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Quaternary basin sedimentation and geodynamics in SW Peloponnese (Greece) and late stage uplift of Taygetos Mt.

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ABSTRACT In this paper we study the geodynamic phenomena that affected SW Peloponnese (Greece) during the Quaternary and we propose a neotectonic evolution for this region from Late Pliocene and onwards. As SW Peloponnese is located at the western part of the Aegean arc system, a few kilometers away from the Hellenic trench, it represents one of the most tectonically and seismically active areas of the Africa-Eurasia collision zone. The study area is a part of the External Hellenides where regions dominated by extensional tectonics transit to regions with compressional tectonic regime. Thus, it offers the opportunity to understand how this transition in different stretching regimes is expressed and study the interplay of extension and compression in basin evolution. These different regimes interact or superimpose each other producing several geodynamic phenomena such as seismicity, normal faults, crustal uplift, folds and detachment faults. More specifically, we are focusing on the following aspects: i) the sedimentation of the various Quaternary basins occurring in the study area, ii) the extensional and compressional structures affecting the study area since the Upper Miocene, iii) the neotectonic configuration of SW Peloponnese, iv) the neotectonic evolution of SW Peloponnese during the Quaternary in conjunction with the kinematic evolution of the various macrostructures and their marginal faults. Moreover special attention is given to the interpretation of the occurrence of large size metamorphic pebbles in terrestrial deposits located at the southern part of Pylia Peninsula originated from the metamorphic rocks of Taygetos Mt. as nowadays Messiniakos Gulf (a deep subgraben) isolates these two uplifted landmasses.

Key words: SW Peloponnesus, neotectonic evolution, basin sedimentation, Taygetos Mt.

1. Introduction

Western Peloponnese lies at the western margin of the Aegean continental block and consist the main feature in the western part of the Aegean arc system (Hellenic Arc). The Hellenic Arc is a convergence zone, associated with northward subduction of the African plate beneath the Aegean (Fig. 1a). The western part of Peloponnese is located in one of the most tectonically and seismically active areas of the Africa-Eurasia collision zone. Intense seismicity, normal faulting and crustal uplift are only a few of the several geodynamic phenomena dominating the evolution of the region.

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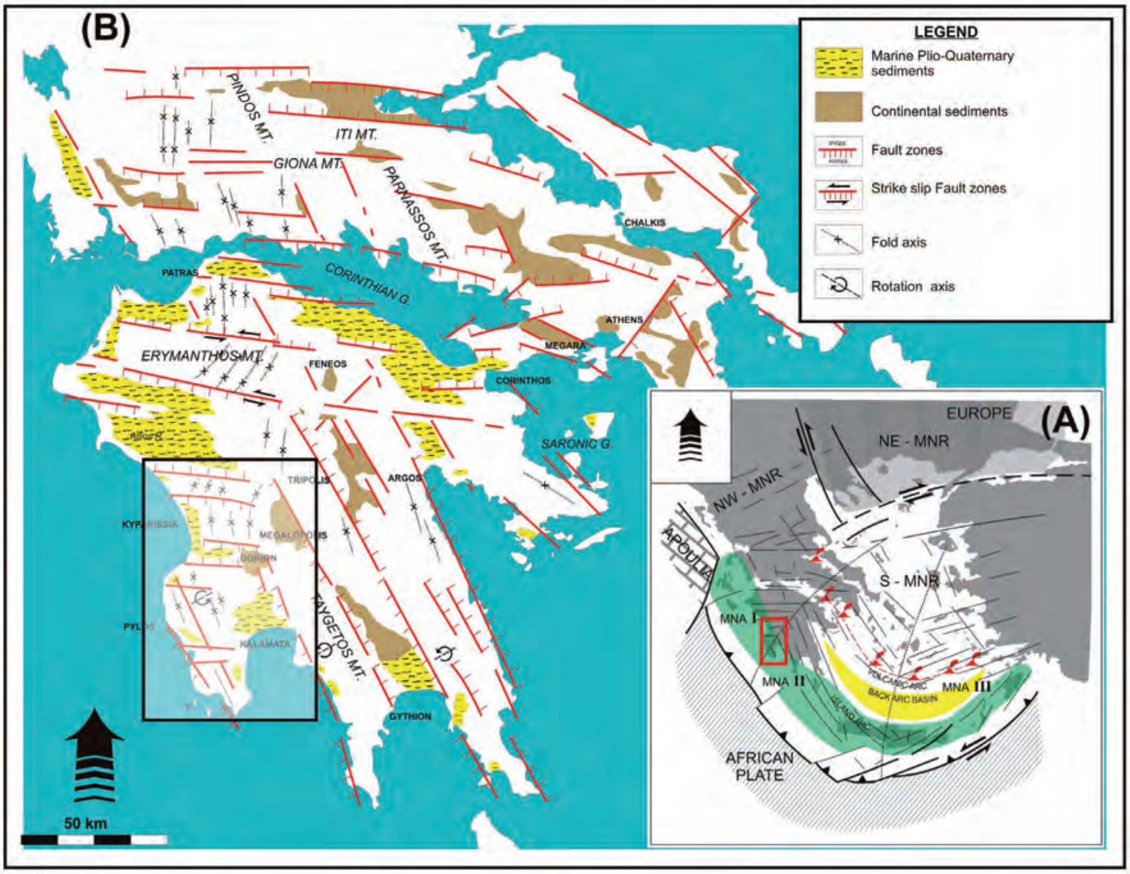


Fig. 1 - The three morpho-neotectonic regions (MNR) (A) as well as the 3 sectors and the main marginal fault zones of the post-Alpine basins in the southern continental Greece (B) [modified by Mariolakos and Fountoulis in Mariolakos *et al*. (2004), from Mariolakos and Papanikolaou (1981, 1987)].

The relief of SW Peloponnese is characterized by the presence of several mountains and drainage basins trending mainly NNW-SSE and E-W (Figs. 1b and 2). The general trend of the major watershed is NNW-SSE, dividing at the eastern part the Alfios drainage network, which flows from SSE towards NNW, from the drainage network of central Messinia, which flows from NNW towards SSE to the Messiniakos Gulf (Fig. 2). At the western part the drainage network flows mainly from east to west towards the Ionian Sea.

The study area is west of Mainalon and Taygetos mountains (Figs. 1 and 2) forming a 80 km long segment of the south-western Peloponnese, trending N-NW and paralleling the Hellenic trench, about 70-80 km to the west (Fig. 1a). The general trend of tectonic features in the area is NNW-SSE, following the geometry of the fold and thrust belt of the External Hellenides (Philippson, 1898; Aubouin *et al*., 1961; Aubouin, 1977; Jacobshagen, 1979) and of the modern Hellenic Arc and trench system (McKenzie, 1972; LePichon and Angelier, 1979). However, the study area (Fig. 1b) is a complex multi-fractured neotectonic macrostructure, which is characterized by the presence of large grabens and horsts bounded by wide fault zones, trending approximately NNW-SSE and E-W (Mariolakos and Papanikolaou, 1981, 1987; Mariolakos *et al*., 1985; Fountoulis, 1994; Fountoulis and Mariolakos, 2008).

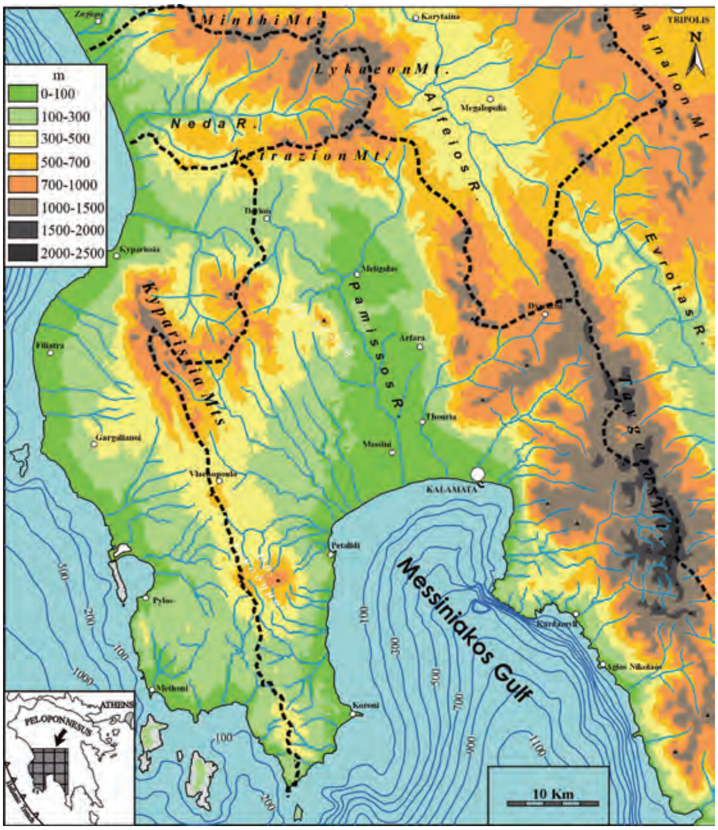


Fig. 2 - Map depicting the relief, the drainage network, the watershed and the isobaths of the study area.

The external part of western Peloponnese is dominated by extensional tectonics as expressed by the E-W oriented normal faults at its central-western part. It has to be mentioned that at the eastern parts of the study area the main fault zones strike NNW-SSE (Fig. 1). The compressional tectonic regime in the depth is not in contradiction with the extensional features observed on shallow structures on the upper plate. Many geophysical, geological, morphotectonic and paleomagnetic investigations were conducted in western Peloponnese, dealing with the crustal structure and the seismic hazard of this area. Mariolakos and Papanikolaou (1981, 1987) and Fountoulis and Mariolakos (2008), claim that the deformation pattern of the area is not only a result of axial extension but also a result of simple shear and torsion.

Extended outcrops of Pleistocene marine sediments are observed within the grabens along the coastal zone up to 15-20 km distance from the coastline and at altitudes of several hundred metres (Kowalczyk *et al.*, 1975; Kelletat *et al.*, 1978; Kowalczyk and Winter, 1979; Fountoulis, 1994; Fountoulis and Moraiti, 1994, 1998; Mariolakos *et al.*, 1994, 1998; Athanassas, 2010). This demonstrates the occurrence of significant vertical tectonic movements during the Quaternary, which permitted the deposition of marine Quaternary sediments in subsided areas and their subsequent uplift and erosion at their present state. At the inner (eastern) parts of the study area Pleistocene lacustrine and continental deposits have filled in the basins. Sequences of uplifted and well-developed marine terraces occupy extensive segments of the south coast, revealing a steady motif of uplift throughout the Pliocene and Pleistocene (Kelletat *et al.*, 1976). Specifically, in the south-western corner of the country, a set of stranded platforms have been mapped (Fig. 1), capped by shallow marine to brackish sediments. Former (Frydas, 1990; Marcopoulou-Diacantoni *et al.*, 1990; Fountoulis and Moraiti, 1994; Kourampas and Robertson, 2000) and current paleontological investigation set the beginning of sedimentation in the Early Pleistocene. Nannoplankton populations of Gephyrocapsa, Pseudoemiliania lacunosa, and Calcidiscus leptoporus, all correlated with the MNN-19e biozone (Rio *et al.*, 1990), bound the chronology of the Quaternary outcrops as far back as 1.6-0.95 Ma.

The extensional character of the tectonism observed near the front of the Hellenic Arc and at a few tens of kilometres above the subducting slab of the east Mediterranean basin provides an exceptional opportunity to study the closely spaced transition from compressional to extensional deformation in an evolving arc (Papanikolaou *et al.*, 2007; Papanikolaou, 2010).

Ductile structures (folds) also occur within the Plio-Quaternary deposits of south-western Peloponnese (Mariolakos and Fountoulis, 1990; Fountoulis, 1994; Fountoulis and Mariolakos, 2008).

The importance of the presence of pebbles coming from the metamorphic rocks occurring in Taygetos and Mainalon mountains in the Quaternary deposits has to be underlined. The present day morphology and drainage pattern does not permit us to give an easy explanation on the occurrence of these metamorphic pebbles in the Quaternary sediments, as the source area (Taygetos and Mainalon mountains) are located in most cases out and far from current drainage basins in which the conglomerates bearing metamorphic pebbles occur. The above mentioned in combination with the Quaternary sediments facies, the source area, the kinematics of the tectonic blocks, the reactivation of the fault zones that bound the tectonic blocks, permits us to understand the paleogeographic evolution and the active tectonics of SW Peloponnese during the Quaternary.

2. Geology and neotectonics

2.1. General

In SW Peloponnese, the following four Alpine geotectonic units from the lower to the upper occur (Fig. 3): a) the Mani Unit which consisting of crystalline carbonates to marbles and metaflysch, occurring mainly in Taygetos Mt.; b) the Arna Unit consisting of metamorphic rocks (schists, phyllites and quartzites) occurring also in Taygetos Mt.; c) the Gavrovo-Tripolis Unit consisting of Mesozoic to upper Eocene neritic carbonates, and Upper Eocene – Oligocene flysch; d) the Pindos Unit consisting of Mesozoic pelagic sediments and Paleogene flysch and e) a small outcrop of the more external Ionian nappe occurring at the western part of Lapithas Mt. near the Zacharo town at the north (Fountoulis and Lekkas, 1991). It has to be mentioned that thick molassic conglomerates of Miocene age occur at the front of the Pindos nappe in the area south of Kyparissia (Mariolakos *et al.*, 1998; Papanikolaou *et al.*, 2007) (Fig. 3).

From the geotectonic point of view, Mani Unit represents the relatively autochthonous unit of the region, which sequentially has been overthrusted by the Arna (phyllite-quartzite s.s.), Tripolis and Pindos units. More specifically the Arna unit has very limited outcrops and they mainly occur in the northern part of Taygetos Mt. (Fig. 3). Southern and western of the present occurrence, the Arna Unit seems to wedge between the underlying Mani and the overlying Gavrovo-Tripolis units as there is no occurrence in the Koskaraka River valley (Fig. 4), nor at the high relief slopes of

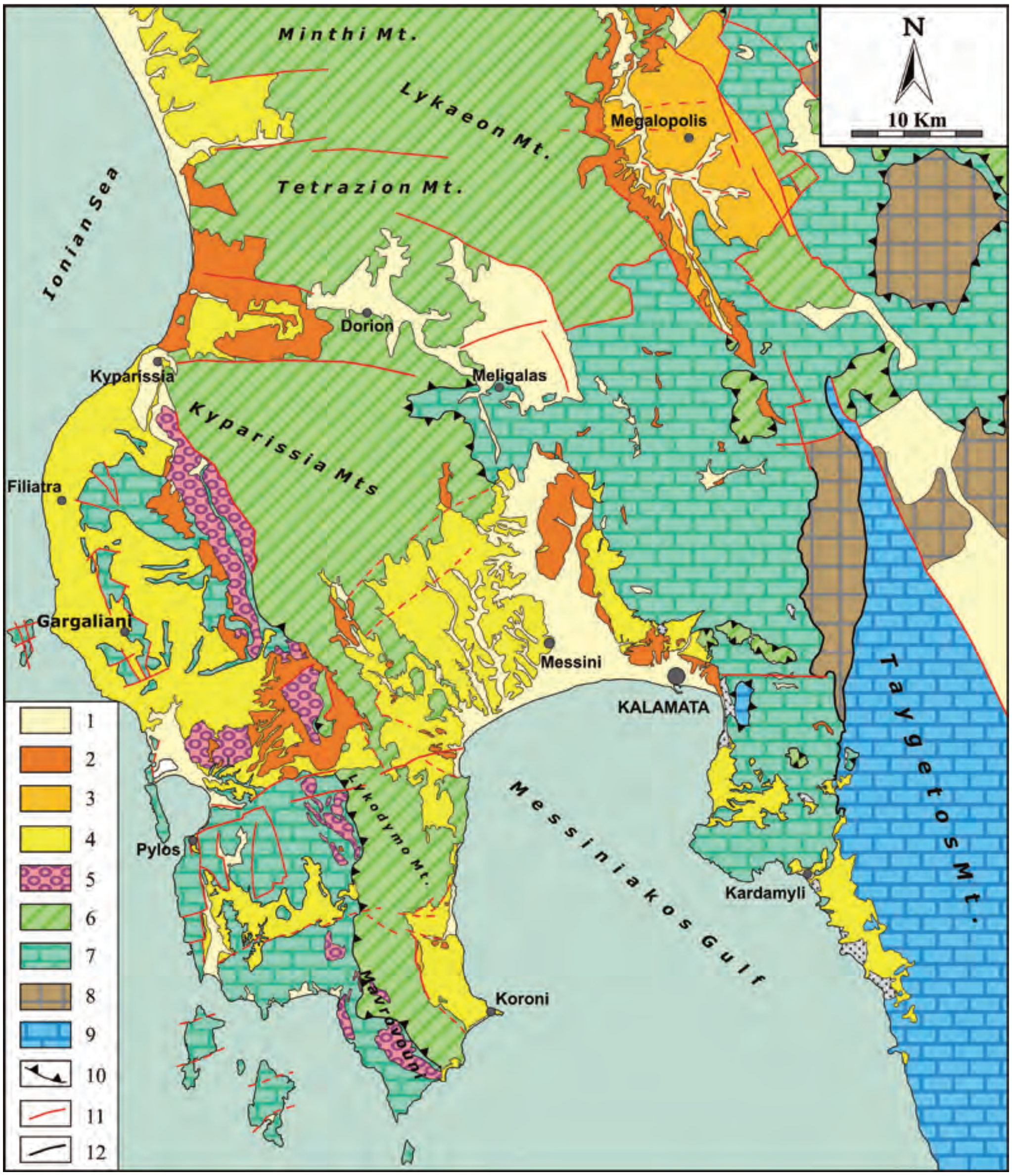


Fig. 3 - Map depicting the geotectonic units and the post Alpine deposits of the study area. 1: Holocene deposits; 2: continental deposits; 3: lacustrine deposits; 4: marine deposits; 5: conglomerates of Messinia (molasse); 6: Pindos Unit; 7: Gavrovo-Tripolis Unit; 8: Arna Unit (Phyllites-Quartzites); 9: Mani Unit; 10: thrust; 11: fault zone; 12: detachment fault zone [based on Fountoulis (1994), Mariolakos *et al.* (1998), Ladas (2000); neotectonic maps scale 1:100,000 central-western Peloponnese, Filiatra, and Pylia, sheets, correspondingly, modified].

Kalathion Mt. In both cases only Tripolis Unit formations occur lying overthrusted directly over Mani Unit formations.

The conglomerates of Messinia are a special noteworthy case. They are cohesive, polymictic with pebbles deriving from the limestones, the radiolarites, and the flysch of the Pindos Unit. They have been deposited mainly over the flysch of the Gavrovo-Tripolis Unit, but even over the Pindos Unit formations. Their thickness in some locations is up to 700 m.

2.2. Quaternary basin sedimentation

A number of Quaternary basins occur in the study area (Table 1). Seven of these basins have been filled in mainly with shallow marine sediments, two of them with continental deposits and one of them with lacustrine deposits.

In most of the marine basins the sedimentation started in Early Pleistocene and lasted all through the Lower Pleistocene (Mariolakos *et al.*, 1994). In some of them, like Zaharo and Neda, sedimentation last until 270 ka B.P. (Fountoulis, 1994; Fountoulis and Moraiti, 1998; Papanikolaou *et al.*, 2007; Fountoulis and Mariolakos, 2008), while the marine sedimentation in Gargalianoi - Filiatra basin ended before 125 ka (Kourampas and Robertson, 2000; Athanassas, 2010).

Table 1 - The main Quaternary basins of south-western Peloponnese (Greece).

	Basin	Sedimentary Facies
1	Zacharo	Marine
2	Neda	
3	Kyparissia	
4	Gargalianoi-Filiatra	
5	Pylos - Kalamata	
6	Koroni	
7	Kampos	
8	Ano Messinia	Continental
9	Dorion	
10	Megalopolis	Lacustrine

In the study area, the Quaternary deposits can be distinguished by their facies in marine, continental (terrestrial) and lacustrine. The lacustrine deposits occur mainly in the Megalopolis basin with a thickness of some hundreds of metres (Vinken, 1965). The Quaternary marine deposits occur in all post Alpine basins of SW Peloponnese, except the Megalopolis (lacustrine), Dorion and Ano Messinia (terrestrial) basins. They mainly consist of marls, sandstones and conglomerates and they have been deposited uncomformably mainly on the paleorelief of the Alpine basement. Thickness varies from basin to basin, being more than 200 m in many sites. The continental deposits consist of oligomictic or polymictic conglomerates, terra rossa, and red siliceous sandstones and sands. They have been deposited unconformably over either the Pleistocene marine deposits, or on the Alpine basement. The age of deposition of every formation varies from Middle to Upper Pleistocene. Following, the main deposits of every basin is presented.

The Zaharo basin lies between Lapithas and Minthi mountains (Fig. 5), is small E-W elongated typical graben that is bounded by the Arini fault zone in the south and the Kaiafa-Koumouthekra fault zone in the north, and is divided in two sub-basins (eastern and western). Post-Alpine sediment formations have filled the Zaharo graben (Hageman, 1977; Streif, 1977, 1978; Mitropoulos *et al.*, 1982; Kamberis, 1987; Fountoulis, 1994). Hageman (1977) considered the sediments in the eastern Zaharo graben (south of Platiana village) to represent the basal post-Alpine strata within the Pyrgos-Olympia basin and correlated the lacustrine deposits with the base of the Makrisio stage sequence (Upper Pliocene) in Megalopolis basin. Kamberis (1987) considers them to be the corresponding Plio-Pleistocene deposits of Vounargo formation in the Pyrgos-Olympia basin. Fountoulis (1994) and Fountoulis and Mariolakos (2008), divided the graben into two sub-basins, the eastern and the western one. No marine post-Alpine sediments (Upper Pliocene - Early Pleistocene) crop out in the eastern sub-basin, while the western one has been filled in with marine and lacustrine post Alpine sediments (Late Pliocene - Early Pleistocene) (Fountoulis, 1994; Papanikolaou *et al.*, 2007; Fountoulis and Mariolakos, 2008). At Lapithas Mt. (776 m), the previously mentioned Pindos Unit, cropping out all the way southwards to Kyparissia, is uplifted and the underlying more external Gavrovo and Ionian units crop out (Fountoulis and Lekkas, 1991).

The Neda basin formed between the Tetrazio and Minthi mountain horsts (Fig. 5), is small E-W elongated typical graben that is bounded by the Neda fault zone in the south and Lepreon fault zone in the north, and has been filled in with post-Alpine deposits distinguished into the following formations: 1) the Elaia formation occurring at the SW part of the basin, consists of Late Pliocene cohesive conglomerates, the pebbles of which come exclusively from the formations of Pindos Unit (Fountoulis, 1994; Fountoulis and Moraiti, 1998), and 2) the Neda formation, which is an Early Pleistocene shallow marine formation, consists of marls, sands, sandy marls and polymictic conglomerates alternations (Fountoulis, 1994; Fountoulis and Moraiti, 1998). The Neda formation is divided into 2 parts, the lower consisting mainly of sandstones, marls, sandy marls and polymict conglomerates alternations and the upper part consisting exclusively of polymict conglomerates the pebbles of which come mainly from Pindos and Gavrovo - Tripolis units. In some locations there are pebbles (less than 1%) from the metamorphic rocks (Phyllites - Quartzites) of Arna Unit (Fountoulis, 1994). The size of the pebbles varies and becomes smaller in a direction from east to west, indicating that the pebbles' source area should be located to the east of the Neda basin. Based on nannoplankton the sedimentation started on Early Pleistocene and finished in Upper Middle Pleistocene [0.27 Ma, Fountoulis (1994) and Fountoulis and Moraiti (1998); site 9 in Fig. 4].

The Kyparissia basin formed between the Tetrazio and Kyparissia mountain horsts (Fig. 5), is small E-W elongated typical graben that is bounded by the Kyparissia fault zone in the south and Kalo Nero fault zone in the north contains, at its base, upper Miocene - Lower Pliocene marine sediments covered unconformably by Pliocene oligomictic conglomerates and Pleistocene marine sediments (Fountoulis, 1994; Fountoulis and Moraiti, 1994). The upper marine sequence is of Lower Pleistocene age and it is not similar to other post-Alpine deposits of Neda and Zacharo basins as the pebbles of the conglomerates come exclusively from the Pindos Unit formations. The total thickness of the marine Pleistocene sediments in this basin is 150 m. However, in neighbouring basins to the north, thickness exceeds 300 m towards the Neda and Zacharo basins (Fountoulis, 1994; Fountoulis and Moraiti, 1994, 1998). Late Pleistocene (Tyrrhenian) and Holocene marine sediments crop out along the coastline from Kyparissia to Kalo Nero (Fountoulis, 1994; Fountoulis and Moraiti, 1994; Mariolakos *et al.*, 1998; Papanikolaou *et al.*, 2007; Athanassas, 2010). This outcrop shows that significant uplift has occurred during the last few tens of thousands of years in this neotectonic graben.

The Gargalianoi-Filiatra basin has been filled in with marls, marly sandstones, marly limestones, sandstones and conglomerates. The Pleistocene age of these sediments is based on

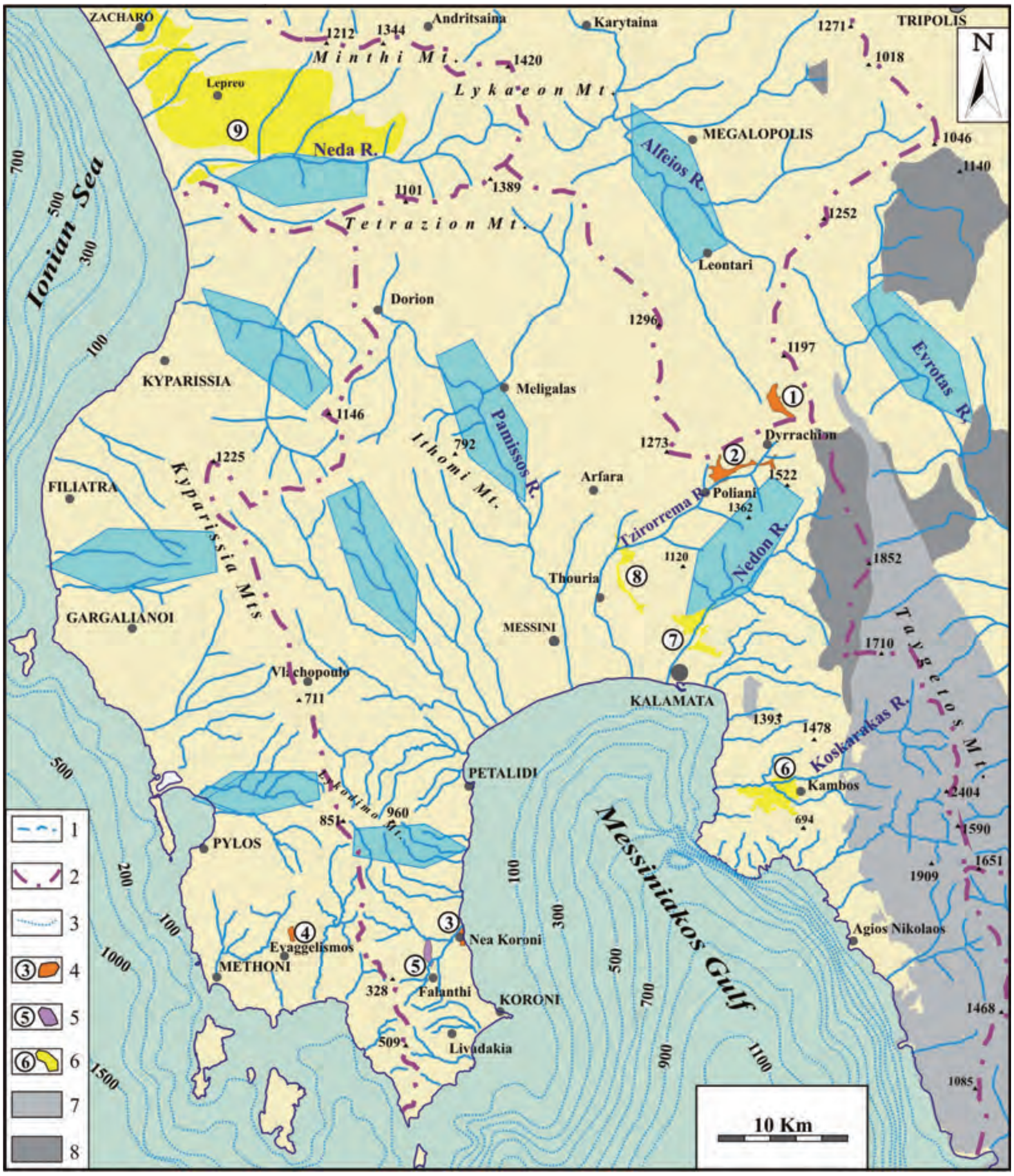


Fig. 4 - Map depicting the main current drainage divides and drainage networks, as well as the locations of post-Alpine formation outcrops that contain pebbles originating from metamorphic rocks, within SW Peloponnese. The large arrows indicate the mean direction of the drainage basin. 1: drainage network; 2: watershed; 3: isobaths; 4: locations of continental deposits; 5: locations of lacustrine deposits; 6: locations of marine deposits; 7: Mani Unit; 8: Arna Unit.

micro and macro fauna (Mariolakos *et al.*, 1994). Athanassas (2010) based on TT-OSL study of the uppermost horizons of the sediments suggests two marine terraces, the age of the older is 225 ka and the age of the younger is 125 ka at the western part of the basin.

The Pylos - Kalamata basin formed between the Kyparissia and Lykodimo Mt. horsts (Fig. 5), contains, at its base, Late Pliocene marls with intercalations of sands and sandstones and the sedimentation continued in Early Pleistocene with sands, marly sands and polymictic conglomerates. At the eastern margin of the Pylos - Kalamata basin, in the broader Kalamata

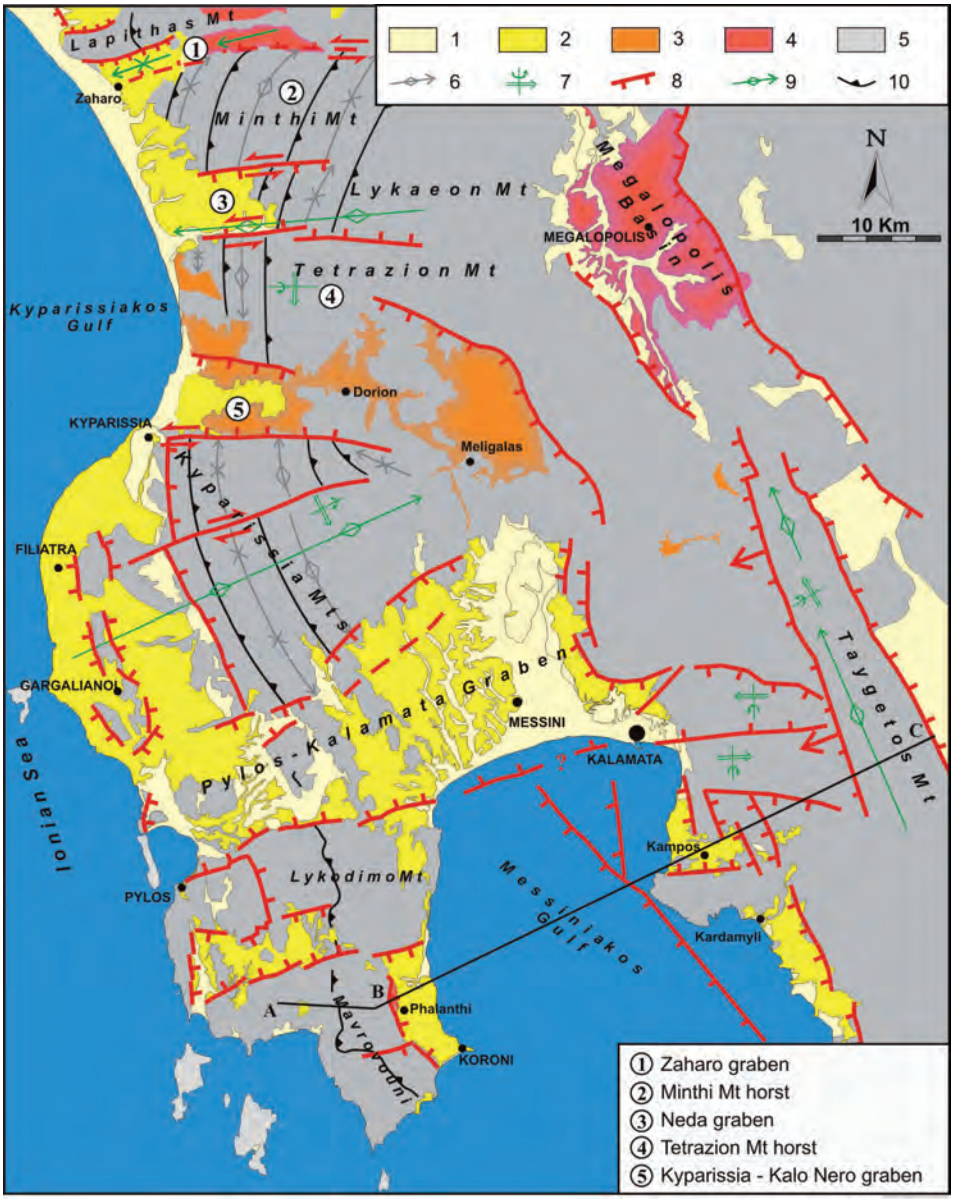


Fig. 5 - Schematic map showing the neotectonic regime of SW Peloponnese. 1: Holocene deposits; 2: Early Pleistocene marine deposits; 3: Plio-Pleistocene continental deposits; 4: Plio-Pleistocene lacustrine deposits; 5: Alpine basement; 6: dominant plunge of Alpine fold axes; 7: rotational axis; 8: neotectonic fault zone; 9: neotectonic fold axis; 10: thrust [after Fountoulis and Mariolakos (2008), submarine faults from Pavlakis *et al.* (1989)].

area, the Pleistocene deposits consist of shallow marine sandstones and polymictic conglomerates. In some cases, these have been deposited unconformably on the older Lower Pleistocene marine deposits. Between the older marine deposits and the younger ones there is intercalation of palaeosoils in several locations. The higher elevation of this formation is up to 340 m. The pebbles of the conglomerates originate mainly from rocks of Tripolis and Arna (metamorphics) units and less from the Pindos Unit. The size of the pebbles is up to 20 cm (site 8 in Fig. 4), while the thickness of the formation is up to 100 m. In (site 7 in Fig. 4), the size of the metamorphic pebbles varies from few centimetres up to 50 cm, whereas cobbles up to 1 m size coming from the Arna Unit occur.

The Kambos depression and Kitries Bay area (site 6 in Fig. 4) are located at the NE margin of the Messiniakos Gulf, some 10 km to the south of Kalamata city has filled in with continental and marine sediments (conglomerates, sandstones and marls) correspondingly. The presence of the *Hyalinea balthica* in the base of the marine succession indicates that their age is at least Early Pleistocene. The continental Pleistocene deposits consist of oligomictic conglomerates with pebbles coming exclusively from the marbles of Mani Unit. They occur in the Kambos basin, overlying conformably the Lower Pleistocene marine deposits (mainly marls). 360 m is the highest elevation of this formation's outcrops (Mariolakos *et al*., 1993).

In Megalopolis basin (Fig. 5), sedimentation starts in Late Pliocene with lacustrine deposits (Makrision formation) and conglomerates (Trilofon formation) occurring mainly at the eastern margins of the basin (Vinken, 1965; Fountoulis, 1994). Red siliciclastic deposits of Lower Pleistocene age [Apiditsa formation: Vinken (1965)] occur mainly at the western margins of the basin consisting of pebbles coming from sandstones, radiolarites, phyllites and quartzites in terra rossa matrix. This formation has been deposited on the Late Pliocene lacustrine deposits (Makrision and Trilofon formations but even on the paleorelief of the Alpine basement during the 1st glacial period in Pleistocene (Vinken, 1965). At the southernmost area of the Megalopolis basin, the formation consists exclusively of pebbles coming from the metamorphic rocks of Arna Unit, the present location of which is further to the south and outside the present watershed of Megalopolis basin [site 1 in Fig. 4: Fountoulis (1994)]. The mean elevation of this occurrence is about 800 m. Younger lacustrine sediments are deposited on the older formations (Vinken, 1965).

Polymictic conglomerates have filled in the Poliani polje (site 2 in Fig. 4). The pebbles of the deposits come from the metamorphic rocks of Arna Unit, as well as the carbonates and clastics of the Tripolis and Pindos units. The size and the percentage of the metamorphic pebbles decrease from east to west, indicating that the source area is located eastern of the Poliani polje (Taygetos Mt.). Hence, at the most eastern entrance of the polje the size of the pebbles is 50 cm, while at its exit at the west decreases to 7-8 cm, the mean elevation of the polje is 660 m, and the thickness of the conglomerates it is up to 90 m (Mariolakos *et al*., 1999). According to Mariolakos *et al*. (1999) the deposition of the conglomerates started at Lower Pleistocene and continues up to present.

In Nea Koroni (site 3 in Fig. 4) and in the coastal area around the Nea Koroni village, polymictic conglomerates occur with thickness of few metres, deposited unconformably over the shallow marine Pleistocene deposits (Mariolakos *et al*., 2001). The pebbles originate mainly from the carbonates and the clastics of the Tripolis Unit, the carbonates, the clastics and the radiolarites of the Pindos Unit, the phyllites-quartzites of Arna Unit and the marbles of Mani Unit. The size of the pebbles varies and in some sites large size pebbles with diameter 50-60 cm occur, coming from the phyllites-quartzites of the Arna Unit.

It is worth mentioned that in Falanthi located NW of Koroni (site 5 in Fig. 4) polymictic conglomerates also occur at the base of the Pliocene sediments bearing lignites. These polymictic conglomerates consist of pebbles coming from Pindos, Tripolis, Arna (phyllites - quartzites) and Mani (marbles) units (Mariolakos *et al*., 2001). The facie of these deposits is lacustrine and the age Pliocene (Koutsouveli, 1987). The size of the metamorphic pebbles is not larger than 15 cm

(Mariolakos *et al.*, 2001). As the size of the metamorphic pebbles in the Pleistocene deposits is much larger than in Pliocene ones, hence it is difficult to relate the Pliocene with the Pleistocene metamorphic pebbles.

In Evaggelismos (site 4 in Fig. 4) there is an occurrence of polymictic conglomerates lying unconformably on the paleorelief created on the flysch clastics of Tripolis Unit, at the inner parts of the Pylia Peninsula close to the Evaggelismos village (Fytrolakis, 1971). The lithologic composition and the size of the pebbles of the conglomerates that comes from the metamorphics is the same as in Nea Koroni and we assume that the deposition took place at the same period as the Nea Koroni formation (Pleistocene). Kiskyras (1959) reports that same conglomerates lying unconformably on the Pindos Unit paleorelief occur west of Koroni close to the Livadakia village (Figs. 4 and 5).

2.3. Neotectonics

The Hellenic territory can be divided in three morpho-neotectonic regions [Mariolakos and Fountoulis, 2004 (Fig. 1a)]. More specifically, Mariolakos and Papanikolaou (1981, 1987) and Mariolakos *et al.* (1985), divided the southern morpho-neotectonic region in three sectors (I, II, III) according to the general direction of the morphologic and the neotectonic major fracture zones prevailing in each region (Fig. 1b). These morphoneotectonic structures consist of successive neotectonic multi-fractured, more or less parallel, horsts and grabens (Fig. 1b).

Most of the extensional structures were previously thought of as the original thrust contact between the Pindos and Tripolis units. However, Fountoulis *et al.* (2010) based on geological mapping and the cross-cutting relationships among these structures indicate that these are SW-dipping faults: their dip, in other words, is towards the arc and they downthrow the original Pindos thrust by a few tens or hundreds of metres each. On the other hand, this fault system is truncated by E-W to ENE-WSW faults, which relay fault activity to the eastern boundary of the Megalopolis basin and the western one of the Taygetos Mt. The whole extensional fault architecture has resulted in the Pindos thrust stepping down from altitudes higher than 1000 m in Mainalon, and Taygetos to negative heights in Messinia (e.g., Kalamata area) and southern Ilia; and the gradual disappearance of the Phyllite-Quartzite metamorphics of Mainalon and Taygetos Mt. towards the west.

All these extensional structures form the eastern boundary of the Megalopolis-Lycaion-Minthi-Tetrazion (MeLyMiTe) and the Pyrgos tectonic depressions, which in turn are separated by the E-W Lapithas horst, at the western end of which the Ionian Unit crops out. The northern and southern boundaries of these tectonic depressions are controlled by oblique-normal faults, perpendicular to the eastern boundary. The throw of these faults increases towards the west and the interplay of all these faults has led to the composite deformation pattern of the MeLyMiTe, which displays extension on its flanks and compression in its centre.

The combination of these extensional faults (which may reach down to the Ionian decollement) with the low-angle floor thrusts of the Pindos, Tripolis and Ionian units leads to additional ENE-WSW shortening, normal to the Hellenic Arc, west of the Peloponnese.

Skourtsos and Fountoulis (2010) suggest a two-stage model for the extensional evolution of south Peloponnese. Upper Miocene-Lower Pliocene extension was accommodated by several mappable brittle detachment faults that exhibit top to the SW-WSW on Taygetos Mt.. The hanging-wall of the detachment comprises of non-metamorphic rocks of the Tripolis and Pindos units. Since Upper Pliocene further exhumation of the metamorphic rocks has resulted in the formation of high-angle normal faults overprinting Neogene extensional structures and cut for first time the lowest Mani (Plattenkalk) Unit. This new fault system tilted the earlier extensional structures and produced an ENE-WSW coaxial deformation. Syntectonic sedimentation within the newly formed basins in Messinia was mainly marine. In Messinia Peninsula, the western sector of transect, Plio-Quaternary movement on NNW-SSE high-angle normal faults and ENE-WSW (to E-W) transverse faults was superposed on an earlier Lower Miocene imbricate thrust stack as it documented by the fact that sedimentation of the west Hellenic flysch continued into the Aquitanian. Marine sedimentation in this area was dominated by the interplay between tectonic uplift and global sea-level change. The central and western sectors are separated from Messinia Peninsula by a series of parallel foreland dipping high-angle normal faults form staircase geometry and control the Messinian half graben. Uplifted Late Pleistocene marine terraces on the footwall of these faults and recent seismic activity suggest that uplift of the internal sectors of transect is still active.

More specifically, the neotectonic structure of SW Peloponnese comprises a number of successive horsts and grabens, extending from the Lapithas Mt. horst towards the south to the Lykodimo Mt. horst [Fountoulis, 1994; Ladas, 2000; Fountoulis and Mariolakos, 2008; see also the Neotectonic Map of Pyrgos and Tropea at scale 1:100,000 (Lekkas *et al*., 1992) and the Neotectonic Map of Filiatra at scale 1:100,000 (Mariolakos *et al*., 1998)]. The main of them from south to north are (Fig. 5): 1) the Lykodimo Mt. horst; 2) the Vlachopoulo - Kalamata graben; the Kyparissia Mts. composite morphotectonic structure; 3) the MEgalopolis LYkaeon MInthi - TEtrazio Composite Morphotectonic Structure (MELYMITE CMS); 4) the Lapithas Mt. horst; 5) the Taygetos Mt. horst, and 6) the Mainalon Mt. horst. Within the major MELYMITE CMS the following second order macrostructures occur: 1) the Zaharo tectonic graben; 2) the Minthi Mt. horst; 3) the Neda tectonic graben; 4) the Tetrazion Mt. horst; 5) the Kyparissia - Kalo Nero tectonic graben; 6) the Lykaeon Mt. horst; and 7) the Megalopolis tectonic graben. All the macrostructures located at the western part of the studied area strike E-W, except the Lykaeon Mt. horst and the Megalopolis graben that strike NNW-SSE.

Concerning the regional neotectonic characteristics of the study area, it should be mentioned that the marginal fault zones of the neotectonic macrostructures strike mainly E-W in the western part and NNW-SSE in the eastern part (Fountoulis, 1994). Their vertical throw is not constant along their whole length, but usually it increases from east to west concerning the E-W fault zones and from south to north concerning the NNW-SSE fault zones. The faults that constitute a fault zone have an en echelon arrangement. They do not have a constant strike along their whole length. The striations observed on fault surfaces, in most cases show a horizontal component of either right lateral component of movement or left lateral (Mariolakos *et al*., 1989a; Fountoulis, 1994; Fountoulis and Mariolakos, 2008).

It is remarkable that folds occur not only in the Alpine formations but also in the post-Alpine ones. Neotectonic folds occur at various scales and types of symmetry and mainly follow a WSW-ENE trend. So, based on the study of the neotectonic structures of SW Peloponnese in various scales, it is concluded that the neotectonic deformation of the area is not of a simple brittle type but of brittle-ductile type, as ductile structures [open folds with big curvature radius occur (Mariolakos and Fountoulis, 1990; Fountoulis, 1994; Fountoulis and Mariolakos, 2008)].

Furthermore, the kinematic evolution of every macrostructure is complicated, as rotation and tilting of the blocks around horizontal axes is observed, resulting differential uplift or subsidence rates at the edges of the blocks (Mariolakos *et al*., 1989b; Mariolakos and Fountoulis, 1994).

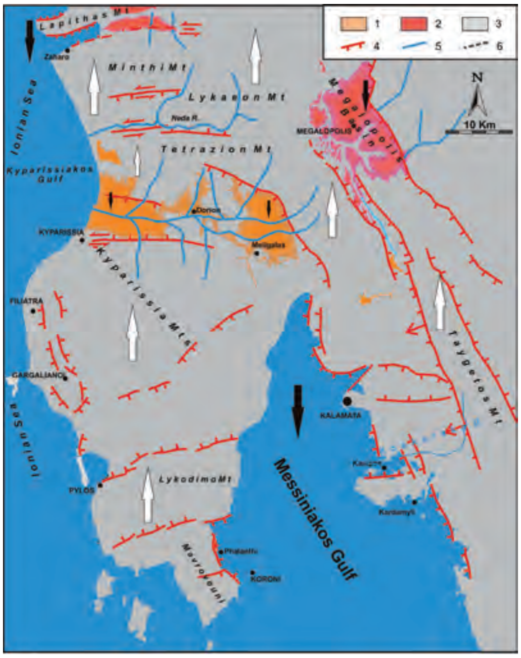
The interpretation of the neotectonic evolution of the Quaternary basins and more specifically the kinematic is more complicated, as the marine Quaternary deposits occur in different altitudes at the margins of the basins.

Mariolakos *et al*. (1994) give their kinematic interpretation for the southern part of SW Peloponnese after the end of Early Pleistocene, based on the facies, the thickness, the present day geographical distribution and elevation of the Pleistocene marine sediments. They suggest that the area is dominated by two large neotectonic units, the Kyparissia Mts. and the Taygetos Mt., which from the kinematic point of view, they function as "tectonic dipoles" having undergone a very complicated rotation around more or less E-W and NNW-SSE trending horizontal axes, that is: i) the Taygetos Mt. has been rotated towards NNW and in the same time towards E and ii) the Kyparissia Mts. have been rotated towards SSE. At the same time towards east, a more detailed study on the neotectonic deformation of the area located between Gargalianoi - Filiatra and Taygetos Mt., 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 and Fountoulis, 1991).

3. Quaternary paleogeographic evolution and geodynamics

The neotectonic pattern presented in the previous sections demonstrates that important changes have occurred during Quaternary in SW Peloponnese. We consider that the morphogenetic procedures should have started after the end of the thrust movements. The overall Quaternary paleogeography of the area can be better visualized in five snapshots (Fig. 6). The greater part of the study area remained land until Late Pliocene as only some parts were submerged (Fig. 6a). The already formed basins of Zacharo, Kyparissia - Kalo Nero, Pylos - Kalamata and Megalopolis were filled in with sediments, while their marginal fault zones were activated, as well as the detachment fault of western Taygetos due to Taygetos Mt. uplift. The Megalopolis basin transformed to a closed geomorphological and hydrological system (lake) resulting the deposition of the Late Pliocene lacustrine deposits. The Neda drainage network continues its evolution. There is no evidence of episodic communication with the Megalopolis basin, while in eastern Zacharo basin lacustrine sediments are deposited and in western Zacharo lagoonal ones. In Neda and Kyparissia - Kalo Nero basins there is deposition of continental deposits. Marine sedimentation took place in parts of Pylos - Kalamata basin.

During Early Pleistocene - Middle Pleistocene time, marine sedimentation progressed landwards for 10-20 km along the western coastal area, followed afterward by a second period of the sea regression to about its present-day stand. This shoreline relocation appears to be related to a first N-S phase of extension of the coastal area, the creation of the E-W oriented horsts and grabens, with a subsidence of several hundred metres and deposition of the Lower Pleistocene marine sediments in the grabens (Fig. 6b). The marine sedimentation continued during Lower Pleistocene, which is the period that the sea transgression reached its greatest extent. More specifically, the already formed basins of Kyparissia - Kalo Nero, Pylos - Kalamata and the western part of Zacharo basin were filled in with marine sediments, while their marginal



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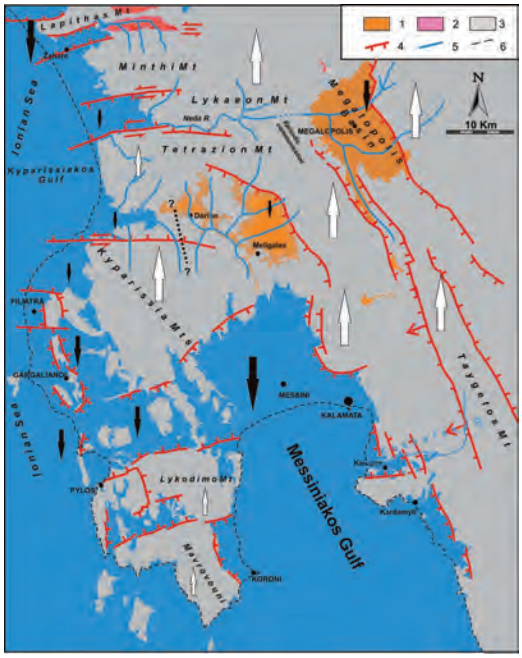

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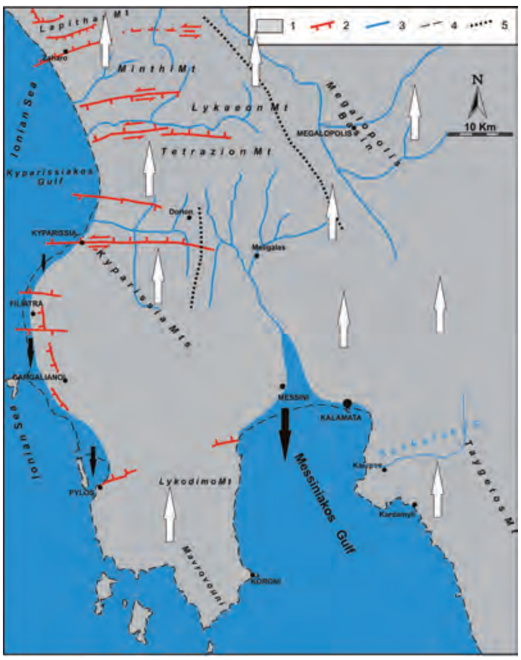

c

Fig. 6 - Neotectonic evolution, white and black arrows indicate uplift and subsidence respectively: a) Late Pliocene, 1 = continental deposits, 2 = lacustrine deposits, 3 = erosional surface, 4 = active fault, 5 = possible drainage network, 6 = recent coastline; b) Early - Middle Pleistocene, 1 = continental deposits, 2 = lacustrine deposits, 3 = erosional surface, 4 = active fault, 5 = possible drainage network, 6 = recent coastline; c) 225 ka 1 = continental deposits, 2 = lacustrine deposits, 3 = erosional surface, 4 = active fault, 5 = possible drainage network, 6 = recent coastline, 7 = possible water divide; d) 125 ka, 1 = erosional surface, 2 = active fault, 3 = possible drainage network, 4 = recent coastline, 5 = possible water divide; e) Present, 1 = erosional surface, 2 = active fault, 3 = possible drainage network, 4 = water divide.

fault zones were activated, as well as the detachment fault of western Taygetos due to Taygetos Mt. uplift. In Dorion, Ano Messinia and Megalopolis basins continental sedimentation took place and the Ano Messinia and Dorion basin have common evolution as they constitute a uniform closed geomorphologic system. There is no evidence for communication between the Megalopolis and Ano Messinia basins, as the Megalopolis basin continued to be a closed geomorphological and hydrological system (lake), which periodically communicates with

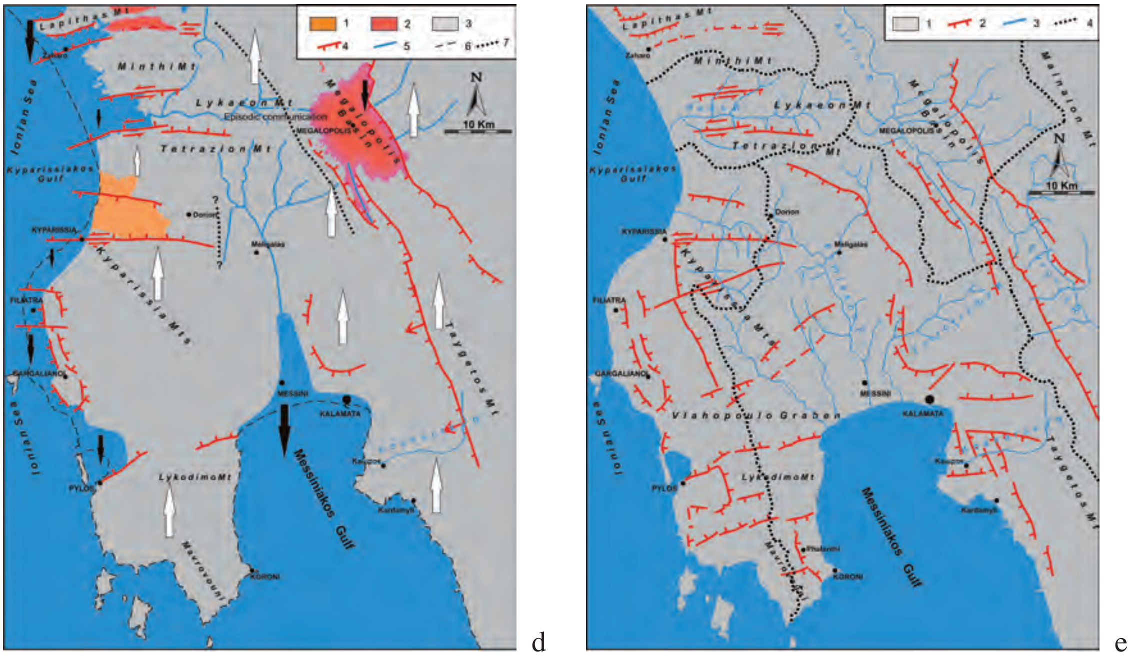

Fig. 6 - continued.

the Neda basin. It has to be stressed that the sedimentation was continuous during the period between the Late Pliocene and the Latest Early Pleistocene (Marcopoulou-Diacantoni *et al.*, 1988).

Next stage in the neotectonic evolution of the area is the period 250 ka ago (Fig. 6c), a regional uplift regime affects the area, but the uplift rates differ from place to place. The deposition of marine sediments continues only in the Neda basin and in the western parts of Zacharo, Pylos - Kalamata basins, as well as in the Filiatra - Gargalianoi coastal area. On the contrary, the continental deposits were deposited in the Kyparissia - Kalo Nero basin, while lacustrine sediments were deposited in the Megalopolis basin. At the same time there is activation of the marginal faults of the basins that are filled in with sediments. The differentiation of the uplift rates resulted in the change of drainage of Dorion and Ano Messinia basins. The Megalopolis basin continued to be a closed geomorphological and hydrological system (lake), which periodically communicated with the Neda basin.

From 250 ka and onwards, the regime of subsidence turns to regime of uplifting, (Figs. 6d and 6e). There is no evidence whether this change was gradual or abrupt. It is possible that there was a period of relative constancy during which, the basins filled up with sediments gradually, resulting in their shallowing.

More specifically, following the former interglacial period that is a period of high sea surface ceases soon after 200 ka, succeeded by cold climate conditions and a massive regression which lasted for following 75,000 years. Sea-level drops by 125 m, laying large portions of the formerly submarine areas open to aerial conditions. Already existing faults influence the drainage network, favouring the development of major streams westbound. Substantial uplift came about that cold stage, altering the relief so that rising sea-level to encounter a different paleogeography by the advent of the last interglacial at 125,000 ago (Fig. 6d). Then, large part of the earlier interglacial remains emerged, undergoing erosion in aerial conditions. Moreover, large torrents have lengthened westwards by a few more kilometres. This cycle is characterized by increased tectonic activity, with the creation of new faults (around Marathopolis) and recrudescence of older faults. Pleogeographic situation indicate that neotectonic regime diagnosed in older phases progresses in the same manner, incurring constant uplift. It is also reckoned that major E-W faults activating in the area chiefly control the development of basic torrents.

The uplift rates since Middle Pleistocene are up to 0.6 mm/yr (Mariolakos *et al.*, 1994; Papanikolaou *et al.*, 2007). It has to be stressed that the western edge of the Kyparissia Mts. is uplifted since Middle Pleistocene with a mean rate of 0.6 mm/yr, while the western edge of the Taygetos Mt. is uplifted with a mean rate of 0.55 mm/yr (Mariolakos *et al.*, 1994). Comparable uplift rates have been obtained also in the Corinth rift, mainly by coral dating on marine terraces, which show a variation from 0.3 mm/yr in the east (Collier *et al.*, 1992) to 1.2-1.8 mm/yr in the central part (Armijo *et al.*, 1996) and to 0.7-0.8 mm/yr in the west (Houghton *et al.*, 2003).

Based on the detailed study of the Quaternary deposits and active tectonics of SW Peloponnese we regard as very significant evidence for the interpretation of the paleogeographic evolution, the occurrence within the Quaternary sediments independently to their facies, of pebbles originated from the metamorphic rocks of Arna (phyllites - quartzites) and Mani (marbles) geotectonic units.

Concerning the post Alpine deposits containing metamorphic pebbles and located at the eastern margins of Pylos - Kalamata basin (major Kalamata area sites 7 and 8), as well as in the Kambos basin (site 6) it is obvious that the drainage network of Nedon, Tzirorrema and Koskaraka rivers is responsible for the transportation of the pebbles. The Tzirorrema drainage network is responsible for the transportation and deposition of the metamorphic pebbles in the Poliani's polje (site 2). The recent watershed of the abovementioned drainage network includes occurrences of the metamorphics (see Fig. 4).

According to Fountoulis (1994), the small size metamorphic pebbles occurring within the Neda basin Pleistocene sediments (site 9) come from areas located eastern of the basin (major Megalopolis area). These pebbles, during flood periods passed from the Megalopolis basin and deposited in the Neda basin for the time period Early Pleistocene - Middle Pleistocene. Hence, there was episodic communication between the Neda and Megalopolis drainage networks when large floods occurred in the Megalopolis basin. It has to be mentioned that the lower part of the recent watershed is at 1000 m altitude and the Pleistocene marine deposits have been uplifted more than 400 m. This means that during the sedimentation period (Early to Middle Pleistocene) the lower part of the watershed should be no higher than 600 m.

It is of special interest the occurrence of metamorphic pebbles in deposits located at the southern part of Pylos Peninsula, because of the location of Messiniakos Gulf, which isolates the Pylos Peninsula from the occurrences of the metamorphic rocks in Taygetos Mt.

In southern Pylia we located two major sites 3 and 4, in which metamorphic pebbles included in terrestrial sediments deposited on the paleorelief of marine Pleistocene sediments (Nea Koroni) and on the paleorelief of flysch clastic sediments (Ladas, 2000). Consequently, the deposition of the terrestrial deposits including metamorphic pebbles over the marine Early Pleistocene sediments in Nea Koroni area should have taken place after the end of the Early Pleistocene, the

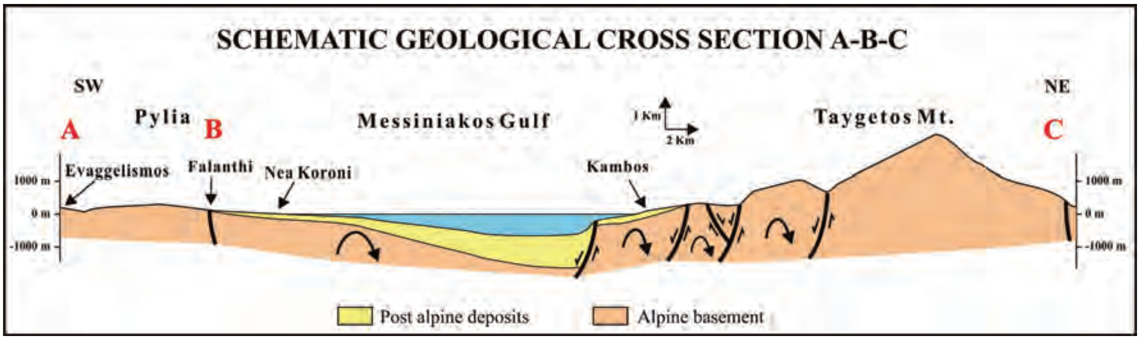

Fig. 7 - Schematic geological cross-section A-B-C across Pylia Messiniakos Gulf and Taygetos Mt. The location is depicted in Fig. 5.

period during which the Pylia Peninsula and generally SW Peloponnese is under uplift regime (Fountoulis, 1994; Fountoulis and Moraiti 1994, 1998; Mariolakos *et al.*, 1994; Papanikolaou *et al.*, 2007).

In order to better understand the paleogeographic evolution of SW Peloponnese a schematic geological cross section (Fig. 7) was compiled from the Pylia Peninsula to Taygetos Mt. Messiniakos Gulf consist a half graben system with a very active eastern margin, (Papanikolaou *et al.*, 1988; Pavlakis *et al.*, 1989) responsible for the creation and evolution of the gulf, while in its western margin there is no marginal fault and the width of the continental shelf is double (about 4 km) compared to the eastern one (less than 2 km). Hence, the greater part of the gulf, together with at least the western part of the Pylia Peninsula, constitute a tectonic block. This block with the Taygetos block behave, from the kinematic point of view, as tectonic dipoles tilting towards ENE around NNW-SSE horizontal axes. This kinematic behavior of the tectonic dipoles is reinforced by the fact that in the western margin of the Koroni basin (Falanthi area) the Early Pleistocene marine sediments have been uplifted up to 180 m, while the thickness of the Plio-Pleistocene deposits in the Messiniakos Gulf is more than 750 m and this maximum thickness is observed near to its eastern margin (Papanikolaou *et al.*, 1988; Pavlakis *et al.*, 1989). In the same time, the Early Pleistocene marine deposits (eastern margin of Messiniakos Gulf and its prolongation onshore) have been uplifted in Kambos area at 360 m (Mariolakos *et al.*, 1993), in Ano Amfia at 460 m (Marcopoulou - Diacantoni *et al.*, 1988), while in the Evrotas basin (east of Taygetos Mt.) have been uplifted at less than 300 m. In conclusion the Taygetos tectonic block has been tilted less than the Messiniakos Gulf block towards ENE.

It is also obvious the difference of the Early Pleistocene marine deposits uplift in the Koroni basin (180 m) compared to that in the Lykodimo Mt. horst, in the eastern slopes of which remnants of Early Pleistocene marine deposits occur at altitudes of 360 m. It has to be mentioned that the pebbles coming from the metamorphic rocks only in southern Pylia occur, while there is no occurrence in northern Pylia. The northern boundary of their spread is the E-W oriented fault zone between the Koroni basin and the Lykodimo Mt. horst (Fig. 5). This fault zone has an horizontal component of movement (Ladas *et al.*, 2004), and played a significant role in the deposition of the metamorphic pebbles only in the southern part of Pylia.

Taking into account that: a) the deposition of the continental deposits containing metamorphic pebbles in southern Pylia took place after the of the Lower Pleistocene marine sedimentation probably during or after Middle Pleistocene and b) the closer to Pylia occurrence of metamorphics is located in NW Taygetos Mt., we believe that the transportation of the metamorphic pebbles from Taygetos Mt. to southern Pylia should have done from NE towards SW by high energy torrents, able to transport even boulders. This is reinforced by the evidence that the higher altitudes in which the post Alpine deposits containing metamorphic pebbles occur, are decreasing from NE towards SW. So, higher elevation is 800 m (site 1 in Fig. 4), then in Poliani polje is 660 m (site 2 in Fig. 4), in the Pleistocene marine deposits is 360 m (sites 7 and 8 in Fig. 4), and in southern Pylia is 0-120 m (sites 3, 4 and 5 in Fig. 4).

We believe that the deposition of the metamorphic pebbles in the Koroni basin and generally in southern Pylia should have taken place at the time period Middle - Upper Pleistocene, that is a period in which the north-western part of Messiniakos Gulf was land due to active tectonics and climatic changes (sea level fall) and its relief was higher than the relief of some parts of southern Pylia.

We assume that there were water divides that form the direction of flow from NE to SW that is from Taygetos Mt. to Pylia. Thus, the kinematic regime of the two distinguished faulted blocks during should have been different during the deposition time. The uplift rate of the eastern margin of Messiniakos Gulf was higher than the western one and the uplift of central Taygetos much higher than the uplift rate of the Pylia Peninsula.

4. Discussion

The paleogeographic change between Early and Middle Pleistocene (Figs. 6c, 6d, 6e) marks an interruption of the general N-S extension and subsidence that produced the post-Alpine basins of western Peloponnese from a new phase of general uplift of the area. Although N-S extension continues since Middle Pleistocene it does not cause any subsidence below sea-level and instead the lower Pleistocene marine sediments have been uplifted and sometimes their transgressive unconformity on the Alpine basement, too. In some cases like Kyparissia, Filiatra - Gargalianoi coast even Late Pleistocene and Holocene marine sediments are uplifted.

The N-NW trending fault zones along the Taygetos Mt. horst and the Megalopolis basin form an arc parallel structure accommodating uplift of the crust above the Hellenic subduction zone, which lies at a depth of about 15-20 km (McKenzie, 1972; LePichon and Angelier, 1979; Laigle *et al.*, 2002). This N-NW tectonic trend characterizes the fold and thrust belt of the external Hellenides throughout the Neogene but also compressional structures during the Early Pleistocene as in the case of the Ionian islands and more especially of Cephalonia (Mercier *et al.*, 1972; Underhill, 1989) and Kylini Peninsula (Dufaure, 1977; Kowalczyk and Winter, 1979). Nevertheless, this active fault zones have the characteristics of a normal fault accommodating E-W extension, similar to the dominant faults of S-SE Peloponnese bordering the Quaternary lignite basin of Megalopolis, the grabens of rivers Pamissos and Evrotas and the horsts of the Messinia Peninsula, Taygetos Mt. / Mani Peninsula and Parnon Mt. (domain II of Mariolakos and Papanikolaou, 1981, 1987). These arc parallel normal faults accommodate back-arc type extension of the Hellenic Arc during the Plio-Quaternary period.

The E-W trending faults along the central-western Peloponnese are Plio-Quaternary structures accommodating N-S extension of this upper plate crust (Fig. 1a), similar to that observed in the Corinth rift and in central Greece (domain I of Mariolakos and Papanikolaou, 1981, 1987; Armijo *et al*., 1996; Roberts, 1996). This N-S extension of the upper plate in the Aegean area cannot be interpreted in terms of the Hellenic Arc dynamics (in the west) but must be balanced by N-S shortening in the Hellenic trench (in the south) within the Africa and Europe convergence system (e.g., Jackson and McKenzie, 1988). E-W extension indicated by the roughly N-S normal faulting observed in the southern Peloponnese and Crete has been interpreted as the result of the beginning of collision of the southern part of the Hellenic margin with the northern margin of Africa (Armijo *et al*., 1992). However, NNW-SSE faults prevail in the southern part of the Aegean area (southern Peloponnese, western Crete) with an ENE-WSW extension [domain II on Fig. 1b: Mariolakos and Papanikolaou (1981, 1987)]. This E-W extension at the southern part of the Hellenic Arc has been proposed to be the result of the beginning of collision of the southern part of the Hellenic margin with the northern margin of Africa (Armijo *et al*., 1992). On the contrary, E-W faults in western Peloponnese have been interpreted as collision structures of the northern part of the Hellenic Arc whereas N-S faults in southern Peloponnese and Kythera have been interpreted as subduction related structures (Lyberis and Lallemant, 1985) assuming a transition from subduction in the south to collision in the north along the arc. However, E-W normal to oblique-slip normal faults can be produced by a transtensional stressfield, which results from the development during Pliocene time of the central Hellenic shear zone (Papanikolaou and Royden, 2007). This zone develops between the northern continental Greece, which converges at a slow rate of 10 mm/yr to the south towards Africa, and the southern Peloponnese and central-southern Aegean area, which converges at a high rate of 40 mm/yr to the SW (e.g., McClusky *et al*., 2000).

Thus, the deformation observed in SW Peloponnese comprises of: i) NNW-SSE extensional structures along the axis of the Taygetos Mt. horst as well as the Megalopolis basin and Messiniakos Gulf, representing arc parallel structures related to the back-arc extension and ii) E-W extensional structures along the coastal zone of SW Peloponnese, related to a transtentional stressfield of the central Hellenic shear zone (Papanikolaou *et al*., 2007).

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