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# Magnitude versus faults' surface parameters: quantitative relationships from the Aegean Region

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**Abstract**

Historical and seismotectonic data from the broader Aegean Region have been collected and all possible information relative to ground deformation associated to earthquakes that hit the area have been re-evaluated. All events associated to co-seismic surface faulting have been selected and further investigated, while geomorphologic and geological criteria have been used to recognise and characterise the seismogenic faults associated to these 'morphogenic earthquakes' (sensu [Bull. INQUA 16 (1993) 24]). In particular, in order to perform seismic hazard analyses, we compiled a list of all earthquakes where the *surface rupture length* (SRL), the *maximum vertical displacement* (MVD) or the *average displacement* (AD) is available. We thus obtained reliable values of these source parameters for 36 earthquakes, of which 26 occurred during the 20th century, 6 in the 19th century and the 3 remaining earlier. Magnitude versus SRL and MVD has been compiled for estimating empirical relationships. The calculated regression equations are: $Ms = 0.90 \cdot \log(SRL) + 5.48$ and $Ms = 0.59 \cdot \log(MVD) + 6.75$, showing good correlation coefficients equal to 0.84 and 0.82, respectively. Co-seismic fault rupture lengths and especially maximum displacements in the Aegean Region have systematically lower values than the same parameters worldwide, but are similar to those of the Eastern Mediterranean–Middle East region. The envelopes of our diagrams are also calculated and discussed for estimating the worst-case scenario. Furthermore, for all investigated seismogenic structures, based on several geological criteria, we measured the 'geological' fault length (GFL), which is the total length of the neotectonic faults showing cumulative recent activity. We then compared SRL with GFL and their ratio shows a clear bimodal distribution with a major peak at 0.8–1.0, indicating that about 50% of the investigated earthquakes ruptured almost the entire fault length, while a second peak around the value of 0.5 is clearly related to a segmentation process of longer neotectonic structures. Further implications of this distribution are also discussed. Eventually, from the distribution of GFL versus magnitude we also infer an important geological threshold for the occurrence of 'morphogenic earthquakes' at about 5.5 degrees.


*Keywords:* Seismic hazard; Morphogenic earthquakes; Seismotectonics; Surface rupture

## 1. Introduction

The Aegean Region is characterised by intense seismic activity due to the rapidly deforming broader Eastern Mediterranean area and its complex neotectonic structure. Earthquakes occur nearly everywhere.

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It is generally accepted that the tectonic regime affecting the Aegean is broadly extensional (e.g. McKenzie, 1978; Mercier et al., 1989; Papazachos, 1990), and dominated by normal faulting. Exception is the North Anatolian Fault–North Aegean Trough fault system and few secondary branches, which are characterised by transtensional or even transcurrent tectonics (Pavlides et al., 1990; Pavlides and Caputo, 1994).

Although the Aegean Region has one of the longest and densest records of historical seismicity in the world (e.g. Galanopoulos, 1981; Guidoboni, 1989; Guidoboni et al., 1994; Papazachos and Papazachou, 1997; Pavlides, 1996a), available knowledge about co-seismic surface ruptures in the region is fairly limited (Ambraseys and Jackson, 1990; Pavlides et al., 2000; Goldsworthy et al., 2002). From the 550 BC Sparta earthquake (?), which is possibly the oldest known seismic event in Greece, there are good descriptions for more than 500 moderate and strong earthquakes. Nevertheless, the historical database is generally incomplete, especially for some specific areas, such as the Cyclades Islands (Central–Southern Aegean), Thrace and Western Macedonia (Northern Greece) or for historical periods of socio-economical decline. The area investigated in the present note is the broader Aegean Region represented by Greece and the territories of the neighbouring countries.

The best information for surface ruptures associated to morphogenic earthquakes (sensu Caputo, 1993) mainly comes from earthquakes occurred on land during the 20th century. According to Caputo (1993), a 'morphogenic' earthquake is an event pro-

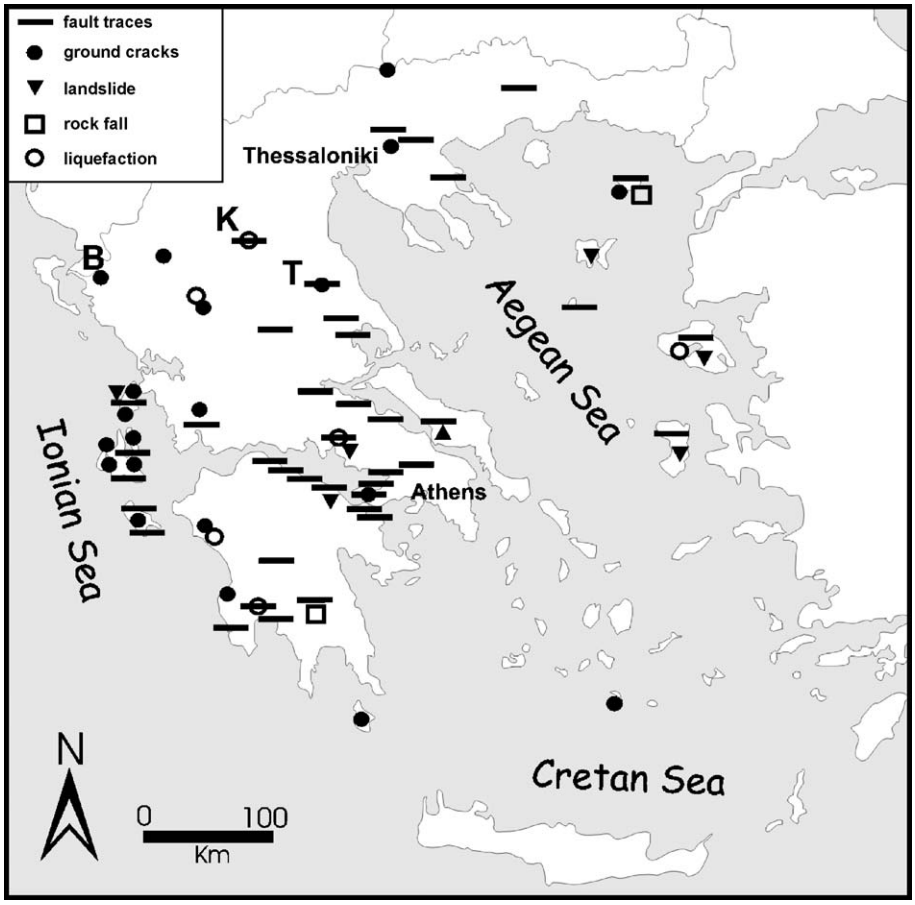


Fig. 1. Map of Greece showing all surface phenomena (fault traces, ground cracks, liquefaction and landslides), not necessarily of tectonic origin, associated to historical earthquakes. Data from Table 1. K = Kozani town; T = Tyrnavos Basin; B = Butrint site.

ducing co-seismic permanent surface deformation directly associated to the seismogenic fault. In contrast, our knowledge of the offshore active faults of the Aegean Region is extremely limited because many epicentres are located under the sea.

Instrumental data cover less than 100 years and the historical information can be considered rather complete from the 19th century onwards. Where earthquake repeat times are less than 100 or 200 years such information might be adequate. However, palaeoseismological researches in Greece have shown that the average recurrence interval of specific active faults is commonly longer than 500 years and usually some thousands years (Pavlides, 1996b; Pantosti et al., 1996; Chatzipetros et al., 1998; Caputo et al., 1998). But the lack of contemporary or historical seismic activity can indicate 'hazard' rather than safety. A typical example is represented by the Kozani-Grevena region, Northern Greece (Fig. 1), which was considered a relatively stable area of low seismicity because of a lack of moderate-large earthquakes till before the event which shocked the area on May 13, 1995 with a magnitude of 6.6 (e.g. Pavlides et al., 1995; Meyer et al., 1996; Pavlides and King, 1998). Quiescence of known active faults probably indicates a seismic risk because strain is accumulating.

The error in appreciating the seismic hazard of the Kozani–Grevena area prior to the 1995 event is certainly not unique. Indeed, destructive and 'unexpected' earthquakes have occurred in the last 20 years in many areas of low seismicity around the word. In most cases, the fault or the fault system responsible for these events was unidentified and the potential danger underestimated. Based on the experience gained with these events, the geological structures affecting these areas and responsible of the 'unexpected' seismicity, in most cases could have been easily identified and recognised as active structures if specific investigations would have been performed earlier. In fact, the occurrence and the evidence of several morphotectonic features documenting a pre-historical activity are common in these areas.

Following this line of thinking, Caputo (1995) suggests the occurrence of a seismic gap in Northern Thessaly, Central Greece, mainly based on geological, stratigraphic, morphotectonic and structural investigations of the major faults bordering the Middle-Late Quaternary Tyrnavos Basin (Fig. 1) showing evidences of recent tectonic activity though without any significant historical record. Indeed, recent palaeoseismological, geophysical (high resolution seismic reflection, ground penetrating radar and geoelectrical tomographies) and archaeological researches confirm the occurrence of Holocene reactivations of the major faults bordering the basin (Caputo et al., 1998; Caputo and Helly, 2000; Oliveto et al., 2001; Caputo et al., 2003).

## 2. The geological contribution to seismic hazard assessment analysis

It is important to note that it is difficult to assign a given earthquake to a particular fault even with modern data. For historical events, morphotectonic and other geological data are important, but mainly clear literature evidence and especially palaeoseismological proof are crucial. There are plenty of examples worldwide and within the Aegean Region. In particular, for moderate events it is often not clear if the propagation directly reached the surface, thus generating co-seismic ruptures (i.e. morphogenic earthquakes; Caputo, 1993). Examples are the 1995 Aigion (Ms = 6.2; Bernard et al., 1997; Koukouvelas et al., 2001) and the 1999 Athens earthquakes (Ms = 5.9; Papadopoulos et al., 2002; Pavlides et al., 2002). But in some cases even strong earthquakes like the 1995 Kozani–Grevena (Ms = 6.6) earthquakes are debated in the Scientific Community. For example, Pavlides et al. (1995) and Mountrakis et al. (1998) accepted a 27-km-long co-seismic surface fault and a maximum displacement of 20 cm. In contrast, Meyer et al. (1996) propose a 12-km-long fault trace and a maximum displacement of 4 cm, while Hatzfeld et al. (1997a) assume as the seismogenic structure another geological fault lying 10 km Southern of the observed ground ruptures, which are conversely interpreted as gravitational features.

Another example of contentious earthquake is the Kresna–Krupnic (Bulgaria) seismic crisis attributed to a single event ($M = 7.0$ or 7.8) or two separate events (e.g. Papadopoulos et al., 1998). Co-seismic rupture has been assumed 60 km long by Papazachos and Papazachou (1997), while typical fault traces probably did not exceed 15–17 km or at most 30 km (Grigorova and Palieva, 1968; Dobrev and Avramova-Tacheva, 2000; Ambraseys, 2001; Meyer et al., 2002).

Similar is the case of the 1894 Atalanti (Central Greece) double event, where a 50-km-long continuous surface fault is suggested (Richter, 1958; Papazachos and Papazachou, 1997). In contrast, recent studies and a reappraisal of Skufos' (1894) report have shown that the co-seismic traces were not continuous and were shorter than 30 km (Pavlides et al., 2000; Pantosti et al., 2001).

On the other hand, even well studied co-seismic faults, as for example the 1968 Agios Efstratios ($M = 7.0$; Northern Aegean; Pavlides and Tranos, 1991) cannot be used in our study, because the fault trace (3 km) is confined within a small island. The observed fault traces are too short for a Ms = 7.0 surface earthquake, while it certainly extends much longer on the sea bottom.

In addition, literature data are commonly fragmented, poor and not sufficient to allow a clear correlation between ground deformation and a specific fault. The example of the Helike and Aegion faults (western Gulf of Corinth) is typical. Indeed, up to 13 historical earthquakes are possibly associated with these faults or with offshore ones. Most probably the 373 BC and the 1402 AD events reactivated these structures, but neither palaeoseismological researches (Koukouvelas et al., 2001; Pavlides et al., 2001b), nor the archaeological information (Soter and Katsonopoulou, 1999) have shown any clear correlation with those historical events. Instead, two unknown earthquakes of the last 2000 years are possible according to recent palaeoseismological data (Pantosti et al., 2002; Pavlides et al., in press)

In conclusion, from a long list and among many papers and catalogues describing ground co-seismic deformation, only few pre-instrumental earthquakes can be directly correlated to a particular fault and even in fewer cases information about the surface rupture length, the maximum displacement and the average displacement is reliable. A fundamental problem in all statistical analyses is the accuracy of the data. Indeed, the seismic ground deformation associated with faults is usually investigated and described by different authors who are not necessarily scientists, especially for older historical events. Therefore, we based on the evidence of the last two decades and we re-examined most of the pre-existing information for co-seismic fault traces. Following the re-appraisal of the available data, we compiled a second list of events for which the surface rupture length and/or the maximum displacement are available (Table 2).

The principal aim of the present research is to perform some quantitative analyses on the revised data and give numerical results useful for Seismic Hazard Assessment. Most publications concerning the Seismic Hazard Assessment, either in the Mediterranean region or worldwide, deal with increasingly more refined mathematical techniques yielded on the statement concerning the likelihood of some descriptive earthquake parameters occurring within some future time span, like 10, 50 years or so. They were based mainly on seismological data.

The search for geological evidence of past earthquakes is focused on the identification of fault source parameters, such as neotectonic fault length, seismic rupture length, displacement, segmentation and palaeoseismic history. All these parameters are compiled and used to develop empirical relationships versus earthquake magnitude. The nature of ground deformation caused by earthquakes is quite complex but for Seismic Hazard it is necessary to attempt a statistical estimate of the magnitude of future earthquakes. Indeed, the surface ruptures associated with historical or instrumentally recorded earthquakes are crucial in studies for Seismic Hazard. As a first approach for Seismic Hazard Assessment in Greece, the worldwide magnitude to fault displacement relationships of Wells and Coppersmith (1994) have been commonly used, though other equations have been also proposed (Bonilla et al., 1984; Kiratzi et al., 1985; Strom and Nikonov, 1997; Ambraseys and Jackson, 1998; Pavlides et al., 2000).

The quality of the geological data in the present note is referred to well-mapped, identified and segmented neotectonic faults (active or capable, possible or potentially active) and their palaeoseismic history. The revision is based on our own experience of the last 20 years or so, from the Greek territory, Western Anatolia (Turkey), SW Bulgaria and Albania, as well as from the literature. Although in the broader Aegean Region there is information of ground deformation for more than 100 historic events, in our opinion only few of the historical earthquakes can be directly associated with confidence to specific and well-identified faults. Papazachos et al. (2001) have attempted to correlate 567 mainly historic earthquakes (5th century BC–2000) with 159 faults in the same region. We believe

that their attempt is in some case not documented, meaningless or even risky.

## 3. Ground deformation induced by earthquakes in Greece

In the present note, all deformational phenomena affecting the surface and triggered by historical earthquakes have been revised. A list of about 60 events has been compiled (Table 1) specifying when surface faulting, ground cracking, liquefaction, land slides and any other topographic change of the ground level has occurred associated to moderate and strong historical and instrumental earthquakes. In this first list are reported only Greek events (Fig. 1).

Table 1 is a preliminary catalogue of earthquakes associated to ground deformation mainly based on Guidoboni et al. (1994), Ambraseys and Finkel (1995), Ambraseys and White (1997), Ambraseys and Jackson (1990, 1998), Papazachos and Papazachou (1997), Goldsworthy et al. (2002). Several other published researches (see specific references for each event in Table 1) as well as unpublished data (oral information and our original field observations) have been used. We are well aware that the database is certainly incomplete and in some case it could be controversial. For this reason, some crucial problems and related controversies on selected earthquakes are discussed in more detail in Appendix A.

The first example, circa 550 BC Sparta (Southern Peloponnese) earthquake is a strong destructive event of magnitude 7.0 according to Papazachos and Papazachou (1997), but Ambraseys (1997) suggests that this event is a historic doublet of the 464 BC earthquake. As a result of the earthquake, the ground opened in the Sparta area (Laconia district, Peloponnese) and ridges from Mount Taygetus were turned off. The earthquake is possibly associated to the Sparta–Taygetus active normal fault, NNW–SSE trending and NE dipping (Armijo et al., 1991; Guidoboni et al., 1994; Ambraseys, 1997).

Similarly, the 373 BC Helice earthquake (Northern Peloponnese Corinth Gulf) is possibly associated with the homonym fault, but up to now there is no clear evidence. The area is controlled by at least four active faults namely the Mammousia, the Helice, the Aegion and an offshore unnamed structure (Koukouvelas and Doutsos, 1996; Koukouvelas, 1998). These faults are associated with historical events. For example, the Eastern Helice Fault segment is possibly associated to the 23 AD (Koukouvelas et al., 2001; Pavlides et al., in press) and to the 1861 earthquakes (Schmidt, 1867, Mouyaris et al., 1992) and the Aegion Fault itself with the 1995 earthquake (Koukouvelas, 1998). Unfortunately, up to now the important 373 BC event has not been attributed to any specific seismogenic structure.

Archaeological evidence also brings additional information for co-seismic ground fractures (e.g. Stiros, 1988; Pavlides et al., 2001a). For example, archaeoseismological researches suggest that at least two major earthquakes struck the town of Butrint (SW Albania; Fig. 1), the first one during late Roman–early Byzantine times (2nd–3rd century) and the second one during the Byzantine period (12th century). The most distinctive evidence for surface ruptures are found in the theatre and in the Tower Gate (Hasani, 1989; Pavlides et al., 2001a). This seismic event was responsible for the destruction of Butrint after the 3rd century AD. The oldest known historical event, which occurred between the late 3rd and the 4th centuries close to Butrint, is that of 358 AD. Accordingly, Butrint can be considered the macroseismic epicentre of the 358 AD strong earthquake, where primary destruction due to seismic fault damages is documented. The maximum Intensity of this event can then be re-considered as IX to X, because of the seismic cracks that affected the buildings' foundations and the floor of the terraced seating of the theatre. In addition, the epicentre of the poorly known seismic event of the 1153 AD could be assumed in the Butrint area, though with a larger degree of uncertainty (Hasani, 1989; Pavlides et al., 2001a).

## 4. Surface deformation associated to historical earthquakes

Co-seismic surface ruptures in the broader Aegean Region are generally associated with moderate-large earthquakes connected with faults extending at depth for 10–20 km and with relevant offsets. The length of seismogenic fault traces commonly does not exceed 15–25 km, while secondary ground deformation (cracks, landslides, liquefaction, etc.) can be widespread. The identification of the seismogenic fault is

Table 1
Large Greek earthquakes associated with surface phenomena

| Date | Affected area | Size | Intensity | Phenomena | References |
|---|---|---|---|---|---|
| 550 BC | Sparta (Peloponnese) | $M \cong 7.0$ | $I_0 = IX$ | rock falls from Taygetus Mt | Gu89 |
| 469–464 BC | Sparta (Peloponnese) | $M > 7.0$ | $I_0 = XI$ | fault traces (?), rock falls | Gu89, Ar91 |
| 426 BC | Atalanti–Scarfea (Phthyotida, Central Greece) | $M \cong 7.0$ | $I_0 = IX$ | fault traces (?) | Gu89, PP97 |
| 373 BC | Helice–Bura (Achaia–Corinthian Gulf) | $M \cong 7.0$ | $I_0 = IX$ | possible fault traces | Gu89, PP97 |
| 330 BC | Lemnos (Northern Aegean Islands) | $M \cong 7.0$ | $I_0 = X$ | land (volcanic) subsidence or landslides (?) | Gu89, PP97 |
| 199–198 BC | Chalkida (Eubea, Central Greece) | $M \cong 7.0$ | $I_0 = X$ | fault traces (?), volcanic eruption (?) | PP97 |
| 551 AD | Patras–Corinthos (Peloponnesse–Phokida) | $M > 7.0$ | $I_0 = IX$ | fault traces, landslides | Gu89, PP97 |
| 1612 | Leukada (Ionian Islands) | $M = 6.6$ | $I_0 = X$ | fault traces and ground cracks | PP97 |
| 1622 May 5 | Zakynthos (Ionian Islands) | $M > 6.5$ | $I_0 = IX$ | fault traces | PP97 |
| 1630 July 2 | Leucada (Ionian Islands) | $M > 6.5$ | $I_0 = X$ | fault traces | PP97 |
| 1633 November 5 | Zakynthos (Ionian Islands) | $M > 6.5$ | $I_0 = X$ | fault traces | PP97 |
| 1636 September 30 | Cephalonia (Ionian Islands) | $M > 7.0$ | $I_0 = X$ | fault traces | PP97 |
| 1650 October 7 | Santorini Volcano (Southern Aegean) | $M \cong 6.2$ | $I_0 = VIII$ | ground cracks | Ak |
| 1714 August 28 | Cephalonia (Ionian Islands) | $M > 6.0$ | $I_0 = VIII$ | ground cracks | PP97 |
| 1740 October 4 | Lamia (Central Greece) | $M > 6.0$ | $I_0 = VIII$ | fault traces | AJ98, PP97 |
| 1759 June 22 | Thessaloniki (Macedonia, Northern Greece) | $M = 6.5$ | $I_0 = IX$ | fault traces | PP97 |
| 1767 July 22 | Cephalonia (Ionian Islands) | $M > 7.0$ | $I_0 = X$ | ground cracks (the longest 100 m, heave 1 m) | PP97 |
| 1820 December 29 | Zakynthos (Ionian Islands) | $M > 6.5$ | $I_0 = IX$ | ground cracks | PP97 |
| 1829 May 5 | Drama (Macedonia, Northern Greece) | $M > 7.0$ | $I_0 = X$ | fault traces (?) | AJ98, PP97 |
| 1846 June 11 | Messinia (SW Peloponnese) | $M = 6.5$ | $I_0 = X$ | fault traces, liquefaction (?) | PP97 |
| 1858 February 21 | Corinthos (Peloponnese) | $M > 6.5$ | $I_0 = X$ | fault traces | PP97 |
| 1860 August 6 | Samothrace (Northern Aegean) | $M > 6.5$ | $I_0 = VII$ | fault traces | PP97 |
| 1861 December 26 | Valimitika (Northern Peloponnese) | $M = 6.7$ | $I_0 = X$ | fault traces (E–W normal faulting) | AJ98, Ko01, Sh67 |
| 1867 February 27 | Cephalonia (Ionian Islands) | $M > 7.0$ | $I_0 = X$ | fault traces | PP97 |
| 1867 March 7 | Lesbos (Northern Aegean Islands) | $M > 6.5$ | $I_0 = X$ | fault traces, landslides and liquefaction | PP97, Fy99, Ka78 |
| 1870 August 1 | Fokis (Central Greece) | $M > 6.7$ | $I_0 = IX$ | fault traces, landslides and liquefaction | AJ98, AP89, PP97 |
| 1872 February 11 | Saghiada (Epirus, NW Greece) | $M = 6.1$ | $I_0 = IX$ | ground cracks | PP97 |
| 1881 April 3 | Chios (Central Aegean Islands) | $M = 6.4$ | $I_0 = XI$ | fault traces and landslides | PP97 |
| 1886 August 27 | Marathoupolis (SW Peloponnese) | $M = 7.5$ | $I_0 = X$ | fault traces | PP97 |
| 1893 February 9 | Samothrace (NE Aegean Islands) | $M = 6.5$ | $I_0 = IX$ | ground cracks and rock falls | PP97 |
| 1894 April 27 | Atalanti–Thermopylae (Central Greece) | $M = 7.0$ | $I_0 = X$ | fault traces (NW–SE normal faulting) | AJ90, PP97, Sk94 |
| 1902 July 5 | Thessaloniki–Mygdonia basin (Northern Greece) | $M = 6.6$ | $I_0 = IX$ | ground cracks | PP97 |
| 1903 August 11 | Kythera Island (Western Cretan Sea) | $M = 7.9$ | $I_0 = XI$ | ground cracks | PP97 |
| 1912 January 24 | Cephalonia (Ionian Islands) | $M = 6.8$ | $I_0 = X$ | ground cracks | PP97 |
| 1914 November 27 | Leucada (Ionian Islands) | $M = 6.3$ | $I_0 = IX$ | ground cracks and landslides | PP97 |
| 1915 August 7 | Ithaka (Ionian Islands) | $M = 6.7$ | $I_0 = IX$ | ground cracks | PP97 |

Table 1 (*continued*)

| Date | Affected area | Size | Intensity | Phenomena | References |
|---|---|---|---|---|---|
| 1921 September 13 | Akarnania (Central Greece) | $M=6.0$ | $I_0=$ VIII | ground cracks | PP97 |
| 1928 April 22 | Corinthos (Peloponnese) | $M=6.3$ | $I_0=$ IX | ground cracks and fault traces (?) | PP97 |
| 1931 March 8 | Valandovo (Macedonia, Northern Greece) | $M=6.7$ | $I_0=$ X | ground cracks | PP97 |
| 1932 September 26 | Eastern Chalkidiki (Northern Greece) | $M=6.9$ | $I_0=$ X | fault traces (E–W normal faulting) | Fl33, Ma33, PT91 |
| 1938 July 20 | Oropos (Central Greece) | $M=6.0$ | $I_0=$ VIII | ground ruptures, liquefaction | PP97, Go02 |
| 1941 March 1 | Larissa (Thessaly, Central Greece) | $M=6.3$ | $I_0=$ VIII | fault traces (?) and ground cracks | AJ90, Ca94 |
| 1948 April 22 | Leucada (Ionian Islands) | $M=6.5$ | $I_0=$ IX | ground cracks | PP97 |
| 1953 September 5 | Sousaki (Corinth area, Central Greece) | $M=6.2$ | $I_0=$ | fault traces | St95 |
| 1954 April 30 | Sophades (Thessaly, Central Greece) | $M=7.0$ | $I_0=$ IX+ | fault traces (NW–SE and E–W normal faulting) | AJ90, PM86, PM87 |
| 1957 March 8 | Velestino (Thessaly, Central Greece) | $M=6.5$ | $I_0=$ IX+ | fault traces | AJ90 |
| 1965 April 4 | Arcadia (Peloponnese) | $M=6.1$ | $I_0=$ X | ground cracks | PP97 |
| 1966 September 1 | Megalopolis (Peloponnese) | $M=6.0$ | $I_0=$ VIII | fault traces | AJ98, Am67 |
| 1968 February 19 | Agios Efstratios Island (Northern Aegean Sea) | $M=7.1$ | $I_0=$ IX | fault traces (strike-slip faulting, NE–SW) | PT91 |
| 1978 June 20 | Mygdonia (Macedonia, Northern Greece) | $M=6.5$ | $I_0=$ VIII+ | fault traces (mainly E–W normal faulting) | Pa80, Me79, Mo92 |
| 1980 July 9 | Magnesia (Southern Thessaly, Central Greece) | $M=6.4$ | $I_0=$ VIII+ | fault traces ( E–W normal faulting) | Pa80, Pa83 |
| 1981 February 24 | East Corinthian Gulf (Central Greece) | $M=6.7$ | $I_0=$ IX | fault traces (E–W normal faulting) | Pa95, Ja82 |
| 1981 March 4 | Boeotia (Central Greece) | $M=6.4$ | $I_0=$ IX+ | fault traces (E–W normal faulting) | Pa95, Ja82 |
| 1983 March 16 | Akarnania (Western-Central Greece) | $M=5.4$ | $I_0=$ | fault traces (sinistral strike slip faulting WNW–ESE) | Pa95, Ko90 |
| 1986 September 13 | Kalamata (SW Peloponnese) | $M=6.0$ | $I_0=$ IX | fault traces (NNW–SSE normal faulting) | Pa95, Ma89 |
| 1993 March 26 | Pyrgos (NW Peloponnese) | $M=5.2$ | $I_0=$ | ground cracks, fault traces (E–W) and liquefaction | Ko96 |
| 1995 May 13 | Kozani–Grevena (Macedonia, NW Greece) | $M=6.6$ | $I_0=$ IX | fault traces (ENE normal faulting) and liquefaction | Mo98, Pa95 |
| 1995 June 15 | Aegion (Northern Peloponnese) | $M=6.2$ | $I_0=$ VIII | fault traces (E–W normal faulting) | KD96 |
| 1996 July 26 | Konitsa (Epirus, Western Greece) | $M=5.7$ | $I_0=$ VIII | ground ruptures | Go02 |
| 1999 September 7 | Athens (Attics, Central Greece) | $M=5.9$ | $I_0=$ VII | fault traces | Pa01 |

Large earthquakes of Greece associated to fault traces, ground cracks and other surface phenomena, not necessarily of tectonic origin. The epicentral area is shown in Fig. 1. The complete list of references is reported in Table 2.

crucial for better understanding the seismic process and to evaluate more reliably the seismic hazard of the area.

Earthquakes in Greece with magnitudes larger than 6.0 are commonly associated to morphogenic faults, thus producing direct surface faulting (e.g. Ambraseys and Jackson, 1990). However, information is also available in the literature for seismic fault traces associated to smaller magnitude events, like the Ms = 5.8 1953 Corinth (Central Greece; Stiros, 1995), the Ms = 5.4 1983 Akarnania (Western Greece; Koukis et al., 1990), the Ms = 4.6 1970 Yali (southeastern Aege-

an; Stiros and Vougioukalakis, 1996) and the Ms = 5.0 1989 Patras (Peloponnese; Doutsos and Poulimenos, 1992) earthquakes. Reported surface rupture lengths for these events are 3.0, 2.5, 0.6 and 1.0–5.0 km, respectively, while the maximum displacement ranges between 1 and 10 cm. The latter two events are neglected because they are probably associated to secondary deformational phenomena and not typically co-seismic. The 1983 Akarnania earthquake is neglected too because it was clearly associated to a predominant component of strike-slip faulting, while in our database we considered only dip-slip normal events.

The first well mapped co-seismic ruptures in Greece and worldwide as well, and the corresponding scientific description of the surface deformational phenomena appeared after the Helice–Valimitika (Gulf of Corinth) 1861 and the Atalanti (Lokrida, Central Greece) 1894 earthquakes by Schmidt (1867) and Skufos (1894), respectively. However, systematic studies, urged by the understanding of the importance of active faulting and the related seismically induced surface phenomena, begun only after the 1978 Thessaloniki event ($M = 6.5$). They were followed by the 1980 Volos ($M = 6.5$), the 1981 Gulf of Corinth ($M = 6.7$ and

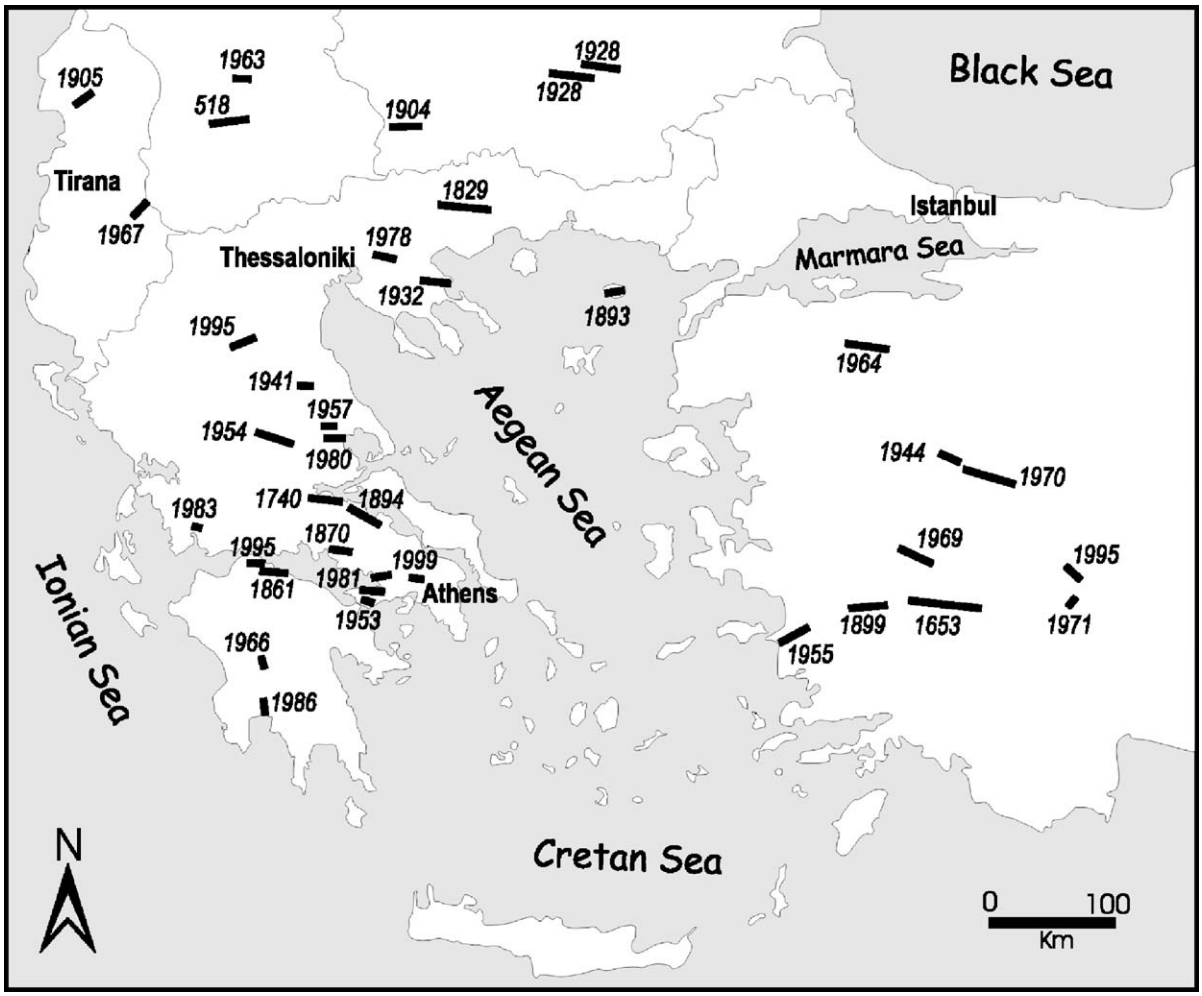


Fig. 2. Map of the broader Aegean Region showing the distribution of all the morphogenic earthquakes (sensu Caputo, 1993) and corresponding co-seismic ruptures considered in the present research. Numbers refer to the year of the seismic event; the bars are oriented parallel to the fault traces, while the length of the bars is roughly proportional to SRL. Data from Table 2.

6.4), the 1986 Kalamata ($M = 6.0$) and the 1995 Kozani–Grevena ($M = 6.6$) earthquakes among others (Fig. 2).

A reappraisal of existing data and additional evidence derived from recent geological investigations illustrate the often complex association between seismic activity and fault geometry that characterise the tectonic framework in mainland Greece. Morphotectonic, structural, seismotectonic and remote sensing analyses emphasise the complex character of the seismogenic zones (Pavlides, 1993; Pavlides and King, 1998), which were previously interpreted as having a relatively simple seismotectonic setting. Structural heterogeneity leads to complex fault geometry and generates a variety of geometrical barriers to seismic rupture propagation. This last aspect of the phenomenon will be further discussed in a later section.

From the list of Table 1, all earthquakes which are clearly associated with surface failures that we assume they can be directly correlated to a seismogenic structure were selected (Table 2). Accordingly, we further investigated these events trying to quantify some important seismotectonic parameters like the surface rupture length (SRL), the maximum displacement and the average displacement (AD).

As above mentioned, the Aegean Region is largely affected by an extensional tectonic regime which mainly produces and activates dip-slip normal faults (e.g. Mercier et al., 1989; Caputo and Pavlides, 1993; Goldsworthy et al., 2002). More than 100 normal faults or fault segments are well known and studied. Therefore, because the bulk of the events for which data are available are associated to normal faulting, we neglected some cases of morphogenic strike-slip faulting and after careful inspection of the data, we thus considered the maximum vertical displacement (MVD). This parameter is listed in Table 2.

Moreover, in order to enlarge the database, we added all possible earthquakes characterised by extensional faulting which occurred in the broader Aegean Region, including the Southern Balkan peninsula and Western Anatolia (Table 2 and Fig. 2). However, also in this case though some recent events are well documented we neglected them because clearly associated to an important or even a prevailing strike-slip component of motion. Among these are the important 1912 Callipolis (Gelibolu), the 1953 Yenice and the 1997 Izmit–Kocaeli and Duzce earthquakes.

In Fig. 2, are represented the locations of the well documented co-seismic fault ruptures listed in Table 2. Most of them are associated with well-known neotectonic and active faults further discussed in a later section of this note.

## 5. Empirical relationships of earthquake magnitude, SRL and MVD

Because surface faulting due to morphogenic earthquakes is of importance to all modern studies of seismotectonics, seismicity and especially seismic hazard assessment, quantitative empirical relationships between earthquake magnitude and fault parameters have been obtained and are discussed here after.

It is important to stress that in this study we consider only the data obtained from geological information, that is relative to well-documented surface ruptures associated with seismic events, both in terms of total length and maximum measured displacement. All other earthquakes weakly linked with specific faults or that produced only secondary gravitational ground deformation have been neglected from this study.

For practical use of geological data in earthquake applied studies, it is noteworthy to obtain a relationship of earthquake magnitude (Ms) versus surface rupture length (SRL) and maximum vertical displacement (MVD). As above mentioned, due to the not homogeneity of the sources, in several cases more than one value is available. As a first attempt, we considered a range of possible values (Pavlides et al., 2000), while for the present research, we further investigated several earthquakes and eventually choose a unique value for each event. Corresponding data and diagrams are represented in Figs. 3 and 4 for Ms/SRL and Ms/MVD, respectively.

Based on these data, we then calculated regressions of Ms on SRL and of Ms on MVD, respectively

$$\mathrm{Ms} = 0.90 \cdot \log(\mathrm{SRL}) + 5.48 \quad (1)$$

$$\mathrm{Ms} = 0.59 \cdot \log(\mathrm{MVD}) + 6.75 \quad (2)$$

Table 2
Seismotectonic parameters of normal faulting earthquakes from the Aegean Region

| No. | Date | Locality | Fault | Country | Ms | $I_{max}$ | GFL (km) | SRL (km) | MVD (m) | AD (m) | Q ### | Ref. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1 | 518 | Scupus (Skopje) | – | FY | 7.0 | X | – | 44 | 3.60 | – | _CC | Am71, Gu89, PP97 |
| 2 | 1653.02.23 | Menderes | Buyuk Menderes E | TR | 7.1 | X | 80 | 70 | 3.00 | – | CCC | AJ98, Am88, PP97, Sa92 |
| 3 | 1740.10.04 | Lamia | Kallidromo | GR | 6.6 | VIII | 28 | 25 | – | – | AC_ | AJ98, PP97 |
| 4 | 1829.05.05 | Drama | Xanthi | GR | 7.2 | X | 50 | 50 | – | – | BB_ | AJ98, PP97 |
| 5 | 1861.12.26 | Valimitika | Helice | GR | 6.6 | X | 26 | >15 | 1.00 | 0.60 | BBB | AJ98, Ko01, PP97, Sh67 |
| 6 | 1870.08.01 | Fokis | Delphi | GR | 6.7 | IX | 28 | 20 | 1.00 | – | ABC | AJ98, AP89 |
| 7 | 1893.02.09 | Samothrace | Therma | GR | 6.8 | IX | >18 | >10 | – | – | BC_ | PP97 |
| 8 | 1894.04.27 | Atalanti | Atalanti | GR | 6.9 | X | 35 | 30 | 1.50 | 0.25 | AAA | AJ90, Sk94 |
| 9 | 1899.09.20 | Aydin | Buyuk Menderes W | TR | 6.9 | IX | 100 | 40 | 1.00 | – | BBB | AJ98, PP97 |
| 10 | 1904.04.04 | Kresna | Krupnik | BG | 7.2 | X | 27 | 25 | 2.00 | 0.10 | CCB | AJ98, DA00 |
| 11 | 1905.06.01 | Shkodra (Scutari) | Scutari–Pec | AL | 6.6 | IX | – | >10 | 1.00 | – | _CC | AJ98, KS80, PP97, SK80 |
| 12 | 1928.04.14 | Chirpan–Plovdiv | E Plovdiv | BG | 6.8 | IX | 43 | 38 | 0.50 | – | CBA | AJ98, Ja45, PP97, Ri58, Sh97 |
| 13 | 1928.04.18 | Popovitsa–Plovdiv | W Plovdiv | BG | 7.0 | X | 55 | 50 | 3.00 | – | CBA | AJ98, Ja45, PP97, Ri58, Sh97 |
| 14 | 1932.09.26 | Ierissos | Stratoni | GR | 6.9 | X | >25 | 20 | 1.80 | 0.30 | CBB | AJ98, Fl33, GG53, PT91 |
| 15 | 1941.03.01 | Larissa | Asmaki | GR | 6.3 | VIII | 14 | 7 | – | – | AB_ | AJ90, Ca94, Ca95, PP97 |
| 16 | 1944.06.25 | Saphane | Saphane | TR | 6.1 | VIII | 22 | 18 | 0.30 | 0.10 | BCC | AJ98, Am88, PP97 |
| 17 | 1953.09.05 | Sousaki | Sousaki | GR | 5.8 | – | – | >3 | 0.08 | – | _CC | St95 |
| 18 | 1954.04.30 | Sophades | Domokos | GR | 6.7 | IX+ | 35 | 30 | 0.90 | 0.30 | AAA | AJ90, Ca90, Ca95, PM86, PM87 |
| 19 | 1955.07.16 | Buyuk Menderes | Buyuk Menderes E | TR | 6.9 | VIII | 38 | 35 | – | – | BC_ | PP97, Sa92 |
| 20 | 1957.03.08 | Velestino | Righeo | GR | 6.5 | IX+ | 24 | >8 | 0.20 | – | BCB | AJ90, Ca90, Ca95 |
| 21 | 1963.07.26 | Skopje | – | FY | 6.1 | IX | 8 | 6 | 0.10 | – | CCC | PP97 |
| 22 | 1964.10.06 | Manyas | Manyas | TR | 6.8 | IX | 45 | 40 | 0.10 | – | BBB | AJ98, Sa92 |
| 23 | 1966.09.01 | Megalopolis | Megalopolis | GR | 5.6 | VIII | 20 | 2 | 0.05 | – | BBB | AJ98, Am67 |
| 24 | 1967.11.30 | Dibra | Okshtuni–Përvalla | AL | 6.6 | IX | 23 | 16 | 0.50 | – | CCC | AJ98, SK80, Ta93 |
| 25 | 1969.03.28 | Alasehir | North Menders Massif | TR | 6.5 | VIII | 50 | 35 | 0.80 | – | BAC | Am88, AJ98, EJ85, KA69 |
| 26 | 1970.03.28 | Gediz | Murat ? | TR | 7.1 | IX | 45 | 40 | 2.30 | – | BAA | AT72, PP97, Ta71 |
| 27 | 1971.05.12 | Burdur | Burdur | TR | 6.2 | VIII | 16 | >4 | 0.30 | 0.14 | CCB | AJ98, Am88, PP97 |
| 28 | 1978.06.20 | Thessaloniki | Mygdonia | GR | 6.5 | VIII | 17 | 15 | 0.25 | 0.08 | AAA | Me79, Mo92, Pa80, Pa96 |
| 29 | 1980.07.09 | Volos | Nea Anchialos | GR | 6.5 | VIII | 20 | >8 | 0.10 | 0.10 | ABA | AJ90, Ca90, Ca96, Pa83 |
| 30 | 1981.02.24 | Alkyonides | Perachora–Pisia | GR | 6.7 | IX | 33 | 15 | 0.80 | 0.30 | BAA | Ja82, Pa82 |

Table 2 (*continued*)

| No. | Date | Locality | Fault | Country | Ms | $I_{max}$ | GFL (km) | SRL (km) | MVD (m) | AD (m) | Q ### | Ref. |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 31 | 1981.03.04 | Alkyonides | Kaparelli | GR | 6.4 | IX+ | >22 | >12 | 0.70 | 0.60 | BBA | Ja82 |
| 32 | 1986.09.13 | Kalamata | Kalamata | GR | 6.0 | IX | 15 | 10 | 0.10 | 0.05 | CBB | Ma89, Pa88, PP97 |
| 33 | 1995.05.13 | Kozani | Paleochori | GR | 6.6 | IX | 39 | 27 | 0.20 | 0.06 | ABA | Ha97, Mo98, Pa95, Ch98 |
| 34 | 1995.06.15 | Aegion | Aegion | GR | 6.4 | VIII | 12 | 7 | 0.07 | 0.01 | BAA | KD96 |
| 35 | 1995.10.01 | Dinar | Dinar | TR | 6.2 | IX | 12 | 11 | 0.30 | 0.12 | AAA | EB96, Ko00 |
| 36 | 1999.09.07 | Athens | Fili | GR | 5.9 | VII | 15 | 8 | 0.06 | – | AAA | Pa01 |

Seismotectonic parameters of selected morphogenic earthquakes (sensu Caputo, 1993) which occurred in the broader Aegean Region (see Fig. 2) associated to fault traces. Locality = affected region; fault = seismogenic fault; country: (GR = Greece, BL = Bulgaria, TR = Turkey, AL = Albania, FY = Former Yugoslavian Republic of Macedonia); Ms = magnitude; $I_{max}$: maximum intensity; GFL: geological fault length; SRL: total length of surface ruptures; MVD: maximum vertical displacement; Q: quality factor for GFL, SRL and MVD, respectively (A = good; C = fair); Ref. references [AJ90: Ambraseys and Jackson, 1990; AJ98: Ambraseys and Jackson, 1998; Ak: Akylas (ms in Greek); Am67: Ambraseys, 1967; Am71: Ambraseys, 1971; Am88: Ambraseys, 1988; AP89: Ambraseys and Pantelopoulos, 1989; Ar91: Armijo et al., 1991; AT72: Ambraseys and Tchalenko, 1972; Ca90: Caputo, 1990; Ca94: Caputo et al., 1994; Ca95: Caputo, 1995; Ca96: Caputo, 1996; Ch98: Chatzipetros et al., 1998; DA00: Dobrev and Avramova-Tacheva, 2000; EB96: Eyidoğan and Barka, 1996; EJ85: Eyidoğan and Jackson, 1985; Fl33: Floras, 1933; Fy99: Fytikas et al., 1999; GG53: Georgalas and Galanopoulos, 1953; Go02: Goldsworthy et al., 2002; Gu89: Guidoboni, 1989; Ha97: Hatzfeld et al., 1997b; Ja45: Jankof, 1945; Ja82: Jackson et al., 1982; KA69: Ketin and Abdüsselamoğlu, 1969; Ka78: Kabouris, 1978; KD96: Koukouvelas and Doutsos, 1996; Ko01: Koukouvelas et al., 2001; Ko90: Koukis et al., 1990; Ko96: Koukouvelas et al., 1996; Ko00: Koral, 2000; KS80: Koçiaj and Sulstarova, 1980; Ma33: Maravelakis, 1933; Ma89: Mariolakos et al., 1989; Me79: Mercier et al., 1979; Mo98: Mountrakis et al., 1998; Mo92: Mountrakis et al., 1992; Pa01: Papadopoulos et al., 2002; Pa80: Papazachos et al., 1980; Pa82: Papazachos et al., 1982; Pa83: Papazachos et al., 1983; Pa88: Papazachos et al., 1988; Pa95: Pavlides et al., 1995; Pa96: Pavlides, 1996b; PM86: Papastamatiou and Mouyaris, 1986a; PM87: Papastamatiou and Mouyaris, 1986b; PP97: Papazachos and Papazachou, 1997; PT91: Pavlides and Tranos, 1991; Ri58: Richter, 1958; Sa92: Şaroğlu et al., 1992; Sh67: Schmidt, 1867; Sh97 : Shanov, 1997; SK80: Sulstarova and Koçiaj, 1980; Sk94: Skufos, 1894; St95: Stiros, 1995; Ta71: Taşdemiroğlou, 1971; Ta93: Tagari, 1993].

The correlation coefficient is 0.84 for the Ms/SRL regression equation and 0.82 for the Ms/MVD regression equation relative to 36 and 31 data, respectively. The confidence value is larger than 99% for both equations, while the corresponding 95% curves are represented in Figs. 3 and 4, respectively.

We also calculated the regression curves of the two geological parameters (SRL and MVD) on Ms

$$\log(\mathrm{SRL}) = 0.78 \cdot \mathrm{Ms} - 3.93 \quad (3)$$

$$\log(\mathrm{MVD}) = 1.14 \cdot \mathrm{Ms} - 7.82 \quad (4)$$

Any further attempt to improve the preliminary database concerning the average displacement (AD) proposed by Pavlides et al. (2000) was not successful. The data are still few and, above all, they appear to be quite scattered if plotted versus magnitude. Any attempt of calculating a regression equation returned a low correlation coefficient, therefore meaningless. We therefore decided to avoid presenting and further discussing these data.

By comparing the above equations with similar ones proposed by other authors, similarities, as well as differences are observed (Figs. 5 and 6). For example, the curves proposed by Wells and Coppersmith (1994) based on a worldwide data set are generally lower than ours, while more similar seem to be the regressions equations proposed by Ambraseys and Jackson (1998) for the Eastern Mediterranean–Middle East Region. The latter similar pattern probably reflects the partial overlap of their database with our and more comparable seismotectonic behaviour during crustal deformation of the two investigated areas. In contrast, if the normal faulting process occurs in different geodynamic conditions (Wells and Coppersmith, 1994) it is reasonable that the Ms to SRL and the Ms to MVD relationships are different.

In addition, Papazachos and Papazachou (1989) carried out a similar statistical analysis for the Aegean Region by calculating empirical relationships between fault length and surface magnitude (Ms) for 21 shallow earthquakes and between the displacement and the

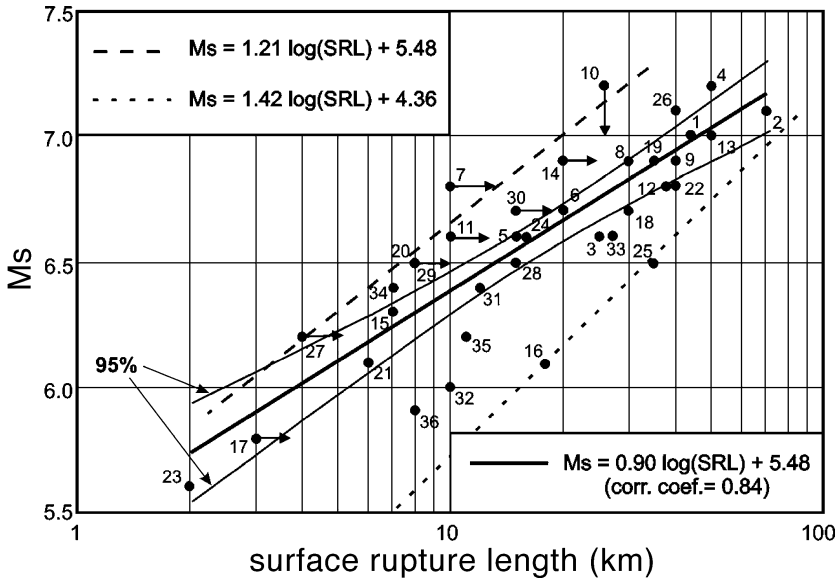


Fig. 3. Magnitude (Ms) versus surface rupture length (SRL). Numbers refer to the events listed in Table 2. In the diagram are also represented the best-fit curve and the corresponding equation (bottom right), the 95% confidence curves and the lower and upper envelopes with the corresponding equations (top left). Arrows associated with some events are discussed in the text.

surface magnitude for 15 events. Due to the lack of the inverse formulas (i.e. Ms on the length and on displacement), we recalculated the same relationships based on their data (their Table 7.4). The curves, represented in Figs. 5 and 6, show important differences probably due to the fact that their data set is not

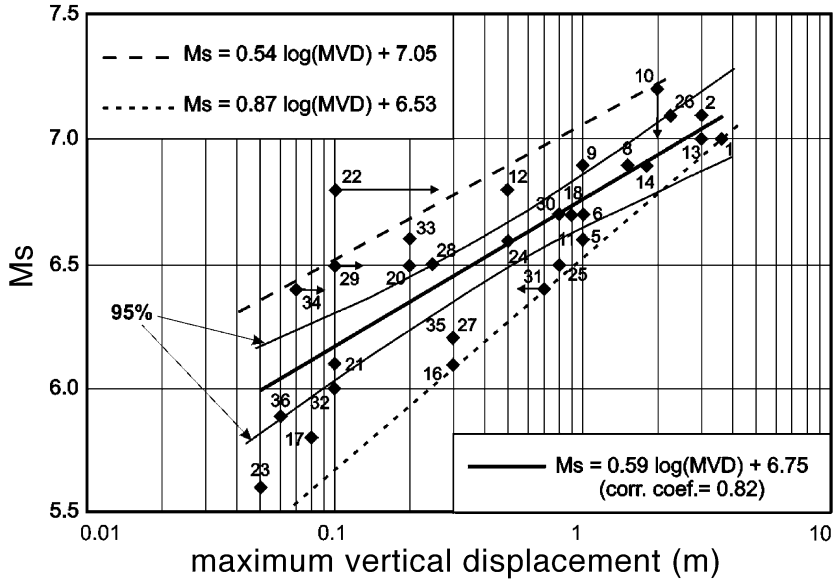


Fig. 4. Magnitude (Ms) versus maximum vertical displacement (MVD). Numbers refer to the events listed in Table 2. In the diagram are also represented the best-fit curve and the corresponding equation (bottom right), the 95% confidence curves and the lower and upper envelopes with the corresponding equations (top left). Arrows associated with some events are discussed in the text.

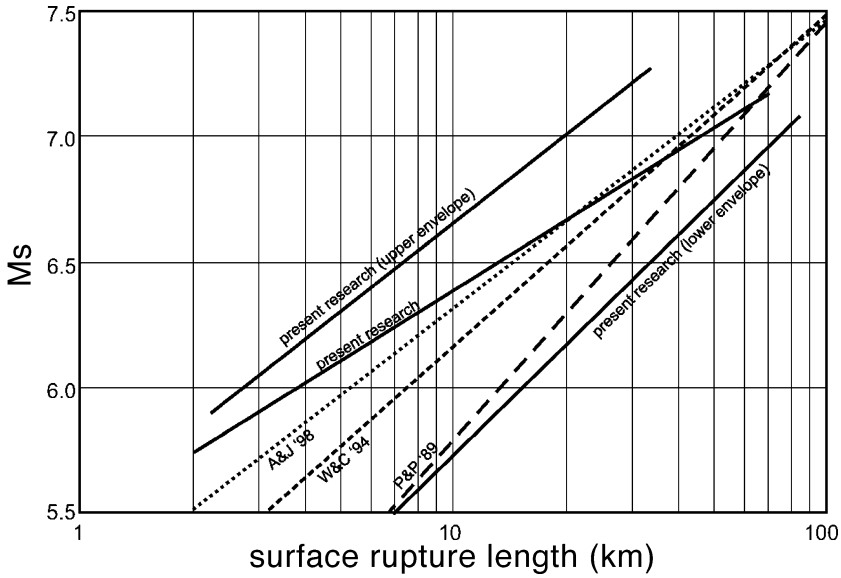


Fig. 5. Comparison between regression curves for Ms versus SRL, proposed by Wells and Coppersmith (1994) for normal faulting from worldwide data (W&C'94) and by Ambraseys and Jackson (1998) for the Mediterranean region (A&J'98). The curve from Papazachos and Papazachou (1989) has been recalculated according to their data because of a typographic error in the given equations (P&P'89). The best-fit regression curve as well as the upper and lower envelopes based on this research is also represented.

homogeneous. Indeed, values were assumed not only on the basis of field observations of surface fault traces (viz. geological information) but also from the spatial distribution of aftershocks (viz. seismological information), which could be related to subsurface rupture length. Moreover, Papazachos and Papazachou (1989) included in the same list, earthquakes associated to both normal and strike-slip faulting.

In order to better understand the general distribution of our data set and to discuss about possible causes for the few misfit cases, it is assumed that the magnitude values are basically correct though it is clear that some events need a careful re-evaluation, especially for the older ones. However, we consider such assumption generally correct because possible errors about magnitude are probably limited. The only exception could be the 1904 Kresna earthquake (no. 10), where it is still debated whether it was a unique 7.2 event or two smaller sub-events. If the latter were the real case, then the magnitude should be reduced accordingly and the datum would be closer to the best-fit equations (Eqs. (1) and (2)) (Figs. 3 and 4).

If we carefully observe the diagrams, we can try to understand and possibly justify why geological information for some earthquakes gives values which are higher or lower than average. For example, some of the earthquakes falling outside the 95% confidence curves are related to events occurring offshore or near the coast. Consequently, two possible effects could have occurred. Firstly, it is obvious that at least part of the surface ruptures associated to such earthquakes could have continued offshore, therefore remaining completely undetected by even accurate field surveys carried out soon after the seismic event. In addition, the superficial maximum displacement could have occurred offshore thus remaining undetected.

Secondly, it is also evident that sediments deposited in a coastal environment are highly saturated in water content. Consequently, surface deformation phenomena are generally hampered by these unfavourable lithological characteristics and the recorded displacements can thus be larger than 'normal'. However, due to the possible occurrence of strong site effects, the same geological conditions could also affect the estimate of magnitude especially if obtained from macroseismic data.

Additional to the near coast events, also the Manyas 1964 earthquake (no. 22) shows a large misfit

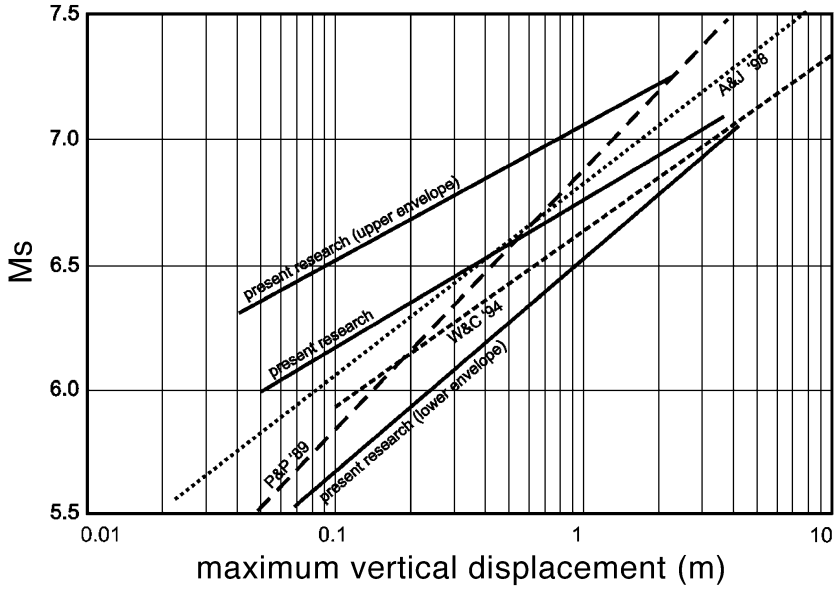


Fig. 6. Comparison between regression curves for Ms versus MVD, proposed by Wells and Coppersmith (1994) for normal faulting from worldwide data (W&C'94) and by Ambraseys and Jackson (1998) for the Mediterranean region (A&J'98). The curve from Papazachos and Papazachou (1989) has been recalculated according to their data because of a typographic error in the given equations (P&P'89). The best-fit regression curve as well as the upper and lower envelopes based on this research is also represented.

of MVD. This is possibly due to some strike-slip component occurred during faulting which probably remained undetected thus reducing the measured value of maximum displacement. If this is the case and because we considering only dip-slip normal faults, this datum should be neglected.

According to these possible local effects, some indicative arrows relative to these earthquakes have been added in Figs. 3 and 4. Further indicative arrows have been also added in all cases where according to the literature data the accepted SRL, as listed in Table 2, should be considered as a minimum value (Fig. 3, event nos. 11, 20 and 27).

Additionally, a different analytical approach was also applied to our database by taking into account the upper envelopes of the Ms-SRL and the Ms-MVD diagrams. This is to estimate the worst-case relationships, a useful parameter in Earthquake Engineering and for Deterministic Seismic Hazard Analyses (see Strom and Nikonov, 1997). Indeed, these curves indicate the maximum expected magnitude, given a geologically determined surface rupture or a palaeoseismologically observed displacement. The upper envelopes are empirically traced in Figs. 3 and 4 also taking into account the possible correction of some of the data as above discussed. The corresponding equations of these attemptive curves are:

$$\mathrm{Ms} = 1.00 \cdot \log(\mathrm{SRL}) + 5.60 \quad (5)$$

$$\mathrm{Ms} = 0.48 \cdot \log(\mathrm{MVD}) + 7.04 \quad (6)$$

It is known, from instrumental data, that most earthquakes in mainland Greece are of magnitude up to 6.9–7.0. As a result, Eqs. (5) and (6) seem to be more appropriate for assessing the effects of a potential earthquake. Both, best-fit curves and upper envelopes could be useful in Seismic Hazard Assessment studies. If the meaning of the average curves are obvious and they do not need any further comment, when dealing with a deterministic approach to evaluate the worst case scenario, the use of the upper envelopes is certainly more appropriate.

Using the same criterion, in Figs. 3 and 4, we also added the lower envelopes thus delimiting the range

of possible values for the different parameters. The corresponding equations are

$$Ms = 1.42 \cdot \log(SRL) + 4.36 \quad (7)$$

$$Ms = 0.87 \cdot \log(MVD) + 6.53 \quad (8)$$

In addition, these curves offer additional information in Earthquake Engineering studies. A typical case is the inverse problem. In fact, given a magnitude based on macroseismic or instrumental data and searching for surface ruptures and vertical displacements following dedicated geological, morphotectonic and palaeoseismological investigations we could estimate the maximum expected values. Moreover, based on geological information and using both upper and lower envelopes, we can more correctly evaluate the possible range of magnitudes associated to past earthquakes for which no instrumental nor historical data are available. Indeed, this is a frequent problem faced when dealing with palaeoseismological data documenting the occurrence of past morphogenic events that are missing in the catalogues due to the lack of information or because much older than the historical record.

## 6. Empirical relationships based on GFL

Most of the events listed in Table 2 are associated to well-recognised seismogenic structures. Accordingly based on dedicated field surveys and on literature data (see references in Table 2), the length of the geological fault (GFL) was estimated, that is those tectonic structures showing evidences of recent activity comparable to that observed or described along the segments where surface ruptures have been documented for a specific earthquake. In other terms, a geological fault represents the cumulative effects of several morphogenic earthquakes.

The GFL has been plotted against magnitude in Fig. 7. In this case, we do not calculate a best-fitting equation considering it geologically meaningless because neotectonic faults could be the result of several earthquakes that could have re-activated different segments or they could have been partially or totally

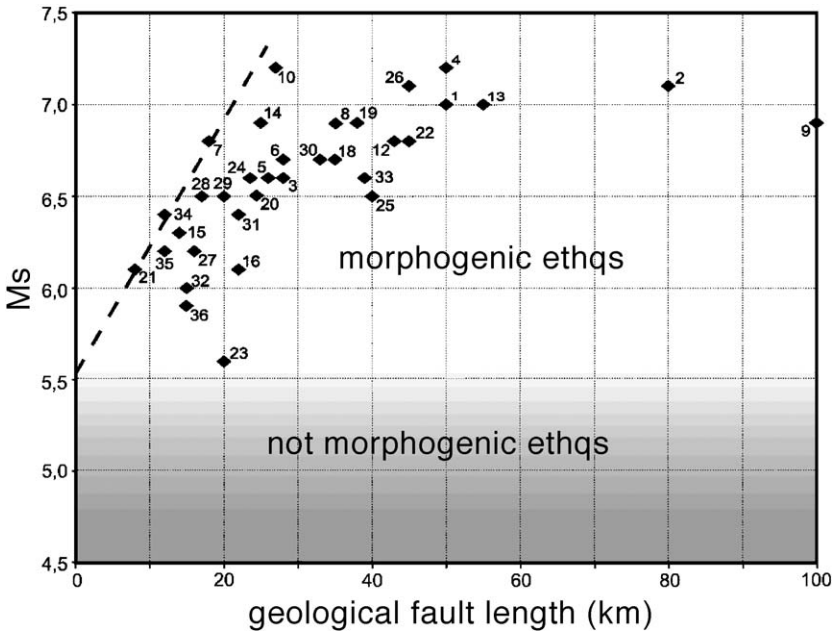


Fig. 7. Magnitude (Ms) versus geological fault length (GFL). Data from Table 2. The dashed line represents a possible lower length limit for geological faults associated to normal faulting as a function of the maximum expected magnitude. The intercept of the curve with the magnitude axis probably represents the lower threshold for morphogenic earthquakes (sensu Caputo, 1993). Further discussion in text.

overlapping in space. In contrast, in Fig. 7 we traced the possible lower limit (dashed line) for neotectonic structures associated to normal active faulting as a function of the maximum expected magnitude. In other terms, a fault capable of generating a moderate to strong earthquake, say larger than 5.5 degrees, should have a clear plurikilometric surficial expression detectable by a dedicated geological survey. This magnitude limit is also in agreement with the choice made by Valensise and Pantosti (2001) for the catalogue of seismogenic sources of Italy.

On the other hand, Ms = 5.5 probably represents a physical threshold separating morphogenic from not-morphogenic earthquakes as defined by Caputo (1993). In practice, all normal faults generating earthquakes lower than about 5.5 degrees cannot produce surficial ruptures directly related to the seismogenic fault. Consequently, these tectonic structures are not suitable for palaeoseismological researches and should be investigated with alternative approaches.

Because in the Aegean Region no significant contribution of creeping has never been detected along normal faults, the observed geological fault traces (GFL) and relative offsets are mostly due to several past earthquakes. Therefore, searching at the surface with geological tools seismogenic sources capable of generating earthquakes with magnitude lower than 5.5 is certainly unpromising. However, according to Figs. 3 and 7, earthquakes up to magnitudes 5.7–5.8 are associated to few kilometres of GFL. In a region like the Aegean, where inherited crustal structures affecting the surface are widespread, the recognition with confidence of 'symptomatic' features of recent tectonic activity solely based on a geological survey could be again a difficult task.

To better understand the propagation process and the possible segmentation of the active faults, we also calculated the ratio between the surface rupture length (SRL) and the geological fault length (GFL). The histogram of this ratio is represented in Fig. 8. In contrast with an expected Gaussian-like distribution, the occurrence of two peaks is evident. The major peak is between 0.8 and 1 and a second peak is around 0.5. The first peak clearly emphasises that almost the half of the investigated earthquakes propagated and ruptured almost the entire fault length. For these seismogenic faults, the seismic potential in terms of maximum expected magnitude has been probably reached and therefore it can be safely considered for seismic hazard assessment analyses.

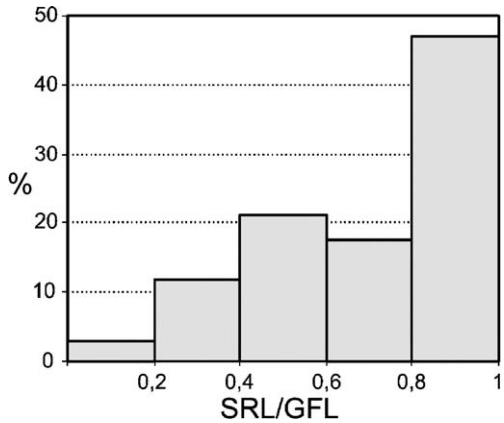


Fig. 8. Statistical distribution of the SRL to GFL ratio. The major peak around 0.8–1 indicates that the half of the historical earthquakes ruptured almost the entire neotectonic fault length as documented from geological, morphotectonic and palaeoseismological surveys. The second peak around the 0.5 value emphasises the existence of segmented faults which have not been entirely activated during past earthquakes.

In contrast, the second peak indicates that for more than 20% of the neotectonic and seismogenic faults the documented historical and instrumental earthquakes re-activated only one half of the tectonic structure thus implying the occurrence of structural barriers and important segment boundaries.

Moreover, it is worth to note that more than 50% of the earthquakes, which were documented here, did not propagate throughout the entire geological structure. This seismotectonic behaviour has obviously strong implications for the seismic hazard of a region because even when past earthquakes that occurred on a seismogenic fault are well documented, this does not assure that a future event will replicate in size the previous one. Eventually, this aspect definitely disproves the validity of the 'characteristic earthquake' model proposed by Schwartz and Coppersmith (1984), at least for the normal faults of the Aegean Region.

This apparently anomalous behaviour could be due not only to the fact that the stress field continuously varies in space and time at all scales (Caputo, 2001), but also because the sealing conditions of fractures at the time of the seismic rupture are commonly different

from those of the previous event. Therefore, the initial rupture on the seismic plane could occur with different mechanical and stress conditions while the latter can be unpredictable both in magnitude and orientation of the principal stress axes.

## 7. Concluding remarks

It is clear that most crustal earthquakes produce permanent and recognisable effects on the geological landscape and this can potentially enable geologists to quantitatively infer the degree of seismic activity in a region (see also Pavlides, 1996a; Pavlides and King, 1998).

The present research has confirmed the clear and straightforward relationships between three important seismological and geological parameters of active faults, such as the magnitude (Ms), the surface rupture length (SRL) and the maximum vertical displacement (MVD) of normal faults. Beyond the numerous difficulties in collecting a statistically sufficient database, it should be noted that the magnitude window of our data set is almost 2 degrees, spanning from 5.6 to 7.2.

Although further historical researches will certainly improve and expand the compiled catalogue, we guess that the limits of the magnitude window for morphogenic earthquakes in the Aegean Region, at least as concerns normal faulting, have been reached. In fact, earthquakes with higher magnitude values and associated to this kind of focal mechanism are not documented even worldwide (e.g. Wells and Coppersmith, 1994). This 'high-magnitude cut-off value' (Ms = 7.2) is also probably due to the limited crustal thickness, characterising the investigated area and consequently the associated active fault length and the possible seismic stress drop.

On the other hand, according to our results, seismic events with a magnitude lower than a certain value, say about 5.6–5.7 degrees, can produce very limited surface effects not necessarily co-seismic. In fact, according to our regression curves, maximum vertical displacements are limited to 1–2 cm, therefore easily destroyed within few days from their formation and often not distinguishable from other features of ground deformation. In addition, the calculated surface rupture length is only 1–1.5 km, with the consequent increased possibility that the seismogenic surface rupture crosses an inhabited or densely vegetated region. In both cases, the probability that these co-seismic features remain undetected during a field survey and thus historically unrecorded exponentially increases.

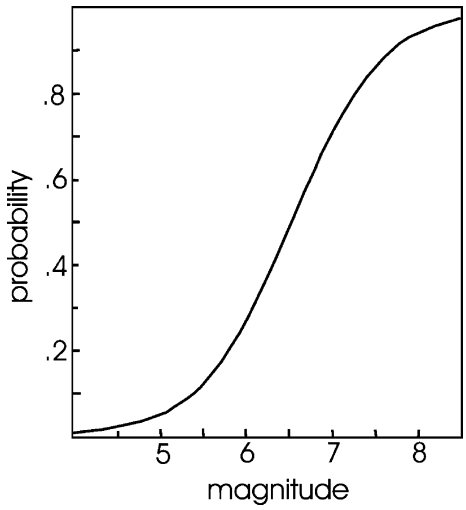


Fig. 9. Probability of producing surface ruptures versus magnitude for earthquakes associated to normal faulting according to Coppersmith and Youngs (2000).

From a statistical point of view, the 'low-magnitude cut-off value' (Ms = 5.5), as inferred from our diagrams and empirical relationships, can be interpreted in terms of 'probability of surface rupture' as a function of the magnitude of an earthquake. Pezzopane and Dawson (1996, in Coppersmith and Youngs, 2000), based on more than hundred extensional seismic events, has recently proposed this approach. These authors suggest that for an earthquake of magnitude 5.5 the probability to produce surface ruptures is about 10% (Fig. 9). Alternatively, this limit could be referred to the probability that these features can be geologically observed and detected.

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## Appendix A

In this appendix, selected earthquakes from Table 2 are described in more detail, while particular problems and criteria for defining specific parameters are discussed. Information for earthquakes and related ground deformation from Western Anatolia (Turkey), which belong to the broader Aegean active tectonic realm, were mainly based on the following sources: Map of Active Faults of Turkey (Şaroğlu et al., 1992; Taşdemiroğlou, 1971; Ambraseys, 1988; Guidoboni, 1989; Guidoboni et al., 1994; Ambraseys and Finkel, 1995; Barka and Kadinsky-Cade, 1988; Westaway, 1990, 1992; Barka and Reilinger, 1997; Ambraseys and White, 1997; Papazachos and Papazachou, 1997; Pinar, 1998; Ambraseys and Jackson, 1998; Wright et al., 1999, our field trips experience in the region and personal communications of many colleagues).

### *A.1. 518 Scupus (no. 1 in Table 2)*

Although Scupus (Skopje) 518 AD is one of the first historical earthquakes, clear evidence of surface faulting is documented. The event is reported by the Latin writer Marcellinus (Guidoboni, 1989; Guidoboni et al., 1994; Papazachos and Papazachou, 1997; Ambraseys and Jackson, 1998) that described a yawning chasm 44 km long and 3.6 m wide. We had two problems for adding this event to our database. The first is due to the great uncertainty of the magnitude. However, the value of $M = 7.0$ macroseismically inferred by Papazachos and Papazachou (1997) nicely fits our regression function (Fig. 3) and even a possible $\pm 0.3$ degrees would leave the earthquake within the two envelopes. The second problem is relative to the orientation and the exact location of the surface ruptures. According to the local geology, we tentatively assumed a roughly E–W-oriented seismogenic structure. The parameters SRL and MD have been also successfully tested in our preliminary similar empirical relationships published in Pavlides et al. (2000).

### *A.2. 1740 Lamia (no. 3)*

From sources cited by Papazachos and Papazachou (1997) and Ambraseys and Jackson (1998), some Greek reports translating the Pocoque's narration (e.g. D. Adam, "Lokrida" 2002) and our own geological observations in the macroseismic area, following the Pegoraro (1972) and Philip (1974) studies too, it seems that the October 5, 1740 strong earthquake, that ruined the town of Lamia, is associated to the WNW–ESE-trending Kallidromo Fault. Ground ruptures followed a WNW–ESE to NW–SE direction from Hypati to Reginion villages for about 25 km.

### *A.3. 1829 Drama (no. 4)*

The case of the 1829 double event is interesting because the two shocks are located in a region (Thrace, NE Greece), which is believed to be relatively aseismic, according to instrumental data. It is likely that the April 11, 1829 earthquake (M about 7.0) was associated to the western segment of the fault bordering the Drama Basin, while the May 5, 1829 ($M$>7.0), was associated to the eastern segment of the Xanthi Fault, also known as Thrace Fault or Kavala–Xanthi–Komotini Fault. Further specific researches should be performed to reduce the degree of uncertainty for this event.

### *A.4. 1861 Valimitika (no. 5)*

The 1861, Valimitika earthquake in Northern Peloponnese (Gulf of Corinth) is a doubtful case in spite of adequate field investigation and the first scientific report in Greece compiled by Schmidt (1867; Fig. A1; see also Richter, 1958). The 1861 event was associated with 13–15-km-long well-mapped surface rupture (Schmidt, 1867) and it possibly extends offshore eastward into the Gulf of Corinth (Fig. A1) This rupture was concentrated along the prominent fault scarp of the Helike Fault (ca. 26 km long), that is the co-seismic ruptures of the 1861 event did not re-

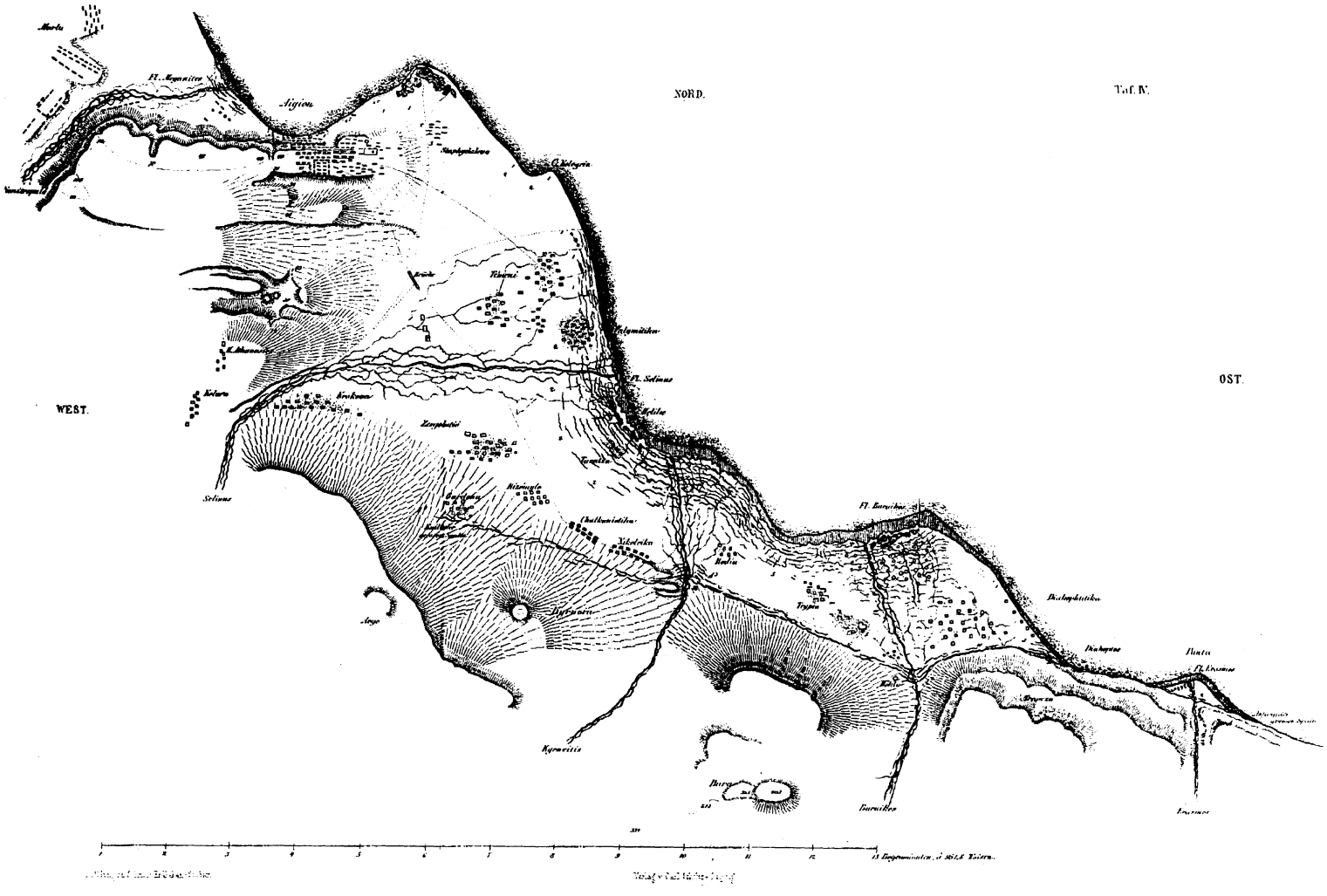


Fig. A1. Reproduction of the original map of Schmidt (1867) relative to the macroseismic area of the 1861 Valimitica (Northern Peloponnese) earthquake.

activated the whole length of the neotectonic structure as it is geologically defined. At least three strong earthquakes are possibly related to a reactivation of this fault, that is the 373 BC, the 1402 AD and the 1861, which is the only certain event. In contrast, palaeoseismological researches (Koukouvelas et al., 2001; Pavlides et al., 2001b) have shown the occurrence of a morphogenic earthquake during the Roman period (23 AD?), while another event missing in the historical catalogues is associated to the eastern segment of the fault. The 1861 earthquake has been also detected during the palaeoseismological investigation (Pantosti et al., 2002; Pavlides et al., in press).

### *A.5. 1870 Fokis (no. 6)*

There is historical evidence, described mainly by Ambraseys and Pantelopoulos (1989), that the earthquakes of 1870 (Arachova-Delphi, Central Greece; Ms between 5.3 and 6.7 ± 0.3) are associated with faulting. According to these authors, a 5- to 6-km-long series of surface breaks terminating with 2-m throw is the only evidence for possible co-seismic surface ruptures. But following their description and fragments of the original sources, it is concluded that the ground ruptures were extending E–W for a distance of about 18–20 km, not continuously but following more or less the 28-km-long series of the known active fault 'segments' or 'strands' crossing the area. The maximum vertical displacement along the contact between bedrock and alluvium was probably no more than 1 m. All these features have been considered as induced by the co-seismic surface rupture process. Rock falls, landslides and liquefaction phenomena were also observed and described.

### *A.6. 1894 Atalanti (no. 8)*

Two major shocks struck the Lokris region in 1894 (Central Greece) near Martino and Atalanti villages (Fig. A2). This double event is the second case of scientifically described and interpreted faulting phenomena associated to an earthquake (Skufos, 1894; see also Mitsopoulos, 1895; Richter, 1958).

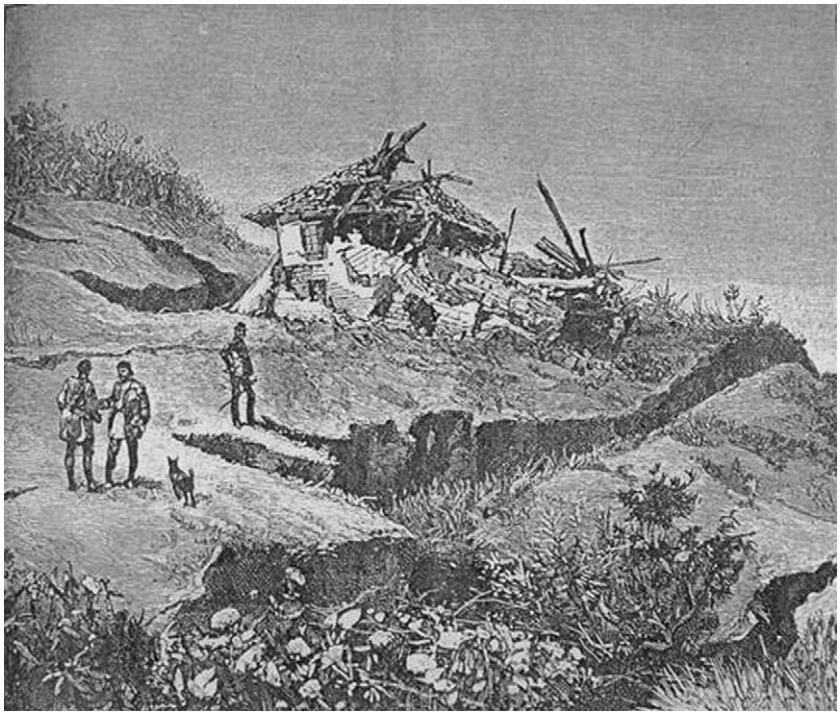

Fig. A2. Ground deformation caused by the 1894 Atalanti (Lokrida, Central Greece) earthquake in a picture of the time (from Science Illustr. 7 July 1894, p. 105; J.T. Kozak collection no. KZ484).

Several authors (Richter, 1958; Ambraseys and Jackson, 1990; Papazachos and Papazachou, 1997) accept a total rupture length of about 50–60 km, including the gaps between the numerous ground ruptures. However, after re-evaluating the original manuscript of Skufos (1894), we can tentatively separate the effects of the two events and assume that the seismic cracks of the major shock (April 27, 1894) were distributed for a distance less than 30 km (possibly 25 km) or according to Pantosti et al. (2001) 32 km. The total length of the geological fault recognised in the field is about 35 km. Other surface deformation described further NNW towards Agios Konstantinos is likely to be landslides and other gravitational phenomena rather than *seismogenetic* features (i.e., seismically induced). This interpretation seems to be confirmed by our preliminary results of a palaeoseismological investigation. The complex geometry of the fault with bends and step-overs appears to be controlled by inherited structures (Ganas et al., 1998; Pantosti et al., 2001). The displacement along most of the fissures was only 10 to 20 cm or at most 30 cm when crossing the Cretaceous limestone. The coastal subsidence in the immediate hanging-wall of the fault is possibly 0.5 m at a site located 300 m north of the fault (Cundy et al., 2000). Although the down-throw values of 1 to 1.5 m, which occurred within the alluvial plain, were possibly due to a combined effect of direct co-seismic ruptures and compaction processes, we accept them as the maximum displacement observed for this event. In contrast, higher values of displacement up to 2 or even 4 m, have been here attributed to gravitational phenomena.

### *A.7. 1904 Kresna (no. 10)*

The 1904 Kresna earthquake (Struma Valley, SW Bulgaria) is believed to be the strongest event in the European–Mediterranean region of the 20th century, because it is referred to in the literature as $M = 7.8$ with 40 to 60 km long fault traces (Papazachos and Papazachou, 1997). However, a recent re-evaluation accepts a lower magnitude of 7.1 to 7.2 for the main shock and Ms = 6.8 to 6.9 for the main foreshock, which occurred 23 min earlier (Eck and Stoyanov, 1995; Ranguelov et al., 2000; Ambraseys, 2001, Meyer et al., 2002). The earthquake caused extensive damage, while the geological effects included landslides, rock-falls, liquefaction, changes in spring water and stream flow and surface faulting (Hoernes, 1904; Grigorova and Palieva, 1968). The main fault trace oriented E–W to NE–SW has been estimated 16–17 km long by Dobrev and Avramova-Tacheva (2000), while Ambraseys and Jackson (1998) suggest a 25-km-long rupture. Forty kilometres west of Krupnik in the modern Former Yugoslavian Republic of Macedonia (FYROM) territory, damages and surface fault ruptures have been described similar to those observed in Krupnik (10–12-km-long fault traces). However, in between there is no sufficient information for co-seismic ruptures, while it is also difficult to distinguish between the ground effects induced by the foreshock and those associated by the main shock. The reported maximum vertical displacement is as much as 1.5 to 2 m at Krupnik confirmed by recent field surveys (Shanov and Pavlides, 2000; Meyer et al., 2002; Shanov, personal communication). As a consequence of the seismogenetic scarps, the Struma River was dammed at Krupnik (Zagorchev, 1995; Shanov et al., 1999; Shanov and Pavlides, 2000).

### *A.8. 1905 Shkodra (no. 11)*

The Shkodra (Scutari) earthquake of June 1, 1905 was a strong event of Northern Albania followed by a great number of aftershocks. Data from Mihajlovic (1951) and from Koçiaj and Sulstarova (1980, including newspapers and the Tirana Seismological Centre collection) indicate a maximum epicentral intensity of IX degrees of the MSK-64 scale, while Papazachos and Papazachou (1997) attribute a magnitude of 6.6. Ground fissures occurred associated with a neotectonic fault extending for about 10 km in a NE–SW direction (Koçiaj and Sulstarova, 1980). The maximum vertical displacement observed along the preserved fault scarps is about 1 m.

### *A.9. 1928 Plovdiv (nos. 12 and 13)*

On April 14 and 18, 1928, two earthquakes (Ms = 6.8 and 7.0) hit central Bulgaria (Plovdiv), giving clear long co-seismic surface ruptures. Geologically, the shocked region is a depression in a mountainous area bounded by two antithetic normal faults. The distance between the two structures is

about 15 km. According to Jankof (1945), during the foreshock of April 14, the northern fault segment was re-activated for about 38 km, while during the second earthquake, which occurred on April 18, the southern one was re-activated, showing a surface rupture of about 50 km and a maximum displacement of about 3 m (Richter, 1958; Glavcheva, 1984; Eck and Stoyanov, 1995; Papazachos and Papazachou, 1997).

*A.10. 1932 Ierissos (no. 14)*

On September 26, 1932, a strong earthquake occurred in Eastern Chalkidiki peninsula, Northern Greece, with a magnitude equal to 6.9, according to Ambraseys and Jackson (1998) or 7.0 following Papazachos and Papazachou (1997). The earthquake destroyed the town of Ierissos and the nearby villages (Stratoni, Stratoniki and Stagira) with IX–X intensities of the Sieberg–Mercalli scale. Strong aftershocks with magnitudes up to 6.3 followed. According to the original descriptions (Floras, 1933; Maravelakis, 1933; Georgalas and Galanopoulos, 1953) and our original observations, the co-seismic fault traces have been entirely mapped at scale 1:5000 (Pavlides and Tranos, 1991; unpublished map of the TVX Mining). These structures are represented by E–W- to ESE–WNW-trending segments generally following the neotectonic fault. The maximum displacement shows vertical offsets of 1–2 m, though this value was probably enlarged by shaking and gravitational forces with displacements as high as 4 m. Accurate measurements by the Greek Navy (Hydrographic Service, 1932) have shown 1–1.40 m maximum co-seismic displacement at the Ierissos Gulf, close to fault, or according to Floras (1933) measurements of about 1.80 m. We thus accept this last value as the maximum co-seismic displacement (MVD) and an average displacement (AD) of 0.30 m (see also Ambraseys and Jackson, 1998; Pavlides et al., 2000). The total observed surface rupture length was less than 20 km. The seismogenic structure is represented by the E–W-trending Stratoni Fault which extends for more than 15 km on land and according to bathymetric maps possibly 10–20 km eastwards offshore. It is a typical normal active fault, striking N110°, dipping 60–50° SSW and with almost dip-slip striations (70–80° pitch). Palaeostress analysis on slickensides shows that the regional stress pattern is extensional with N–S trending $\sigma_3$ principal axis. The estimated geological fault length is certainly larger than 25 km and possibly up to 35–40 km.

*A.11. 1941 Larissa (no. 15)*

The macroseismically determined epicentre of the 1941, Larissa earthquake ($M = 6.1$) has been tentatively located some kilometres NE of Larissa town, Thessaly plain (Galanopoulos, 1950). Indeed, in the bedrock NE of Larissa a series of ruptures striking NW–SE have been reported after the earthquake and tentatively assumed to be the seismogenic fault by Ambraseys and Jackson (1990). However, numerous ground fractures also occurred for about 7 km along the recently recognised E–W-trending Asmaki Fault (Caputo et al., 1994), which is characterised by about 1-m-high morphological scarp. Accordingly, we assume this tectonic structure to be the seismogenic fault associated to the 1941 Larissa earthquake.

*A.12. 1953 Sousaki (no. 17)*

The earthquake occurred on September 5, 1953 (Ms = 5.6) shaking Loutraki, the Isthmus of Corinth and the Megara peninsula (Central Greece). The broader Sousaki region is affected by E–W to NW–SE normal faults, but their geomorphologic expression is not well developed. The co-seismic rupture was nearly 3 km long showing a maximum displacement of about 8 cm. The surface ruptures were not clearly associated to any known or typical neotectonic fault. It possibly extends offshore into the sea (Saronikos Gulf), so the geological fault could be considered slightly longer than the co-seismic surface ruptures (Stiros, 1995).

*A.13. 1954 Sophades (no. 18)*

Another important earthquake, which occurred in the western Thessaly plain (Central Greece), is the April 30, 1954 Sophades event. Papazachos and Papazachou (1997) propose a magnitude of 7.0, but following a systematic revision of the Aegean earthquake magnitudes, Ambraseys and Jackson (1990) suggest an Ms of 6.7. According of Papastamatiou and Mouyaris (1986b), the surface rupture was only 2–3 km. However, a reappraisal of the original field notes of

Giannis Papastamatiou (in Papastamatiou and Mouyaris, 1986a) allowed to infer about 30 km of co-seismic fault trace (Ambraseys and Jackson, 1990; Pavlides, 1993). Minor fissures were also reported within the alluvial plain, but they are probably unrelated to the seismogenic structure. Further morphotectonic scarplets possibly associated with the earthquake were also re-examined in the field (Caputo, 1990; Caputo and Pavlides, 1993; Pavlides, 1993) confirming a probable total length of about 30 km including long gaps between the various fissures.

### *A.14. 1957 Velestino (no. 20)*

This is an insufficiently studied event, which occurred on March 8, 1957 with a magnitude Ms = 6.5. The only seismotectonic information comes from Ambraseys and Jackson (1990) reporting some breaks for more than 1 km, crossing the carbonate bedrock, and about 500-m-long cracks in Quaternary alluvium deposits. Taking into account the referred dislocation of the railway further East, a total length of more than 8 km is a more realistic value. The occurrence of a major morphotectonic escarpment and other structural data suggest the existence of a seismogenic structure possibly longer than 28 km (Caputo, 1990, 1995; Caputo and Pavlides, 1993).

### *A.15. 1966 Megalopolis (no. 23)*

Ambraseys (1967) describes two strong earthquakes in Peloponnese (April 5, 1965; Ms = 6.0 and September 1, 1966; Ms = 5.5) along the Neogene–Quaternary Megalopolis valley. According to Papazachos and Papazachou (1997) magnitudes are 6.1 and 6.0, respectively. In the epicentral area, almost all shale-sandstone formations showed signs of co-seismic movements, while of equal importance were landslide phenomena. Ambraseys (1967) shows a line of the most intense ground movements within alluvial deposits about 20 km long, but not clear evidence of faulting. In the present work, we accept the fault parameters and magnitude of Ambraseys and Jackson (1998).

### *A.16. 1967 Dibra (no. 24)*

The November 30, 1967 earthquake (Ms = 6.6) is a typical event associated to surface ruptures occurred in a NE–SW direction with a vertical displacement of 0.5 m and a total length of 16 km (Sulstarova and Koç iaj, 1980). It was an oblique-slip fault with significant normal component and secondary right-lateral strike-slip component. In Tagari (1993), are included field photographs of the ruptures, while some of them have been studied from the Tirana Seismological Centre collection. The seismogenic fault segment, about 23 km long, belongs to the Elbasan Fault (Vlora–Elbasan–Dibra) which is associated in the literature to a right-lateral strike-slip kinematics. Nevertheless, our field observations and measurements show a prevailingly normal component of motion along this structure.

### *A.17. 1978 Thessaloniki (no. 28)*

Co-seismic fault traces (length and displacement) for the Thessaloniki June 20, 1978 (Ms = 6.5) earthquake have been taken from Papazachos et al. (1979), Carver and Bollinger (1981), Mercier et al. (1979, 1983), Soufleris et al. (1982), Mercier and Carey-Gailhardis (1983), Mountrakis et al. (1983, 1992) and Pavlides (1993, 1996b). The main fault that was reactivated during the 1978 sequence was a 15-km-long structure bordering southwards the Mygdonia Basin, ENE of Thessaloniki. Additional surface ruptures from the same fault line (parallel strands) or linked and antithetic secondary activated structures have a total length of less than 20 km, a maximum co-seismic displacement of 20–25 cm and an average one 5–10 cm; Fig. A3). Other ground deformational features have been neglected because attributed to gravitational and liquefaction phenomena or sympathetic faulting.

### *A.18. 1980 Volos (no. 29)*

For the 1980 Volos event (Ms = 6.5) in southern Thessaly, clear evidence of seismic ruptures (2, 1.2 and 0.2 km) were observed by Papazachos et al. (1983) and Tsatsanifos (1987) soon after the earthquake thus documenting an E–W trending ~ 8 long co-seismic fault (including gaps) and 10 cm of maximum displacement. Unfortunately, only a small segment of the morphogenic structure affected the coastal plain, while the fault continues under the waters of the Pagasitikos Gulf (Caputo, 1996). Consequently, the listed surface

Fig. A3. Surface rupture showing about 10 cm of vertical displacement and about 10 cm of heave that occurred as a consequence of the 1978 Thessaloniki (Macedonia, Northern Greece) earthquake near Stives village.

rupture length of 8 km (Table 2) certainly represents not only a minimum value, but also the MVD could have occurred offshore.

### *A.19. 1981 Alkyonides (nos. 30 and 31)*

Co-seismic fault traces occurred in the eastern Gulf of Corinth during the 1981 seismic sequence. It consisted of three principal shocks, that is those affecting the Alkyonides–Perachora peninsula on February 24 (Ms = 6.7) and on February 25 (Ms = 6.4) and the Kaparelli area on March 3 (Ms = 6.4). In the present research, we considered only those co-seismic features directly associated with pre-existing neotectonic faults either crossing the bedrock or along the bedrock–sediments contact. Other ground deformational features, including displaced recent sediments, have been neglected because attributed to gravitational and liquefaction phenomena or sympathetic faulting. According to Papazachos et al. (1982, 1984), Jackson et al. (1982), King et al. (1985) and IGME (unpublished data) during the first two shocks three segments of surface ruptures occurred. A total length of 12–15 km was observed in the northern Perachora peninsula (Pissia–Alepochori). Dip-slip normal component displacement was ranging from 10 to 80 cm. After the third event (March 4, 1981, Ms = 6.4), continuous co-seismic traces appeared along a typical normal fault, antithetic to the former one, for more than 12 km with a maximum displacement of 70 cm (Fig. A4). They possibly continued southwest under the Corinth Gulf.

Fig. A4. Surface rupture induced by the 1981 Alkyonides (Corinth Gulf, Central Greece) earthquake occurred near Kaparelli village along the Kaparelli Fault and separating loose Quaternary deposits from the carbonate bedrock. Local vertical displacement is about 70 cm, corresponding to the maximum observed value.

*A.20. 1986 Kalamata (no. 32)*

The 1986 Kalamata (southern Peloponnese) seismic sequence with a major shock of Ms = 6.0 (Papazachos and Papazachou, 1997) generated some typical co-seismic traces along a N–S-trending neotectonic fault for 7–8 km. However, surface ruptures have been observed also in a northern segment of the same structure, therefore suggesting a possible total length of about 10 km and a vertical displacement of 10 cm (Papazachos et al., 1988; Mariolakos et al., 1989; Pavlides, 1993).

*A.21. 1995 Kozani–Grevena (no. 33)*

The May 13, 1995 earthquake was an unexpected event occurred in a region of the Western Macedonia (Greece) characterised by very low instrumental seismicity and no historical events reported in the catalogues. The information about the ground deformation has been collected from several papers appeared on a Special Issue of the Journal of Geodynamics ("Results of the May 13, 1995 Kozani–Grevena Earthquake", edited by Pavlides and King, volume 26, issues 2–4, 1998). Further data are from Pavlides et al. (1995), Meyer et al. (1996) and Hatzfeld et al. (1997a,b), Clarke et al. (1997). Continuous typical co-seismic ruptures extend for less than 10 km in Palaeochori–Sarakina villages. Other shorter ruptures occur along the same fault line near the Rymnio village, to the East, and the Feli village, to the West (Mountrakis et al., 1998), while some minor antithetic co-seismic ruptures appeared. The total surface rupture length here considered is thus about 27 to 30 km, while the maximum vertical displacement is 12–20 cm (Fig. A5) and the average less is than 10 cm. They belong to a system of NE–SW trending and NW-dipping normal faults (Servia–Polyfytos–Aliakmon), while during the 1995 event only the western segment was activated (Aliakmon Fault).

*A.22. 1995 Aegion (no. 34)*

Ground deformation parameters for the 1995 Aegion Ms = 6.4 earthquake based on Koukouvelas and Doutsos (1996) and Koukouvelas (1998). The E–W trending and N-dipping Aegion Fault is an active structure about 10 to 12 long, activated for less than 7 km, with an average displacement 0 (heave only) to 1 cm and maximum 6–7 cm. The seismogenic fault of this event is believed to be a low angle deep structure (Bernard et al., 1997).

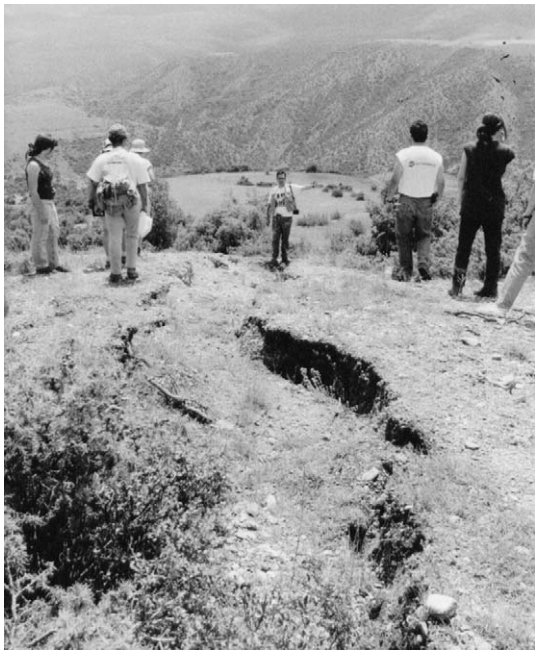

Fig. A5. Surface rupture induced by the 1995 Kozani (Macedonia, Northern Greece) earthquake occurred near Sarakina village and affecting Neogene molasse deposits of the Mesohellenic Trough. Local vertical displacement is about 10 cm.

*A.23. 1999 Athens (no. 36)*

Due to the moderate size of the Athens earthquake (Ms = 5.9), no typical, continuous, co-seismic ruptures were found. This makes difficult to determine which was the seismogenic fault, without taking into account other evidences. Field observations mainly (neotectonic and morphotectonic) undertaken soon after the earthquake, LANDSAT satellite imageries, macroseismic data, focal mechanisms, aftershock distribution and SAR interferometry (Kontoes et al., 2000), identify the seismogenic structure in the Fili (or Phyle) Fault running along the foothills of the Parnitha Mountain, West of Athens (Pavlides et al., 1999, 2002) This structure is less than 15 km long, has a WNW–ESE strike and dips to the SW. The total surface ruptures possibly extends for 8 km, while

aftershock distribution indicate a longer 20–30 km seismogenic structure at depth (Papadopoulos et al., 2000; Pavlides et al., 2002) though its geological surface expression (GFL) is no more than 15 km. The observed maximum vertical displacement is 6 cm, on the bedrock (Jurassic limestone), while the average displacement is less than 1 cm though not necessarily co-seismic.

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