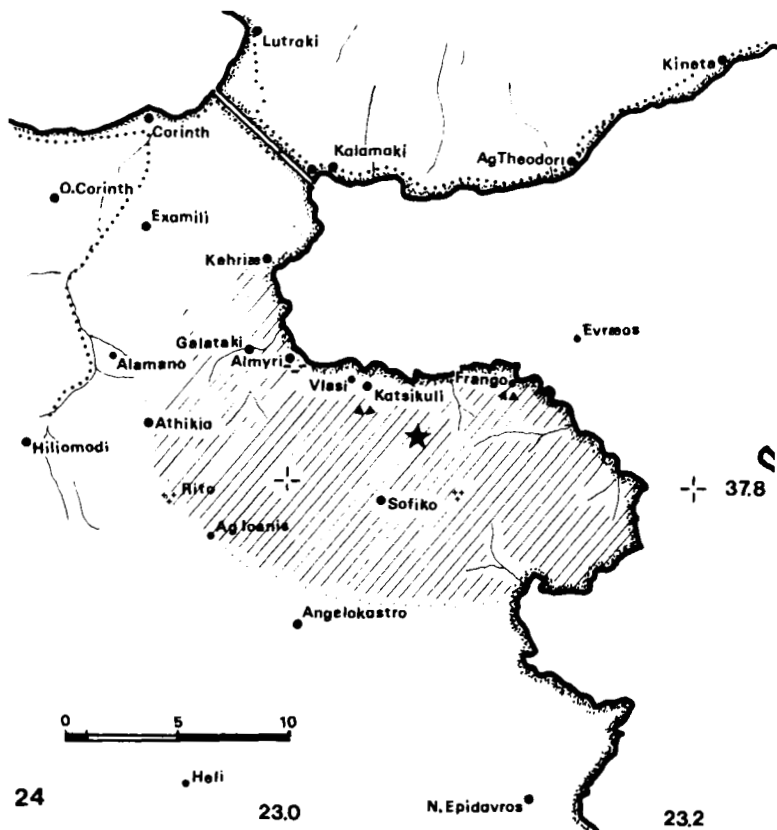
Figure A24. 1930 April 17; SE Corinth earthquake (M_s 5.9).

damage, enhanced by the Corinth earthquake two years earlier, was widespread but minor.

P; *Critikos (1930); *Critikos (1932a,b); *Shebalin *et al.* (1974)

1938 July 20; Oropos (*Ms* 6.1)

This earthquake caused considerable damage in the north part of Attiki province, particularly in the region between Oinofyta–Kakosalesi–Kapandriti and the coast to the north, Fig. A26. In all, about 8000 houses were destroyed or damaged beyond repair and 18 people were killed and 180 injured. The shock caused damage to the railway line between Schimatari and Kiurka. Railway buildings and water tanks were damaged and the masonry bridge at the 63 km mark was shattered, disrupting railway communication between Kakosalesi and Oinofyta. In Skala Oropou, at Halkustsi and near Palatia beach deposits liquefied and long cracks in the ground were formed running parallel with the coast, in places associated with mud volcanoes. The depth of the shoal water that fringes the coast of Halkutsi increased noticeably. In contrast, it is reported that rocks emerged off the coast of Kalamos and that the salt springs at Ag. Apostoli dried up temporarily. Ground cracks 30–50 m long, running intermittently for some distance, were reported from south of Oropos with the valley side (north) downthrown by a few tens of centimetres. There was no damage to the railway tunnel near Kiurka but near here ground cracks were reported that extended for a few hundreds of metres beyond the northern portal of the tunnel. We have assumed normal faulting (290:45: − 070) parallel to young faults observed between Oropos and Milesi, as well as between Palatia and Ag. Apostoli. The slip vector is taken to be parallel to other faults in the north Evia Gulf.

The earthquake caused some concern to the authorities because of its association by some experts with the impounding of the Marathon reservoir, a lake retained by a 35 m high arch dam about 15 km SE of the epicentre, completed seven years earlier.

A strong aftershock on July 27 of $M_s = 4.7$, caused additional damage to Kalosalesi (mod. Avlon) and to Oropos, followed by minor shocks that were reported from the same region. The main shock was not felt very far, Fig. A25, and it was of short duration.

P; Negreponte (1938); Galanopoulos (1966)

1941 March 1; Larisa (*Ms* 6.1)

This earthquake occurred in the plain of Larisa, Fig. A27, and caused extensive damage to the villages along the east margin of the plain, Fig. A28. In the epicentral area more than 250 houses collapsed completely and about 900 were damaged beyond repair, without casualties. This was due to
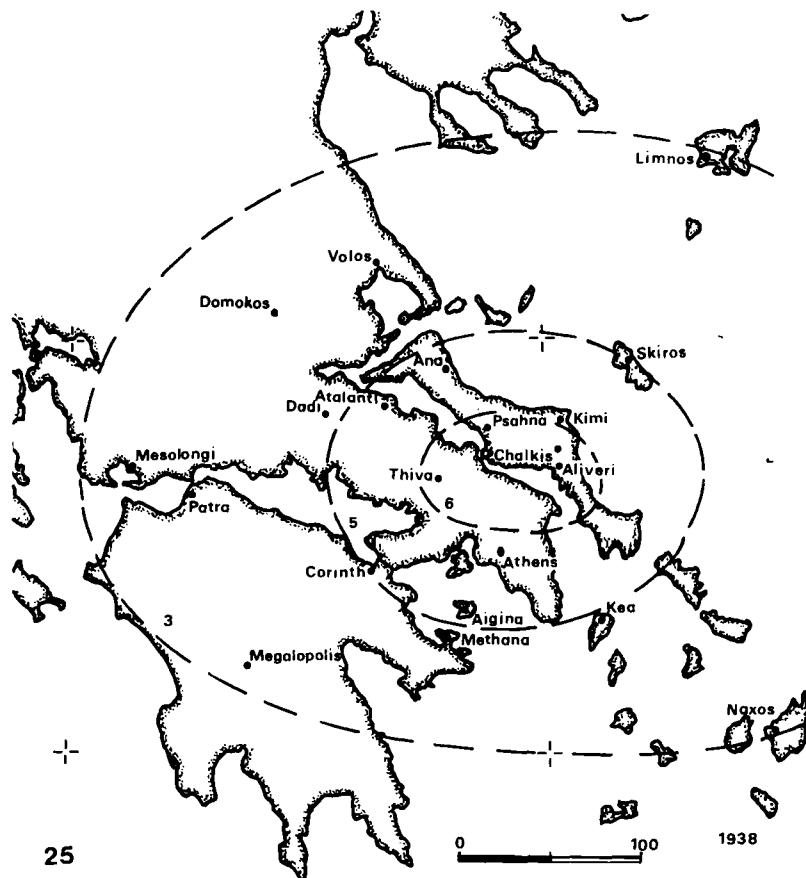
Figure A25. 1938 July 20; Oropos earthquake (M_s 6.1).
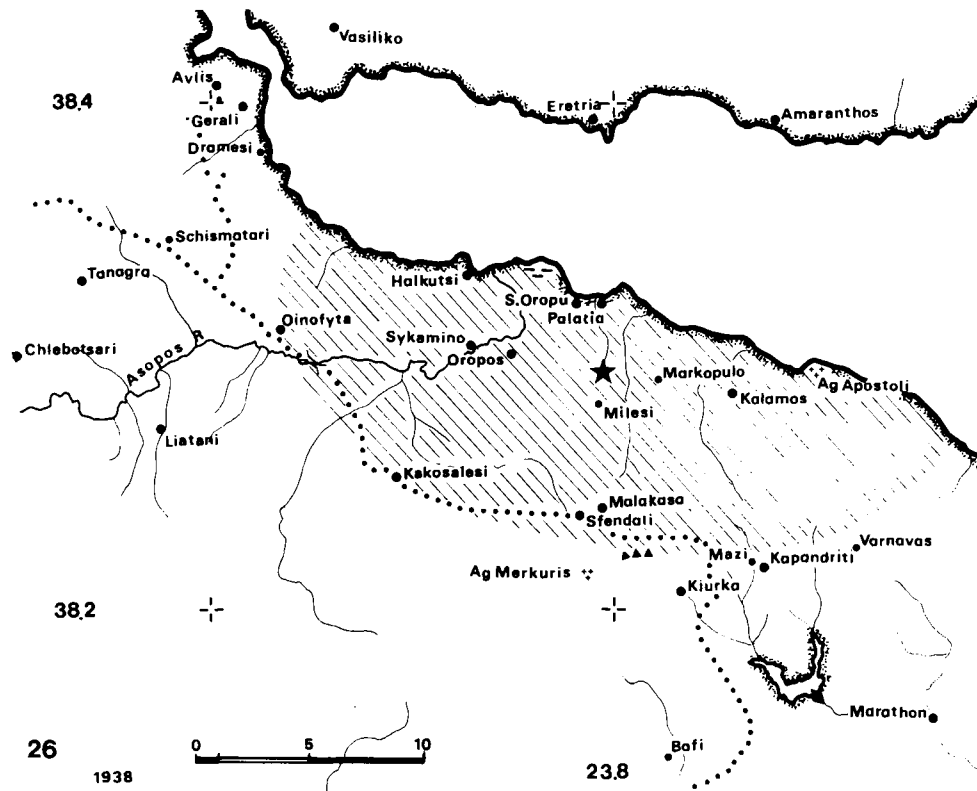
Figure A26. 1938 July 20; Oropos earthquake (M_s 6.1).

Figure A27. 1941 March 1; Larisa earthquake (M_s 6.1).

the use of timber framed constructions that resisted the shock without collapse in many villages. The shock caused damage to the railway and telephone lines as well as to the irrigation system. Widespread liquefaction of the ground, foundation failures and slides added to the damage. Along the river Pinios and its tributaries the ground slumped and cracks appeared along banks running for many kilometres.

The earthquake occurred during the Greek–Italian war, shortly after a series of air raids on Larisa, a town with a population of over 30 000, which was heavily damaged. The combined effects of the air raids and earthquake resulted in the destruction of 600 and the damage of 3500 houses that made 10 000 people homeless. About 40 people were killed in the earthquake and over 100 were injured. Liquefaction of the ground was reported from many places in the town as well as the development of high artesian pressures that resulted in the damage of deep walls.

The earthquake produced widespread ground fracturing in the valley floor, all of it associated with ground failures. A series of ruptures in crystalline limestones and schists east of Toivasi, striking N–150–E reported after the earthquake, are the only features found on rock, not necessarily of tectonic origin.

Within 12 hr, the earthquake was followed by three aftershocks of magnitude less than 5.1 that concluded the sequence.

Tentatively, we assume normal faulting (150:45: − 090) parallel to the escarpment bordering the plain of Larisa.

P; *Critikos (1941); *Maravelakis (1944); Galanopoulos (1946); *Galanopoulos (1950); *Shebalin *et al.* (1974)

Figure A28. 1941 March 1; Larisa earthquake (M_s 6.1).

1947 October 6; Mesinia (*Ms* 6.9)

Without significant foreshock of aftershock activity, an earthquake offshore SW Peloponnesus caused widespread damage in the district of Pylos, Figs A29 and A30. The area most seriously damaged extended from Hristiano in the north to Koroni on the coast in the south, an area of a radius of about 30 km. Within this area, however, intensities varied from place to place, from *VI+* to *VIII +* (MSK), without any well-defined pattern. The earthquake destroyed or damaged beyond repair about 5 per cent of the houses in the region and rendered uninhabitable another 16 per cent, that is, causing irreparable damage to 7800 houses injuring 29 people.

The shock triggered landslides within a large area and produced a small sea-wave which at Methoni flooded the coast to a depth of about 60 m inland. The landing places Petalidi, Pylos, Koroni and Kalamata suffered some damage due to slumping, and at Koroni the sea advanced inland permanently. The shock caused liquefaction of the ground along the coast from Petalidi to Nisi and as far as Kalamata to the east that caused damage to the Pamisos irrigation system. There is no evidence of ground deformations of tectonic origin. Maximum intensities, some of them enhanced by local soil conditions, are diffused within an area of about 20 km radius in the SW tip of the Peloponnesus, making it difficult to define the extent of the epicentral region. However, their general distribution suggests that the source extended offshore into the Gulf of

Kalamata where the ISS epicentre has been located, Fig. A30. The overall intensity distribution of this event is similar to earthquakes along the west margin of the Peloponnesus (1898, 1899, 1927). We have assumed the mechanism to be normal faulting (200:45:−080) with the same mechanism as the nearby Kalamata earthquake of 1986 September 12.

Long-period oscillations of the ground, lasting over one minute, were reported from epicentral distances up to 150 km. The shock was felt as far as the Nile Delta in Egypt, SE Italy and east Sicily, but it was not perceptible in northern Greece and in most of the islands in the Aegean.

P; Galanopoulos (1949), *Shebalin *et al.* (1974)

1954 April 30; Sofades (*Ms* 6.7)

Preceded by a small shock a few days earlier, a relatively large magnitude earthquake caused extensive destruction in the valley of Karditsa in central Greece, Figs A31 and A32. The earthquake destroyed about 8500 and damaged 5500 houses, killing 25 and injuring 175 people. Most of the losses occurred in the larger population centres situated on thick, saturated sedimentary deposits and swamp areas. The larger urban centres in the epicentral region of Karditsa (population 22 000) and Farsala (population 6000) suffered 32 and 18 per cent loss of their building stock. In Sofades (population 4300) more than 90 per cent of the houses, mostly old structures, were destroyed or damaged beyond
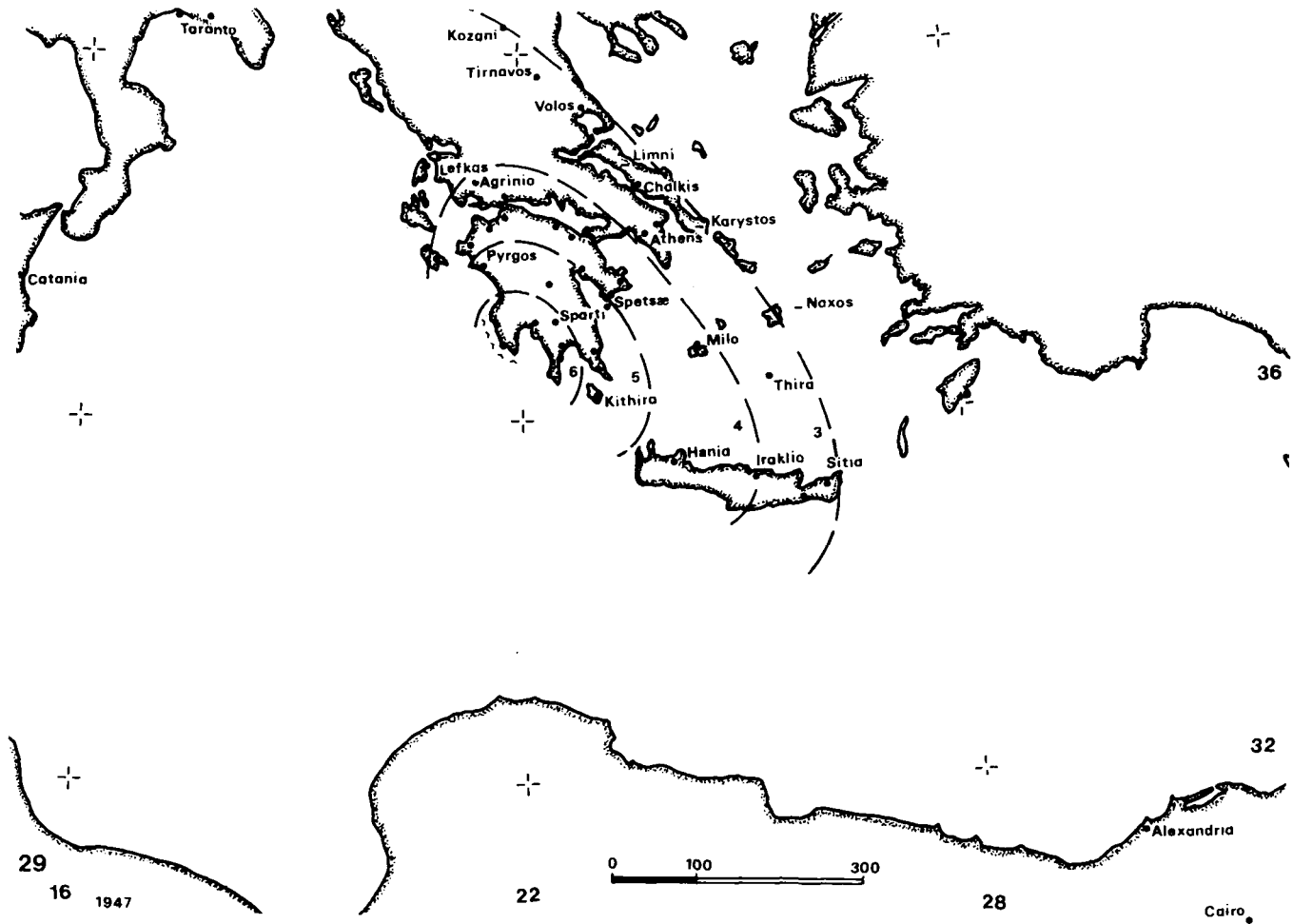
Figure A29. 1947 October 6; Mesinia earthquake (M_s 6.9).

repair. Sporadic damage extended over a large area outside the epicentral region, shown in Fig. A32.

The shock caused widespread liquefaction in the Karditsa valley and in the reclaimed swamps of Xynias–Domokos to the south, as far east as Volos, 50 km to the east of the epicentre, where port installations and a few buildings suffered some damage. The railway line, where it crosses the plain, was damaged; embankments slumped and spread, and railway stations in the valley were shattered. There was no damage to viaducts and bridges carrying the line across the hills. Telephone lines were thrown down but there was no serious damage to the road system, except between Sofades and Stavros, where a large number of culverts were destroyed due to spreading of the fill. Of the few large engineering structures in the epicentral region, grain silos suffered most due to foundation failures, some of them seriously. Most of the public buildings in the towns were rendered uninhabitable.

The earthquake was associated with a normal fault-break striking N–300–E, perhaps up to 30 km long, running discontinuously along the foothills of the SW margin of the valley, extending further to the southeast into the Domokos hills, Fig. A32. This break consists of a number of segments which have been identified in cursory inspections by different authors at different times during the last 35 yr,

probably not all of them associated with this earthquake, and some of them not visible today. The average throw on these breaks is about 30 cm and the maximum reported was 90 cm, with no clear evidence of the sense of lateral motion. Railway tracks broken by the fault movement between Velesiotes and Ekara, and road displacements near Kedros and Lefka, repaired immediately after the event, do not resolve this ambiguity. The first motion fault plane solution by McKenzie (1972) is good enough to confirm normal faulting, but not its orientation. We have assumed a strike parallel to the main topographic escarpment from Loutro to Ekara (300:45: − 090), Fig. A32.

The earthquake was felt throughout Greece and in most parts of Bulgaria and Macedonia, Fig. A32.

P; Papastamatiou & Vetoulis (1955); *Shebalin et al. (1974); Ambraseys (1975); Papastamatiou & Mouyaris (1986a,b)

1955 April 19; Volos (*Ms* 6.2)

This earthquake, preceded by a small shock, occurred in the region of Volos, Figs A33 and A34, in an area already damaged by the shocks of 1954. The main earthquake and the strong aftershock of April 21 affected the coastal strip east of the town, particularly the densely populated urban
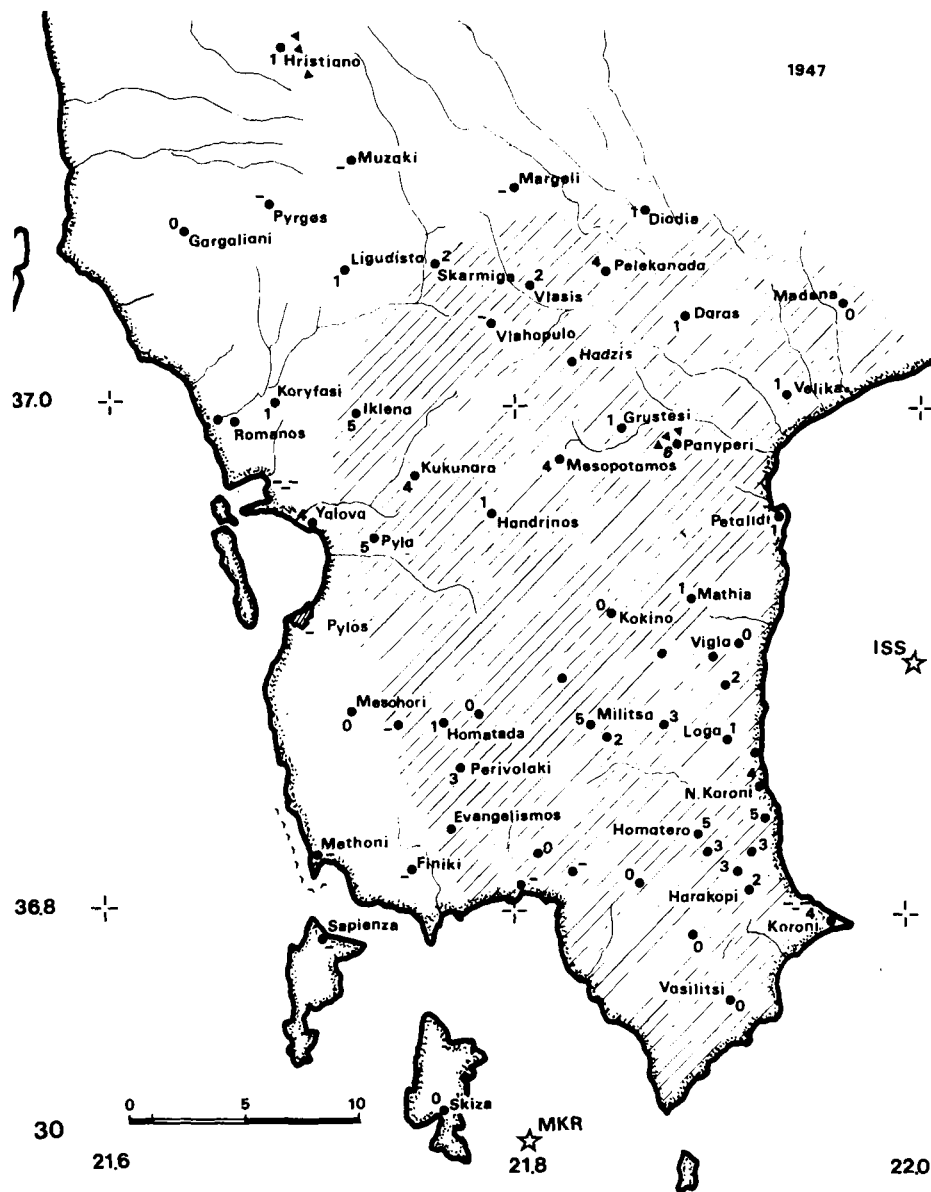
Figure A30. 1947 October 6; Mesinia earthquake (M_s 6.9).

area of Volos where 5 per cent of the houses were destroyed and 60 per cent were damaged, mostly vulnerable constructions or houses built on unstable ground. In all 910 houses collapsed and 13 800 were damaged; six people were killed and 150 injured. There was some damage to the port facilities of Volos, to the oil storage installations, customs house, quay and railway pier. The shock was associated with a seismic sea-wave in the Gulf of Volos that flooded the coast of Pagasae and caused ships to drag on their anchors. We assume normal faulting for this earthquake (090:45: −090) on an E–W plane, based on proximity to the 1980 earthquakes (Papazachos *et al.* 1983; Jackson *et al.* 1982b)

P; *G; Georgalas (1955); *Shebalin *et al.* (1974)

1956 July 9; Amorgos Island (*Ms* 7.2)

A large magnitude earthquake in the Aegean Sea, followed 12 min later by a destructive aftershock that caused

extensive damage to the islands around Amorgos, Fig A35. In all, 530 houses were destroyed and 1500 were damaged beyond repair; 53 people were killed and 120 injured. Much of the destruction was caused by the main aftershock which, it is said, was felt more violently than the main shock.

The earthquake was associated with a large seismic sea-wave that affected the whole of the Aegean Sea. The zero-time wavefront was calculated at a source just off the NE coast of Amorgos Island, very near the instrumental epicentre, Fig. A35. On the south coast of the island wave heights on land reached 30 m, the wave causing some damage to coastal areas up to 100 km away, destroying about 40 small sailing ships and large boats. Fig. A35 shows the wave heights on land (in feet). The sea-wave was probably generated by either a submarine slide in the Amorgos basin, which has a depth of about 500 m, or by faulting of the sea bed. The event was felt over an area rather small for its magnitude and it was followed by aftershocks that continued for almost seven months.
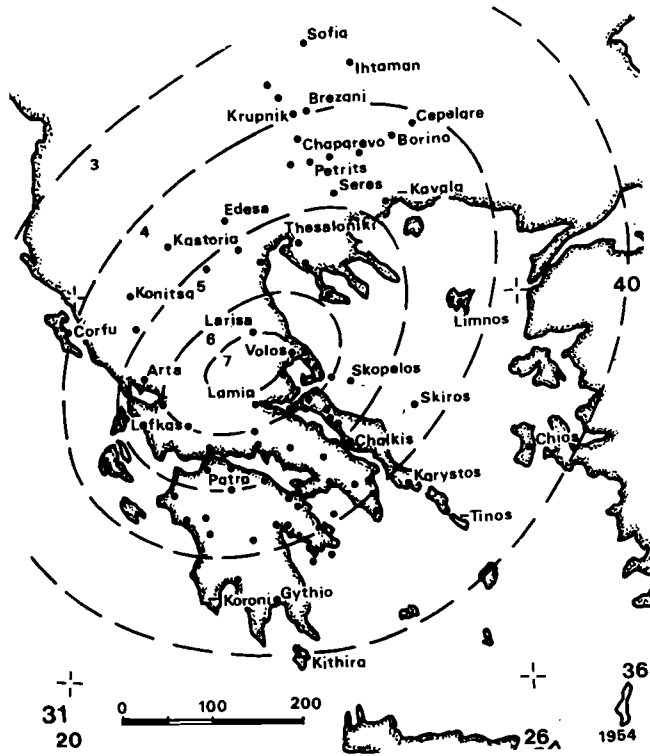
Figure A31. 1954 April 30; Sofades earthquake (M_s 6.7).

The first motion fault plane solution of Shirokova (1972), (060:45: − 090) is consistent with normal faulting on the steep escarpment along the SE coast of Amorgos Island, which bounds the deep Amorgos basin (McKenzie 1978).

P; U; G; Galanopoulos (1956); Ambraseys (1960); McKenzie (1972); *Shebalin *et al.* (1974); Makropoulos & Burton (1981)

1957 March 8; Velestino (*Ms* 6.5, 6.6)

Two shocks of almost the same magnitude 7 min apart caused heavy damage in the region of Velestino between Farsala and Volos, Figs A36 and A37. In this area, already damaged by the shocks of 1954–5, out of 64 000 houses, 3600 collapsed and 6200 were damaged beyond repair. In all three people were killed and 80 injured. Damage statistics define well the epicentral area, Fig. A36, within which rockfalls, landslides and ground failures added to the destruction. A field survey immediately after the event and a number of visits some years later did not produce clear evidence of surface faulting; though a series of breaks about a kilometre long in limestone showing normal faulting north of Aerino and Kokina, and another break, about 500 m long in alluvium between M. Perivolaki and Velestino, the latter not visible in 1966, are attributed by local people to these particular earthquakes. It is said that the railway line between Aerino and Velestino was damaged and repaired
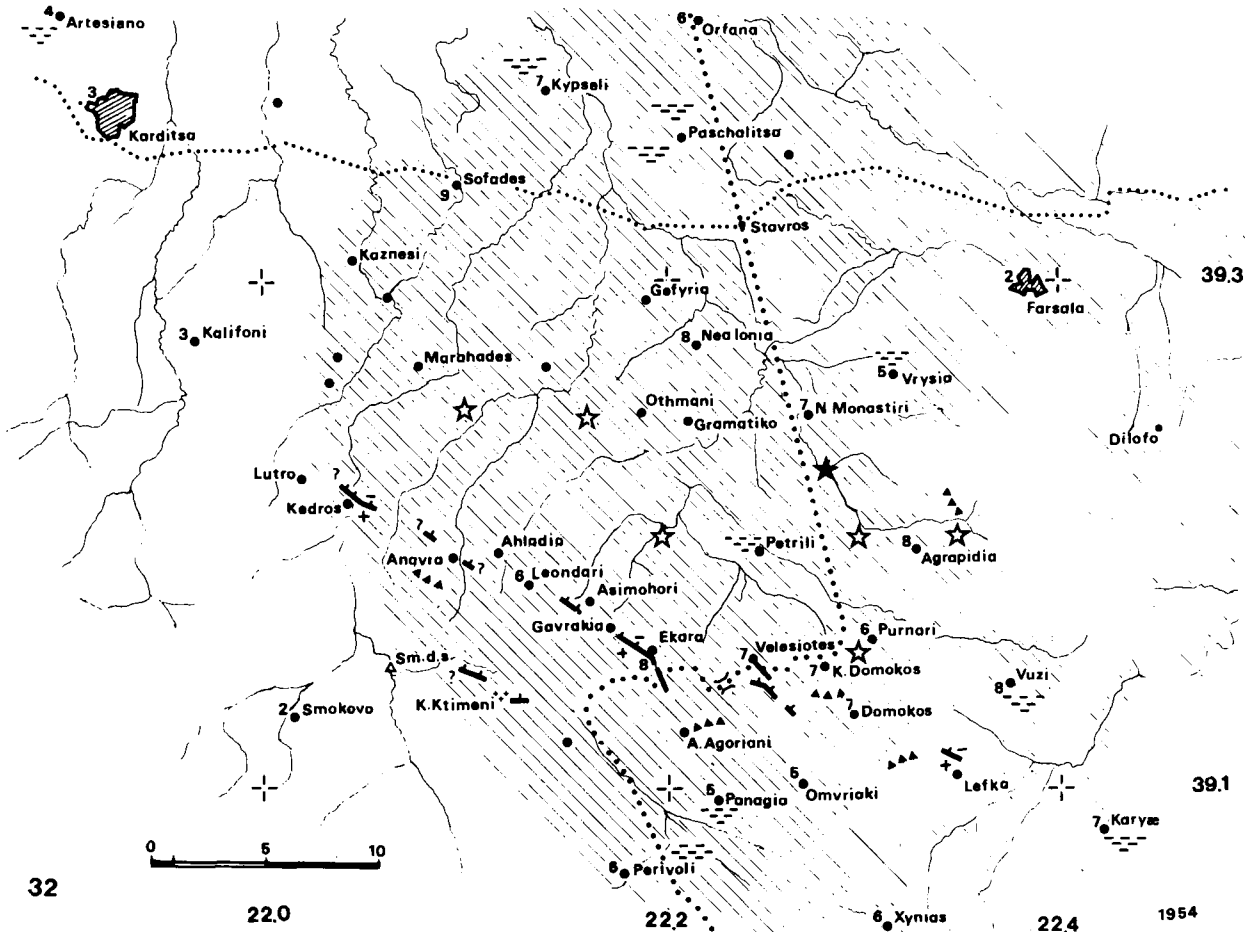
Figure A32. 1954 April 30; Sofades earthquake (M_s 6.7).
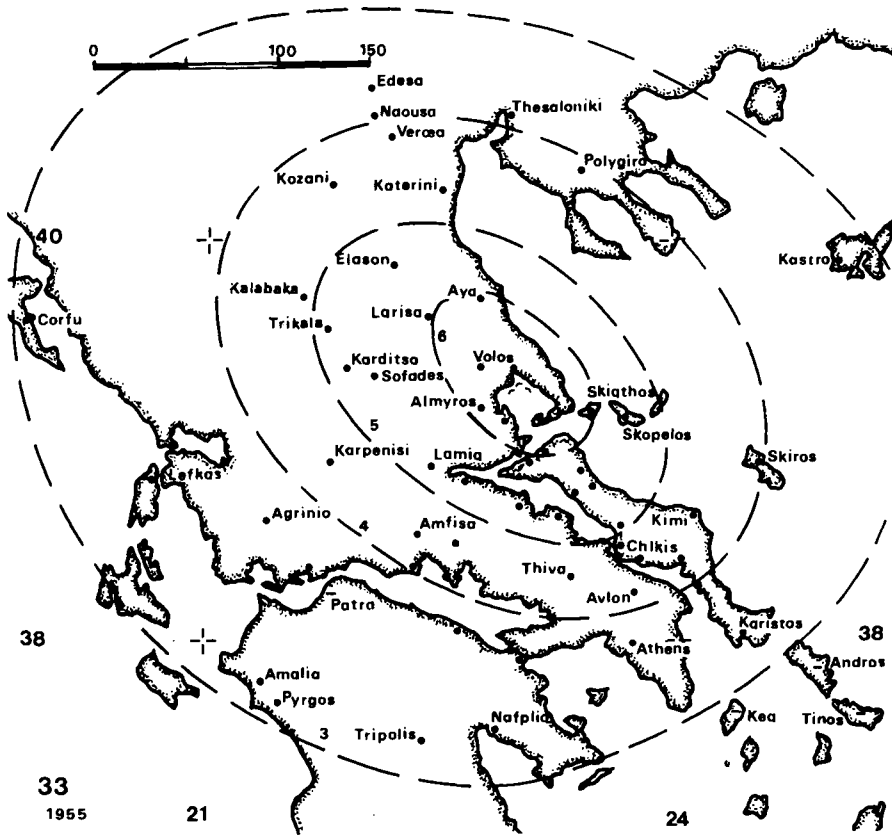
Figure A33. 1955 April 19; Volos earthquake (M_s 6.2).
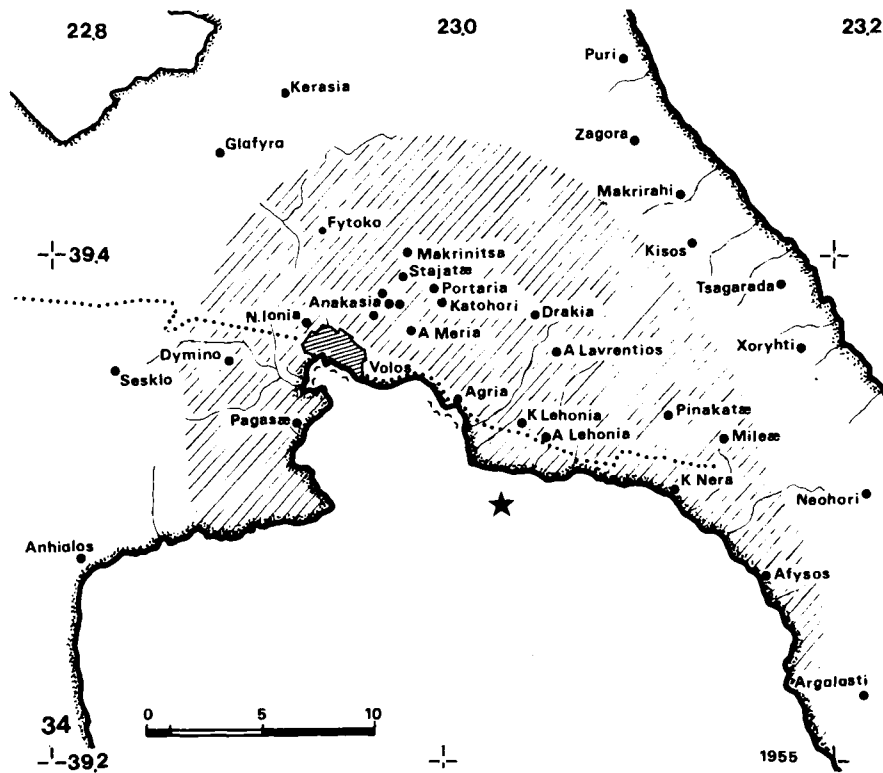
Figure A34. 1955 April 19; Volos earthquake (M_s 6.2).
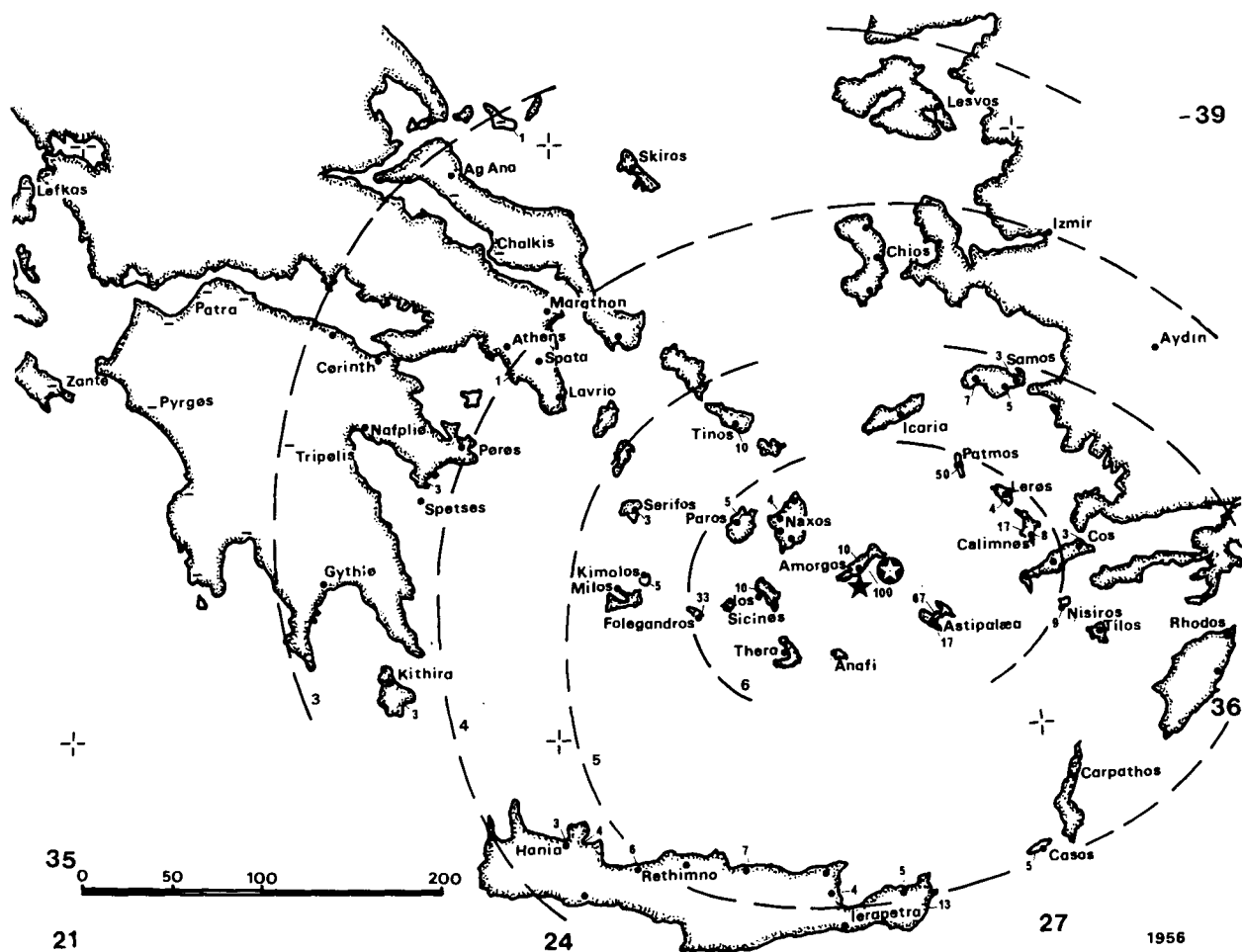
Figure A35. 1956 July 9; Amorgos Island earthquake (M_s 7.2).

after the earthquake, but details are lacking. We have assumed normal faulting (110:45:−090) from the proximity to the 1980 earthquakes. the strike is taken from the trend of prominent topographic escarpments in the epicentral region, and is similar to that of the 1980 sequence.

Damage due to foundation failures was reported from the region west of Lake Karla and in Volos where the landing pier and quay facilities settled excessively and the textile and tobacco factories near Volos were damaged.

Damaging aftershocks continued till late November, affecting mainly the eastern part epicentral region. The main shock was felt as far as south Bulgaria, Macedonia and it was perceptible in Sicily.

P; U; *G; Papastamatiou (1975); *Shebalin et al. (1974); *Papazachos et al. (1982)

1965 March 9; Alonisos Island (Ms 6.5)

An earthquake with an offshore epicentre in the NW Aegean Sea caused considerable damage to the islands of Alonisos and Skopelos. in all about 1940 houses were destroyed or damaged beyond repair but relatively few collapsed completely, killing only two and injuring two people. The shock triggered small landslides but no rockfalls. At Patitiri, on the SW tip of Alonisos and on the

east coast of the islet of Kyra Panagia, the coastline sunk in places by about 30 cm. This was most probably due to slumping of beach deposits. No seismic sea-wave accompanied the earthquake and there was no damage to the submarine cables, Fig. A38. From first motions of McKenzie (1972), the focal mechanism is (040:90:180).

P; U; G; *Shebalin et al. (1974)

1965 March 31; Aetolia (Ms 6.7)

This earthquake caused widespread, but otherwise minor damage in the Peloponnesus and on either side of the Gulf of Corinth. It caused the collapse of only 100 houses and the damage of 900, killing six and injuring 17 people. The shock, which was not preceded or followed by significant seismic activity, was felt throughout Greece and its was perceptible in a narrow, northwest trending zone, along the Adriatic coast of Montenegro and SW Italy. It was reported from isolated places as far northwest as Trieste and Ljubljana, but not from most of the islands of the Aegean or south of Crete, showing an intensity distribution similar to that of 1898, 1899, 1927 and 1947, Fig. A39.

The focal depth of this event calculated by ISC is 45 km. USCGS gives 78 and BCIS 100 km. Also the fact that in spite of the large number of macroseismic observations, Fig.

Figure A36. 1957 March 8; Velestino earthquakes (M_s 6.5 and 6.6).

A39, no well-defined epicentral area can be determined, suggests that this earthquake had a sub-crustal focus. Within isoseismal *VI*, intensities up to *VIII*, reflecting local conditions and contamination from the earthquakes of April 5 and July 6, make it impossible to define a limited epicentral region. From first motions of McKenzie (1972), the focal mechanism is (136:76:080)

P; U; G; Vetoulis (1965); Vetoulis & Eleftheriou (1965); Ambraseys (1967); *Shebalin *et al.* (1974)

1965 April 5: Magalopolis (Ms 5.9)

This earthquake in SW Peloponnesus was not a large magnitude event, Fig. A40. It was preceded or followed by significant shocks and affected the upper reaches of Alfios River, Fig. A41. However, because of its location in a densely populated district with many poorly constructed houses, many of them sited on old landslides, ranks as one of the most damaging for its size. Out of 99 000 houses in

the region, 6970 were destroyed or damaged beyond repair and 6480 were badly damaged. 18 people were killed and 155 were injured. What is unprecedented is that 70 per cent of these losses were due to foundation failures and landslides in shales and flysch deposits, that developed during and after the earthquake. Liquefaction of the ground was observed in a few places in river deposits and caused no damage. Well designed and built houses, bridges and dams suffered no damage. There was no evidence of ground movements due to faulting. From first motions of McKenzie (1972), the fault mechanism is (125:74: − 032).

The shock was felt in Greece but not in the Aegean Islands or northern Greece. Because of the occurrence of this earthquake only 5 days after the intermediate depth earthquake of March 31, the macroseismic effects of the two events are often confused.

P; U; G; Vetoulis (1965); Vetoulis Eleftherioll (1965); Kokinopoulos & Tasios (1966); *Ambraseys (1967); *Shebalin *et al.* (1974); *Papazachos *et al.* (1982)
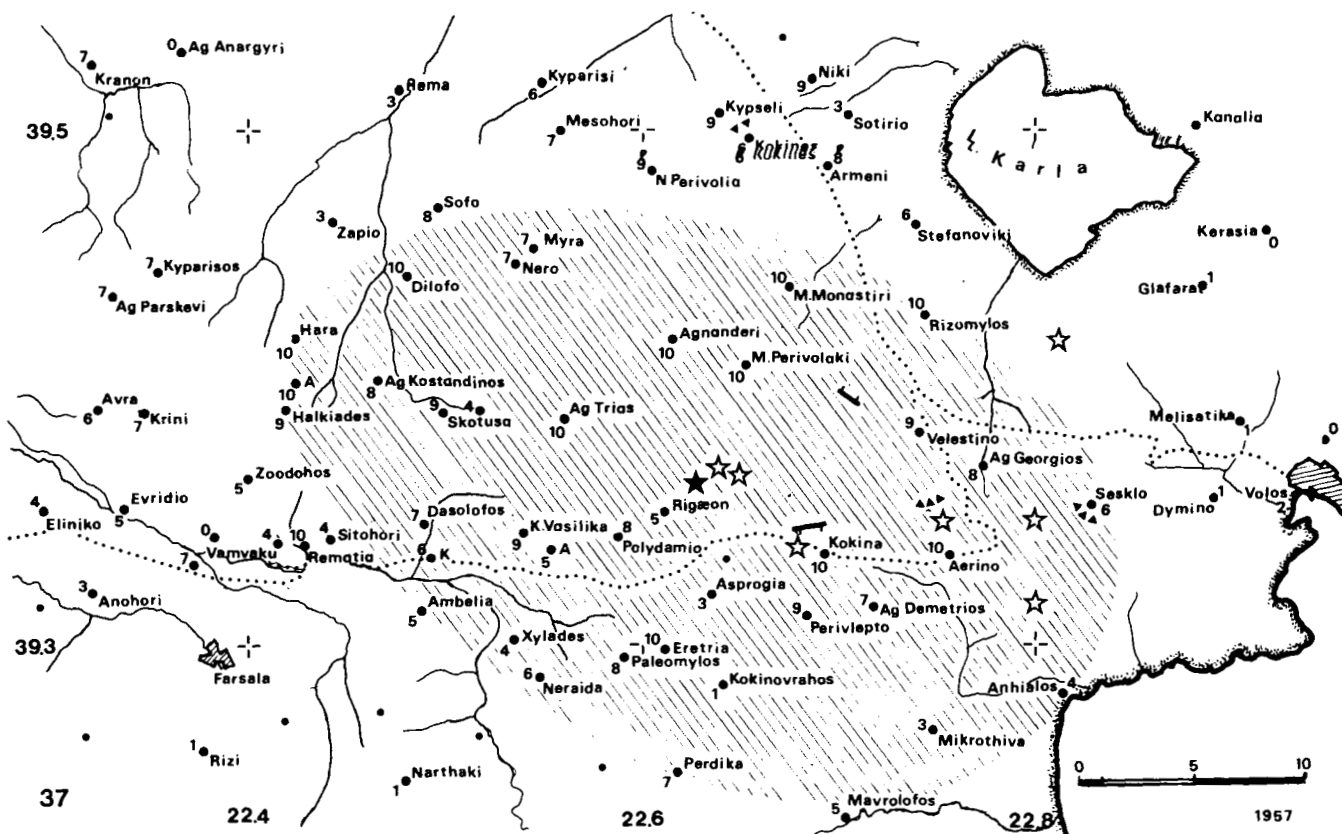
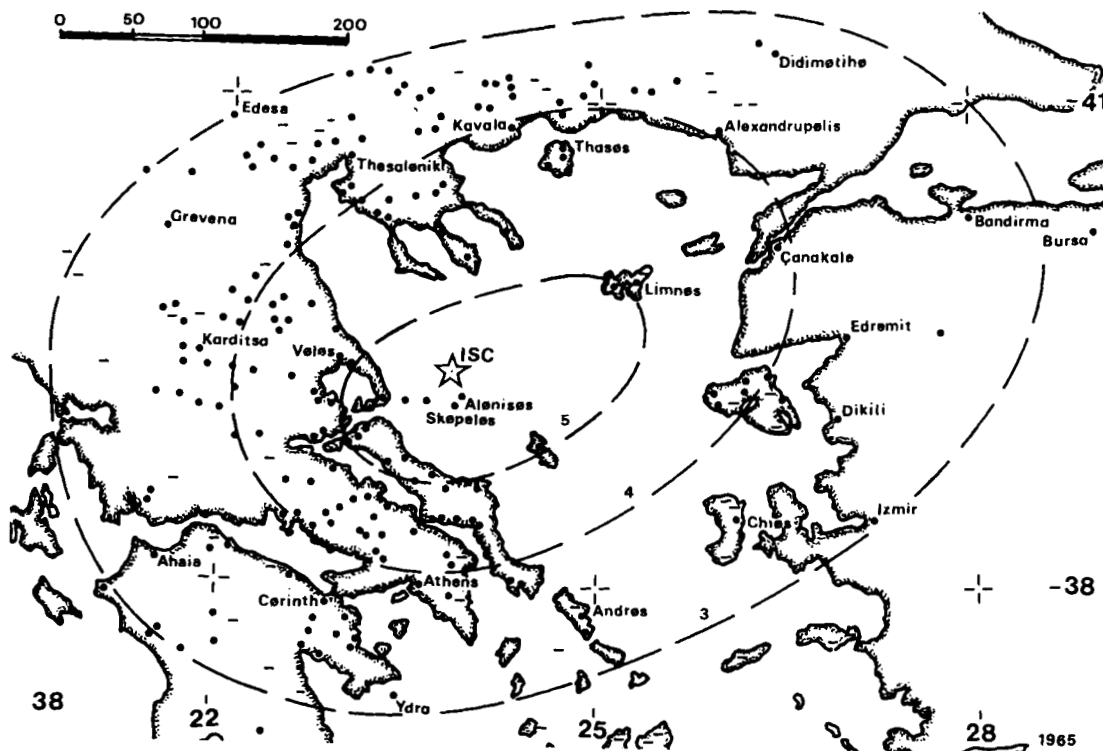**Figure A37.** 1957 March 8; Velestino earthquakes (M_s 6.5 and 6.6).

Figure A38. 1965 March 9; Alonisos Island earthquake (M_s 6.5).

Figure A.39. 1965 March 31; Aetolia earthquake (M_s 6.7).

1965 July 6; Eratini (Ms 6.4)

This earthquake had an offshore epicentre in the Gulf of Corinth. It affected the same area shaken by the intermediate depth earthquake of March 31 and added to the damage, destroying about 300 houses and damaging about 1700 beyond repair on both sides of the Gulf. Damage statistics for this earthquake do not differentiate between the damage caused by the three earthquakes in 1965, but they do confirm that the epicentral area was offshore, Fig. A42.

The shock triggered numerous slides on both sides of the gulf, some of them initiated by the exceptionally wet period that preceded the event and by earlier shocks. A 3 m high sea-wave, caused by a submarine slide off the coast of Eratini, surged on the coast of the Panormos bay drowning one person. In the port of Eratini the height of the wave was 1 m and to the east of Panormos 0.5 m. The wave was not reported from other parts of the coast. There is some evidence that another small wave flooded the coast of Acrata where there has been extensive liquefaction to the northwest of the village. The shock affected the yield of

artesian wells within a radius of 40 km and injured six people. The shock was not followed by an significant sequence of aftershocks.

From first motions of McKenzie (1972), replotted with a crustal velocity at the focus (Jackson & White 1989), the focal mechanism is (090:74: −115). E–W striking normal faulting is in agreement with faults visible on both sides of the Gulf of Corinth.

P; G; U; Ambraseys (1967); Shebalin *et al.* (1974); Papazachos *et al.* (1982)

1966 February 5; Kremasta (Ms 6.2)

This earthquake in Evritania, central Greece, was preceded by few and followed by many small shocks that continued for almost a year. The main aftershock affected a relatively small and mountainous region where out of about 4000 rural houses mostly poorly built, 1130 collapsed killing one and injuring 40 people, Fig. A43. Landslides and rockfalls triggered by the foreshocks, the main shock and aftershocks during a wet period, caused additional damage over an area
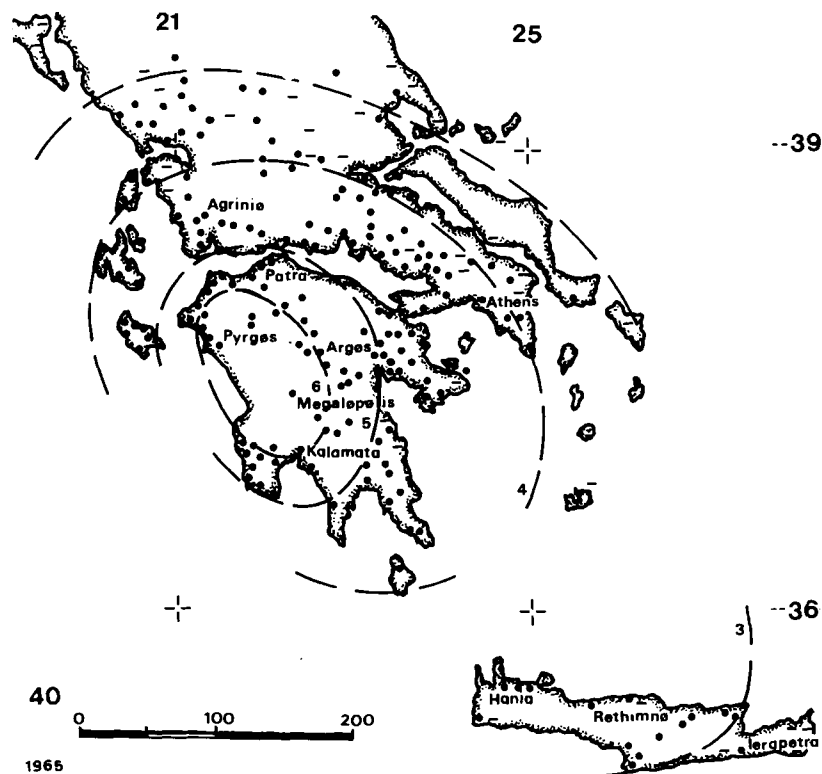
Figure A40. 1965 April 5; Megalopolis earthquake (M_s 5.9).

much larger than the epicentral. Since February 1963, when rainfall exceeded a 70 yr record and caused an unprecedented number of slides in central Greece and northern Peloponesus, even small shocks aggravated the stability of many settlements, i.e. Alestia, Mikros Horio, Vraha and Papadia.

The shock caused no damage to the 125 m high earth dam at Kremasta located about 20 km SE of the epicentre. The impounding of the Kremasta lake began in July 1965 and at the time of the earthquake the reservoir level was 10 m below conservation level. The earthquake is thought to be associated with the filling of the reservoir but there is no conclusive evidence for this.

A cursory survey of the epicentral region six months after the earthquake did not produce any evidence of faulting, either because there was none or because it was in inaccessible ground and was missed. From first motions and some P-waveforms the focal mechanism is (252:66: $-$ 100) (Anderson & Jackson 1987).

A strong aftershock an hour after the main shock was responsible for additional damage in the Kredi–Palaeokatuna area to the west of the main shock and for the collapse of steep banks on the east rim of the Kremasta Lake. A series of ground cracks reported from southeast of Palaeokatuna, running for about 2 km in a direction N–50–E, mainly in limestones, were attributed by local people to these events. A levelling survey of the region of the reservoir area showed no significant movements.

P: U: G: Public Power Corporation (1966); Comninakis *et al.* (1968); Shebalin *et al.* (1974); Papazachos *et al.* (1982); Stein , Wiens & Fujita (1982)

1967 March 4; Skiros (*Ms* 6.8)

An offshore earthquake in the north Aegean was strongly felt in the Islands of Skiros, Ag. Efstratios and Lesvos, Causing no damage. No seismic sea-wave was reported from the region. The earthquake, which was not preceded or followed by significant shocks, was felt over a large area, from the Peloponesus in the south to Sofia and Edirne in the north and from Epirus to Canakale.

From first notions of McKenzie (1972), replotted with a crustal velocity at the focus, the focus mechanism is (098:60: $-$115), Jackson & White (1989). Normal faulting is consistent with motion on the large escarpment bounding the deep (1000 m) basin NE of Skiros.

P; G; *Shebalin *et al.* (1974)

1967 May 1; Tzumerka (*Ms* 6.2)

The earthquake happened in the mountainous region of Tzumerka between the rivers of Arachthos and Acheloos, Fig. A44. Maximum effects were reported from the west side of the Acheloos river between Gardiki, Neraida, Terpnas and Mesohora where a dozen mountain villages were totally destroyed. The shock reactivated old slides in this area along the Acheloos river as well as tens of kilometres away, notably at Megali Goritsa, 30 km north of the epicentral area, as well as west of Tzumerka mountains along the mountains where 1140 houses collapsed killing 15 people. In all, 3800 houses were destroyed or damaged beyond repair, mainly old constructions and houses built on steep slopes and slide areas.

Figure A41. 1965 April 5; Megalopolis earthquake (M_s 5.9).

During the five days following the earthquake numerous small aftershocks added to the damage, mainly in the SE part of the epicentral region.

We could find no evidence of faulting. The focal mechanism (200:56:−040) is well constrained by first motions and *P*-waveforms. (Anderson & Jackson 1987)

P: U: G; Shebalin *et al.* (1974); Papazachos & Comninakis (1982)

1970 April 8; Antikyra (*Ms* 6.2)

This earthquake happened on the east part of the Gulf of Corinth, and it was strongly felt on both the north and south coast of the Gulf, Fig. A45. No house collapsed and in all about 600 houses were damaged, most of them in the densely settled south coast, where two people were injured. The railroad reported changes in roadbed elevation in some areas between Lygia and Lautro due to settlements. Small fissures opened in the ground near Antikyra and rockslides raising clouds of dust were noticed from the arid mountains about 10 km southwest of the village. The distribution of damage suggests that the epicentral region was offshore. From first motions of McKenzie (1972), replotted with a crustal source velocity, the focal mechanism is (090:74: −115). Normal faulting on E–W planes is consistent with surface faulting observed both sides of the Gulf of Corinth.

A strong aftershock caused additional damage, mainly to the north part of the Gulf.

P: G: *Shebalin *et al.* (1974); *Papazachos & Comninakis (1982)
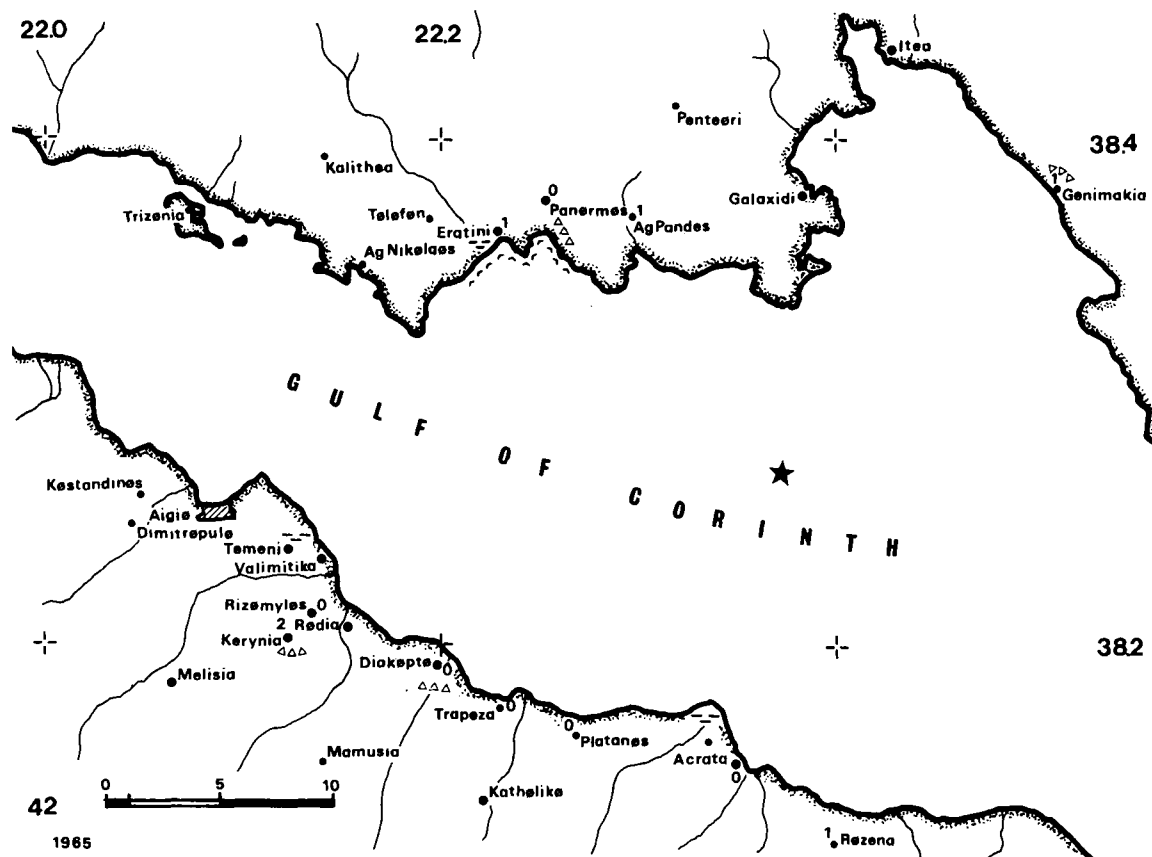
Figure A42. 1965 July 6; Eratini earthquake (M_s 6.4).

1980 July 9; Almyros (*Ms* 6.4, 6.0)

Between the 4th and 9th of July the region souhtwest of Volos in central Greece, was subjected to a series of strong shocks. The main earthquake on July 9 was preceded 1.5 min earlier by a strong foreshock and followed 2.5 min later by an aftershock of almost the same magnitude. These events affected the alluvial plain of Almyros (Fig. A46) where they caused extensive damage. Out of 14 500 houses, 31 per cent were damaged beyond repair or collapsed, mainly old constructions affected by previous earthquakes, in which 23 people were unjured. Much of the damage concentrated in the three larger towns in the region, of Anhialos, Almyros and Surpi where 31, 38 and 7 per cent of the houses were destroyed respectively. Ground motions are described as violent; however, free standing columns in the historical site of old Anhialos were left standing while nearby rural houses were totally ruined. Damage outside the epicentral area was sporadic and in places serious, particularly in villages with ageing building stock.

The earthquake triggered liquefaction of the ground in the region of Anhialos, particularly along the coast that caused some damage to houses and harbourworks. Liquefaction was also observed in the plain of Mikrothiva where deep wells developed abnormally high artesian pressures that for many hours after the earthquakes caused ejection of water. In the plains of Almyros and Anhialos water wells overflowed and excessive differential settlements

caused some damage to the water supply system, to fly-overs of the motorway and to port facilities in Volos.

Surface faulting was reported for these earthquakes from three localities. A 2.5 km long rupture in alluvium, striking N-100-E, with the south side downthrown by an average 20 cm, was found passing through Anhialos. It extends to the west where it joins a pre-existing fault on which there was no evidence of surface movement. A second rupture, about 1.5 km long, striking N-90-E cuts the secondary road from Mikrothiva to Velestino just south of the site of Dexameni, about 200 m north of the village. Here there is no evidence of vertical displacement but only open cracks. However, the formation of small grabens and the distortion of tree roots that straddle the break indicate right-lateral motion of a few centimetres. A water conduit that straddles the break was also deformed at the time of the earthquake but not broken. Two and a half months later the conduit was ruptured, flooding the ground depressions along part of the break, the two broken parts showing a left-lateral offset of more than 10 cm. A third surface rupture in this alluvium, about 1.3 km long, striking E-W, was observed about 1 km to the west of the road that leads north of Mikrothiva. Here the south side is downthrown by about 20 cm, in places the top soil detaching itself from outcropping limestones, which however do not show signs of faulting. Based on surface faulting and first motion the focal mechanism is (270:50: -90).

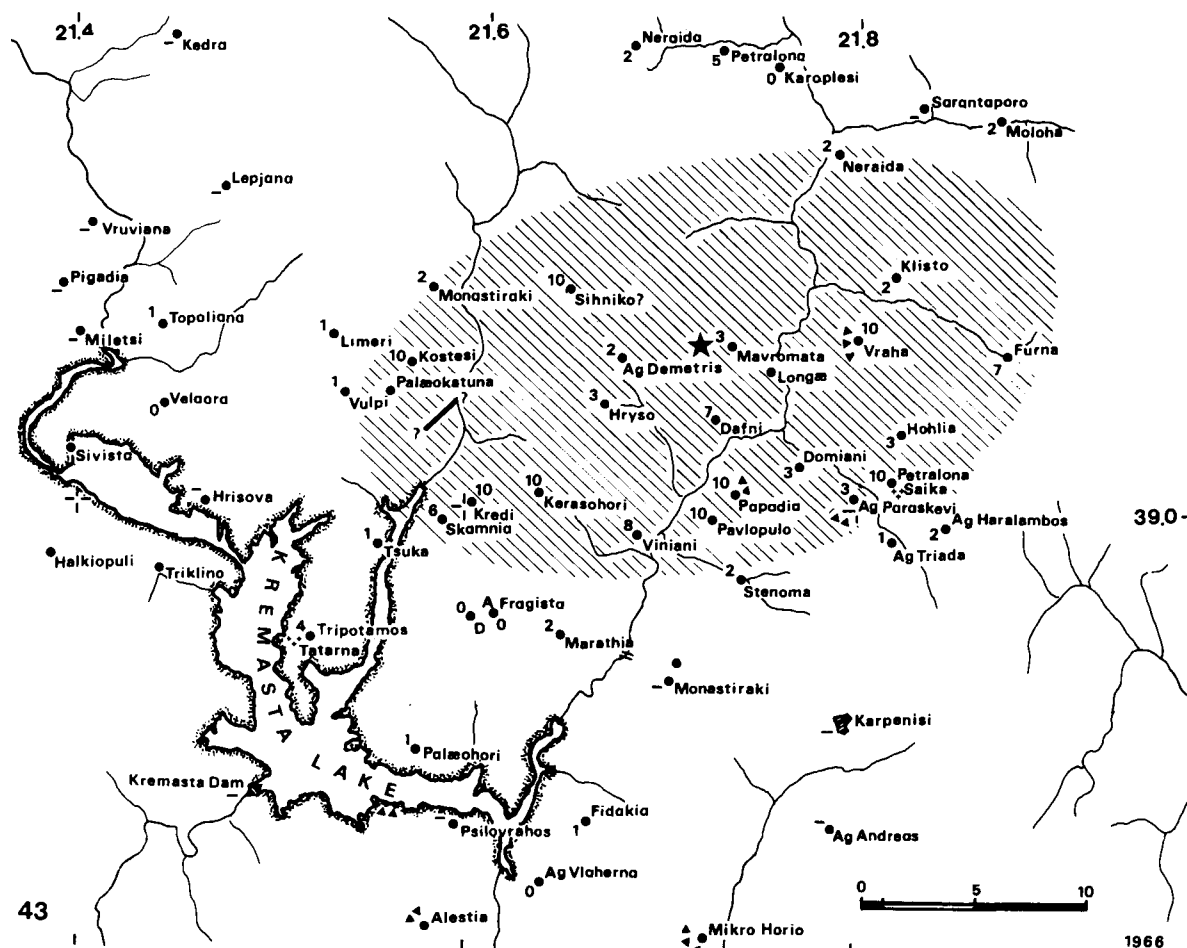The shocks did not set up a seismic sea-wave in the Gulf

Figure A43. 1966 February 5; Kremasta earthquake (M_s 6.2).

of Volos and they were not followed by a long aftershock sequence.

P: G: Papazachos *et al.* (1982); Papazachos *et al.* (1983); Tsatsanifos (1987)

1981 February 24; Alkionides (*Ms* 6.7, 6.4, 6.2)

Without a foreshock the first earthquake affected the easternmost part of the Gulf of Corinth, and was followed five hours later by an equally strong shock, both with offshore epicentres, Fig. A48. These shocks caused extensive damage within a radius of about 25 km, killing 14 and injuring about 500 people. Because of their temporal closeness it is not possible to separate the individual effects of these two events. Sporadic damage was observed up to 100 km from the epicentral region. 70 km to the east, Athens (population 3 000 000) suffered unprecedented damage: about 500 houses collapsed and more than 5000 people were made homeless. A large number of exhibits in archaeological museums were broken and the Parthenon sustained minor damage. In all, about 11 000 houses, 100 hotels and a number of factories were destroyed or damaged beyond repair. The total losses, including the effects of the aftershock of March 4, amounted to £350 million.

The first shock of the February 24 appears to have caused widespread damage throughout the epicentral region Fig. A48). It was felt over a large area (Fig. A47), and it was accompanied by luminous phenomena seen over the Gerania mountains. The second shock early on February 25 caused more serious damage, particularly in the region of the Gerania mountains, but it was not felt as far as the first shock. These earthquakes were associated with normal surface faulting that extended for about 15 km along the north-facing slopes of the Gerania mountains, north-dipping with throws of about 70 cm on average. About 10 m offshore Skinos remains of Roman baths are found at a depth of 1.3 m below sea-level, confirming long-term subsidence of the coast in this area. It is not possible, however, to say how much of the faulting was attributable to each shock. The coast immediately to the north of the break subsided by about 50 cm and there is evidence that the first shock was associated with a small seismic sea-wave reported from Strava, which, however, was not recorded by the tide gauge at Posidonia. There is some evidence that this flooding of the coast may be associated with submarine slumping triggered by the shock in the bay of Porto Germano or west of the islet of Daskalio. It is of interest that the few houses and abandoned monastery on the islets of Daskalio and Zoodoho Pigi both situated in the epicentral area of the
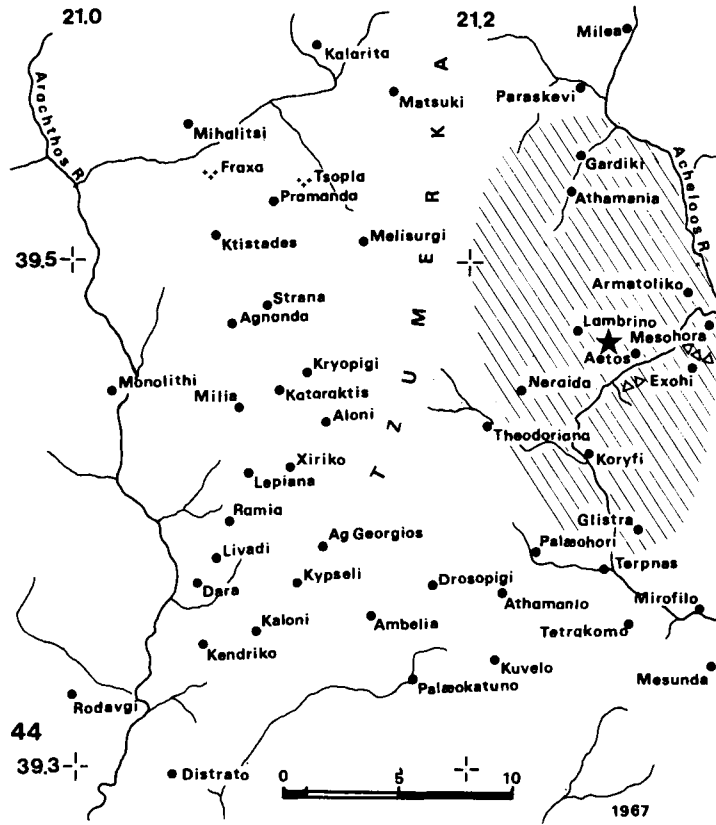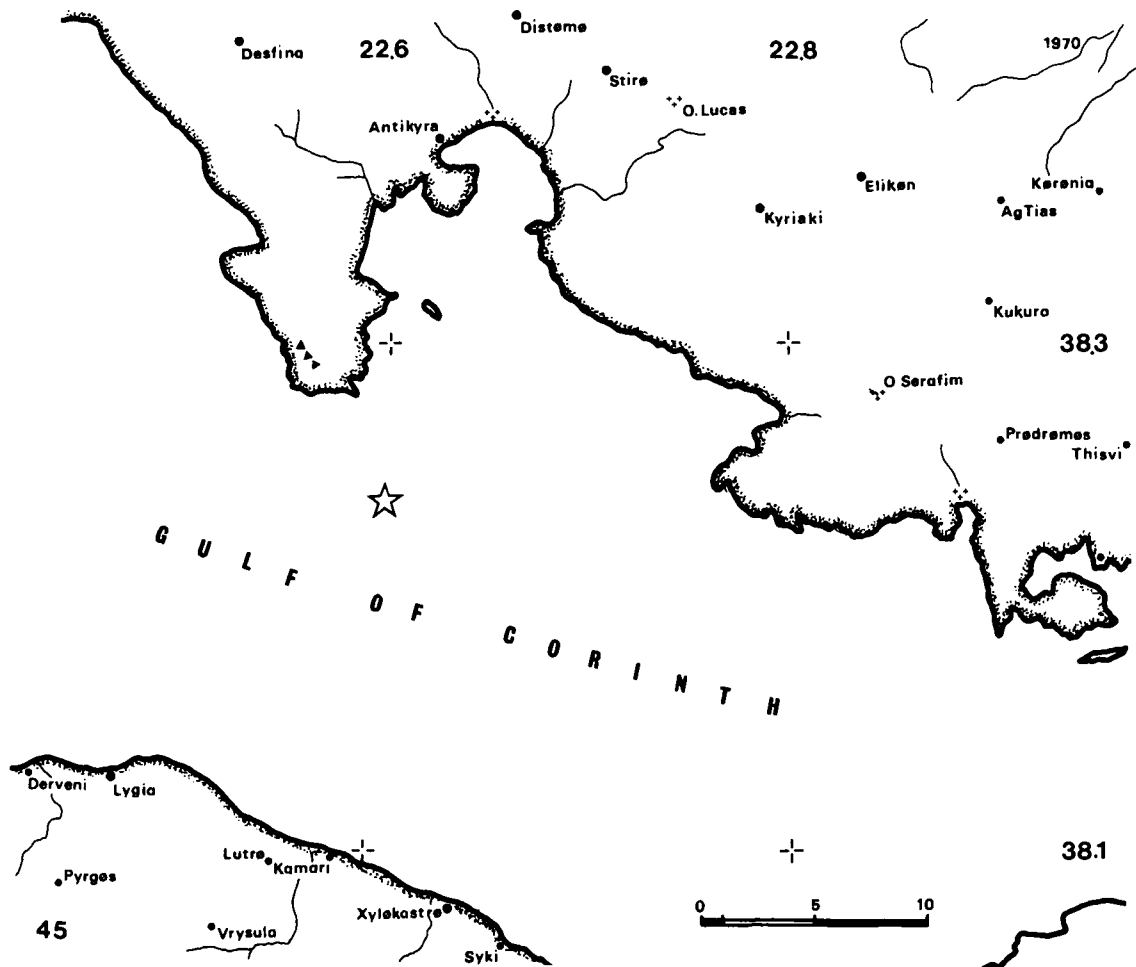
Figure A44. 1967 May 1; Tzumerka earthquake (M_s 6.2).
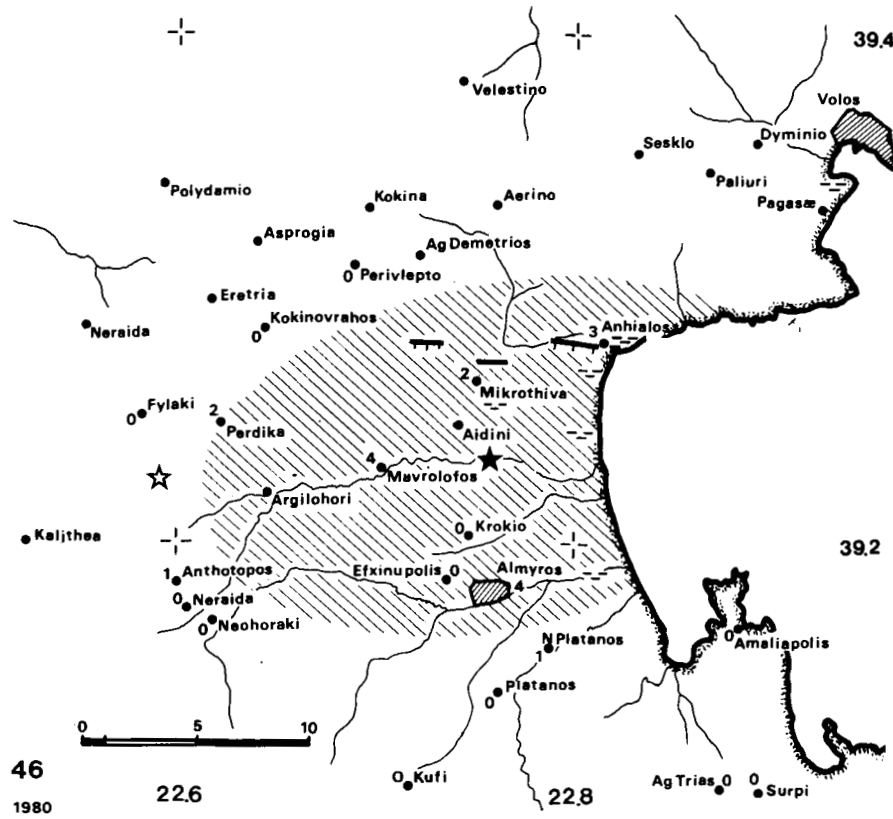

Figure A45. 1970 April 8; Antikyra earthquake (M_s 6.2).

Figure A46. 1980 July 9; Almyros earthquakes (M_s 6.4 and 6.0).

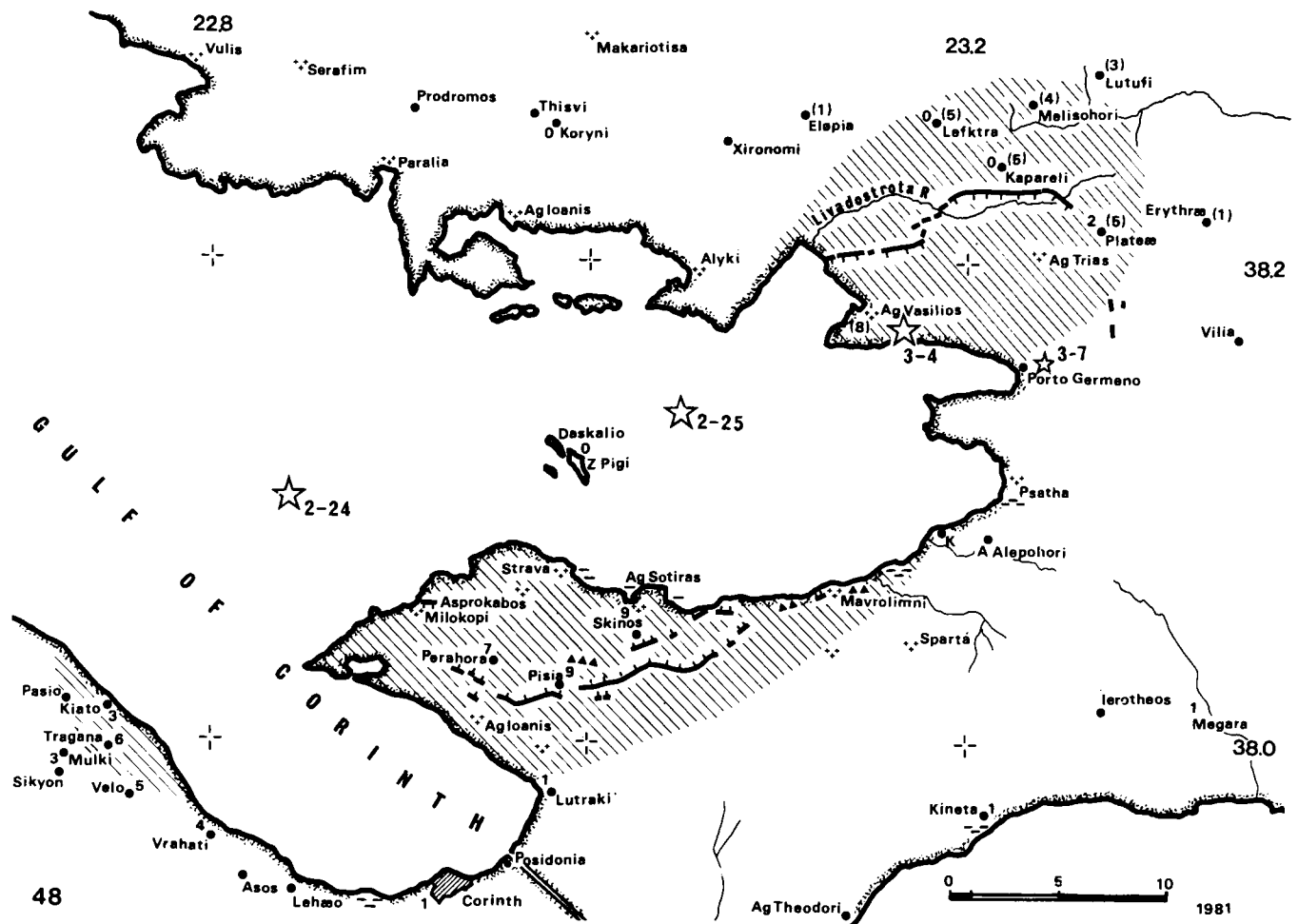**Figure A47.** 1981 February 24; Alkionides sequence (M_s 6.7, 6.4 and 6.2).

Figure A48. 1981 February 24; Alkionides sequence (M_s 6.7, and 6.4 and 6.2).

main aftershock, suffered some damage but they were not destroyed. The earthquakes triggered numerous landslides and rockfalls from steep limestone cliffs and caused the liquefaction of fluvial and beach deposits.

A third shock on March 4, with an epicentral region on the north side of the Gulf of Corinth in the region of Kapareli, enlarged the epicentral area and added to the destruction. A number of villages quite some distance from the epicentre, already damaged by the previous shocks were ruined. No one was killed, but 12 persons were injured. This shock was associated with normal faulting with the south block downthrown that trends E–W, running discontinuously for 12 km from the coast north of Ag. Vasilios to near Platae, and was associated with a sinking of the coastline south of the break. Average throws were about

60 cm. The shock triggered liquefaction in the valley floor of Livadostrata and elsewhere Fig. A48.

From first motions, surface faulting and waveform analyses, the focal mechanism of these earthquakes were (285:40:−070) for February 24, (250:42:−080) for February 25, and (068:47:−082) for March 4.

The aftershock sequence was not long but it did include at least two damaging events in the region of Kapareli.

P: U: Berz (1981); Gurpinar (1981); Khoury (1981); Jackson *et al.* (1982a); Khoury *et al.* (1983; 1984); Papazachos *et al.* (1983); Perissoratis, Mitropoulos & Angelopoulos (1984); Snider *et al.* (1984); King *et al.* (1985); Vita-Finzi & King (1985)