Contents

Foreword

We have tried to present the information in this guide as concisely as possible and not to overwhelm the reader with vast, usually unmanageable amount of data. For this reason, we introduced the entry Key references at the beginning of each chapter, to serve as a rough guideline. We have also tried to make the description of the topics as factual as possible and not to tackle with the innumerable theoretical aspects and possible explanations that may arise after all data and information have been presented. After all, we firmly believe that "heated" and constructive discussions will take place "on site"!

The guide opens with two "orientation" chapters, to make the reader at home with the geodynamic, geological, and neotectonic setting of southern Greece. When more generalized information was needed, we added chapters ("interludes") that acclimate the reader with the topics of the stops to follow. In these interludes information that does not pertain to a particular stop is presented, but it is important for the general understanding of the area to be visited.

The first day will take us to the earthquake-stricken area of Egio, where we will have the chance to discuss various earthquake-related geological effects, together with the overall neotectonic setting of the Gulf of Corinthos. Then, driving along the northwestern coasts of the Peloponnessos, we shall reach Kyllini and then Pyrgos, both of which suffered the blows of Engèlados (the ancient deity responsible for earthquakes in Greece) in the last ten years. The results of diapiric phenomena will also be discussed, in one of the most highly active areas in Europe. Travelling south, we shall admire the tranquillity of an ancient sacred site, Olympia, where the ancient Olympic Games were held.

The second and third days will mostly be spent in the magnificent province of Messinia, both geologically and environmentally speaking. In the second day, we shall examine the neotectonic behaviour of active graben margins and later, the application of neotectonic mapping on a tectonically active area. The morning of the third day will see us in the meizoseismal area of the Kalamata earthquakes of 1986. The topics of this day are various: we shall see the effect and discuss the role of brittle-ductile deformation in neotectonics on Zìmbeli fault; we shall converse on the "obvious" question if it is safe to build on fault surfaces in the picturesque village of Eleohori; and we shall realize the behaviour of a multi-fractured rock mass in an earthquake in Tziròrrema. A bit further to the north, we shall see the application of indirect indicators for the decoding of the neotectonic evolution of a region. The coach will take us then through the centre of the Peloponnessos to its northeast; there lies Kenchreai, an ancient submerged establishment.

After the overnight in Loutraki (whose name means little spa, because of the abundance of mineral water springs) we go onboard to cross the Isthmus canal, a unique experience by all means. Have your cameras ready, because neotectonics is there, everywhere! The bus will pick us up from the western end of the canal and will take us to the Perahora peninsula. All stops there involve the examination of active faults, all of which were reactivated in the 1981 Alkyonides earthquakes.

We have kept the best for the last. This is an enormous engineering project conducted almost 4000 years ago. The protohelladic nation of *Minyes*, aided by their patron Hercules, are there to give us thorough academic lessons on hydraulic engineering!

The evening of the fourth day will see us in Athens, the capital of Greece. Despite its enormousness, there is some underlying beauty waiting for you to discover -if you are not exhausted by then! Even so, pluck up some strength and visit some of the most famous ancient sites in the world that are here especially if you are not due to fly next morning. After all, your hotel is only a stone's throw away from the place where Democracy was born, 2600 years ago!

We sincerely hope that this guide will meet your needs. And, please remember that the best heated scientific discussion can be held over a cool glass of the magnificent Greek wine!

Overview-1

Geology of Peloponnessos and Sterea Hellas: Introduction

Key references: Mariolakos, 1976; Papanikolaou, 1986

Greece forms a very characteristic part of the Alpine System, known as the Hellenic Arc. It represents one of the major mountain chains of the Alpine Alpine-Himalayan System, resulted from the convergence/collision between The Eurasian and the African continental plates.

The morphotectonic direction of the Hellenic Arc in Continental Greece is NNW-SSE, (Fig. 1) bending gradually to E-W between Kythera and Crete. Eventually, the direction becomes NE-SW east of Dodekanissa ("twelve islands") island complex up to Turkey.

The Hellenides comprise a large number of geotectonic units, corresponding to individual nappes; the overall kinematics show a movement directed from the core of the arc in the Aegean Sea towards the periphery, in the Ionian and Libyan Seas.

Two main orogenic cycles have been distinguished in the Hellenides, namely:

- The paleo-Alpine orogeny of Late Jurassic – Early Cretaceous
- The Alpine orogeny, which started in Late Eocene and culminated during Oligocene and Miocene times. However, plate movements with resulting orogenic processes are still active along the present Hellenic Arc and Trench System.

The geotectonic units of the Hellenides can be separated in two groups, namely the internal and the external ones; the former have undergone deformation in both orogenic cycles, while the latter only in the second.

The main geotectonic units distinguished in Peloponnessos and Sterea Hellas and more precisely the area of the field trip are the following:

Parnassos:	Triassic – L. Cretaceous neritic carbonate sequence (interrupted by 3-4 bauxite horizons) and Paleocene – Eocene flysch.
West. Thessalia Beotia:	Continuous sequence from Triassic to Eocene. It is the most internal stratigraphically continuous unit of the Hellenides.

Eastern Greece: It was deformed twice, during the paleo-Alpine and the typical alpine orogeny. It consists of a neritic Triassic –Jurassic sequence (*Subpelagonian unit*) overlain by the obducted ophiolites, the transgressive Upper Cretaceous limestones and the Eocene flysch on top.

Mani: Slightly metamorphosed and intensively deformed isoclinally folded rocks with stratigraphic column similar to that of the Ionian (Triassic-Eocene).

Ionian: Evaporites, neritic carbonates, pelagic limestones and cherts (Triassic-Eocene) and flysch (Upper Eocene-Lower Miocene)

Gavrovo: Neritic carbonate sequence and flysch (Jurassic-Lower Miocene).

Tripolis: Neritic carbonate sequence similar to that of Gavrovo Unit. The base of the stratigraphic column consists of late Paleozoic-Middle Triassic volcanoclastic formation (Tyros beds), slightly metamorphosed to slates-phyllites.

Arna: Metapelites, quartzites and metabasalts (low-grade blueschist metamorphic rocks).

Pindos: Alternation of pelagic limestones, radiolarites, pelites and flysch (Triassic-Eocene)

Concerning the alpine tectonism of the units occurring in Peloponnessos, the following could be mentioned (Papanikolaou, 1984):

The non-metamorphic units of Hellenides are characterized by longitudinal NNW-SSE trending *b*-folds without schistosity, mostly representing flexural-slip folding. The geometry of folding is controlled by the interlayering of competent and incompetent rock units. The rocks probably have not been subjected to load of more than 3-5 km.

The low-grade metamorphic rocks of the Mani sequence in Peloponnessos are characterized by longitudinal NNW-SSE folds with axial-plane schistosity. They have been subjected to a load of about 4-7 km.

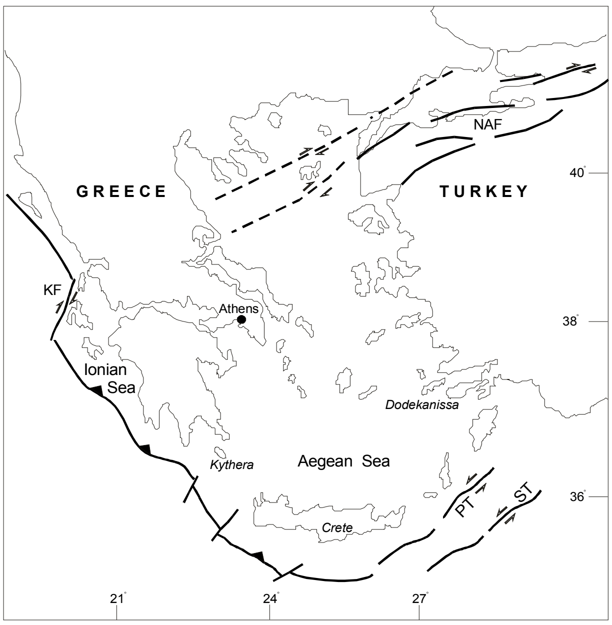


Fig. 1 The Hellenic Arc and Trench System. KF: Kefalonia Fault, NAF: North Anatolian Fault, PT: Plini Trench; ST: Strabo Trench.

Overview-2

Hellenic Territory: Current Geodynamic Regime

Key references: Mariolakos et al., 1985;Mariolakos & Papanikolaou, 1987

The present Hellenic Orogenic Arc is restricted at the southern part of the Hellenic territory, in contrast to all the previous ones, which extended throughout the whole length of the Hellenides.

During the Middle Miocene, a part of the Hellenic arc, still active today, was cut off from the Tethyan chain and since then it followed its own evolution. To the north, this part is bounded by the prolongation of the right-lateral Anatolian fault (Fig. 1). In the northern Aegean region, this fault coincides with the northern limit of the active part of the Hellenic arc, bounding an area termed “Aegean microplate” (McKenzie, 1970; 1972; 1978; Galanopoulos, 1972).

The present geometry of the Hellenic Arc has been developing since the Late Miocene. The back-arc basin and the volcanic arc are restricted in the Aegean plate region.

According to Le Pichon et al. (1981), the present geodynamic regime of the Hellenic arc is characterized by asymmetrical movement; along the Ionian trench, the subduction direction is NE-SW and the regime is pure compression, in accordance with the fault plane solutions while, in the Pliny and Strabo trenches, the direction of movement is composite, featuring a substantial sinistral NNE-SSW horizontal component (Fig. 1). In the back-arc area there are extensional structures also with significant horizontal component of movement.

Many geodynamic models have been proposed for the Hellenic arc and especially for Peloponnessos. These models accept that the latter under extensional stress field, accompanied by graben created by normal faulting in the back arc basin (Ritsema, 1974; McKenzie, 1978; Mercier, 1979; Le Pichon & Angelier; 1979; Dewey & Sengör 1979 and others).

Mariolakos & Papanikolaou (1981a) suggested that marginal fault zones control the configuration of neogene basins (Fig. 2a). These fault zones create an asymmetry to basin morphology and sedimentation. According to them, the Hellenic arc is separated in three large parts (Fig. 2b). In part I the big fault zones have an E – W direction. In part II the direction is

NW - SE and in part III the direction changes to NE - SW. This arrangement shows that only parts II and III have apparent dynamic relation to the Hellenic arc and trench system, while part I has its own peculiarity.

Data on the current deformation pattern of the Hellenic Arc have been provided by numerous researchers, as:

- *in situ* measurements of the stress field in shallow drillings (<10m),(Paquin *et al.*, 1982)
- paleomagnetic investigation of the Neogene and Quaternary sediments (Laj *et al.*, 1982)
- fault plane solutions (McKenzie 1972; 1978, Ritsema, 1974; Drakopoulos & Delibasis, 1982; Papazachos *et al.*, 1984.

Mariolakos & Papanikolaou (1987) combined the results of various geological, seismological and geophysical studies, and proposed a present (active) deformation model (Fig. 3).

Further data that assist in the interpretation of the current deformation regime of the Hellenic territory have been supplied by geodetic measurements (Billiris *et al.*, 1991) and the distribution of earthquake foci (Fig. 4).

More recent investigations from SW Peloponnessos (Messinia) (Mariolakos et al., 1991) showed that the stress field responsible for the neotectonic-active deformation is that of rotational couple (Figs. 5,6) which caused not only brittle but also ductile deformation structures resulting from local transpression and transtension sectors.

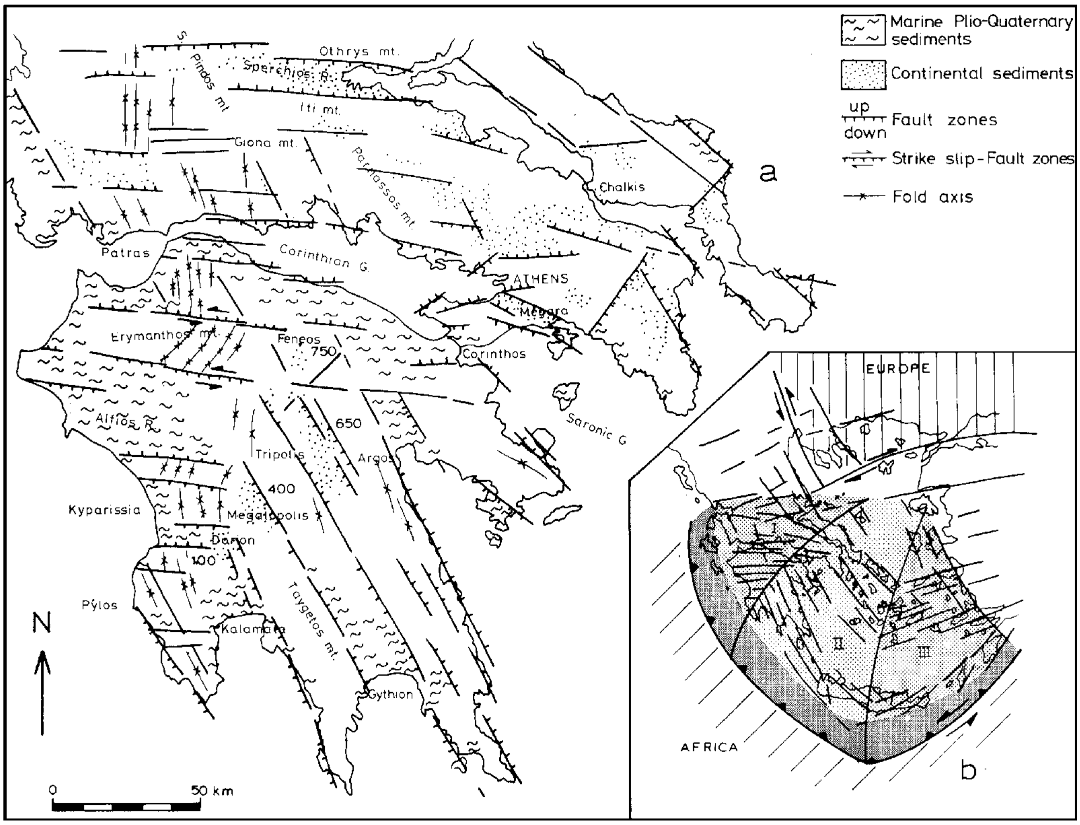


Fig. 2 The neotectonic map of the main marginal fault zones of the post-alpine basins in the southern continental Greece (a) and the neotectonic fault pattern of the Hellenic arc (after Mariolakos *et al.*, 1985).

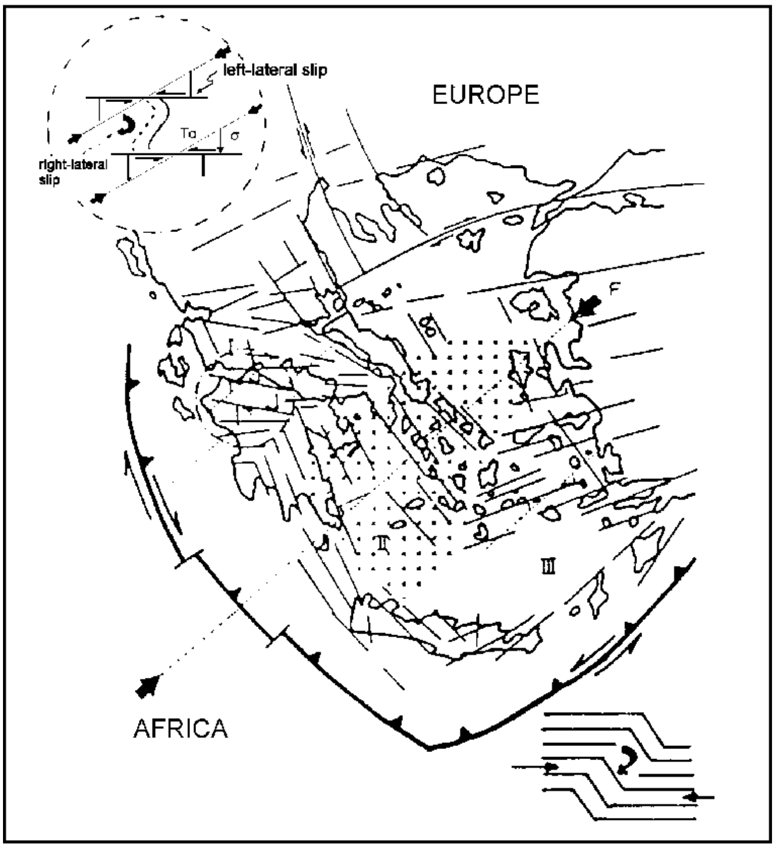


Fig. 3 Analysis of the general stress field F on the faults of the three parts (I, II, III) of Fig. 2.1, into pure (σ) and shearing (Tα) components. In the upper left part of the picture the NW Peloponnessos region is analyzed (under magnification). The faults are developing with a substantial horizontal sinistral component and a dextral rotation of the part in between. The relatively non-seismic Cyclades region is dominated by NW-SE oriented faults. In these faults, the shear component (Tα) of the general stress field is minimum (after Mariolakos & Papanikolaou, 1987).

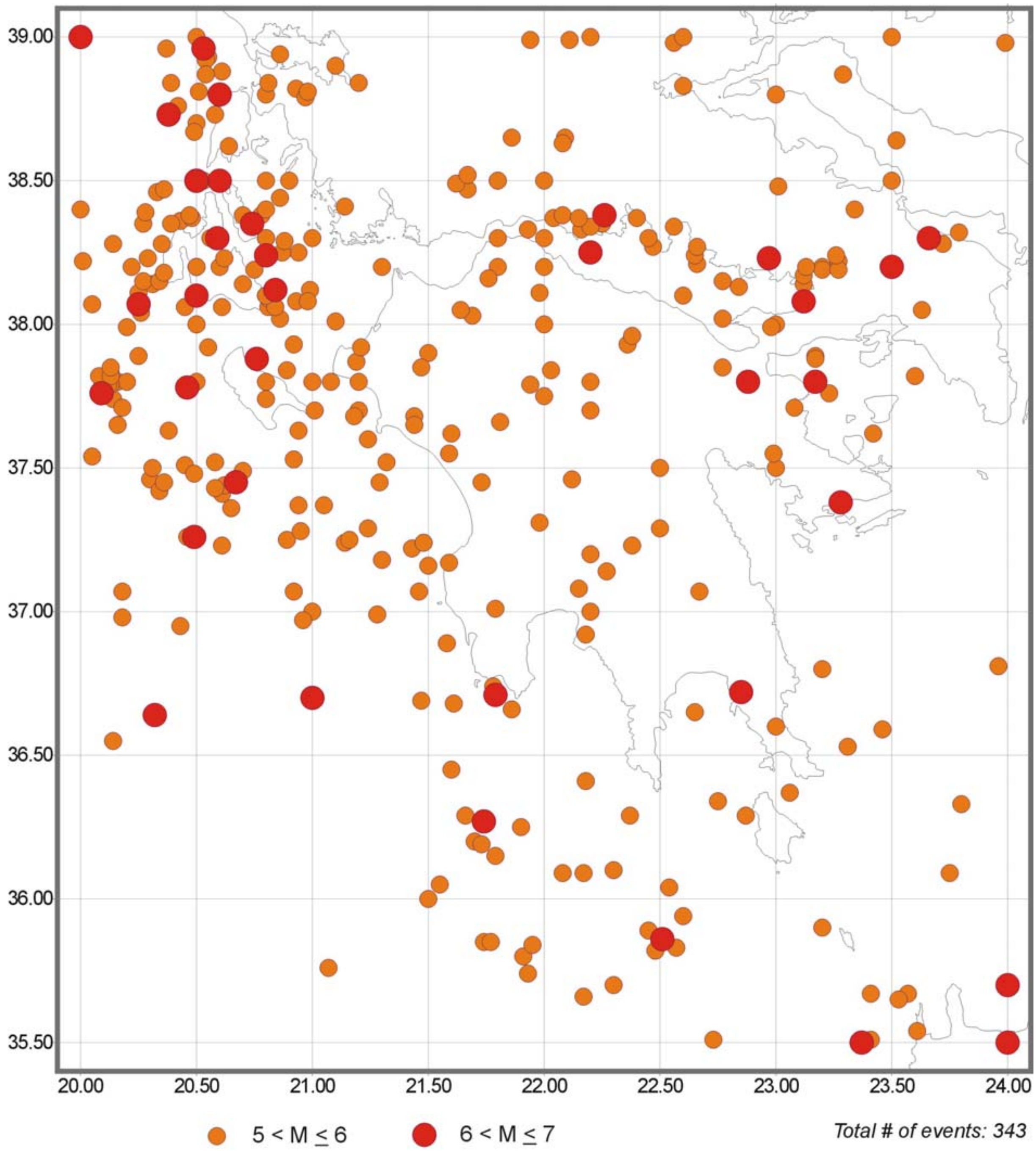


Fig. 4 Earthquake epicentre distribution of the current century (1900-1997) for magnitudes M≥5. Data from the monthly bulletins of NOA and ISC, and from Comninakis & Papazachos (1986).

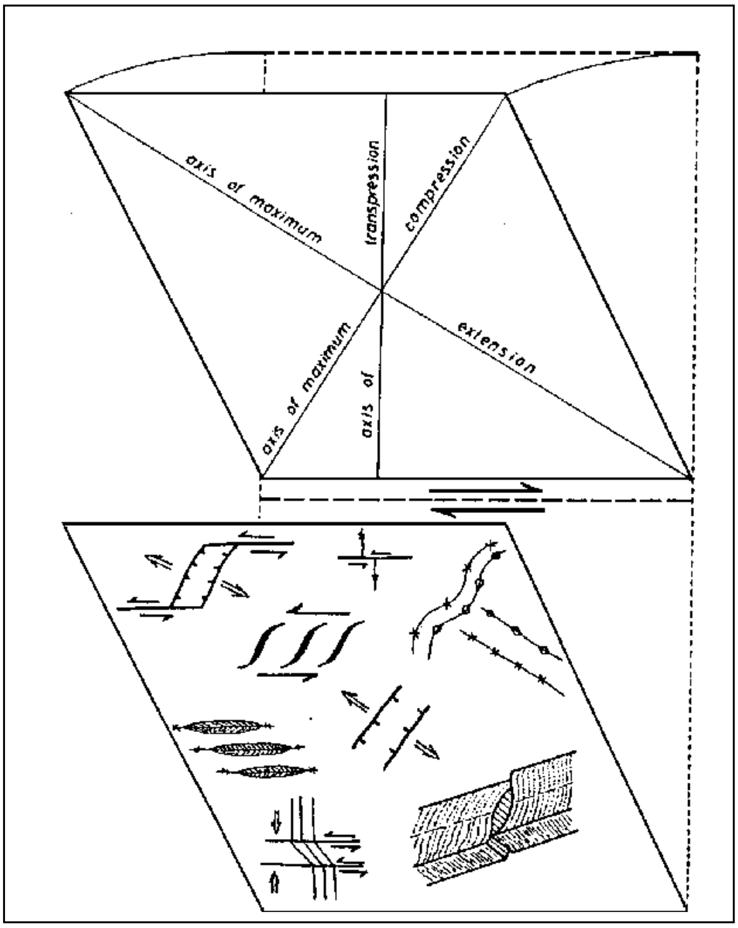


Fig. 5 Torsional deformation and related structures (After Mariolakos *et al.,* 1991a).

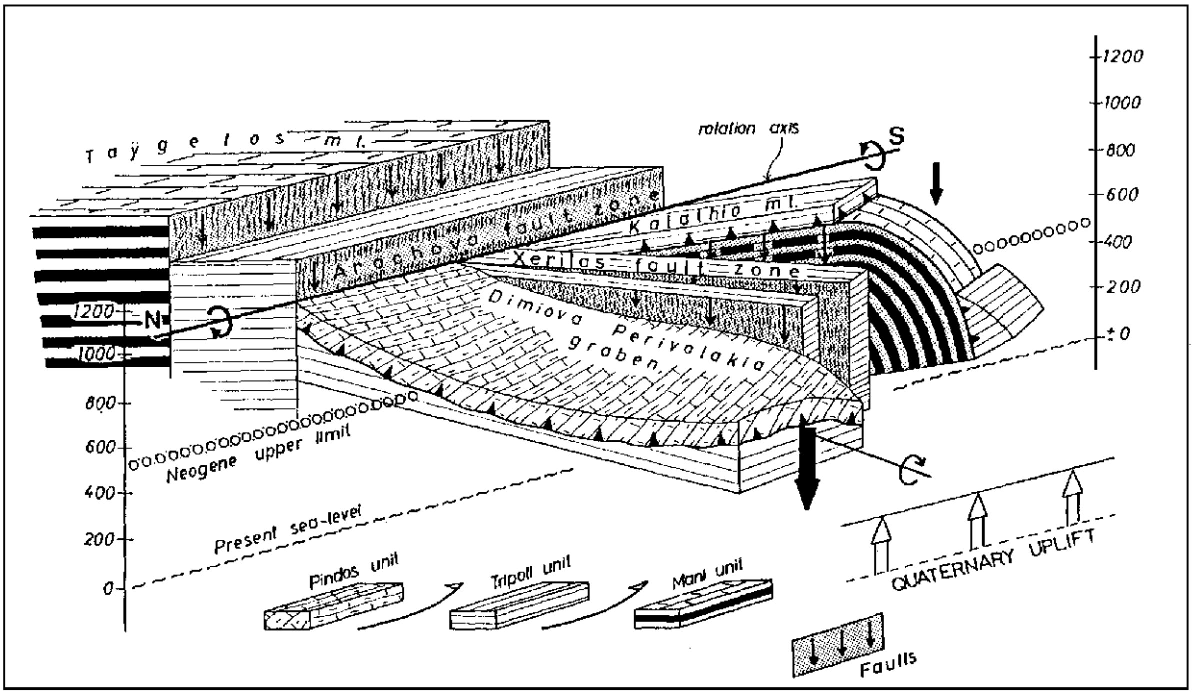


Fig. 6 Torsional deformation pattern between Dimiova-Perivolakia graben and Mt. Kalathion Horst (Messinia, SW Greece) (after Mariolakos *et al.,* 1991a).

Stop A1: Egio

The earthquake of 15 June, 1988 - Related geological effects - Neotectonic implications

Key references: Lekkas et. al., 1996; Mariolakos & Fountoulis, submitted

Central-Western Greece was hit by an Ms=6.1 (National Observatory of Athens, 1995) earthquake on 15 June 1995, at 03:16 local time (Fig. A1). The shock was named "Egio earthquake". The strongest aftershock (Ms=5.4) came 15 minutes later, while the whole aftershock activity lasted for some days. Extensive damage was caused in and around the seaside town of Egio, along the northern coast of the Peloponnessos and to a lesser extent, in Erateini in southern Sterea Hellas. Twenty-six people were buried and killed under the debris of two multi-story buildings, an apartment block in Egio and a hotel in Valimitika, a few kilometers east of the town.

A series of earthquake related surficial effects were produced by the shock (Fig. A2); liquefaction, submarine slides, coastline change (mainly retreat) and seismic fractures and/or ground fissures, together with earthquake-related ground deformation (pop-ups). All these aggravated the damage and caused significant problems along the coastal zone of Egio and at the opposite side of the Gulf, in Erateini.

Geological setting

On the south coast of the Gulf, the surficial outcrops consist of Pliocene and Pleistocene marine deposits and alluvial fan deposits. On the northern coast there are no marine post-alpine sediments; the outcrops consist of mesozoic limestones and flysch, and quaternary fluvial deposits in the depression of Erateini.

Egio

Coastline retreat. It mainly affected the coastal area between the mouths of the Selinountas and Vouraikos. The magnitude of retreat varied between few m. and some tens of metres, as at Cape Trypias.

Seismic Fractures. The seismic fractures that occurred in the vicinity of Egio are located within a zone that develops parallel to the escarpment of Egio. This scarp seems to have been formed by the activity of a (?) normal fault, whose foot-wall consists of loose alluvial and fluvial deposits and its hanging wall comprises L. Pleistocene – Holocene consolidated conglomerates bearing sand and marl intercalations. At first, seismic fractures were observed at the western outskirts of the town and developed mainly in the premises of two factories, the Hellenic Weapon Industry and Kouniniotis. They were a few to some tens of m. away from the escarpment and could be followed for a distance of 2 km. At the western edge of their occurrence (Rododafni) they cut formations corresponding to those of the footwall and their trend was WSW-ENE, while heading eastwards, they cut the recent deposits of Meganeitos river; their trend had shifted to E-W. The fractures cut various constructions founded on these formations, as road paving, supporting walls, embankments and fences. At places, one could observe a small, 2-3 cm. offset (northern-side-down), while locally they were *en echelon* arranged.

Liquefaction. Liquefaction phenomena were located along the coastal region, over an area of 10 km^2, between the mouth Selinountas river and cape Trypias.

Erateini

Seismic fractures. Their orientation bore close relationship to the coastline, and developed either parallel or normal to it. That is to say, in Erateini the two sets of fractures were NW-SW and NE-SW, while in Ag. Spyridon they were N-S and E-W.

Liquefaction. Such effects were very limited in the area.

Shoreline retreat. The magnitude here was quite lower than in the southern coast and did not exceed two or three meters in sites where river and coastal deposits occur. In this point it should be stressed that shoreline displacement was also observed in coastal areas where alpine rocks outcrop.

Pop-up structures. The fracture sets whose strike was normal to the coastline were connected with local transpressional structures and systematically deformed man-made constructions (sidewalk flagstones and sewage pipes). The local maximum compression was always parallel to the shoreline and normal to the local alpine tectonic grain (fold axes and thrusts)

GPS data

The seismogenic deformation, as expressed in the form of crustal displacement, was measured by GPS. Especially for the case of coastline retreat, these data (Bernard *et al.* (in press)) show that it occurred in area where the displacement vectors were normal to the coast (Fig. A1), with direction away from it, while no coastline retreat took place where the displacement vector magnitude was significant, but its direction was parallel to the coast. Furthermore, the measured N-S horizontal displacement is very clear was accompanied by a significant amount of NW-SE displacement (**Fig. A1**).

Offshore data

Papanikolaou et al. (1997) reporting the results of the three oceanographic surveys confirm both the occurrence of a number of submarine slides and a series of mostly en echelon E-W trending faults that bore signs of recent activity. They also suggested that these are gravity faults that create distinct escarpments on the shallow platform of the Gulf.

Thus, attempting to account for active deformation in the area we need to take the following into account:

- Bernard et al. (in press) suggested that the shock was caused by a low-angle, N-dipping detachment fault beneath the Gulf.
- Among the 148 events analyzed by Rigo et al. (1996), about 30% of them corresponded to oblique-slip (strike-slip percentage more than 35%), another 37% contained 10-30% strike-slip, and ‘pure’ dip-slip (s.s less than 10%) accounted for 33% of the whole. The same authors report a N 0° to N 10°E trending T axis for the area of western Corinthiakos, while Bernard et al. (in press) give for the 15 June 1995 event a T axis azimuth of N 10°W. These two options may provide adequate explanation for individual faulting, but seem to fail to account for the overall fault and deformation pattern in and around the Gulf, where the dominant trend of active faults is E-W, them being arranged in a right-stepping en echelon pattern, both along the northern and the southern margins of the Gulf (Papanikolaou et al., 1997). Most of the aftershock epicenters were located west of the main shock epicenter, forming zones more or less parallel to the coastline between Cape Psaromyta and Agios Spyridon, the alpine structural grain and the submarine fault zones.
- Apart from the observed seismic fractures, no sign of seismic faulting was observed either north (Erateini) or south (Egio) of the Gulf.
- South of the Gulf, horst-and-graben structures are present; however these are absent in the north.
- The described pop up like structures in Eratini and Agios Spyridon villages have the same strike (geometry) with the existing tectonic features (fold axes, thrusts).
- On the contrary, the measured with GPS subsidence of the Cape Psaromyta, in which only thin-bedded limestones occur, can be explained by the tectonic deformation of the area.

The suggestion a low-angle detachment in depth, coupled with the right-stepping en echelon pattern of active faults, and the percentage of oblique-slip focal mechanism solutions (Rigo *et a*l. 1996, Mariolakos & Papanikolaou 1987), guide us to the thought that the overall tectonic regime is other than that of pure extension.

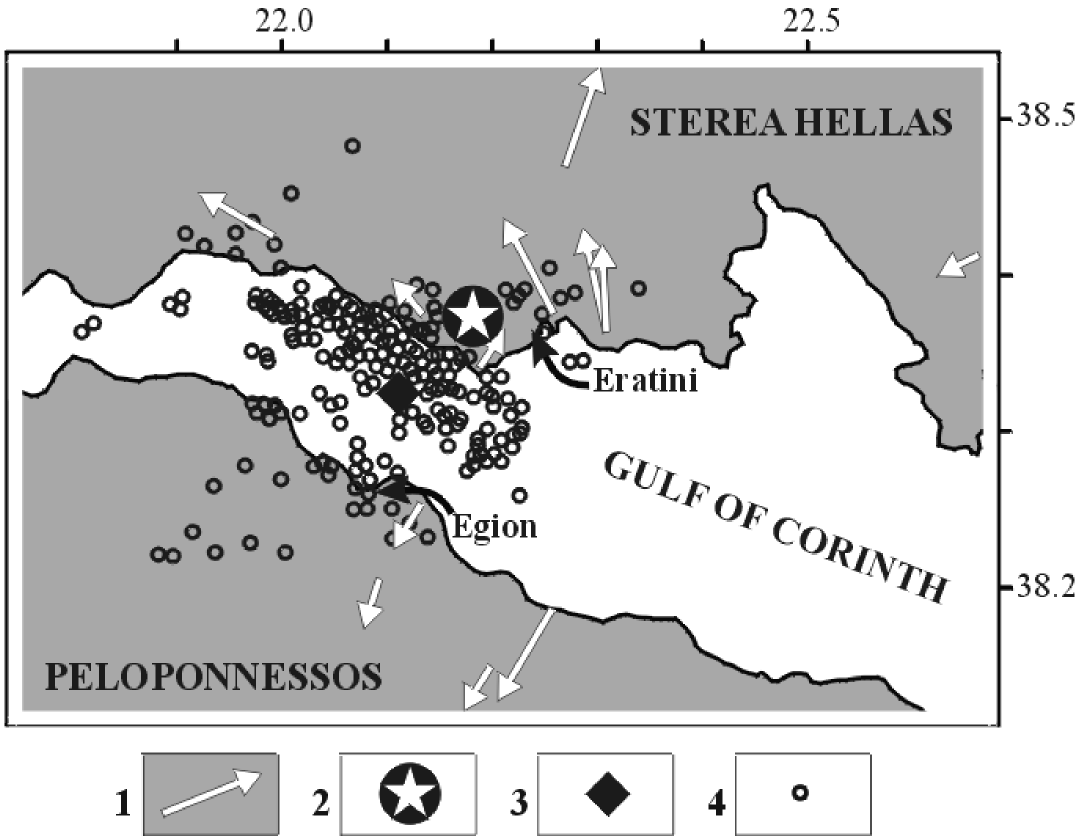
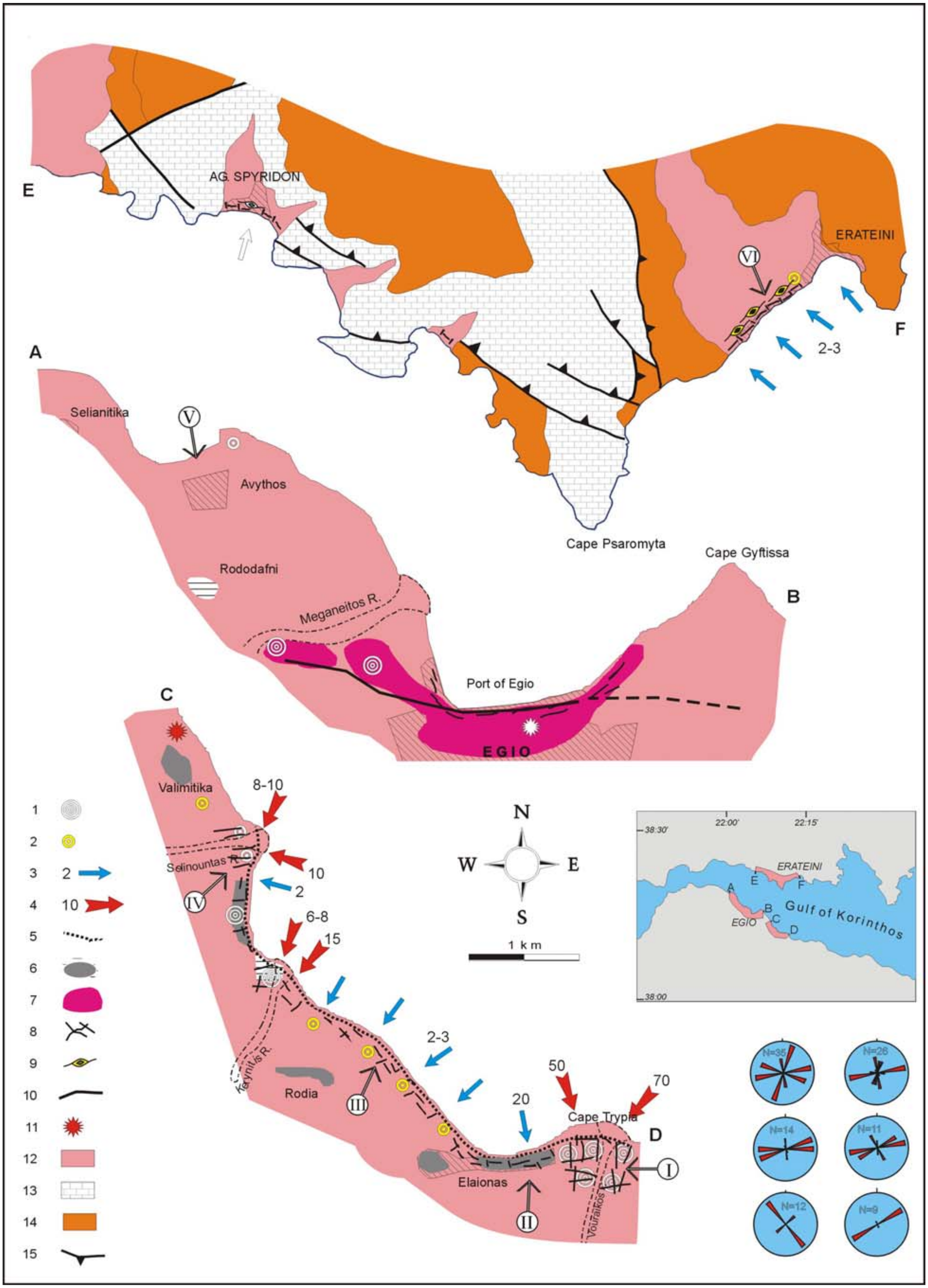


Fig.A1 Earthquake activity (15-6-1995 to 17-7-1955) in Western Gulf of Corinthos (data from NOA, 1995) and the observed horizontal displacement for the 15-6-1995 Egion earthquake (Bernard *et al.*, 1996). 1:horizontal displacement, 2:epicenter of the main shock, 3:main shock, 4: aftershocks (From Mariolakos & Fountoulis (subm.)).

Fig. A2 (following page). Regional map with the occurrences of the earthquake-related surficial effects. 1. Liquefaction (large), 2. Liquefaction (small), 3. coastline retreat (small magnitude) and retreat estimate, 4. coastline retreat (large magnitude) and retreat estimate in metres, 5. post-earthquake shoreline, 6. minor damage areas, 7. major damage areas, 8. ground fissures, 9. pop-up deformation, 10. earthquake fracture, 11. building collapse, 12. post-Alpine sediments, 13. Alpine carbonates 14 Alpine clastics, 15 Thrust. Latin numerals denote the locations of the rose diagrams of fissures. Modified from Lekkas *et al.*, 1996 and Mariolakos & Fountoulis (subm.).


E
AG. SPYRIDON
ERATEINI
VI
F
2-3
A
Selianitika
V
Avythos
Rododafni
Cape Psaromyta
Cape Gyftissa
B
Meganeitos R.
Port of Egio
EGIO
C
Valimitika
8-10
Selinountas R.
10
IV
2
6-8
15
2-3
Kerynitis R.
Rodia
III
20
50
70
Cape Trypia
D
Elaionas
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II
Vouraikos
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4 10
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38:30'
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ERATEINI
Gulf of Korinthos
EGIO
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N=26
N=14
N=11
N=12
N=9

Stop A2: Kyllini

Post-Tyrrhenian uplift- diapir-related brittle-ductile deformation
The earthquake of Oct. 16, 1988

Key references: Mariolakos et al., 1990; Lekkas et al., 1992

NW Peloponnessos is one of the most seismically active areas of Greece due to the fact that it is close to the Hellenic Trench. Consequently, a large number of seismically active neotectonic fault zones, occur in the straits between Zakynthos Island and Kyllini peninsula. According to historical and instrumental records, numerous destructive earthquakes have taken place in the area since 399 BC, most, if not all, of which were shallow (h<20 km) and had high intensities.

Geological setting

The following alpine formations outcrop in Kyllini peninsula (Fig. A2-1)

- Kastro evaporites: They consist mainly of triassic anydrites and outcrop 500 m. west of Kastro village.
- Kastro limestones: They are white thin-bedded and locally medium-bedded calcarenite cretaceous limestones. They occur mainly at the area of Kastro and north of the Kyllini thermal springs.

Both formations belong to the Ionian Unit.

The post-alpine formations can be distinguished, from the oldest to the youngest, into:

- *Psili Rahi formations*: Pliocene conglomerates consisting of mainly siliceous and secondarily carbonatic pebbles. They usually display by graded bedding.
- *Ligia formation*: It overlies the Psili Rahi conglomerates and consists of clays, marls sands and sandstones and outcrops at the largest part of the area; its age is Pliocene-Early Pleistocene.
- *Glossa formation*: It consists of porous marine calcarenites, approximately a few meters thick, overlying the Ligia formation. The age of the formation is Tyrrhenian.

- *Alluvial deposits*: unconsolidated sediments that occur all over the plain.
- *Coastal deposits*: unconsolidated fine (sand) and coarse material (pebbles), which occur along the coast of the Kyllini peninsula.
- *Beachrocks*: cohesive intertidal sandstones.
- *Talus scree and debris*: cohesive or unconsolidated scree, which occur north of the Kyllini thermal springs and around the Kastro village.
- *Dunes*: They consist of unconsolidated sand and they occur along the largest part of southern and northern coast of the peninsula.

Neotectonic Macrostructures

The neotectonic structure of NW Peloponnessos is characterized by the occurrence of graben and horsts bounded by fault zones, which in some cases are visible or concealed (Fig. A2-2). The fault zones have various strikes; the main ones are N-S and NE-SW.

The main characteristic neotectonic macrostructure is the Pyrgos graben, which has been filled with post-alpine sediments. Within the graben, several smaller tectonic blocks (second-order macrostructures) have developed (Fig. A2-3). All these second order neotectonic structures, although their kinematic evolution has not been uniform, are considered to be the result of the same stress field. Some of these are characterized by the occurrence of evaporites and diapiric phenomena, as in the area of Kastro).

Diapiric-related structures

Diapiric movements in the area have caused a suite of neotectonic deformations, mainly exhibited in the form of a large, N-S anticline, whose axis coincides with the main outcrop of the evaporites. This doming is accompanied by significant N-S faulting and has exhumed the alpine formations that underlie the post-alpine deposits. Doming has led to significant drag in the beds of Glossa formation.

Kyllini earthquake

Kyllini earthquake occurred on October 16, 1988; its magnitude was Ms=5.5, and caused extensive damage in Vartholomio and Kyllini More than 1500 buildings were destroyed, while 4.500 buildings were damaged (Max intensity I_o=VII-VIII (MM) while at the southeastern part of Zakynthos the intensity was I_o=V-VI. The epicenter was located in the straits between the Zakynthos and Kyllini. Evaluation of the damage reports by the authorities, showed that the maximum values were observed along an E-W direction.

The Kyllini earthquake caused severe damage to the buildings as well as several other geological effects as fault reactivation (seismic faults), ground ruptures rockfalls, shoreline displacement, liquefaction, sandblows, etc. (Fig. A2-4).

The damage was uniform, both for old and new buildings and for high and low ones. The percentage of severely damaged buildings for each village is dependent on the geological factors and geotechnical factors, namely foundation soil and neotectonic deformation. In many cases, the damage was directly attributable to seismic fractures or to fault reactivation. For instance, in Kastro most of the destroyed buildings lay on the trace of the reactivated fault.

Other factors that may have controlled the damage distribution are the shallow water table in the area and the outcrops of formations prone to liquefaction.

Reactivated faults:

During the earthquakes of October 16, 1988 two faults south-southeastern of Kastro village were reactivated (Fig. A2-4). The fault surfaces bore signs of recent activity, with a throw of 5-20 cm and striated west-dipping slickensides. A narrow band of cataclasite has been formed along the fault. Also, numerous rockfalls occurred along the fault traces. The N-S strike of the reactivated faults follows the neotectonic grain of the major area.

Ground fissures were observed at several sites in the major area of Kyllini peninsula (Fig. A2-4). The most interesting ones have been observed and studied at Bouka (Mariolakos *et al.*, 1991), at Trypito, Paleokastro and Bytineika (Lekkas *et al.*, 1990). Liquefaction and sandblows were also observed in areas very close to the coast (Fig. A2-4).

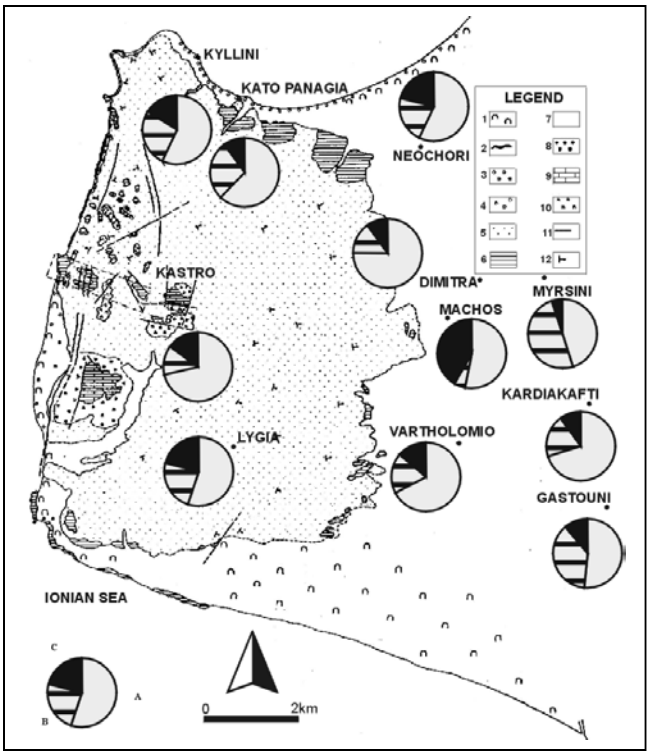


Fig. A2-1 Geological sketch map of Kyllini peninsula demonstrating the percentage of the constructions of each village that were left intact (A), were moderately damaged (B), were totally destroyed (c) (1. Dunes, 2. Scree, 3. Beachrocks, 4. Coastal deposits, 5. Alluvial deposits, 6. Glossa sandstones formation, 7. Lygia formation, 8. Psili Rachi conglomerates formation, 9. Kastro limestones, 10. Evaporites). (After Lekkas *et al.*, 1990).

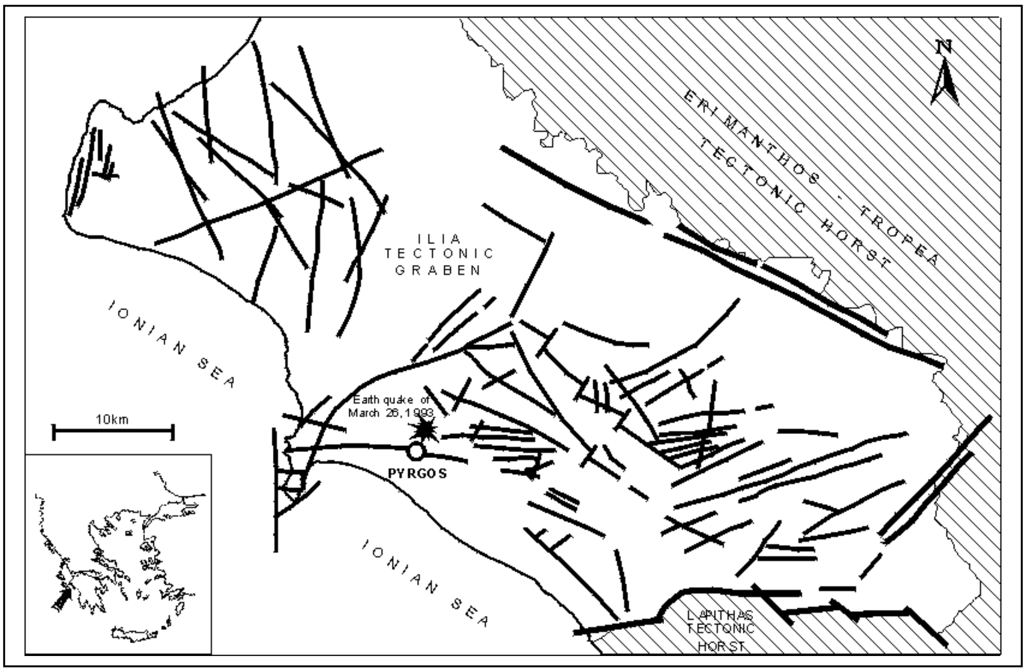


Fig. A2-2. The 1st order neotectonic macrostructures of NW Peloponnessos (after Lekkas *et al.*, 1992).

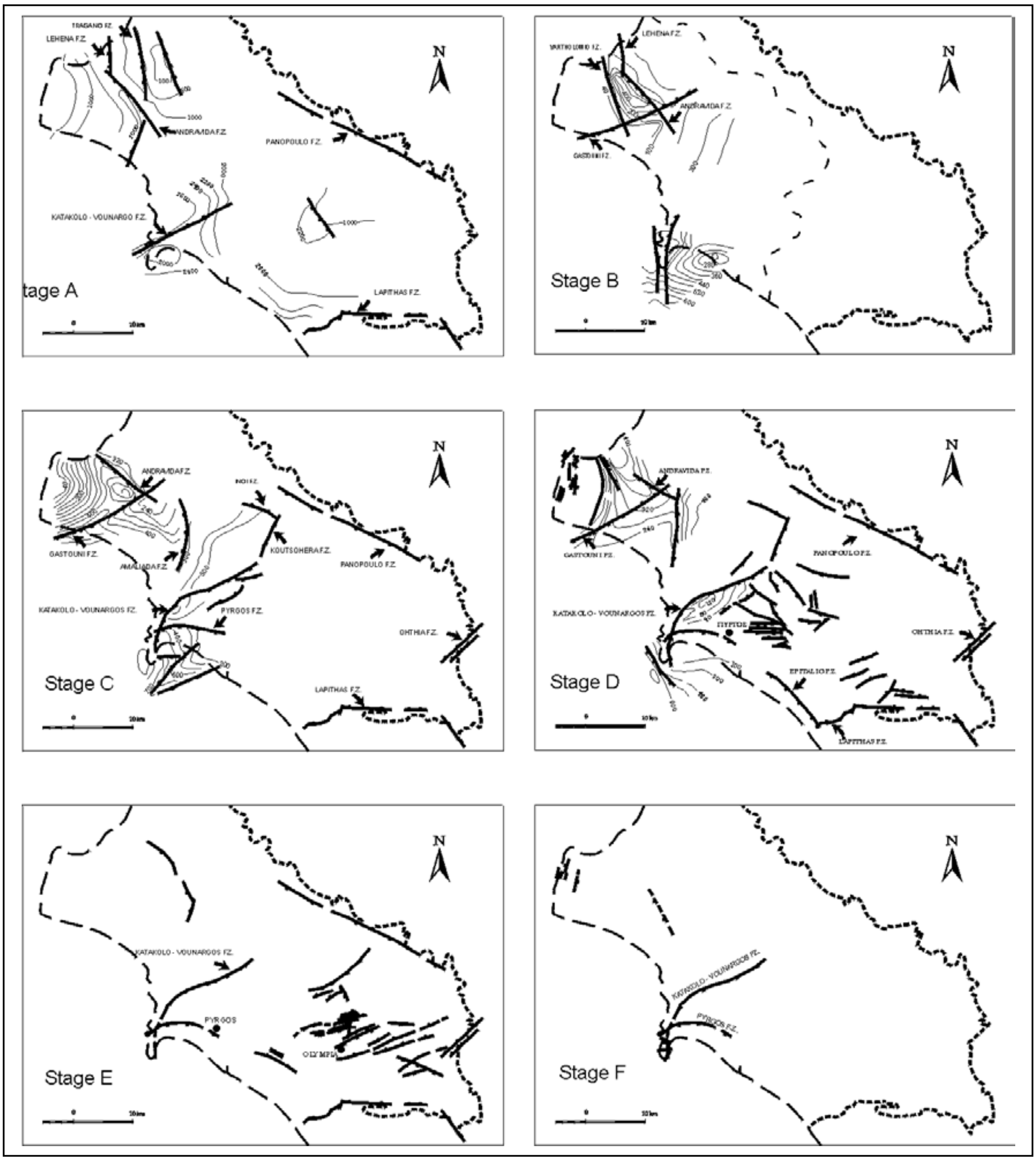


Fig. A2-3 The main stages (A, B, C, D, E, F) of the neotectonic evolution of Pyrgos graben (after Lekkas *et al*., 1992).

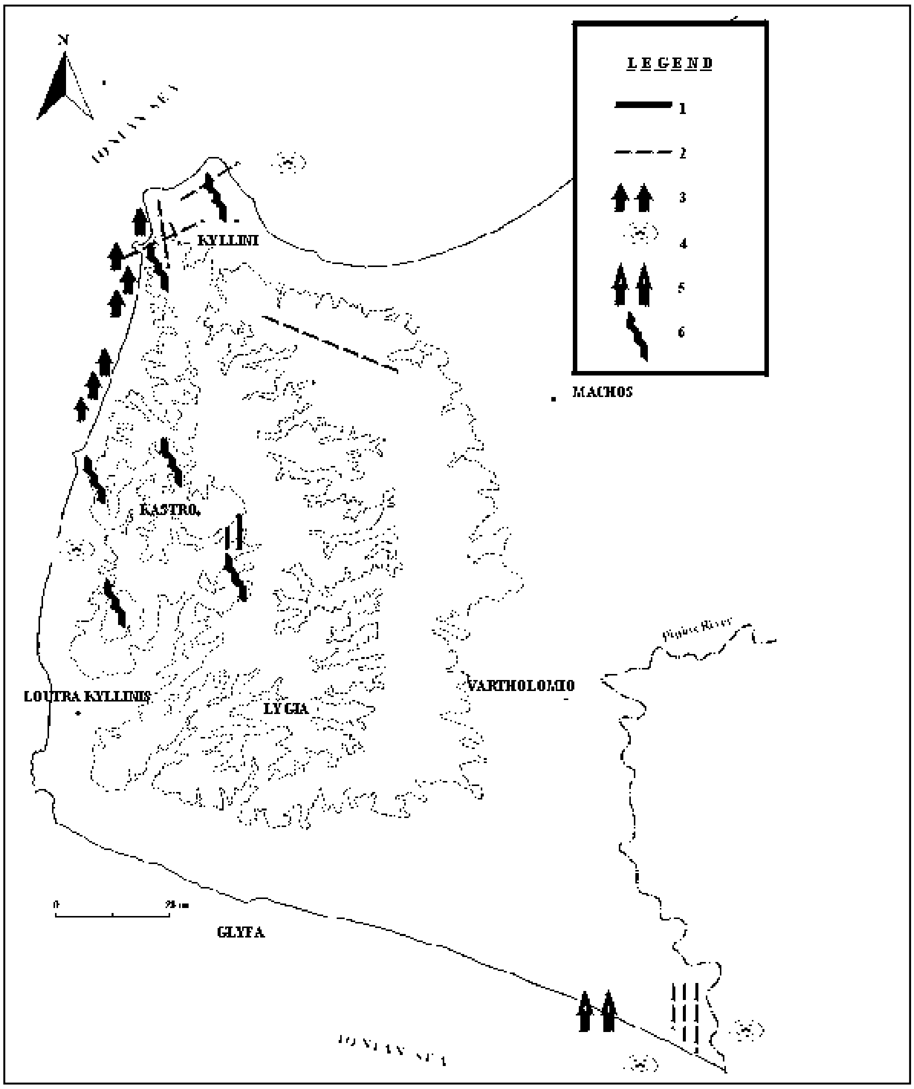


Fig. A2-4 Synthetic map with the macroseismic observations at Kyllini peninsula (1. earthquake fracture, 2. Ground fissure, 6. Rockfalls, 3. Shoreline displacements, 4. Liquefaction, 5. Sandblows (after Lekkas *et al*. 1990).

Stop A3: Pyrgos

Geological - geotechnical conditions of Pyrgos
The Pyrgos earthquake (26 March 1993)

Key References: Lekkas et al., 1992; 1994

Geological setting

The graben of Pyrgos, which extends over a large area, is characterized by complex lithostratigraphical structure and by neotectonic deformation, which has been particularly intense during Holocene (Lekkas *et al*., 1992; 1994).

The outcrops in the broader Pyrgos area consists of the following:

Recent Marsh Deposits. They develop surficially over the other formations at the flat plane part of the area and overlie mainly the alluvial formations, although the contact between them is not well defined very clear. The most important outcrops overlay are north-west of Lapato and east of Aghios Athanasios quarters. They consist of alternations of brown-grayish-brown clays, gray, blue-gray clay silts, clay and silty sands containing abundant organic remnants. Total thickness is up to 5 meters.

Alluvial Formations. They crop out in the flat area of the town of Pyrgos and overlay unconformably the older formations. They comprise of brown to gray soft clays with irregular (both in vertical and lateral sense) intercalations of brown silt and gray-brown sand. They contain numerous floral remnants as well as coarser material (gravel and pebbles). Their thickness does not exceed 12 m.

Erymanthos Formation. It outcrops over a limited area, mainly at Kokkinohoma, Kouvelos and at the Aghioi Pantes cemetery. It is a fossil outcrop of the Pleistocene Erymanthos formation, which comprises mainly conglomerates of terrestrial origins connected with a red-siliceous fine unconsolidated formation. It corresponds to a huge paleo-talus cone with frequent lateral diversification and covers a large part of the graben of Pyrgos. In the city of Pyrgos it is represented by red to brown red clays and yellow brown sandy clays, loose horizontal sandy conglomerates and micro-conglomerates. It overlies unconformably the Vounargo formations and its thickness varies form 2 to 8 meters.

Vounargo Formation. It is perhaps the most characteristic formation that outcrops in the graben of Pyrgos. Its age is Plio-Pleistocene, its thickness is up to 600 meters and comprises continuous intercalation of clays, silts, sandstones, sands and marls with constant diversification both in vertical and lateral sense. At the study area it occurs in the urban region and more specifically at the quarters of Aghia Kyriaki, Agios Nikolaos, and Aghios Athanasios. Two members of the formation can be distinguished with different lithological composition and geotechnical properties.

<u>The upper member</u> (Aghios Athanasios) develops at the quarters of Aghios Athanasios, Keramidhaki, Aghios Spyridon and partially at Lapato and practically covers all the lower parts of the uplands of Pyrgos. It comprises mainly yellow-brown cross-bedded sands and yellow silty sands with certain intercalation of yellow-brown sandy clays. Its thickness exceeds 30 meters.

<u>The lower member</u> (Aghios Nikolaos) occurs at the quarters of Aghios Nikolaos and Aghia Kyriakia and covers most of the uplands of the city of Pyrgos. It comprises fossil-bearing gray-blue-grayish marls which alternate with silty, sandy and clay marls, while locally there are thin intercalation of sand, sandy-silt, lignite horizons and yellow-brown sandstones. Its thickness is more than 80 m.

The formations that occur in the area of Pyrgos have undergone neotectonic deformation and are crossed by a number of faults of East-West mean strike that are part of the Katakolo-Pyrgos fault zone. This fault zone was responsible for the earthquakes of March 26, 1993.

More specifically, the elongated outcrop of the lower member of Agios Nikolaos terminates at the F1 (E-W bearing) fault, south of which the upper (Aghios Athanasios) member of the Vounargo formations occurs. It is a normal fault, accompanied by a morphological discontinuity (to the south of the city main square); its throw is at least 50 meters (estimated from morphotectonic features) and eastwards it branches into two faults.

To the north there is an identical picture with the occurrence of the F2 fault (av. strike NE-SW), which also branches into two minor faults to the east. The F2 fault juxtaposes the outcrops of Aghios Athanasios and Aghios Nikolaos members, while it also crosses some outcrops of Erymanthos formation. Its throw is smaller, (20-30 meters) and all along it, seismic fractures caused by the shock of 26 March 1993 were recognized.

Pyrgos earthquake (26 March 1993)

On March 26, 1993 the major area of Pyrgos was affected by an earthquake of magnitude Ms=5.2 (Figs. A2-2; A3-1). The main shock as well as the aftershock caused extensive damage in Pyrgos and the surrounding area. Also, several other geological effects as liquefaction and sandblows, landslides, fractures, etc. were observed.

Seismic fractures were observed in the following two areas: (a) Pyrgos- Several E-W en echelon seismic fractures were observed at the northeastern part of the city. Their length ranged between some meters and 10 m. They had an opening of 2-5 cm; while no vertical throw was observed, but there was a well-defined right lateral component of movement. Along the occurrence of fractures, the damage was higher, even in new constructions. (b) Lastaiika village- seismic fractures were observed at the southern part of the village. Their strike was E-W and were echelon arranged. Their length varied from some centimeters up to some meters.

High intensities were observed along and on either sides of F1 and F2 that cross the geological formations. These faults belong to the Pyrgos-Katakolo fault zone, segments of which were reactivated in the March 26, 1993 earthquake. Along this fault zone, and more precisely along the F2 fault, seismic fractures were observed and those were the locations where damage was particularly intense. Particularly high intensities were recorded at the areas of Kouvelos and Kokkinohoma, where the fossil outcrops of Erymanthos formation overlie unconformably the Vounargo formation. The low geotechnical properties of these outcrops and their small thickness seem to have been the crucial factors for the magnification of intensities.

Some formations or parts of them have caused local isolated "islets" of high intensity isoseismals north-west of Lapato and east of Aghios Athanasios, where marsh deposits with poor geotechnical properties occur. High intensities were also recorded west of the Aghioi Pantes cemetery. They coincide with the outcrops of loose sands that belong to Aghios Athanasios member of Vounargos formation. At this very location, even earthquake-resistant buildings (i.e. schools) suffered severe damage.

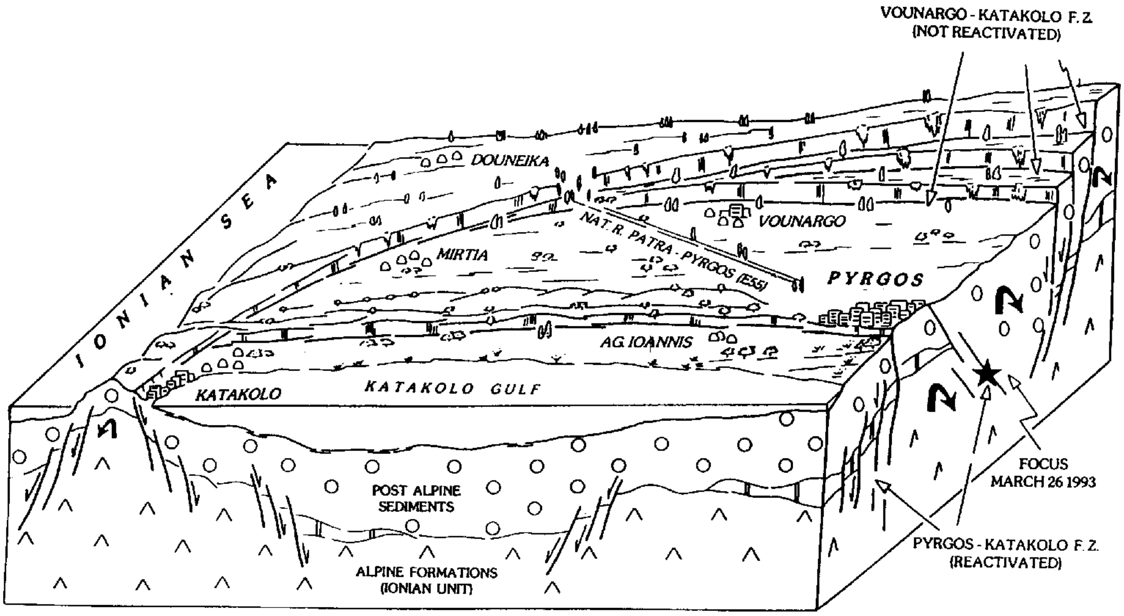


Fig. A3-1 Block-diagram the geological and neotectonic structure of Pyrgos area (after Lekkas *et al.*, 1994).

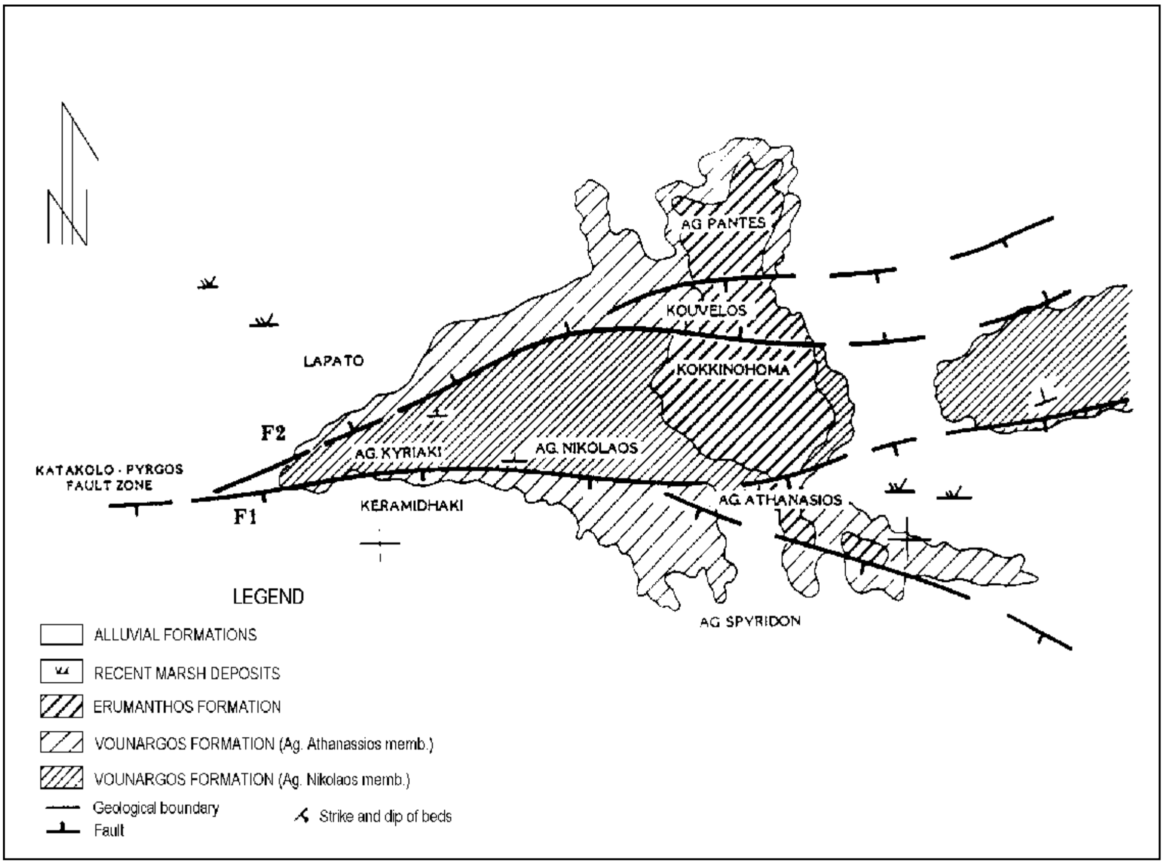


Fig. A3-2 Geological and tectonic map of Pyrgos area (after Lekkas *et al.*, 1994).

Stop A4: Olympia

Archaeology

The archaeological site of Olympia is situated on an active neotectonic graben at the northwestern Peloponnesus. At the foot of the *Kronion* Hill stretches the space formed by the wedge of land between the converging rivers of the Alfios and the Kladeos. This was the site called the Sacred Grove of the *Altis*, regarded as belonging to Zeus in which, in historically recorded times, the most famous of Greek sanctuaries was established.

Formerly it had been a place of worship of pre-Hellenic deities. Every four years, athletic contests were organized here in honor of Zeus, lasting seven days. According to legend, the Olympic Games began after a victory by Pelops against Oenomaos, King of Pissa. Historically, the Olympic Games began in 776 BC Up to the 5th century BC, the sacred enclosure contained the *Heraion* (Temple of *Hera*), the *Prytaneion*, the *Pelopeion* and the *Hippodameion* while, at the foot of the *Kronion* hill the twelve Greek city treasures stood. Outside the enclosure to the West was the Stadium with a 45.000 seating capacity (men only were allowed in). Access to the stadium was along a vaulted passage and, to the South, was the *Bouleuterion* where the Olympic Senate met.

From the 5th century onwards, the sanctuary assumed its final form with the impressive Temple of *Zeus* (peripteral in the Doric style), the Metroon, the Arcades (*Stoa* of Echo and the southern *Stoa*), the Gymnasium and the *Palaestra* the living quarters of the priests, the large *Leonidaion* Hostel and the *Fillippeion*. To the south of the dwellings of the priests, excavations revealed the studio of the sculptor *Phideias* in which he carved the golf and ivory statues of *Zeus*, one of the seven Wonders of the world. Finally, in Roman times, the villa of the Roman emperor *Nero* was added, also the *Exedra* of *Herod Atticus* and Roman baths.

The Olympic Games ceased in 393 AD after the edict issued by *Theodosius the Great*, which forbade all the pagan festivals. They were revived for the first time after 15 centuries, in 1896 in the marble stadium in Athens. Today, an international Olympic Academy functions at Olympia.

Interlude-I

Neotectonic Setting of Messinia
The Kalamata earthquake of 13 September 1986

Key references: Mariolakos et al., 1989; 1991; 1994a, b

The Neotectonic structure of the SW Peloponnessos is characterised by the presence of tectonic grabens and tectonic horsts (1st order structures) striking NNW-SSE and E-W (Mariolakos *et al*., 1987). Such 1st order neotectonic macro-structures are e.g. the Kalamata - Kyparissia graben, the Vlahopoulo graben, the Gargaliani - Filiatra horst, the complex morphotectonic unit of Kyparissia Mts. the complex tectonic horst of Lycodimon Mt. (Fig. I-1). At the margins or inside the 1st order neotectonic macrostructures, neotectonic structures of minor order exist, the strike of which is perpendicular or parallel to the trends of the 1st-order ones. All these macrostructures are dynamically connected, as they have resulted from the same stress field, whereas from the kinematic point of view, there are differences that appear either at the primary stage of their creation or during their evolution (Mariolakos *et al*., 1989).

Such 2nd order neotectonic macro-structures within the Kalamata - Kyparissia graben are the following: (Fig. I-1) (i) Kato Messinia graben; (ii) Meligalas horst; (iii) Ano Messinia graben; (iv) Dorion basin, and (v) Kyparissia - Kalo Nero graben.

Mariolakos *et al*. (1994a) interpreted the kinematic evolution of the Messinian peninsula after the end of the Early Pleistocene, taking into account the phase, and the thickness of the post alpine deposits, and the present-day outcrop altitude of the post alpine marine sediments. According to them, two large neotectonic units dominate, the "*Kyparissia Mts*." and the "*Taygetos Mts*." that function as "tectonic dipoles", rotated around E-W and NNW-SSE axes (Fig. I-2), that is: the former has undergone south-southeastern and simultaneous eastern rotation and the latter has undergone north-northwestern and simultaneous eastern rotation. The *Taygetos M*t. is a horst whereas the *Kyparissia Mts*., despite its high altitude, is, in its greater part, a graben (Mariolakos & Fountoulis, 1991).

A more detailed study on the neotectonic deformation of the area of Taygetos Mt., Kyparissia Mts. and Filiatra horst, has shown that the deformation is not simply of brittle but of brittle-ductile type, as there have been created folds of big curvature radius (Mariolakos & Fountoulis, 1991). Consequently, the Taygetos Mt. should be considered a neotectonic anticlinal multi-faulted mega-horst. On the other hand, Kyparissia Mts. is a more complicated structure, compared to Taygetos Mt.

The morphogenetic procedures started soon after the end of orogenic culmination. The greater part of the area was emerged until Late Pliocene, when it gradually began to submerge in Early Pleistocene. The marine sedimentation continued during Lower Pleistocene, which is the period that the sea transgression reached its greatest extent. Lower Pleistocene has been confirmed in many locations of the Messinia peninsula by the presence mainly of Hyalinea balthica (Koutsouveli 1987; Markopoulou - Diacantoni *et al*., 1989; 1991; Frydas, 1990). From Middle - Pleistocene, the kinematics is reversed, from subsidence uplift. However, there is no evidence if this change was gradual or abrupt. It is possible that there was a period of dynamic equilibrium during there was sediment accumulation in deepening basins. (The palaeoecologic data show that the marine sedimentation was constantly taking place in shallow and warm waters and was continuous during the period between the Late Pliocene and the Latest Early Pleistocene (Markopoulou-Diacantoni *et al*., 1989).

The Kalamata earthquake (13 September 1986)

Kalamata earthquake occurred on September 13, 1986. Its magnitude was M=6.2 R and caused tremendous damage. The location of the epicentre was a controversial matter: others accepted just underneath the city of Kalamata, while others placed it some 10 km to the north. The strongest aftershock took place two days later with 5.6 magnitude. The entire aftershock sequence was confined within a 100-^{km2} triangular-shaped area, which roughly corresponds to the Dimiova - Perivolakia graben, only in which all site effects were observed.

Mariolakos *et al*., (1987b; 1989) studied the related geological effects and drew the following conclusions: (1) the density of the active, possibly active and generally neotectonic faults, seems to be irregular in the broader area (2) faults and the fault zones are rare in the quaternary deposits. Reactivated faults have been observed, but not within the narrow Kalamata area; (3) a large number of small-scale ruptures have been observed in the Neogene formations but not so many as in the quaternary deposits, and (4) numerous faults have been observed in the limestones of Pindos unit, one of which is in Diàsselo, north of Kalamata.

Areas of intense fracturing can be differentiated from others, where faulting and fracturing is sparse. In the former, fractures are so abundant that they are the dominant geological features. Many of them are active and represent seismic faults (a lot of these faults have been reactivated during the second great aftershock of 15/9/86) and are usually accompanied by a tectonic breccia zone. Intensive karstification is observed parallel to the active fracture surfaces.

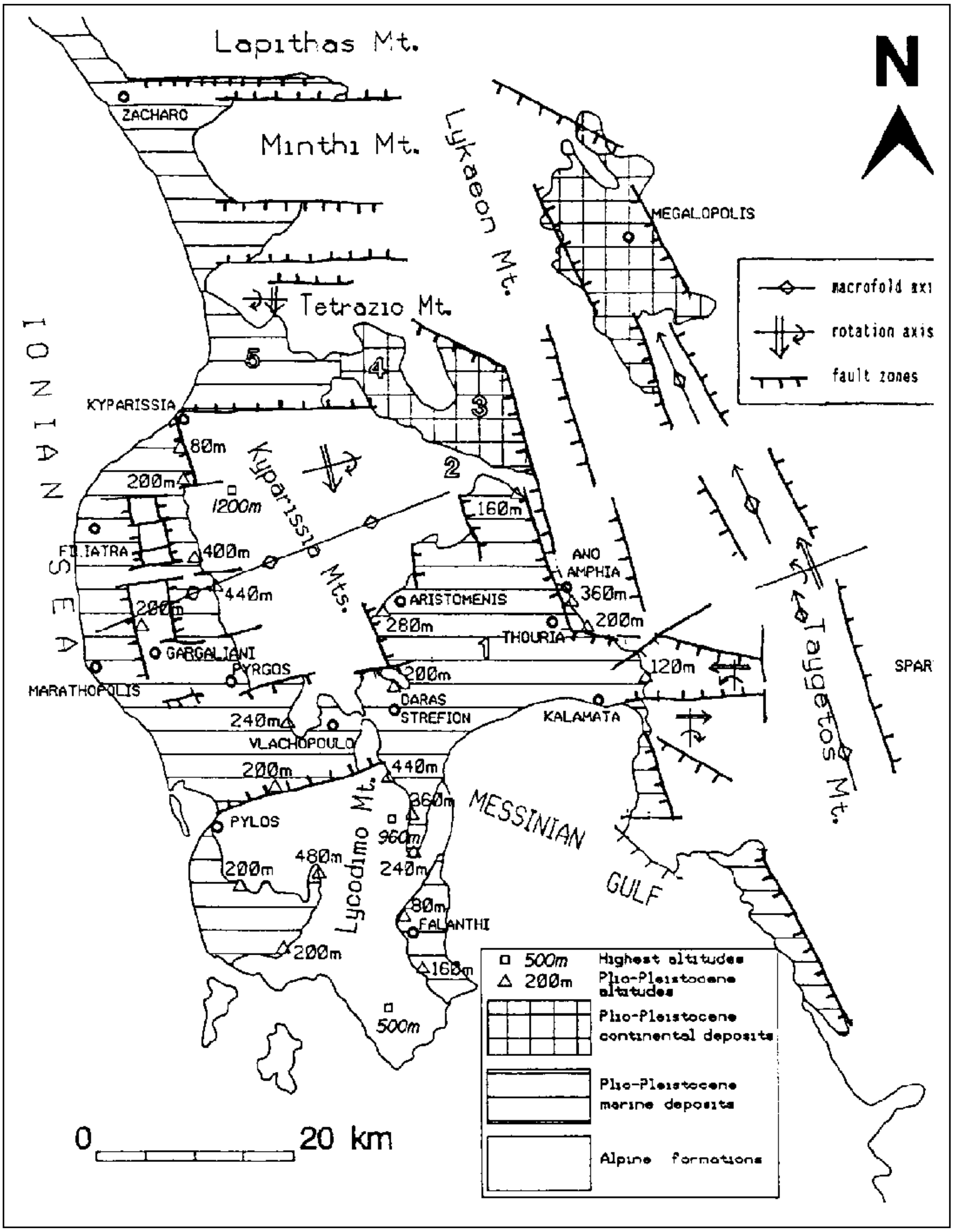


Fig. I-1 1st and 2nd order neotectonic macrostructures. The 2nd order neotectonic macrostructure of Kalamata-Kyparissia graben are : 1: Kato Messinia graben, 2: Meligalas horst, 3: Ano Messinia graben, 4: Dorion basin, 5: Kyparissia-Kalo Nero graben. After Mariolakos *et al.*, 1994a.

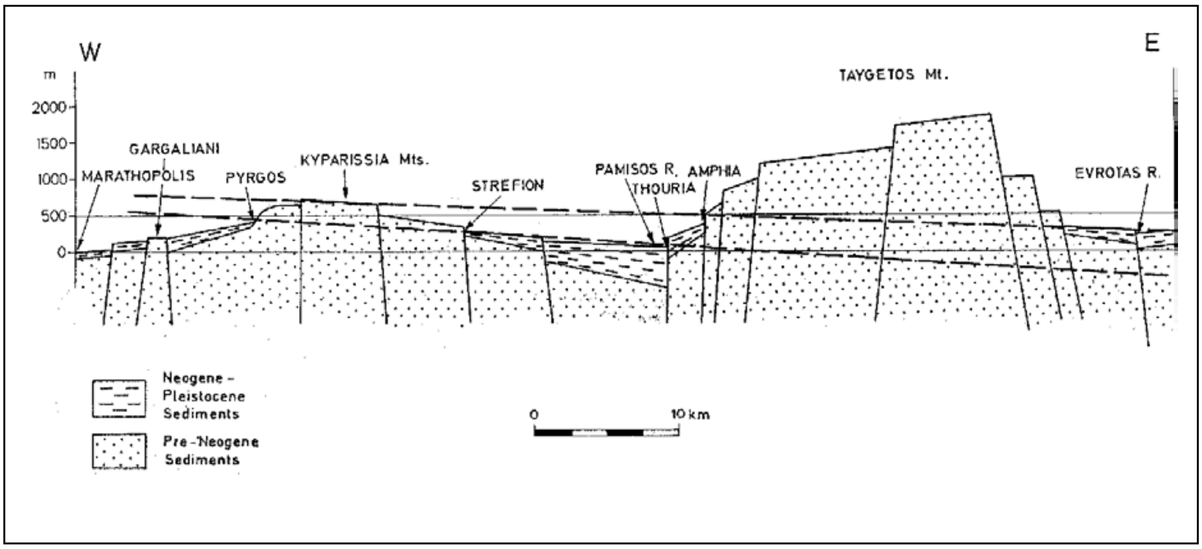


Fig. I-1. Schematic cross-section to show the two active neotectonic macrostructures in SW Peloponnessos. After Mariolakos *et al.*, 1994a.

Stop B1: Pidima fault zone

Paleoseismological sedimentary indicators

Key reference: Mariolakos *et. al.*, 1997

This fault zone forms the eastern margin of Kalamata - Kyparissia graben. Its strike is NNW-SSE and length is approximately 20 km (Fig. I-1). It is a composite, en echelon arranged fault zone; one of its segments is the Amphia- Agrilos fault that has been active at least since the L. Pleistocene, as revealed by the examination of sediment sequences in the area (Fig. B1-1). Successions of marine sediments of E. Pleistocene age are covered by terrestrial deposits, which, in turn, are overlain by a younger marine sequence. This younger marine sequence is again covered by terrestrial deposits. There is no sedimentary transition between the phases, a fact indicating that the abrupt change was of tectonic origin. It is, therefore, suggested that at least two episodes of local tectonically-controlled subsidence took place along this segment of the Pidima fault zone.

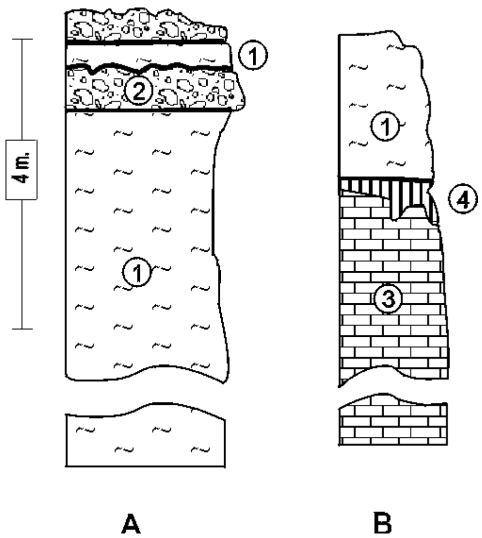


Fig. B1-1. Lithostratigraphic columns of Ano Amphia (A) and Agrilos (B). (1): Pleistocene marine sediments (marls, sands) (2): terrestrial deposits (3) alpine carbonates (4) paleosol.

Stop B2: Kyparissia fault zone
&
Stop B3: Filiatra

Neotectonic mapping: a case study

Key References: Mariolakos et. al., 1988; Mariolakos & Fountoulis, 1991; Fountoulis, 1994

At this stop we will have the chance to review the application of neotectonic mapping (1:100000 scale) on the tectonically active region of southwestern Peloponnessos.

Compilation procedure

The compilation of the neotectonic map involved:

- Collection of all available data on the geology, seismology, engineering geology, and hydrogeological conditions of the area. The data included research papers and publications, geological maps, air-photos and satellite images.
- Initial evaluation of the available data.
- Study of air-photos (1:33,000 scale) and compilation of photogeological map.
- Compilation of 1/50,000 tectonic map.
- Compilation of 1/100,000 geomorphologic - morphotectonic map.
- Field work, which also included the study of the neotectonic deformation of the alpine structures.

Geomorphology

The mapped area is bounded to the north by Mt. Lapithas, to the east by Mt. Lykeo and Ano and Kato Messinia basins, to the south by the morphological depression of Pylos - Velika. The Ionian Sea forms the western boundary of the area.

The drainage network

The drainage network is mostly dendritic, while at certain areas is becomes rectangular. The largest streams are the rivers Neda and Arkadikos (Sellas). The former flows roughly E-W and is asymmetric, as the smaller order catchments in the northern portion are almost double in size compared to their southern counterparts. The latter has a catchment with more peculiarities. The main, 5th order stream initially flows E-W and then bends almost at right angles, to a WNW direction, towards the Gulf of Kyparissia. Locally the network is rectangular. Many smaller-order catchments at the northern part are quite elongated, along a NNE-SSW axis. The southern sub-catchments develop radially. Finally, a 4th-order sub-catchment that occupies almost all the western part of Arkadikos basin has a relatively small main branch, while the 1st and 2nd order ones are very long and have a N-S direction.

As for the other catchments at the south, their two main features are that they are very elongated and that their flow direction changes around the SW and southern part of Arkadikos watershed.

The sheet can be divided into four domains, according to the predominant direction of flow:

- The northern part, comprising river Neda and the northern portion of Arkadikos: flow direction is NNE-SSW and to a lesser extent, NW-SE
- The central part that coincides with the southern-southeastern portion of Arkadikos catchment: flow direction is N-S to NNW-SSE, with few branches flowing NE-SW to E-W.
- The southeastern part; flow direction here is NNW-SSE, with few branches flowing E-W to ENE-WSW.
- The western part (Gargaliani - Filiatra): here the flow is radial, from NW-SE to E-W and then NE-SW (Fig. B2-1).

Morphotectonic study of planation surfaces

Erosional surfaces are not uniform in size. The ones lying at altitudes of 200-400 m and 400-600 m. are sizable enough, while at higher altitudes they are significantly smaller. At the southern part there are large planation surfaces and up to the ‘line’ connecting Lambena village with Filiatra. To the north of this line, these surfaces decrease in size; however at surfaces lying at higher altitudes no such change occurs. Further to the north and up to the valley of Neda River there are large p.s at 200-400 m. altitude. However, north of Neda valley and mostly at the area among Giannitsohori - Figaleia - Andritsena and Zaharo, the p.s. at 200-400m and 400-600 m. altitude are represented by small relics of them, while the higher ones are quite extended, always in comparison to the southern part.

The extended p.s. develop along a SSE direction, form Hristiano to Hora, where there is a change in direction towards the NE. Many of them are elongated, as result of river erosion. Their dips are roughly normal to the above-mentioned directions.

The higher p.s. (600-1200 m.), especially the ones on Mt. Egaleo are clearly elongated along a NNW-SSE direction. Such development cannot be seen at the rest of the area, except for the northern and northwestern parts, where all the surfaces develop along a NW-SE direction. Finally, there is a group of p.s. at 200-400 m., at the mouth of river Neda with a mean NW dip.

Geographical distribution of planation surfaces

Concerning the geographic distribution of p.s., three domains can be distinguished: the southern, northern and eastern one.

Southern domain. It comprises the area south of Arkadikos River. All over the area, there is a gradual decrease of altitude towards the SE. However, from and to the north of the E-W trending Kyparissia - Aetos fault zone, there are only low-altitude p.s., without any medium- or high altitude ones. The kinematical image is that of a tectonic dipole, where the blocks are rotated around a horizontal NNE-SSE trending axis that led to uplift in the northern part and subsidence in the southern.

Northern domain. It lies between rivers Arkadikos and Neda and the p.s. distribution is similar to the one of Kyparissia. Besides an E-W trending fault zone occurs, parallel to the flow direction of Neda. It also concluded that the domain corresponds to a tectonic dipole, rotated around a E-W trending axis.

Eastern domain. No systematic distribution that could lead to tectonically useful results is verifiable for this area.

Western domain. (Fig. B2-1) All p.s. in this area have developed on pleistocene marine deposits. What is important here is that they Their inclinations/gradients are distributed in a fan-like manner, that it that the northern ones have a northwesterly inclination and heading to the south, inclinations becomes gradually westerly and then southwesterly.

Neotectonic macrostructures of the Sheet

The main (1st order) macrostrures of central-western Peloponnessos are: (i) *Kalamata - Kyparissia graben*; (ii) *Vlahopoulo graben*; (iii) *Filiatra depression*; (iv) *Kyparissia Mts. graben*; (v) *Mt. Tetrazio horst*; (vi) *Neda graben* and (vii) *Minthi horst* (Fig. I-1). These structures are bounded by fault zones that have the following main characteristics: (1) the trends of the constituent faults are not stable throughout the margins, and (2) these faults are interrupted by others of different trend; all together, they form conjugate fault sets created under the same stress field.

For these reasons the trend of the grabens is not stable. The confirmation of such geometric features, in combination with other observations, that are not important to be mentioned here, leads to the acceptance of a kinematic, and subsequently dynamic, model, different from the one that could be obtained from the analysis of individual faults, that are apparently normal. The en echelon arrangement of the marginal faults, suggests oblique-normal character, rather than pure normal one; hence the deformation should be attributed to a coupling (?torsional) stress field, as it has been proven by detailed geologic, geomorphologic and neotectonic studies.

The following multifaulted block structures can be distinguished in the study area: (i) *Zaharo graben*; (ii) *Mt Minthi horst*; (iii) *Neda graben*; (iv) *Tetrazio horst*; (v) *Kalo Nero - Dorio - Kyparissia graben*; (vi) *Kyparissia Mts. morphotectonic structure*; (vii) *Filiatra dipression*, and (viii) *Vlahopoulo graben*. It should also be mentioned that in addition to the faults, large-scale neotectonic folds within these fault blocks have been observed, mainly in the Filatra and Neda grabens.

Below follows a description of the neotectonic macrostructures we shall meet in the course of the field trip.

Kalo Nero - Dorio - A. Messinia graben

It forms part of a larger-order structure, the Kalamata - Kyparissia graben and consists of three sub-basins; one of them (Dorio) is not tectonically controlled.

There are two important fault zones within the graben. (a) The NW-SE trending Kato Melpeia - Desyllas f.z. and (b) Kyparissia - Aetos f.z.. The former displays (i) increasing offset towards Desyllas, (ii) intense transverse stream incision and development of talus cones and (iii) highly fractured rocks along the fault surface. This fault zone is characterised as active. The latter has also westward-increasing offset whereas monomictic tectonic breccias has been developed along strike. However, no fault surface can be confirmed nor scree or talus cones develop: in this sense it is regarded as possibly active. Another active fault actually determines the flow of river Peristera; it trends E-W. Also, there are several small faults in the vicinity of the village of Myros.

Filiatra depression

It is a 1st order neotectonic macrostructure and can be broken down into several smaller-scale ones (Fig. B2-2). Its eastern margin consists of NNW-SSE trending, en echelon arranged faults at the west of Mt. Kyparissia; however, its western termination cannot be seen, as it is offshore (in the Ionian Sea).

One important structure within this depression is the Gargaliani-Profitis Ilias sub-horst. Its western margin has a NNW-SSE trend and is bounded by numerous en echelon arranged faults. It is the most prominent topographic discontinuity in the Filiatra depression and a large part of it is covered by successive generations of breccia and scree.

Besides the NNW-SSE trending faults, there are other, E-W trending ones, characterised by prominent topographic discontinuities, developed on the Lower Pleistocene and the alpine formations. Slickensides are rare and are found only where carbonates are cut and there are occurrences of tectonic breccias. These E-W faults disrupt also the NNW-SSE ones. At the southern part, in the mountain built of the Messinia conglomerates, the dominant fault trends are NE-SW and ENE-WSW. Here, striations that show normal-oblique or even reverse-oblique offset were found. The whole area has undergone thorough study and, after Mariolakos & Fountoulis (1991) the deformation type here is of brittle-ductile character and is the result of a rotational couple stress field. This is evidently manifested by the structural contour map (Fig. B2-3) of the substratum/Lower Pleistocene contact, which displays an overall curvature around a N70 -trending, WSW-plunging axis.

Vlahopoulo graben

It occupies the southernmost part of the sheet. It southern margin is formed by a E-W to ENE-WSW trending fault, visible in the vicinity of Pylos.

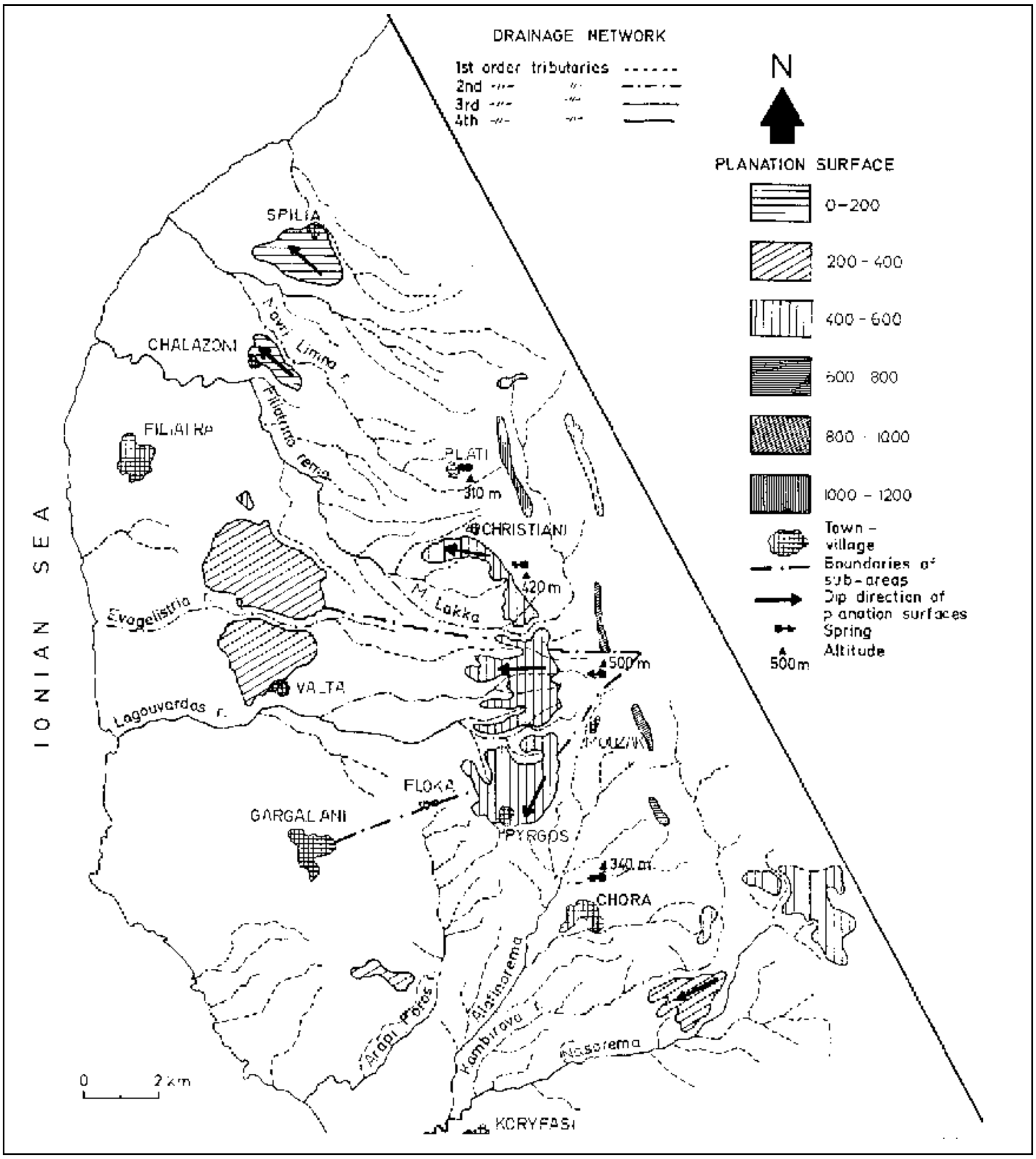


Fig. B2-1 Map to show the planation surfaces and drainage network developed mainly on the Early Pleistocene marine deposits. After Mariolakos & Fountoulis, 1991.

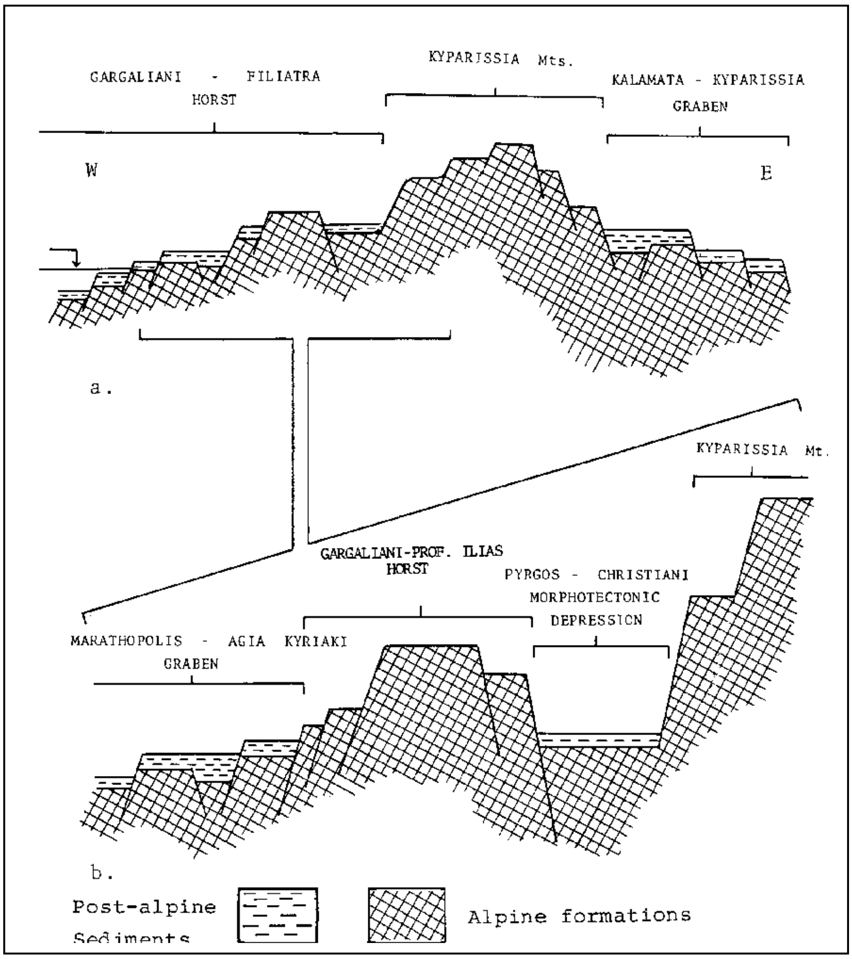


Fig. B2-2. Schematic cross-sections of 1st-order (a) and 2nd-order (b) neotectonic macro-structures. After Mariolakos & Fountoulis, 1991.

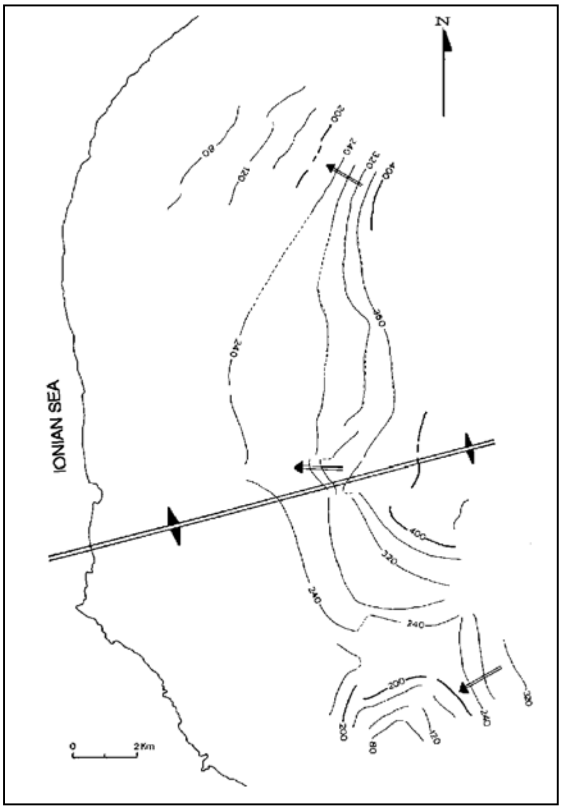


Fig. B2-3. Structural contour map of the Pleistocene deposits/alpine flysch contact. After Mariolakos & Fountoulis, 1991.

Stop C1: Kambos

Kambos fault zone
Active Tectonics - Morphotectonics

Key references: Mariolakos et al., 1989; 1993

Introduction

Kambos depression and Kitries Bay area are located in the Messinia province, some 10km to the southeast from Kalamata (Fig. C1-1). The relief of the area is rather smooth, featuring hills and valleys that constitute the transitional forms towards the purely mountainous region of Taygetos Mt. to the east. On the eastern part of the Kitries Bay, the relief continues its smooth form under the Messinian Gulf water, down to the shelf break, where an abrupt, tectonically induced change in the slope gradient takes place. This markedly differs from the underwater topography of the western part of the Messinian Gulf, in which a gentle topography continues underwater down to 1100 m (Papanikolaou *et al*., 1988). The relatively gentle slopes of the Mantinia-Doli area and the steeper terrain of the southern part of the study area (south of Doli) is another eye-catching difference. This is due to tectonic control over the whole region, as we will see below. The neotectonic macrostructures of the area are depicted in Fig. C1-2.

Neotectonic - Kinematic interpretation

The neotectonic evolution of three of these neotectonic macrostructures (Kitries - Mantinia sub-graben, Vardia - Koka horst, Kambos graben) is shown in Fig. C1-3:

At the very end of the Pliocene the region started gradually to submerge. At the beginning of Early Pleistocene, deposition of shallow -water sediments on a well-formed paleorelief on the carbonates and flysch of the Tripolis geotectonic unit at Kitries - Mantinia sub-graben. At the same time, the Vardia-Koka horst worked as a barrier for the clastic material (mainly pebbles) that came from the Taygetos Mt. horst (which was uplifted and eroded) and started to fill (fig. 3A) the Kambos graben.

The whole area continued gradually to submerge. As a result, the transgression of the sea from the west to the east continued until all the area was sunk below sea-level, up to the Doli and Stavropigi villages (Fig. C3-3B). Tectonic activity and/or climatic conditions provided a great amount of coarse clastic material, which passed over the Vardia-Koka horst and was deposited as alternations of sandstones and oligomictic conglomerates in a shallow - water marine environment, at the Kitries - Mantinia sub-graben. At the same time, in the Kambos graben, oligomictic conglomerates continued to accumulate, partly covering the pre-existing continental deposits.

At approximately the end of the Early Pleistocene, the kinematic regime shifted uplift. This resulted in the regression of the sea and the local exhumation and subsequent erosion of the older, fossilized, pre-Pleistocene relief. The uplift phase was neither abrupt, nor uniform over the whole area.

Kambos graben drainage network consists of two main torrents, both lying at the northern part. These are the Koskarakas torrent (flowing E-W) and a smaller one, (Fig. C1-1). Both of these are deeply incised and are now eroding into the substratum. The incision of the gorges has resulted from the fact that they both flow within an uplifted block and the presence of the Vardia - Koka horst, built up of carbonates of the Tripolis geotectonic unit.

The following mean vertical movements have been calculated, base on the thickness, phase, altitude and age of marine deposits, according to Mariolakos *et al.,* 1993: mean subsidence rate (overall): V_s=0.45 mm/yr, mean uplift rate (Doli area): V_{uD}=0.45 mm/yr; (Mantinia area) V_{uM}=0.37 mm/yr.

The area is a complex multi-fractured neotectonic macrostructure, typical of the transitional neotectonic zones located between two megastructures. In this case, these two megastructures are the Taygetos Mt. horst to the east and the Messinian Gulf graben to the west. It is, in fact, a mosaic of variably sized blocks, which, however, are rather small. The almost total absence of large (and many) faults crossing the marine Pleistocene deposits is evidence for the following:

i. The fragmentation of the study area, as well as the position of the blocks relative to each other, had taken place before the deposition of the marine Pleistocene sediments.

ii. During both the subsidence and the uplift phase, the aforementioned small blocks lost their tectonic independence. As a result, the "mosaic" of blocks behaved as a single macro-block (neotectonic unit).

The area is a typical example of a system (located within marginal neotectonic zones) that behaves as a graben for a long period but, during a later phase, is connected to one, or more, neighbouring blocks and hence shows a totally new kinematic behaviour. Such cases are very frequent in the broader Messinia area: for instance, the present-day Asprochoma - Koutalas horst (Mariolakos *et al*., 1987c), in earlier stages of its neotectonic evolution consisted of at least two blocks. These two blocks behaved as a single unit during the uplift phase. The local stress field, therefore, should have remained the same during the whole Pleistocene. It was the regional kinematic (as a consequence of the dynamic) regime that shifted from subsidence to uplift at the end of the Early Pleistocene, since all the Early Pleistocene marine sediments have been uplifted several hundreds of meters over Messinia (Marcopoulou-Diacantoni *et al*., 1989; 1991).

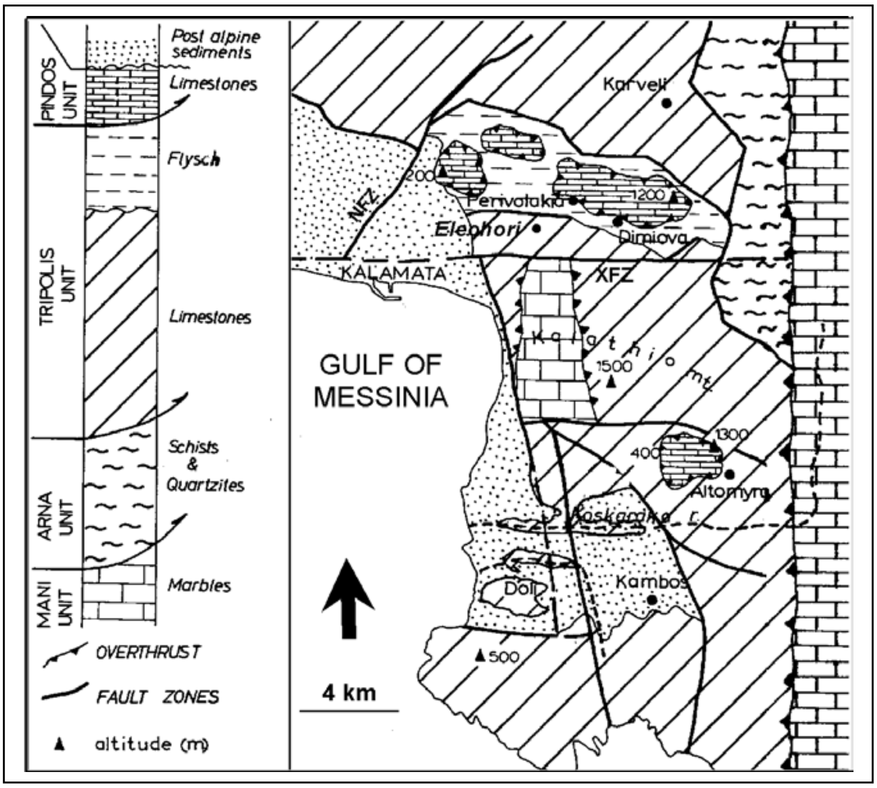


Fig. C1-1 Simplified geological map showing the alpine nappe pile, and post -alpine sediments of the region. After Mariolakos *et al.*, 1993.

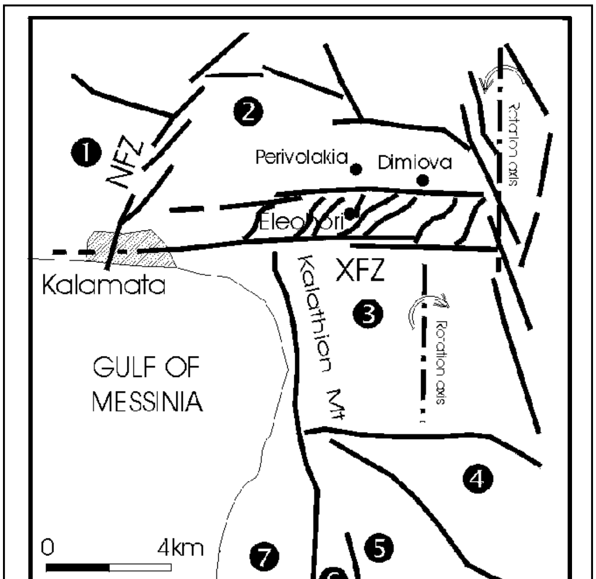


Fig. C1-2 Smaller-order neotectonic macrostructures of the Kato Messinia sub-graben. 1. Asprohoma-Koutalas horst 2. Dimiova-Perivolakia graben, 3.Kalathio horst, 4. Altomyra half-graben, 5. Kambos graben, 6. Vardia-Koka horst, 7. Kitries - Mantineia sub-graben. After Mariolakos & Fountoulis (subm.).

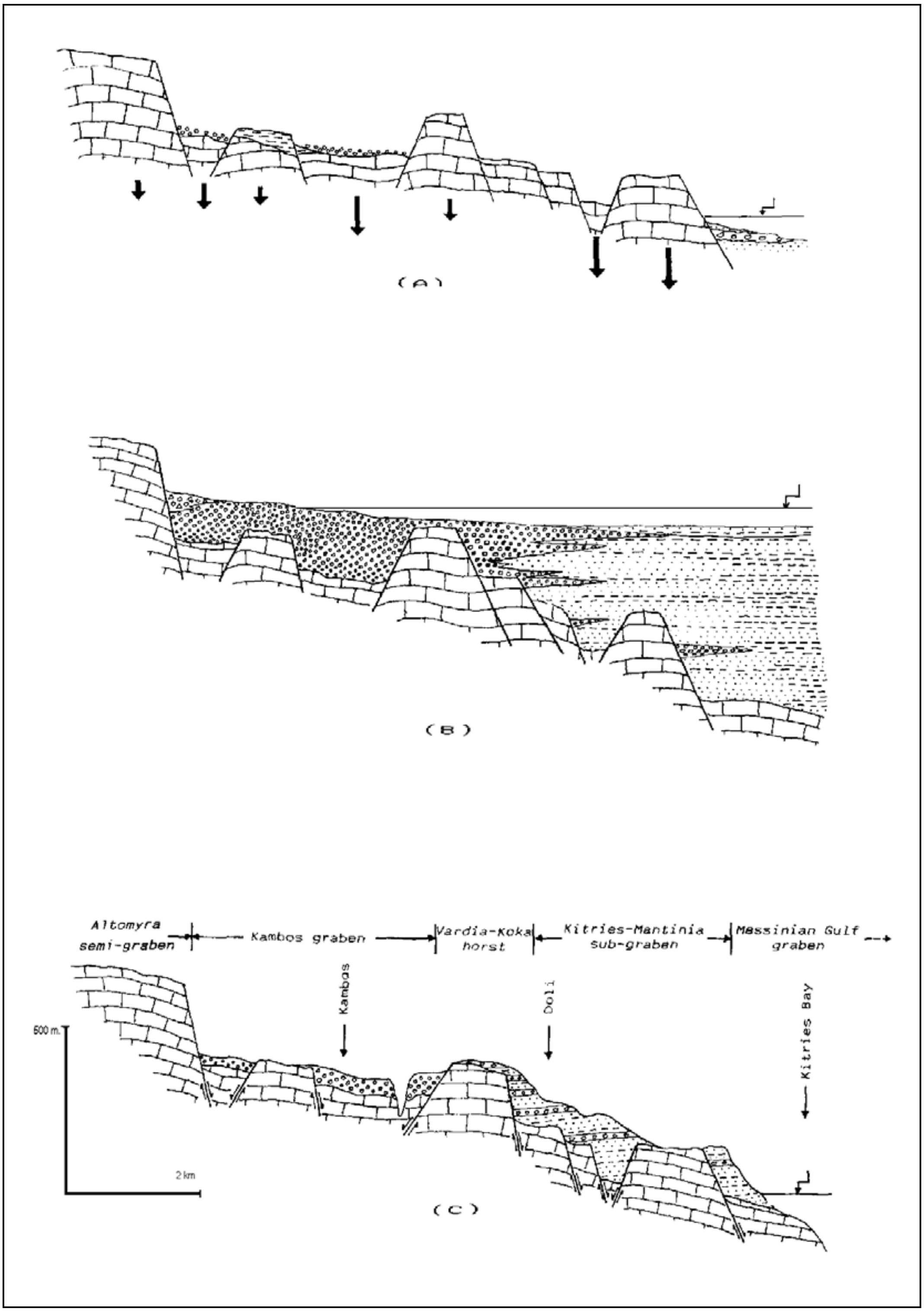


Fig. C1-3 Schematic depiction of the three main stages of the neotectonic evolution of the study area. (A) At the beginning of the deposition of the oldest pleistocene sediments. (B) Filling up with pleistocene marine deposits, when the uplift started. (C) The present geological and morphotectonic status. After Mariolakos *et al.*, 1993.

Stop C2: Zimbeli

Multi-phase brittle-ductile active deformation

Key references: Mariolakos et al., 1987a

The Zimbeli fault is located in the Dimiova-Perivolakia graben and forms the tectonic boundary between the eocene neritic limestones and the flysch of the Tripolis geotectonic unit (Fig. C1-1).

The main characteristic of this NW-dipping fault surface, most of which was uncovered after a landslide, is that its strike isn't constant, ranging from NE-SW to ENE-WSW. This fault surface has been deformed under a local stress field, which differs from the regional one. Its formation is placed at the first stages of flysch sedimentation (U. Eocene) as a synsedimentary fault or even earlier, as it has been described from many places of Peloponnessos (Mariolakos 1976). Since then, it has been repeatedly reactivated, which is proven by successive generations of striations and tectonic breccia sheets. The shape of the fault surface is not plane but undulated.

The main minor structures on the Zimbeli fault surface are the following: (1) striations (2) tectonic breccia sheets, and (3) microfaults that intersect the main fault surface.

Tectonic breccia sheets

The tectonic breccia sheets observed on the fault surface can be distinguished in four different categories, according to their composition and relative age. The oldest tectonic breccia is compact and monomictic and consists exclusively of very small angular particles of the neritic eocene limestones and its thickness does not exceed 5 mm. This breccia can be observed all over the fault surface, but it is better exposed at the southwestern part of the fault surface. It is covered by a thin calcitic sheet (central and southwestern part of the surface) the thickness of which in some places exceeds 0,5 cm, but usually is thinner. This calcitic film is in turn covered by a compact calcitic crust consisting of fragments of eocene limestones and from the previous tectonic breccias and occurs mainly at the central part of the fault surface; its thickness locally exceeds 5 cm.

The relatively younger tectonic breccia is polymictic and consists of flysch and eocene limestones fragments, while in some places of the surface fragments from the older tectonic breccias are present. The size of this angular coarse material varies from 1 to 5 cm. Contrary to the other tectonic breccias, this one is not very compact. Its thickness varies from 5 to 10 cm, and locally is greater than 50 cm.

Striations

They are distributed all over the fault surface. Fig. C2-1 shows the statistic analysis of the measured striations.

The older set (I) seems plunges 25/252. It always occurs beneath the tectonic breccias and the calcitic crusts. The frequency of this set is very low. The next generation (II), whose frequency is relatively low, plunges 48/312. It occurs not only on the eocene limestones, but also on the surface of the first monomictic breccia. The third generation (III) occurs mainly on the central and on the southwestern part of the surface. It is sub-horizontal and its frequency is relatively low.

The next generations of (IV and V) plunge 36/010 and 36/266, respectively. They occur on the surface of the eocene limestones as well as on the first two tectonic breccia sheets. The frequency of set IV is very low. Set V has the highest frequency of all. The problem that arose was whether set V postdates set V or not, as their relationship is not clear.

A sixth set (VI) (14/035) was also detected; however it could not be dated as it did not intersect any other set.

Remarks:

Except set II, all other generations are oblique-slip.

Set III is sub-horizontal.

Sets I, III and V are sinistral, and sets IV and VI are dextral (Fig. C2-1).

Microfaults

Zimbeli fault surface is interrupted by numerous microfaults, most of which postdate the striations and breccia sheets. Their visible length is about 5-6 m. They can be distinguished in several sets or generations according to the following features (a) their size (b) if they have been filled in with calcite or not (c) if they cut one of more tectonic breccias (d) their strike and dip (e) if a fracture set or microfaults are intersected by another or not.

The general characteristics of these fractures based on field observation can be summarized as following.

They are antithetic to the main fault surface. The average dip is approximately 50(iii) The displacement varies from 1 to 15 cm or even more in few cases and they are en echelon arranged. Some of them, especially the larger ones are in the form of gashes, while the smaller ones are usually filled with calcite. Their sense of displacement is not uniform, and some have downthrown the SE block or vice versa.

The "relative age" of some fracture sets couldn't be always determined. Generally speaking, the older fractures are relatively small, have small displacement and cut only the two older tectonic breccias. The younger sets are longer (5-15 m), have a maximum displacement of 3-4 cm and occur in the form of gashes, with a 10-cm in the middle. It is remarkable that these fractures intersect all tectonic breccias. In some places, where more than one set of fractures occurs, their "relative age" can be easily determined. In fig.10 three sets of fractures are distinguished. The oldest one set I, III being youngest.

In Fig. C2-2a the curvature of the Zimbeli fault surface is shown, as well as the magnitude of displacement along it. All these measurements have been made parallel to the fault surface and perpendicular to its strike. In Fig. C2-2b the size and the shape of the "gap" (opening) between the two segments of the largest observed microfault is depicted. As shown in these diagrams, the magnitude of the displacement decreases gradually towards the extremities of the fracture.

Zimbeli fault surface displays its deformation history through successive generations of striations, breccia sheets and microfaults. The following conclusions can be drawn:

The microfaults that dissect the main fault surface define blocks that have undergone slight counter-clockwise rotation around a NE-SW horizontal axis. This rotation can be attributed to a local torsion, which is not readily explained by mechanics of pure shear, that is extension. Note also that the fault surface itself is curved, which corresponds to a 2-3% NE-SW shortening; the combination of these observations suggests the participation of a compressional stress field and a brittle-ductile type of deformation.

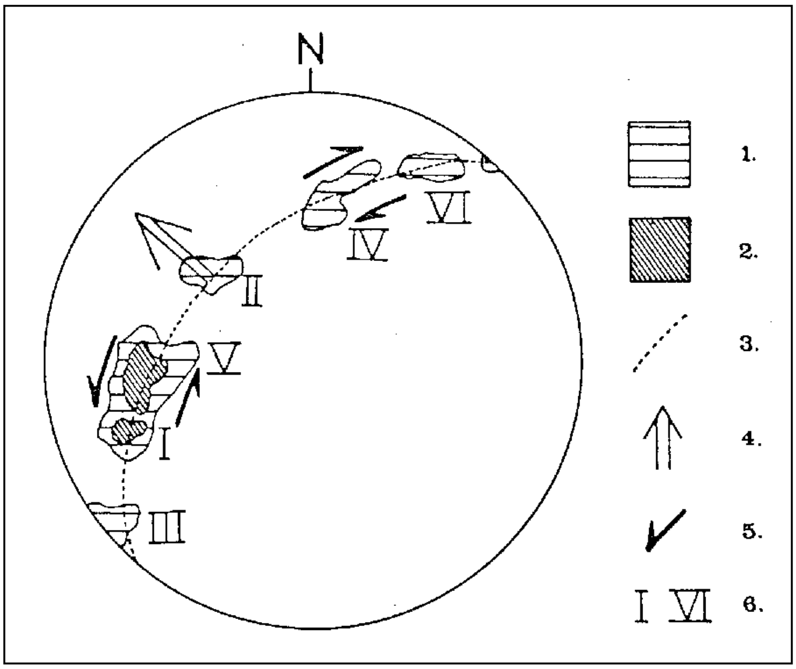


Fig. C2-1 Equal-area (lower hemisphere) projection of striations on the Zimbeli fault. 1: density 10-30%, 2. density >30%, 3. mean fault plane, 4. fault dip direction, 5. strike-slip component, 6. striation sets. After Mariolakos *et al.*, 1987a

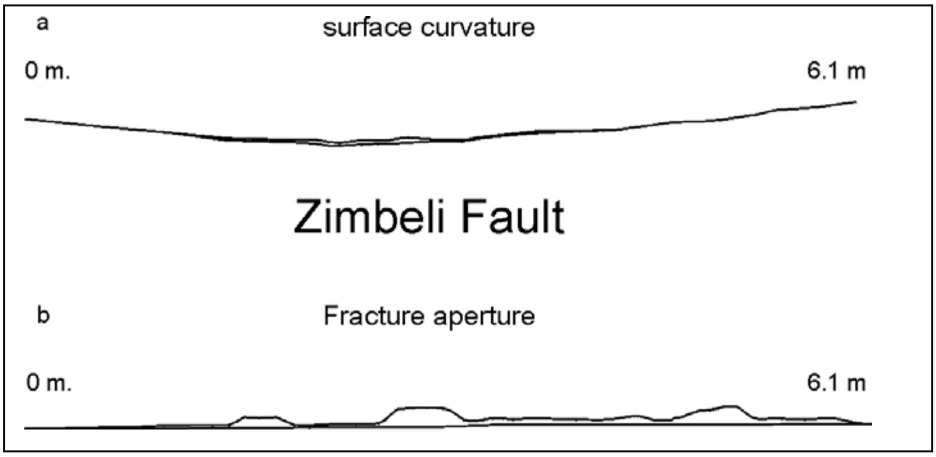


Fig. C2-2 (a) Curvature and displacement magnitude along Zimbeli fault plane - section is normal to fault surface (b) fracture aperture along the fault plane - section is parallel to fault surface. After Mariolakos *et al.*, 1987a.

Stop C3: Eleohori

Earthquake response of constructions - correlation with neotectonic deformation

Key references: Mariolakos et al., 1992a,b; Mariolakos & Fountoulis, submitted

Eleohori village has been built on the highest part of an elongated hilly area at an altitude of 530m. Along the watershed the morphological gradient is very low not exceeding 15%. On the contrary, at the slopes of the hilly area the morphological gradient is very high and in some locations it is more than 120%.

Almost all the houses had been founded on the thick-bedded or unbedded neritic Cretaceous-Eocene carbonates of Tripolis geotectonic unit, the thickness of which is more than 1.000m. Generally, unfractured carbonate rocks are very cohesive with good response to seismic loading. However, this is not usually the case as the carbonate rock-mass is crossed by a big number of different types of densely spaced discontinuities.

Study on the behaviour of the rock mass as a foundation formation in Eleohori showed that:

- both old and new constructions were destroyed
- the distribution of damage bore no relationship to morphological gradient
- the damage was closely interrelated to the occurrence of seismic fractures (Fig. C3-1).
- Surprisingly enough, both of the houses that remained intact after the earthquake, are founded on a fault surface; however, this fault was not reactivated in the Kalamata earthquakes
- Not all fractures were formed by the main shock; a number of them were the result of the main aftershock. In the same way, damage in Eleohori could be distinguished in two phases.
- Note also that the remnant of an ancient wall did not collapse in the earthquake. This is perhaps due to the type of construction

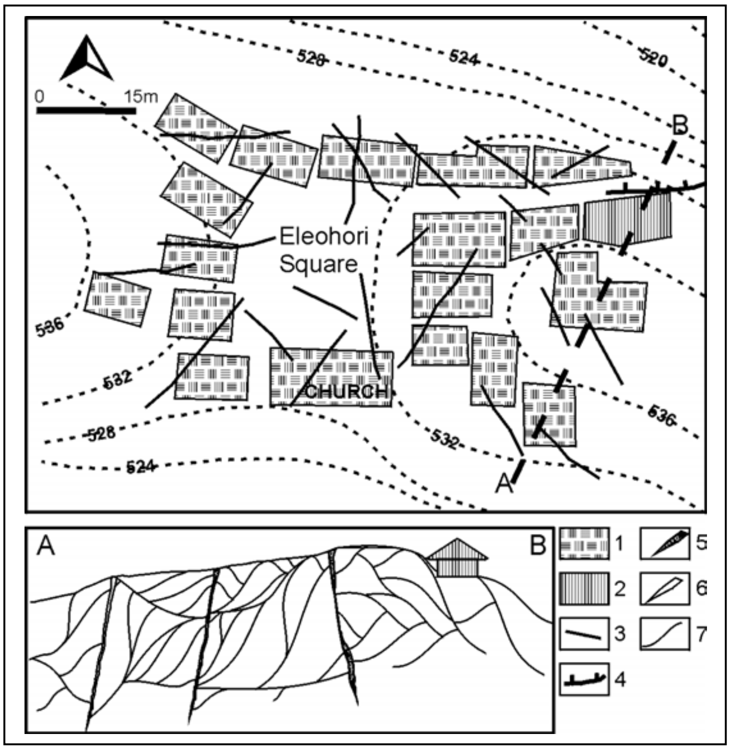


Fig. C3-1 Correlation between the type of fractures (active – inactive) and the damages in Eleohori. 1:Damaged houses, 2:non-damaged house, 3:seismic faults and fractures, 4:inactive fault, 5:active fault reactivated by the Kalamata earthquake, 6:active fault not reactivated by the Kalamata earthquake, 7:inactive curved fault surfaces. After Mariolakos Fountoulis (subm.)

Stop C4: Tzirorrema

Seismic response of rock mass
Neotectonic evolution of a graben margin

Key references: Mariolakos et al., 1987d; Markopoulou et al., 1989

Tzirorrema torrent flows towards the Messinian Bay. During its course, it erodes the Mesozoic calcareous formations of Tripolis Unit, before entering the plain of Kalamata. The degree of incision decreases upstream, and the zero point is just south of Polyani (Stop C5) On our way upstream, we shall stop at two locations. The former displays a part of the noetoectonic evolution of the eastern margin of the Kato Messinia graben, and the latter will provide us with the opportunity to discuss the behaviour of the theoretically compact rock mass in earthquake-induced deformation.

Location I: Thouria

A well-defined paleorelief has been formed on the carbonates of the Tripolis unit. This relief has been covered by the marine pleistocene sediments that filled the Kato Messinia graben. They consist of marls, sands and polymictic conglomerates, and their overall thickness is approximately 400 m. Erosion of these deposits has uncovered the alpine paleorelief. The evolution of this eastern margin was also discussed in Stop B1.

Location II: Tzirorrema torrent

The calcareous rock mass has suffered intense fracturing, part of which is seismically induced, as it could be confirmed in the 1986 Kalamata earthquake. On the mesoscopic scale, we can see that the smaller order fault surfaces are anastomosed and/or bounded by higher-order faults. All faults are s-shaped in cross section and are en echelon arranged, which is evidence of the dynamic and kinematic interdependence between the lower and higher-order faults (Fig. C4-1). Usually the fault surfaces are polished and more than one generation of striae occurs on them. No tectonic breccia or loose zones have developed along these faults.

These surfaces have been formed in the last alpine orogenic stages or in the very early stages of the neotectonic period; hence, they are considered inactive.

Large-scale Y-shaped fault sets (main fault and conjugate antithetic) can be observed on the steep valley gradients These faults have formed in the neotectonic period and most of which are still active. On the intersection of the fault sets with the ground surface, karst forms (sinkholes) have developed. The faults intersect with each other and cut the mass to large rhomboidal blocks. The density of the neotectonic faults is locally so high, that the whole mountain gives the impression of a gigantic loose mass ("macro-breccia"). However, fault density is not uniform, and some compact, more or less intact blocks can be found "floating" in the loose rock mass.

Rockfalls

The area of Tzirorrema was, among others, the locus of numerous rockfalls, induced by the Kalamata earthquake. However, most rockfalls took place not in the main shock, but in the largest aftershock, two days later. What is also noteworthy is that they occurred at locations where the (theoretical) conditions were unfavourable; boulders were displaced and even in areas of practically no relief. Nevertheless, the highest percentage of rockfalls occurred in areas of mean morphological gradient higher than 50%. In all other locations of lower gradient, rockfalls were directly associated with the occurrence of faults (Fig. C4-2). Rockfalls were observed mainly where active reactivated faults exist in combination with a favorable morphology while some meters alongside, where the gradient was the same, no reactivation was observed and the rockfalls significantly limited. All rockfalls had a direct relationship with tectonic loose zones and their geographical distribution is related to the frequency of reactivated tectonic discontinuities, which had a NW strike (Fig. C4-3).

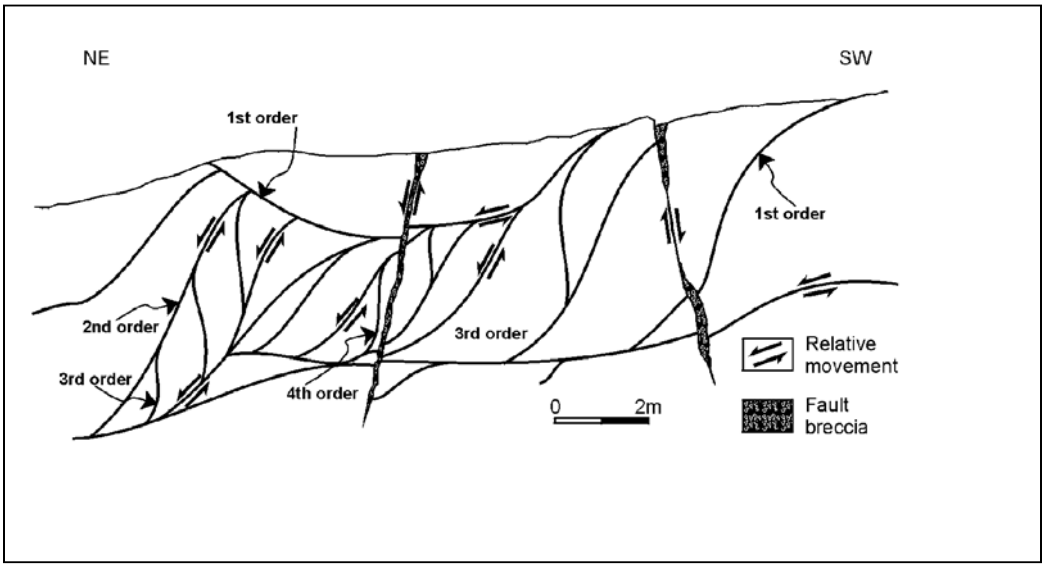


Fig. C4-1 Schematic depiction of older (inactive) neotectonic faults of various order intersected by younger active ones. After Mariolakos & Fountoulis (subm.)

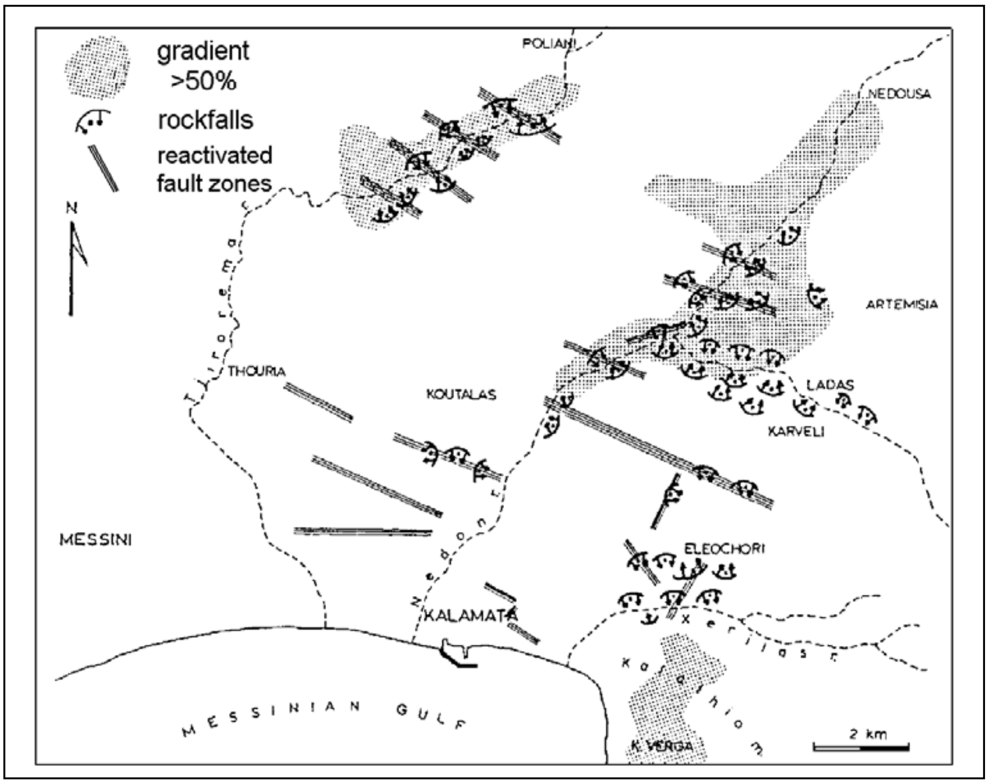


Fig. C4-2 Distribution of rockfalls induced by the Kalamata earthquakes. After Mariolakos *et al.*, 1992a.

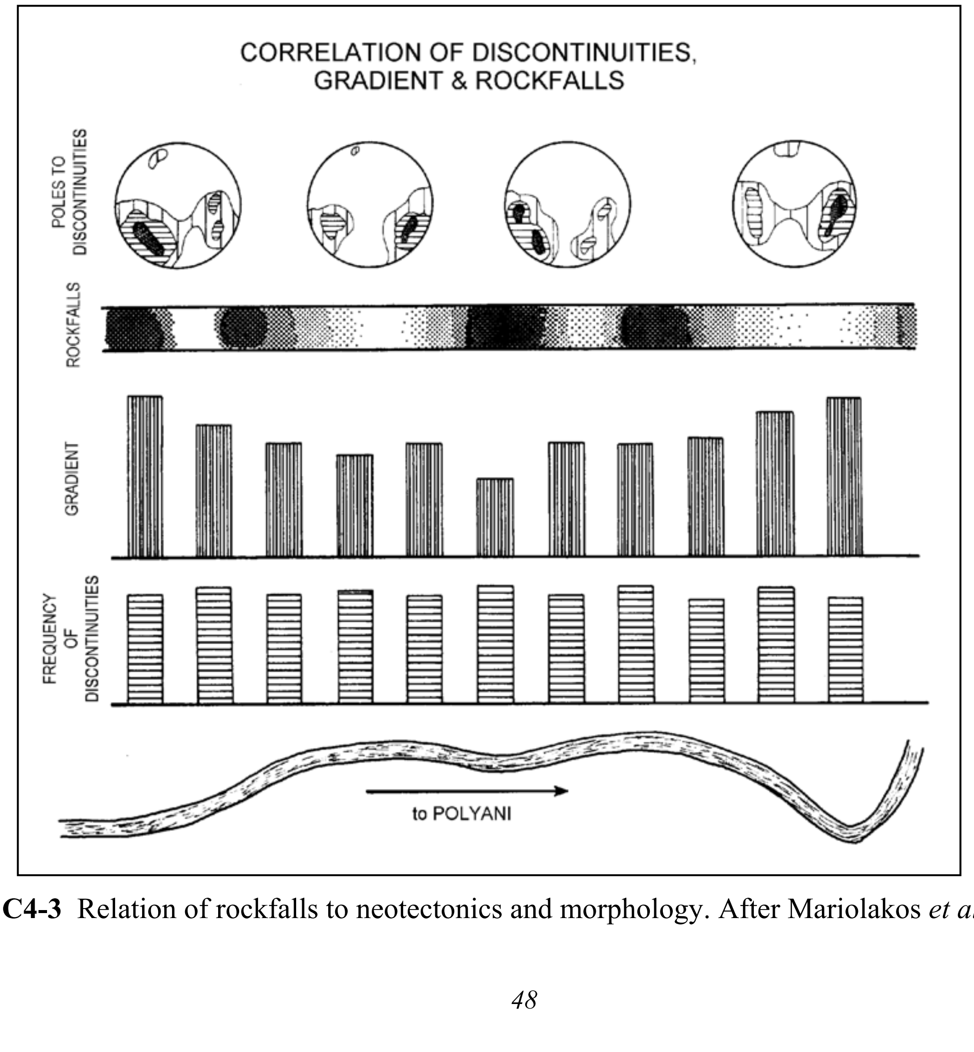


Fig. C4-3 Relation of rockfalls to neotectonics and morphology. After Mariolakos *et al.*, 1987d.

Stop C5: Polyanì

Indirect-indicators of neotectonic evolution

Key reference: Mariolakos et al., 1987e

Lying at an altitude of 650 m., this polje is actually placed between two large grabens, of Kato Messinia in the south and Megalopolis in the north.

This karstic form has developed in cretaceous carbonates and has subsequently been filled with fluvial deposits that comprise silt, clay and conglomerates. The latter display high percentage of metamorphic pebbles that come from Arna Unit (phyllites and quartzites). The thickness of the deposits is locally more than 100m; their age is Middle to Upper Pleistocene and are partially covered by talus deposits, the composition of which markedly differs from that of the underlying sediments, as they consist mainly of fragments of clastic composition (from the eocene flysch of Tripolis Unit).

After examining the distribution and composition of the fluvial deposits in the polje, the following observations could be stated (Fig. C5-1):

- Pebble size decreases downstream the Tzirorrema torrent, which flows through the polje, from 50 cm initially to 8 cm the southern end.
- The ratio of metamorphic to non-metamorphic pebbles decreases significantly downstream. At the northern end of the polje, the conglomerates consist almost exclusively of metamorphic clasts but heading southwards the participation of carbonatic and clastic pebbles increases.
- In three consecutive sampling locations downstream of the polje, (Nos 2, 3, 4 in Fig C5-1) the ratio of metamorphic/non metamorphic pebbles in the pleistocene deposits increases towards the lower members of the deposits.

The evolution stages of the polje are presented in Table 1.

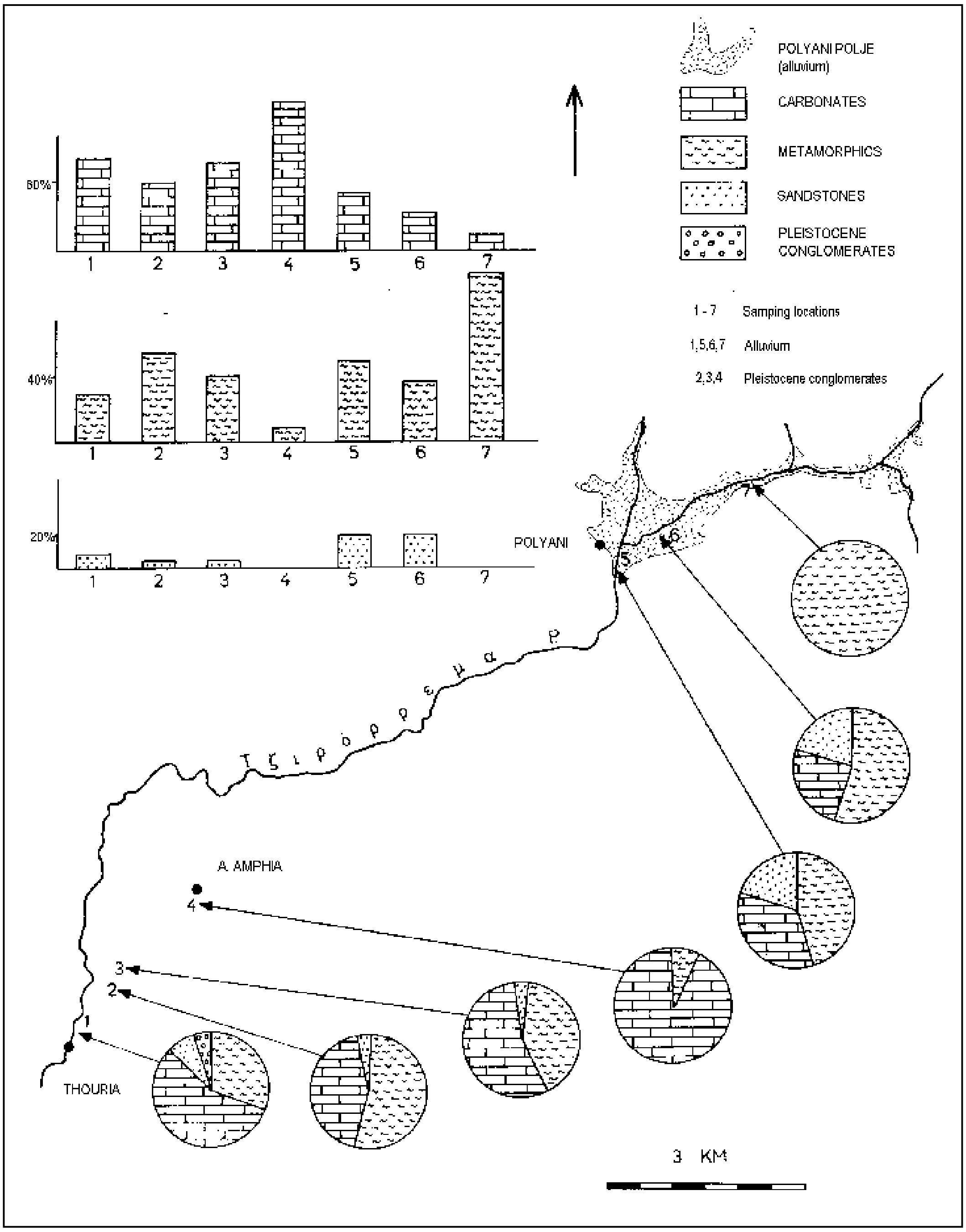


Fig. C5-1 Location sites and percentage of clasts per site. Sites 1, 5, 6, and 7 have been sampled from alluvium; sites 2,3 and 4 have been sampled from the pleistocene marine conglomerates. After Mariolakos *et al.*, 1987e.

Table 1: Evolution stages of Polyani polje.

Stage	Event	Age
VII	present-day condition	Quaternary
VI	Fault reactivation upstream and downstream of the polje Local uplift Onset of incision and transformation to open system Completion of sediment accumulation	
V	Continued influx of (mainly metamorphic) material Closed system, calm periods (red clays)	
IV	Fault reactivation Onset of sediment influx - composition unknown Closed stage - Polyani becomes a polje Closed geomorphological and hydrological system Open hydrogeological system, possibly temporally closed	U. Pleistocene
III	Fault reactivation First stages of polje formation Erosion dominates Open system, at least partially	L. Pleistocene
II	Faults Erosion - formation of tectonically controlled valleys Open hudrological, hydrogeological and geomorphological system	U. Miocene
I	End of tangential tectonics	M. Miocene

Stop C6: Kehreai

Holocene submergence

(*Key references*: Flemming, 1978; Scranton & Ramage, 1967)

Classical, Hellenistic and roman constructions (dated 200 BC to 400 AD) are now submerged: Archaeological excavations revealed remnants of a once prosperous seaside town lying now approximately 2 m. below present sea level. Most of this submergence may be related to seismic activity, perhaps at the end of the 4^{th} century AD.

Interlude II

Neotectonic Evolution of the Isthmus of Corinthos

Key references: Freyberg, 1973; Mariolakos, 1976; Mariolakos & Stiros, 1987

The major area of Corinthos consists of post-alpine formations. It represents a neotectonic graben, bounded by two marginal, E-W striking fault zones, namely the Gerania Mt. fault zone to the North and the of Onia Mt. f.z to the south (Fig. II-1a).

The Isthmus of Corinthos is a narrow strip of land that connects the Peloponnessos to the Hellenic mainland and marks the easternmost limit of the Corinthiakos Gulf. It is located in the above mentioned graben, and consists of a succession of uplifted, well-exposed Pliocene marls unconformably overlain by several cycles of near-shore Pleistocene conglomerates.

Neogene and Quaternary sediments are cut by numerous, NE-SW and E-W striking normal faults. Some of these can be observed along the 6.3-km-long canal, cut at the end of the nineteenth century along the path of Diolkos, an ancient ramp used as a vessel transportation route connecting the Corinthiakos and the Saronikos Gulfs. Because these faults belong to two different sets and dip to the north and south, respectively (section A-A' in Fig.II-2), the Isthmus has been interpreted as a horst; i.e., as a simple, extensional feature (Philippson, 1892).

An alternative interpretation based on the mapping of the unconformity between the Pliocene and Pleistocene deposits (Eutyrrhenian) was given by Freyberg (1973). Detailed geologic mapping and borehole and geophysical data were used for the compilation of this surface, shown in Fig. II-3. The deformation of this originally horizontal surface to its present-day complex shape is presumed to express the Quaternary deformation pattern in the area. Or, as Freyberg (1973) stated, "*The Isthmus...seems to be formed by anti-tilted fault-blocks. The northern system of the block-faulted area has its greatest height in the East and dips below sea-level towards the West ("Tieflage") while the system in the South acts reversely, and between them, a neutral zone, tilted to a lesser degree, exists*" (Fig. II-3). Thus, a NW-SE section shows an apparent horst structure, whereas any section along a southwest-northeast axis would disclose a graben (Fig. II-2). The deformation of the Isthmus, the cover of which consists of clay and sand, does not favor brittle fracturing and is accommodated by numerous parallel, homothetic and antithetic normal faults, most of them active, which impart a pseudoplasticity to the crust in this area.

The area of the Isthmus is tectonically active. In the past 60 years, three earthquakes (1928, 1953, and 1981) have been associated with moderate to minor surface faulting (Fig. II-1a; Mariolakos and Stiros, 1987).

Lithofaga molluscs and beachrock are exposed at a height of about 1.1 m above the present sea level at the ancient (2000-yr-old) harbor constructions of Leheon; they disclose emerging beaches that can be followed up to Corinthos. Freyberg (1973) reported rounded pottery sherds of undetermined age in this area (Neolithic to Roman) cemented in the conglomerates of former shorelines. The latter are disappearing east of Corinthos, while at Possidonia, beachrock covers some fourth century BC constructions of the *Diolkos*, and an ancient platform next to it is cemented to 0.75 m above sea level. Farther north, at Loutraki, the same, probably beachrock structure is at a height of 1.0 m. At the eastern exit of the canal, submerged ruins of the ancient harbor of Schoenous have been identified at the modern site of Kalamaki. There are also data showing that the westernmost portion of the Saronikos Gulf coast is submerging; the harbor of Kehreai (5 km south of Kalamaki) is the best known example.

The long-term (Quaternary) and short-term (late Holocene to present) data discussed above are summarized in Fig. II-4.

According to Mariolakos & Stiros (1987), the Corinthiakos Gulf separates a bulging area (anticline?) to the south and a structural depression to the north, the Isthmus marking their easternmost limit and an area where the associated deformations are minimum and easily observable. The depression to the north may extend as far as the Ionian Islands, which are characterized by pre-Quaternary reverse faults and a Holocene or even older tilt antithetic to that observed in the northern part of the Isthmus (Fig.II-5).

This analysis presumes that back-arc compression is not confined to a narrow zone hardly covering the westernmost end of the Peloponnesos, where minor contraction features have been reported (Mercier *et al*., 1979), but may extend as far as the isthmus. In the northern Peloponnessos, however, this compression should be mild, thus producing only a very long wavelength bulging. This is contradictory to previous concepts that east-west normal faults and focal mechanism solutions of shallow earthquakes signify that this area is under extension (Mercier *et al*., 1979). However, these analyses fail to explain the differential vertical motions along the Corinthiakos Gulf, as well as the antithetic tilting of the Isthmus strata. Furthermore, the fact that the crust in the northern Peloponnessos is up to 2.5 times thicker than in the Aegean, to the east (Makris, 1978), may suggest that the former is an area of crustal shortening.

Tselentis and Makropoulos (1986) computed the deformation tensor in the Corinthiakos Gulf by using large (M>5.5) earthquake fault-plane solutions, and showed that north-south extension is not the dominant mode of deformation in this area. Their data seem to corroborate some north-south as well as some east-west extension, but the corresponding values are very small relative to the dominating deformation (uplift and subsidence of the southern and northern coasts, respectively) and may be insignificant, as no standard deviations are computed. However, some north-south extension, as a consequence of the east-west back-arc compression, is likely to exist. East-west-striking normal faults may originate from this second-order extension, especially since older lines of weakness are reactivated (Sebrier, 1977). However, things must be more complicated, because normal faults and focal mechanisms contain some characteristic strike slip.

Hatzfeld *et al.* (1990) computed 16 focal mechanism solutions of shallow earthquakes that show N-S mostly normal faulting, while few of them are strike-slip. However almost none of the normal mechanisms corresponds to pure dip-slip, but most of them are oblique-slip.

The previous discussion suggests that the well-documented torsional deformation of the Isthmus is taken up by steep, parallel normal faults. Consequently, these normal faults, as well as the normal faults flanking the Corinthiakos Gulf, are not indicative of regional extension, as was previously believed.

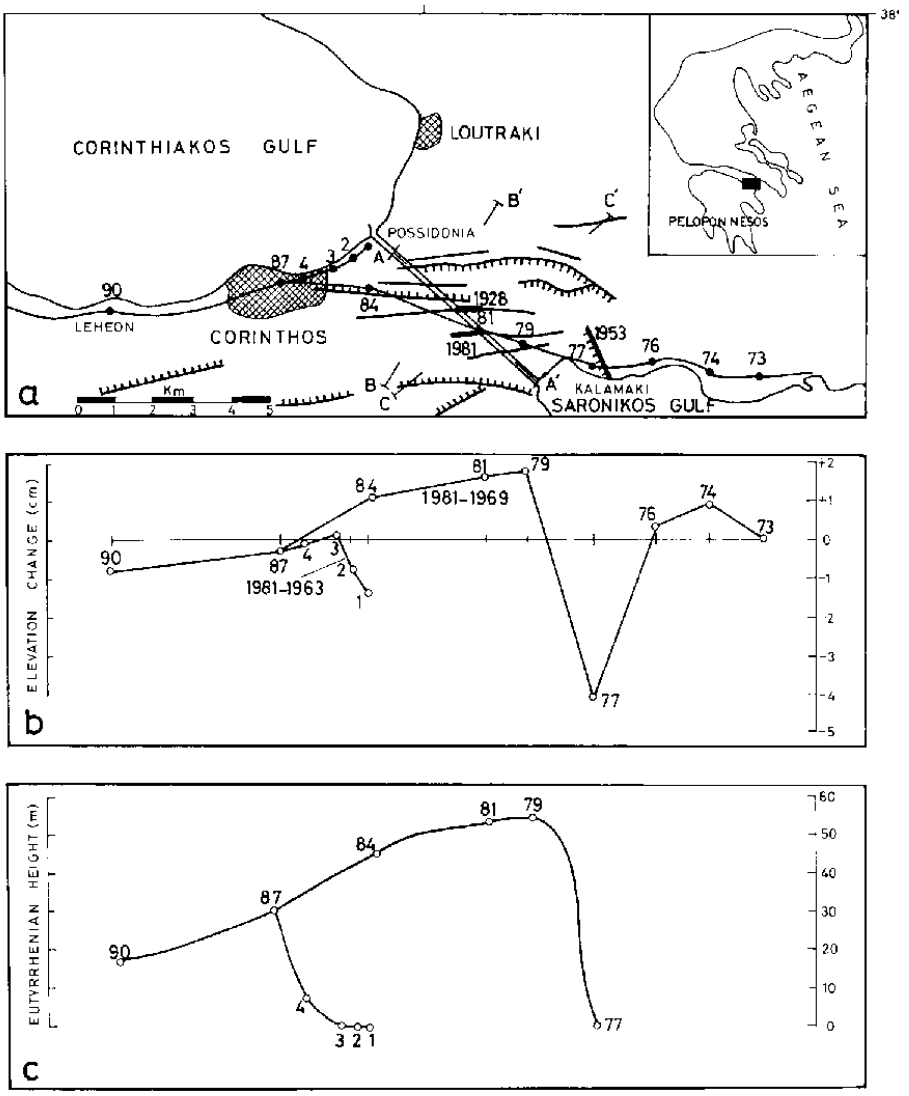


Fig. II-.1 a: Location map. Main faults, seismic faulting (thick lines), leveling benchmarks common in 1969 and 1981 topographic, and locations of cross sections of Fig. 4.2 are shown (after Mariolakos and Stiros, 1987)

b: Ground-surface displacement for the period 1969-1981: Central part of Isthmus (between benchmarks 79 and 84) appears uplifted relative to near-coastal areas. Subsidence of benchmark 77 is consistent with the motion expected at the hanging wall of the faults that ruptured in 1953 and 1981, as well as with a general subsidence tendency along the Saronikos Gulf. During the intersurvey period, the only important earthquakes that affected the Isthmus were the 1981 events. These earthquakes were associated with normal faulting about 10 km north of the study area. Therefore, any motion observed is likely to be associated with local tectonics.

c: Elevation change of the Eutyrrhenian layer along the leveling route. The resemblance of Figure 1b and 1c, presumably showing the present-day and quaternary deformation of the Isthmus, respectively may reveal that the pattern of crustal deformation in this area has not changed since the Early Quaternary.

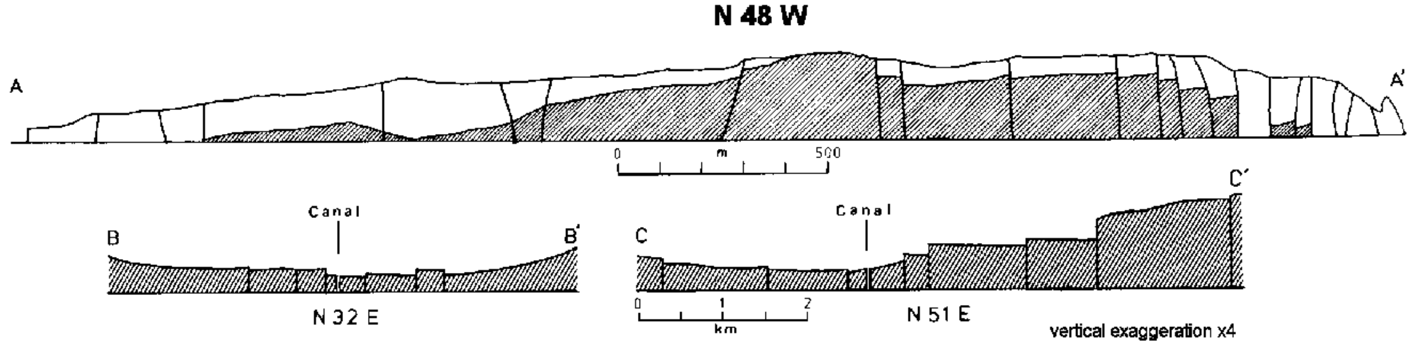


Fig. II-.2 Schematic cross sections of Isthmus (for locations see Figure 4.1). Neogene deposits are shaded, Quaternary deposits are blank. In sections B-B' and C-C' Quaternary cover is too thin to appear at the scale used. (Based on data of Freyberg, 1973).

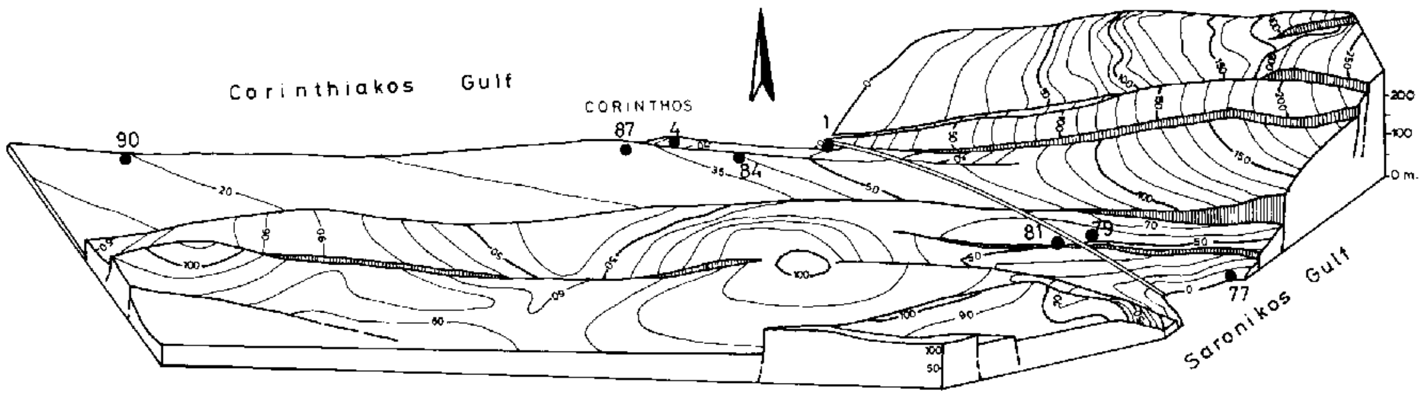


Fig. II-3 Perspective view of contour map of Eutyrrhenian layer height, after Freyberg (1973). Canal and leveling benchmark positions (bold numbers) are also shown (After Mariolakos and Stiros, 1987).

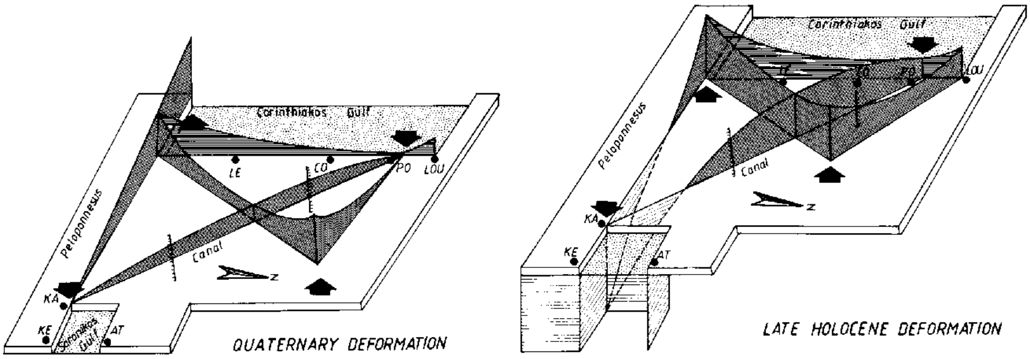


Fig. II-4 Quasi-perspective block diagram of Isthmus deformation (not to scale). Arrows indicate torsional deformation. LE=Leheon; CO=Corinthos; PO=Possidonia; LOU=Loutraki; KA=Kalamaki; KE=Kenchreai; AT=Aghioi Theodori.

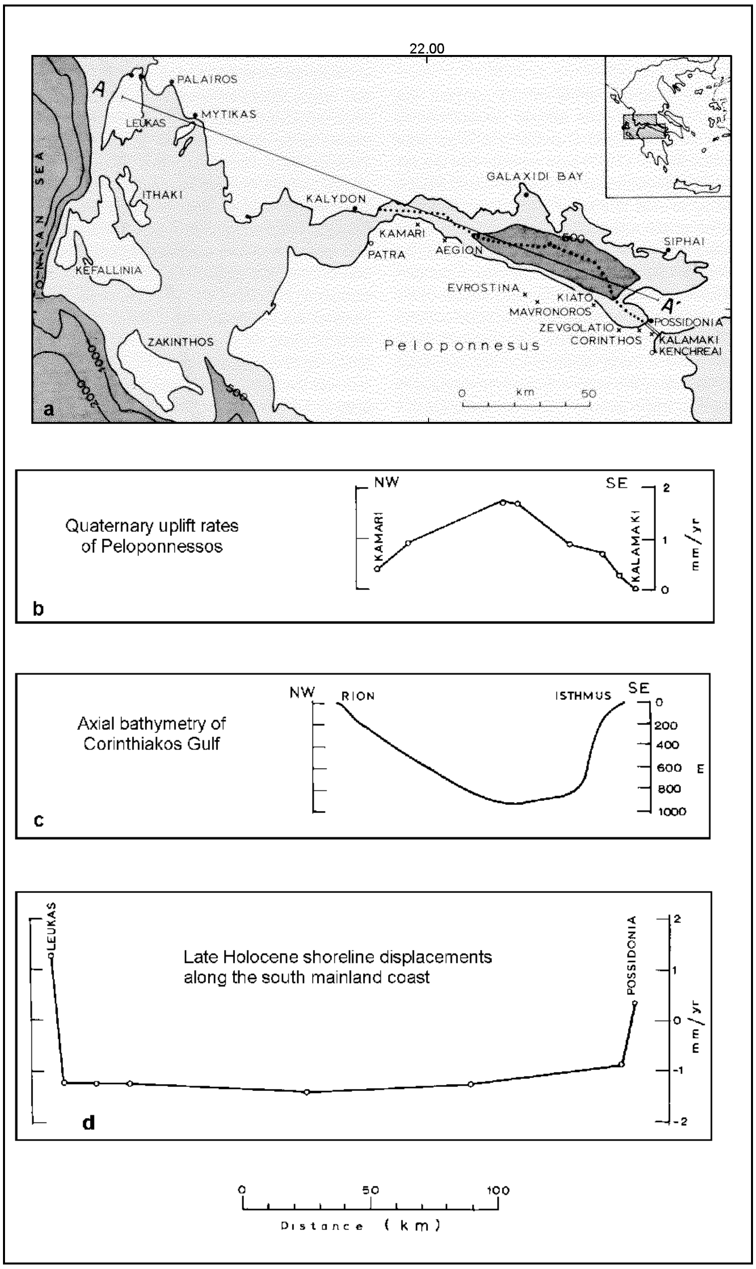


Fig. II.-5 Differential movements and morphological features in Corinthia-kos Gulf combined with deformation of Isthmus. a: location map, b: Qua-ternary uplift rates from sampling locations (marked with ‘x’ in (a)); c: axial bathymetry of Gulf; d: Late Ho-locene shoreline displacement (points correspond to solid circles in (a)) (After Mariolakos and Stiros, 1987).

Stop D1: Isthmus Canal

Ancient Corinthos and Diolkos

Key reference: Mariolakos & Schröder, 1979

A few kilometres west of the modern town, the Ancient town of Corinthos is founded partially on the Neogene formations and partially on the Quaternary deposits (Tyrrenian). The construction material is mainly pleistocene calcareous oolitic sandstone (known as poros) that outcrops near the city of Corinthos.

Diolkos was a ramp connecting Corinthiakos to Saronikos Gulfs, on which the ancient vessels were pulled from one side to the other in order to avoid the circumnavigation of Peloponnessos. (Fig.D1-1). It should be noted that there were people who profited from the existing harbors of *Lechaion* on the Corinthiakos Gulf side and *Kenchreae* on the Saronicos Gulf side and supported the *Diolkos*. However, the passage of ships was becoming more and more difficult as commerce was being developed and warfare necessitated the building of larger vessels.

It was Periandros (600 B.C.), the Tyrant of Corinthos, who first conceived the idea to cut the Isthmus by constructing a canal. He soon, however, gave up when he faced the technical problems involved in such a construction. In addition he was afraid of the wrath of Poseidon to whom the Isthmus area was dedicated.

Three hundred years later, *Dimitrios the Besieger* planned to construct the canal, but was stopped by his engineers who, (as mentioned by Strabo), ascertained that the sea level of the Corinthiakos Gulf was higher than that of the Saronikos Gulf and thus a canal opening could cause the flooding of the coasts of Egina, an island in Saronikos Gulf.

Julius *Caesar* and *Calligula* considered again the question of opening a canal. About a century later when *Nero* came to Corinthos to attend the Isthmia festivities, he took the decision to undertake this project. Indeed, the opening started both from the Corinthiakos and the Saronikos Gulf sides. However, after his return to Rome and his death, all work ceased.

Herodes Atticus who later attempted to continue the project, was ordered to stop. Several hundreds years later, another attempt by the Venetians was condemned to failure.

Following the liberation of Greece, a new study was made but the estimated cost was more than the newly formed State could afford. Finally, half a century later on March 29, 1882 the official foundation ceremony took place. The completed canal was inaugurated on July 25, 1893.

The expert advice of some of the authorities of the time was sought and Ferdinand de Lesseps was consulted. The final design of the cut and the execution of the entire work was undertaken and brought to completion by the Hungarian civil engineer Bela Gerster.

Isthmus Canal - Technical data

Total length 6.300 m of which 540 m comprise the outer port.

The slopes at their base are walled by stones for a length of 3.500 m.

Width at the bottom: 21 m.

Width at the surface of the sea: 24.6 m.

Depth (uniform): 8 m.

Max. height: 79 m.

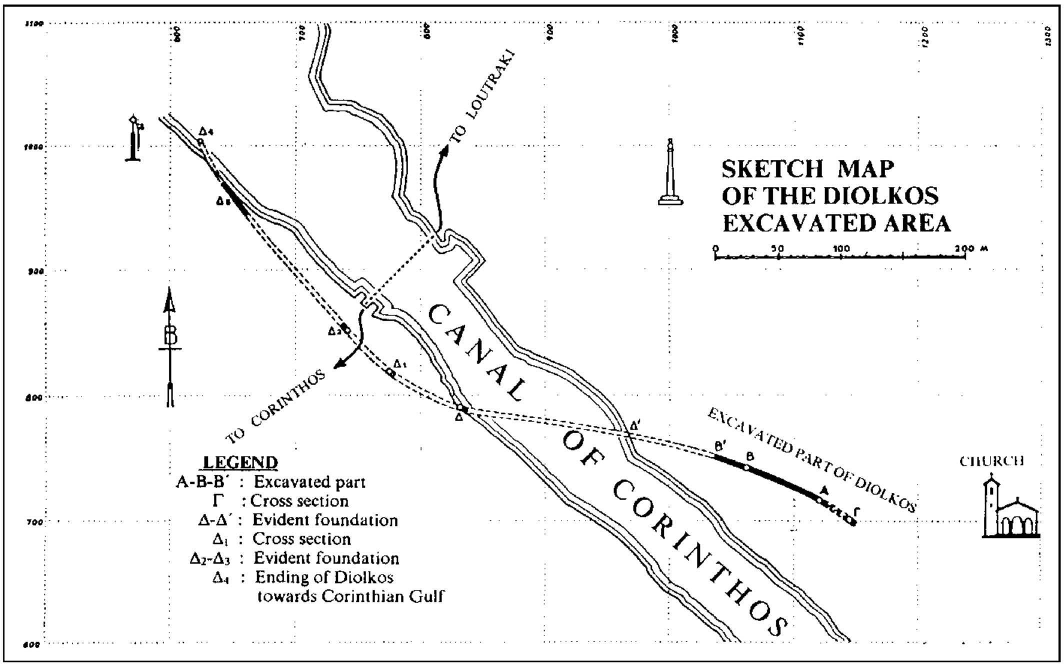


Fig. D1-1 The Isthmus with Canal and Diolkos.

Stop D2: Pissia fault

Fault-Plane morphology - slip plane phenomena

Key reference: Stewart & Hancock, 1988

Pissia fault is located in Perahora peninsula, a few kilometres north of the village of Perahora.

The local tectonic grain comprises a succession of sub-parallel tectonic horsts and grabens, formed by E-W to NE-SW fault zones (Fig. D2-1) The fault forms part of a right-stepping en echelon fault zone, which also include Schinos fault (Stop D3). Segments of this fault zone (Pissia and Schinos faults included) were reactivated in the 1981 Alkyonides earthquake series.

The fault zone cuts neritic Mesozoic limestones, while thick colluvial apron and alluvial fan sediment cover has developed on its hanging wall. The fault surface itself displays a variety of slip-plane phenomena, as corrugations, gutters, pluck holes and comb fractures, while a very thick cataclastic zone has developed on the footwall (Fig. D2-2).

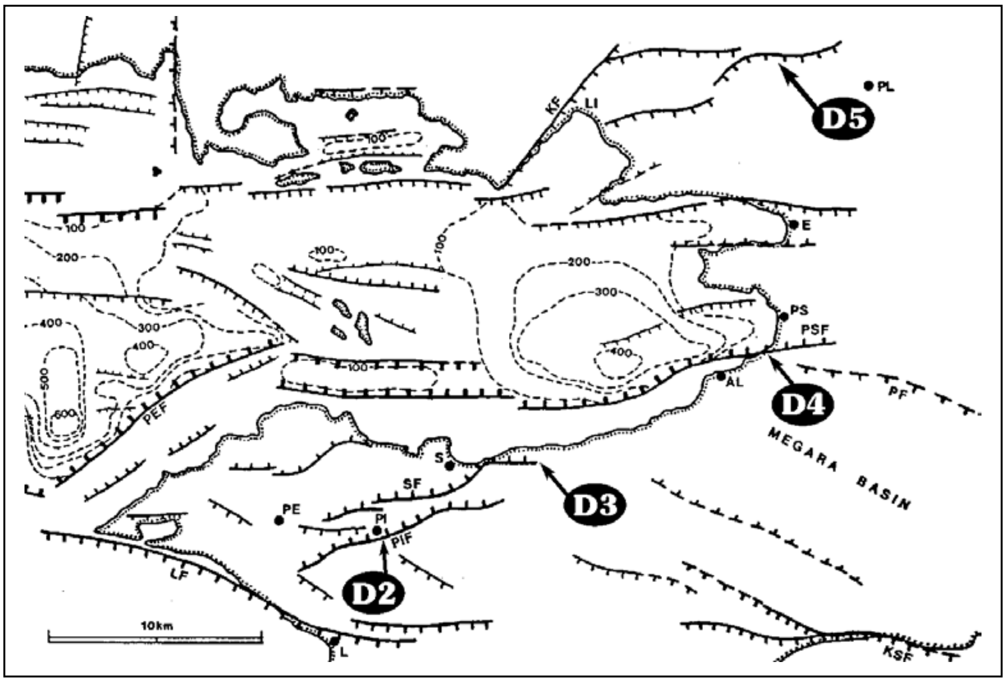


Fig. D2-1 Tectonic map of the Perachora peninsula and the Alkyonides Gulf (after Sakellariou *et al.*, in press). LF: Loutraki fault, PEF: Perahora f., PIF: Pissia f., SF: Schinos f., PSF: Psatha f., KF: Korombili f., KSF: Kakia Skala f., Pateras f. Thickness of fault lines indicates magnitude of vertical throw; also shown are marine sediment thickness contours (dashed). D2, D3, D4 and D5 indicate locations of the stops in the field trip..

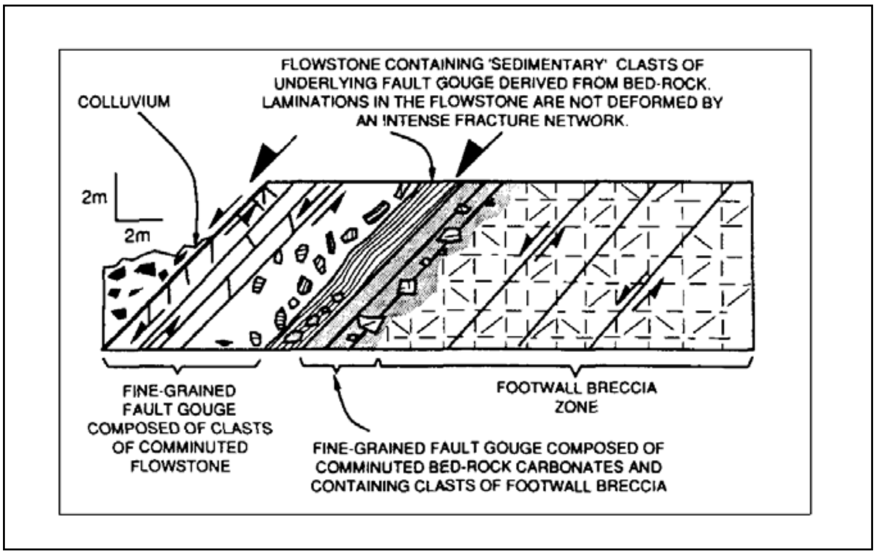


Fig. D2-2. Schematic cross section of the Pissia fault breccia zone (after Roberts & Stewart, 1994).

Stop D3: Schinos fault

Fault reactivation

Kay references: Pandosti et al. 1996

Schinos fault is an E-W to SW-NE trending structure that was reactivated in the 1981 Alkyonides earthquake series, following a pre-existing tectonic line. A series of fractures were produced along its strike, some of which were detected within the Bambakies alluvial fan, which consists exclusively of ophiolitic clasts. The fractures were en echelon arranged, while a small antithetic scarp was also produced.

Trenching investigations in the Bambakies fans suggested that the 5-m cumulative displacement of that has been produced by three earthquake events, since AD 590, with an average recurrence interval of 700 yrs (Fig D3-1).

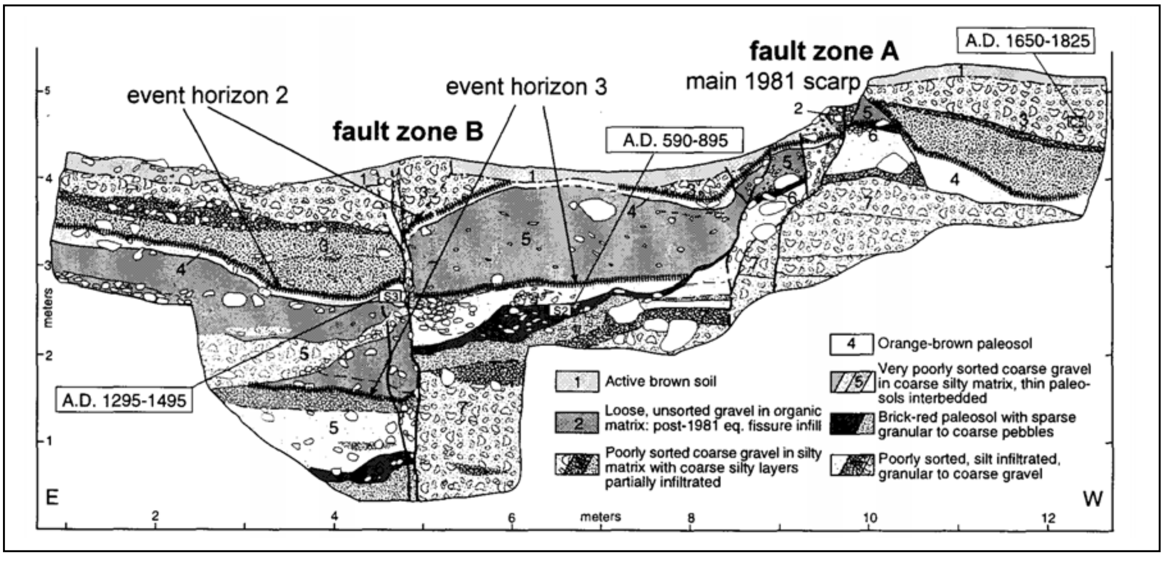


Fig. D3-1. Trench log at the Bambakies fan (after Pandosti *et al*., 1996)

Stop D4: Psatha fault

Active faulting - fault reactivation

Key references: Mariolakos & Fountoulis (submitted)

Psatha fault is located between the easternmost margin of the Gulf of Corinth (Greece) and the westernmost margin of Pateras Mt., It is the southern boundary of the graben of Psatha bay. It strikes NE-SW and cuts the southern marginal fault zone of the Pateras Mt. horst which strikes NW-SE, and forms the boundary between the Pateras Mt. Horst and Megara half graben. It is located in the transition zone between a horst (Pateras Mt.) that consists of neritic limestones of the Sub-Pelagonian alpine geotectonic unit and a graben (Gulf of Corinth) that has been filled with Plio-Quaternary deposits.

Psatha fault is a very interesting case study, as it displays a series of Quaternary reactivation events. In the area of Psatha Bay, the Pleistocene deposits consist mainly of debris cones and alluvial fans. consist of calcareous angular pebbles arranged in layers, slightly consolidated in the upper part whereas in the lower parts are more cohesive. Very often limestone blocks occur within these deposits. The thickness of the deposits is estimated to be more than 60 meters, while the age is most probably Würm.

The shape of the fault surface is not planar but curved. The tectonic breccias observed on the fault could be distinguished at least in two different categories taking into account the lithology of the fragments they consist, the matrix and their relative age. The oldest tectonic breccia is compact and oligomictic and consists exclusively of very small angular particles of the neritic limestones. The thickness of this breccia sheet some centimeters and can be observed all over the fault surface either it is eroded or not and is covered by a thin calcitic film, the thickness of which is usually less than

The relative younger tectonic breccia is polymictic and consists of fragments of post alpine deposits of Plio- Quaternary age (sands, silts, etc.) as well as of neritic limestones, Locally, fragments from the older tectonic breccias are present as well. The size of the angular coarse material varies from 1 to 5 cm, the origin of the cement is from the Plio- Quaternary deposits. The thickness of this breccia is more than 2 meters.

The structure indicating successive reactivation of the fault is the presence of different direction striking and dipping striae. They are distributed all over the non-eroded fault surface regardless if the fault cuts limestones, older or younger tectonic breccias. The older striation set seems to be that of first (I) set, plunging 65/300. It occurs always on the older oligomictic breccias surface with very high frequency.

The next slickenside generation – whose frequency is relatively low – is that of the second (II) set, plunging 56/320. They occur not only on the surface of the oligomictic breccias, but also on the surface of the polymictic breccias and especially on pebbles that come from the limestones.

The third (III) generation of slickensides occurs on fault surfaces cutting the polymictic tectonic breccias, plunging 50/330. The slickensides have been printed especially on pebbles and gravels of the polymictic breccias coming from the neritic limestones.

Taking all the above mentioned into account, which are (1) the curved fault surface, (2) the successive breccias (3) the oblique slickensides (4) the oblique-slip character of the reactivated faults during the seismic activity of 1981 (Mariolakos *et al*., 1981), and (5) morphotectonic studies that were conducted after the earthquake activity of 1981,we believe that the deformation of both the narrow and the major area is not that of pure extension but it is most complicated namely of a rotational couple stress field.

Stop D5: Kaparelli

The Alkyonides earthquakes of 1981
Fault reactivation

Key reference: Mariolakos et al., 1981

The earthquakes of February-March 1981 caused severe damage over a very large area comprising the major part of Corinthia, Viotia and Attica. Displacements along fault zones and along coastlines or creation of new fractures were observed mainly at the Perachora peninsula and along the coastal area of the eastern Corinthiakos Gulf (Loutraki, Alepochori, Psatha, Aegosthena) as well as at the area of Kaparelli, Platees and Erythres.

Four fault zones were reactivated during the earthquake sequence of February-March 1981, namely, Platees-Kaparelli f.z., Kalamaki f.z., Schinos f.z. and Pissia f.z. These four fault zones can actually be grouped in two, en echelon arranged. In this sense, the

The common features of the major fault zones are the following:

- All 4 fault zones show a mean E-W strike.
- The active faults follow older ones; which means older tectonic lines were reactivated.
- All fault zones are accompanied by a large number of microstructures .
- Although it is the E-W trending faults that we have been generally reactivated, sometimes faults of other direction have been reactivated, as in the case of Kaparelli-Platees fault.
- The reactivated faults of normal character, with oblique net slip with a dextral movement.

The ratio of the dip-slip component to the strike-slip component is about 10/3, as it can be detected all along the fault zones. The vertical slip of the faults varies from place to place and it is generally 50-60 cm. The largest amount of vertical slip was 1,60 m. and was recorded at the intersection of two E-W and NW-SE trending faults between Kaparelli and Platees

Although the normal character of the faults prevails, from the dynamic and kinematic point of view they are the result of coupling and shearing which are the local analysis of regional torsional stress-field as this illustrated in Fig. D5-1.

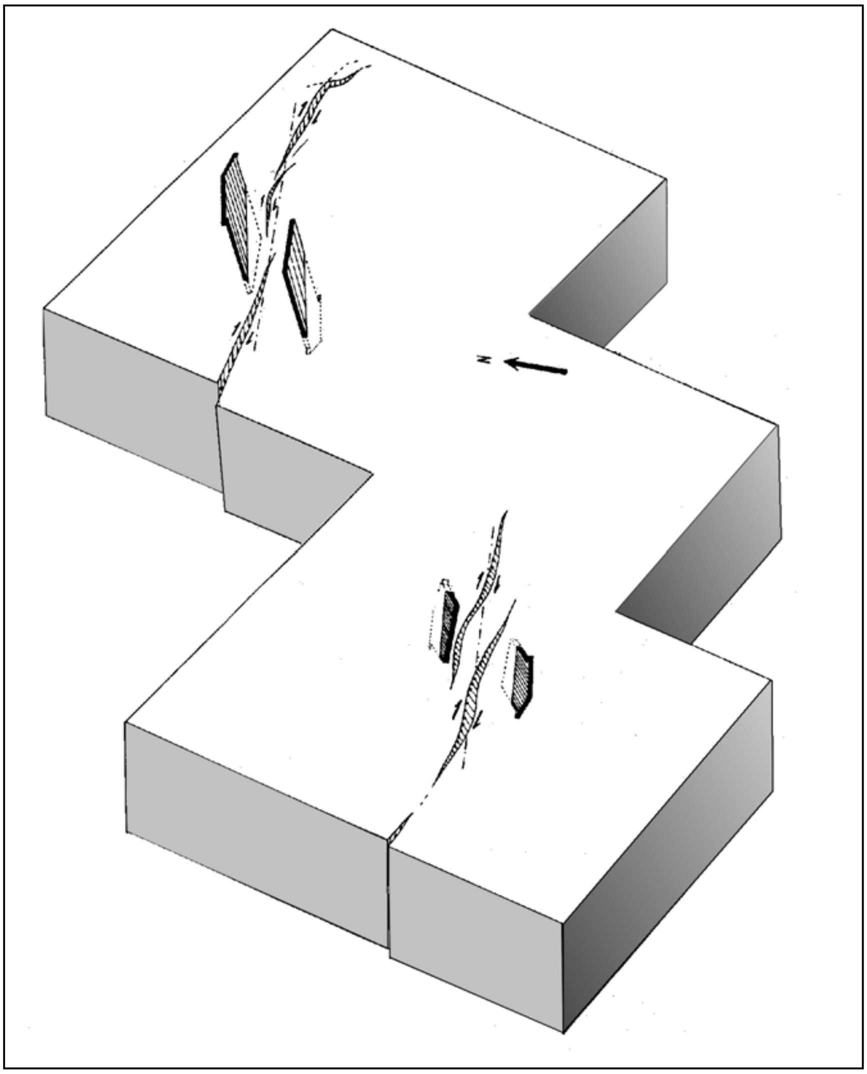
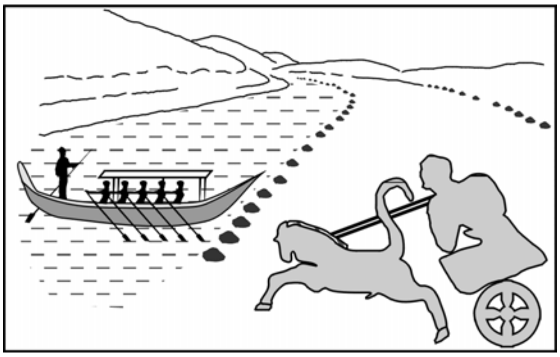


Fig. D5-1 Schematic block diagram (not to scale) to show the relative displacement along the reactivated fault zones and the resulting stress pattern. (after Mariolakos *et al*, 1981)

Stop D6: Orhomenos

Minyan Ancient Flood Prevention Works (c. 1600 BC.)

Key reference: Knauss, 1984



The flood prevention - drainage works in Lake Kopaida are the earliest in Europe, since their construction dates back to 1600 BC. The technique implemented by the ancient people of Minyes is the same as the one currently taught at the Higher Education Institutions all over the world. These works, together with the mines in Lavrio, are the most important archeological sites suitable for the investigation of the technology of the ancient times; however the mining facilities in Lavrio are much younger (c. 800 BC.).

The largest navigable canal of the ancient times (27 km long) was constructed and, according to Knauss, it was used for product transportation (Fig. D6-1). The water level at some parts of it was 1.5 m. above the adjacent areas that were used for cultivation and the embankment was totally impermeable (Fig. D6-2). The concept of the construction of that floodway is totally different from the one applied by a British company at the beginning of this century; the latter bore significant drawbacks, and subsequently numerous problems arose.

After the construction of the canal by the Minyes the water level at some parts of it was 1.5 m. above the ground surface (Fig. D6-2), which was used for crops. The works were totally impermeable, and their stability was excellent. It should be noted that the present national road connecting Kastro and Orhomenos is partially constructed on the embankment.

These works are related to the stage(s) of climatic optimum(s) in the Holocene and any further archaeogeological study should be valuable for the estimation of climatic fluctuations during the last 10 ka. Events as Noah's flood, or the less familiar Deukalion's flood, reflect such climatic conditions, which are, in turn, related to flood-prevention works. Knowledge of such climatic changes is important for the contribution to the prediction of such future changes.

The disappearance of the Minyes and their civilization is connected to the destruction of their works by the thebian Hercules, who sealed the entrance to the sinkhole that funneled the water carried by their canal with a boulder. So the whole area of Kopaida was inundated again, destroying all flood-prevention works, crops, and finally all the towns built on the flanks of the valley. Obviously, the cause of the fall of the boulder was an earthquake. This boulder can nowadays be seen still blocking the entrance to the sinkhole and the site has been proposed for inclusion in the international Geotope list of UNESCO. We should also say that the whole area between the springs of Orhomenos to the town of Kastro (ancient Kopai) and the sinkhole of Neo Kokkino could be developed into a unique Earliest European Flood Prevention Technology Park.

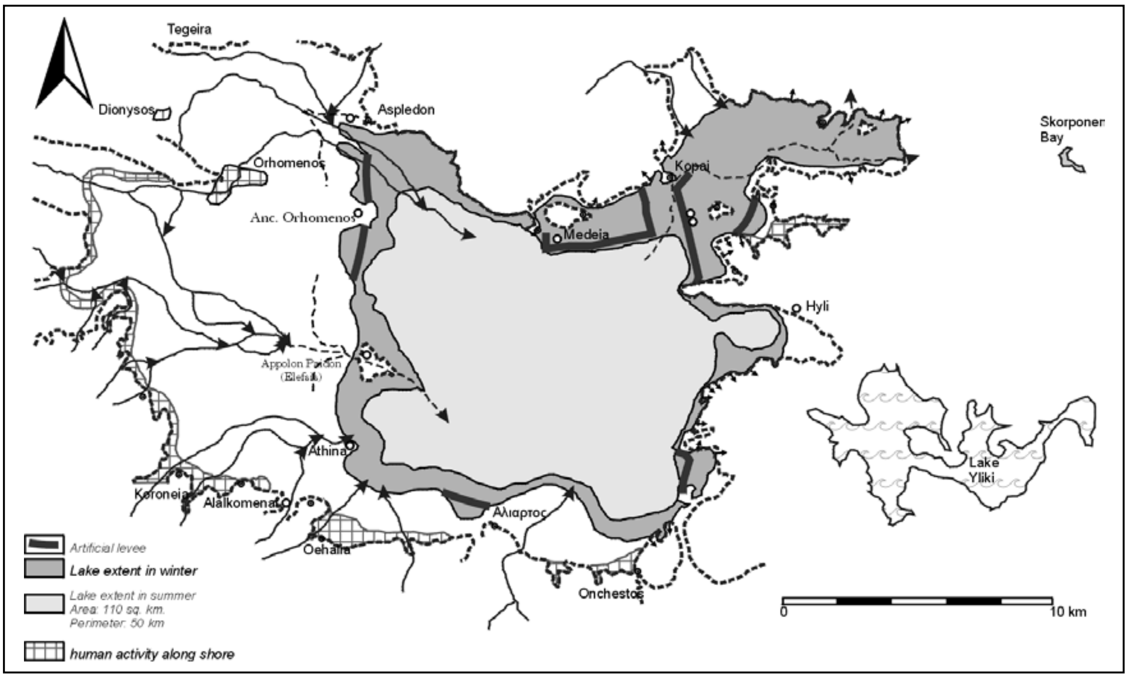


Fig. D6-1 Hystero-Helladic period reconstruction of lake area (dark grey: winter highstand, light grey: summer lowstand). Also shown the artificial levees and extent of ancient towns-hamlets.

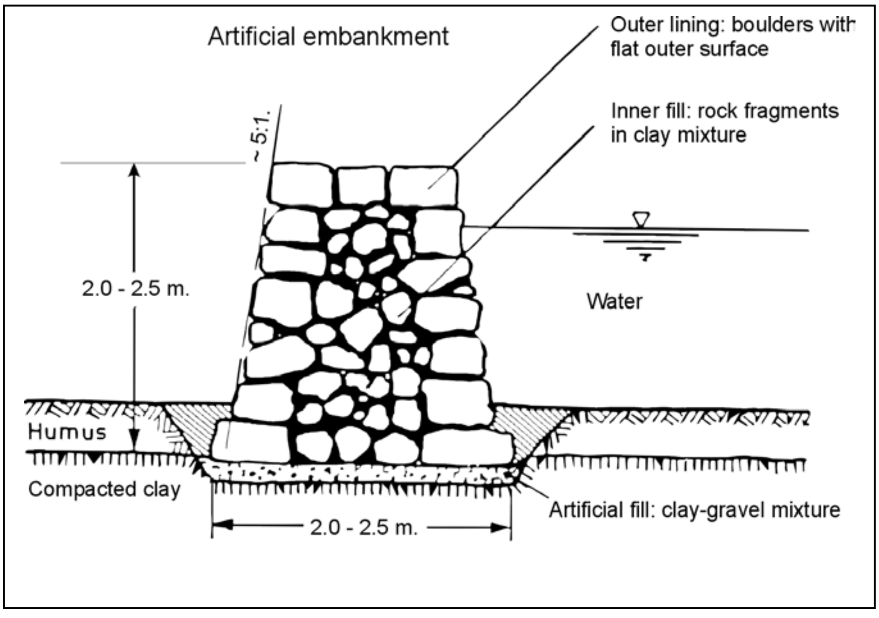


Fig. D6-2 Cross-section of artificial embankment.

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