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Active deformation of a segment of arc: the strait of Kythira, Hellenic arc, Greece

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Abstract—A field analysis of Late Cenozoic faults has been carried out in the southwest Hellenic arc, on the islands of Kythira and Antikythira as well as on both sides of the strait, in southeast Peloponnisos and western Crete. The previously unknown geological structure of Antikythira is briefly described. Together with accurate bathymetric data obtained using a multi-beam echo sounder on submarine fault scarps, the results of studies on land lead to a new interpretation of the neotectonic deformation of this segment of the Hellenic arc. The submersion of the strait is a consequence of extension related to normal faulting, the extension rate being faster than in adjacent segments of the arc. The direction of extension related to Late Pliocene–Quaternary normal faulting is determined from on-land information. The existence of oblique en échelon normal faults indicates that dextral deformation plays a role within the Kythira strait, in connection with the opening of the adjacent Cretan Sea triggered by subduction beneath the Hellenic arc. The whole pattern of fault systems in southeast Peloponnisos, as inferred from teledetection and field studies, is consistent with the clockwise rotation that has been independently proposed and verified by palaeomagnetic studies. The Kythira strait thus provides an example of a complex transform–extensional deformation and rotation between two major segments of external arc.

INTRODUCTION

RECENT geodynamic studies have shown that the Aegean region underwent large-scale extension during the Late Cenozoic, thus resulting in the subsidence of the Aegean Sea and more especially the Cretan Sea basin, north of the Cretan segment of the Hellenic arc (McKenzie 1978, Le Pichon & Angelier 1979, 1981). Neotectonic studies, including a quantitative estimate of the amount of extension, have demonstrated, however, that in the largest segments of the outer arc like Crete, the coefficient of linear extension along the axis of the arc is small (approximately 1.1), whereas the coefficient of transverse linear extension relative to the axis of the arc reaches 1.3–1.4 (Angelier 1979a). One cannot extrapolate the estimate of the longitudinal component of extension to the whole of the outer arc that bounds the Cretan Sea to the south, southwest and southeast, because it would be difficult to understand how this moderate lengthening of the perimeter could accommodate the widespread extension of the Cretan Sea basin. This suggests that more intense deformation has occurred in the narrowest and lowest segments of the outer arc, that is the straits east and west of Crete. In addition, palaeomagnetic studies now demonstrate that the Peloponnisos segment of the outer arc has undergone a 25–30° clockwise rotation whereas the Cretan segment has not been affected by significant rotation (Laj *et al.* 1982). This difference obviously implies a particular pattern of deformation in the portion of the outer arc that connects Crete and Peloponnisos. Thus we have especially studied the Late Cenozoic tectonics of the Kythira strait.

Within the southwestern Hellenic outer arc, Peloponnisos and Crete are connected by a 100 km long NNW–SSE ridge (Fig. 1). At the axis of this submarine high, most depths are shallower than 150 m; two islands, Kythira and Antikythira, and numerous islets are present. This narrow segment of the Hellenic arc is bounded by major scarps on both sides, and it separates the 3000–5000 m deep Hellenic trench to the west from the 1000–2000 m deep Cretan Sea basin to the east (Fig. 2).

It has been possible to identify Mesozoic and Palaeogene elements of some major Hellenic nappe units on the islands of the Kythira strait (Theodoropoulos 1973 for Kythira; this paper for Antikythira). Such observations confirm the continuity of the Alpine belt of the Hellenides between Peloponnisos and Crete. Figure 1 summarizes the regional distribution of the main elements of the nappe pile which was formed before Late Serravallian–Early Tortonian times (Aubouin 1973, Jacobshagen *et al.* 1978). Clarifying the pre-Late Miocene nappe structure in the Kythira strait is necessary in order to analyse the subsequent Late Cenozoic tectonics of the area, and this will be done in the next section. In addition, we shall briefly describe the geological structure of Antikythira (Figs. 3 and 4), which was previously almost unknown.

Approximately 13–12 Ma ago, both the subduction presently active beneath the Hellenic arc, as shown by seismicity (e.g. McKenzie 1972), and the widespread extension of the Aegean basin had begun (Le Pichon & Angelier 1979). These processes resulted in the subsidence of the Cretan Sea (McKenzie 1978) and the uplift of the Hellenic arc (Angelier & Le Pichon 1980).

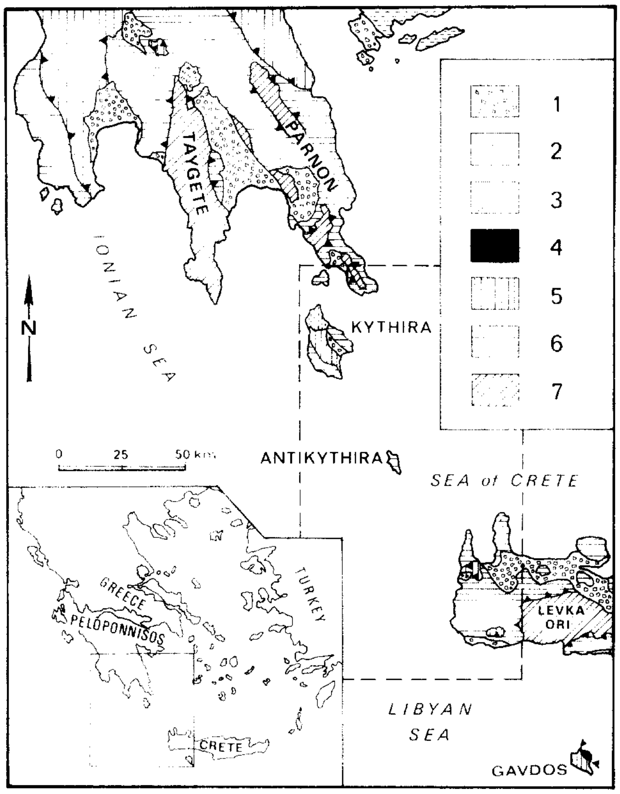


Fig. 1. Alpine nappe structure of the southwestern Hellenic arc (modified after Aubouin *et al.* 1976). 1, Neogene; 2, pre-Neogene Argolide series; 3, metamorphic terranes of Kythira; 4, ophiolitic nappe (Gavdos); 5, Pindos nappe; 6, Gavrovo-Tripolitza nappe (*sensu lato*); 7, lower metamorphic units attributed to the Ionian zone. Dashed line: location of Figs. 2, 6, 7 and 8.

The analysis of faults that affect both the whole of the Alpine nappe pile and the unconformable molassic sediments of Late Cenozoic age enables one to define the mechanisms and the geometry of the extensional deformation (Angelier 1979a, Mercier *et al.* 1979, Mercier 1981). However, a striking difference exists between the southern segment of the Hellenic arc and the western one. To the south, both the Cretan Sea basin and the island of Crete have undergone significant extension, with numerous normal faults and large vertical movements. As a result, the various geological units are distributed in apparent confusion within a complex mosaic of fault blocks. Thus, reconstructing the pre-Late Miocene structure of the Hellenic nappe system has been especially difficult in the southern Aegean arc (e.g. Aubouin *et al.* 1976). In contrast, the Late Cenozoic fault pattern is much less dense and most throws are smaller in the western portion of the arc that includes western and northern Peloponnisos, as well as mainland Greece: as a consequence the present structural pattern of the Alpine nappe system is much more regular (Dercourt in Fantinet 1977). Both the transition zone between these different regions and the western limit of the extensional Cretan Sea basin are close to the Kythira strait. In the last sections of this paper, the mechanisms and the geometry of the Late Cenozoic deformation of the area will be discussed on the basis of field analyses of faults exposed on land and interpretation of submarine morphology which incorporates new, accurate bathymetric data. They will then be compared with the kinematic reconstruction of the deformation of the whole of Aegea since 13–12 Ma.

ALPINE STRUCTURE OF THE KYTHIRA STRAIT

Scattered pre-Neogene outcrops within the Kythira strait area belong to several major nappe units of the external zones of the Hellenides. The continuity of the Alpine belt between Peloponnisos (Dercourt 1964) and Crete (Bonneau 1976) has been demonstrated while stratigraphic and tectonic correlations with the main units that had been classically described in mainland Greece were established (Aubouin 1973, Jacobshagen *et al.* 1978).

In southern Peloponnisos, the metamorphic series of the lower unit are exposed in the large tectonic windows of the Taygete and Parnon massifs, below the Gavrovo–Tripolitza nappe (Fig. 1). These formations, including strata of Palaeogene age, probably belong to the Ionian zone (Thiébault 1974), which is the lowest outcropping unit of the Alpine nappe system in the whole of the southern Aegean arc (Aubouin *et al.* 1976). Both the Gavrovo–Tripolitza nappe and the overlying Pindos nappe also comprise sediments of Palaeogene age.

On Kythira, the metamorphic terranes that are exposed in the northern portion of the island (Theodoropoulos 1973) belong either to the Ionian zone or more likely to the Phyllite series, which is probably composed of both the lower Gavrovo–Tripolitza units and the upper part of the Ionian series (Lekkas 1980). The wide pre-Neogene outcrops of central and southern Kythira are elements of the Gavrovo–Tripolitza and Pindos nappes (Fig. 1; Theodoropoulos 1973).

On Antikythira, the pre-Neogene formations belong to the Gavrovo–Tripolitza zone while some strata of the Pindos nappe are exposed in the neighbouring islet of Pori (this paper; Fig. 3).

Along the western coast of Crete, the Phyllite series and the limestones of the Gavrovo–Tripolitza units are widely exposed, under small remnants of the Pindos nappe (Fig. 1). The metamorphic formations attributed to the Ionian zone are visible in the tectonic window of the Lefka Ori massif, in the same structural position as in southern Peloponnisos. They represent the autochthon to the Alpine nappe system on Crete (Bonneau 1973) as well as in the whole of the southern Hellenides (Aubouin *et al.* 1976).

STRUCTURE OF ANTIKYTHIRA ISLAND

Antikythira is an island 10 km long, 3 km wide and 378 m high, situated on the crest of the ridge between Kythira and Crete. Numerous large normal faults cut both the pre-Neogene formations and the unconformable Miocene sediments. Late Cenozoic faults thus control the morphology of this island which is elongate

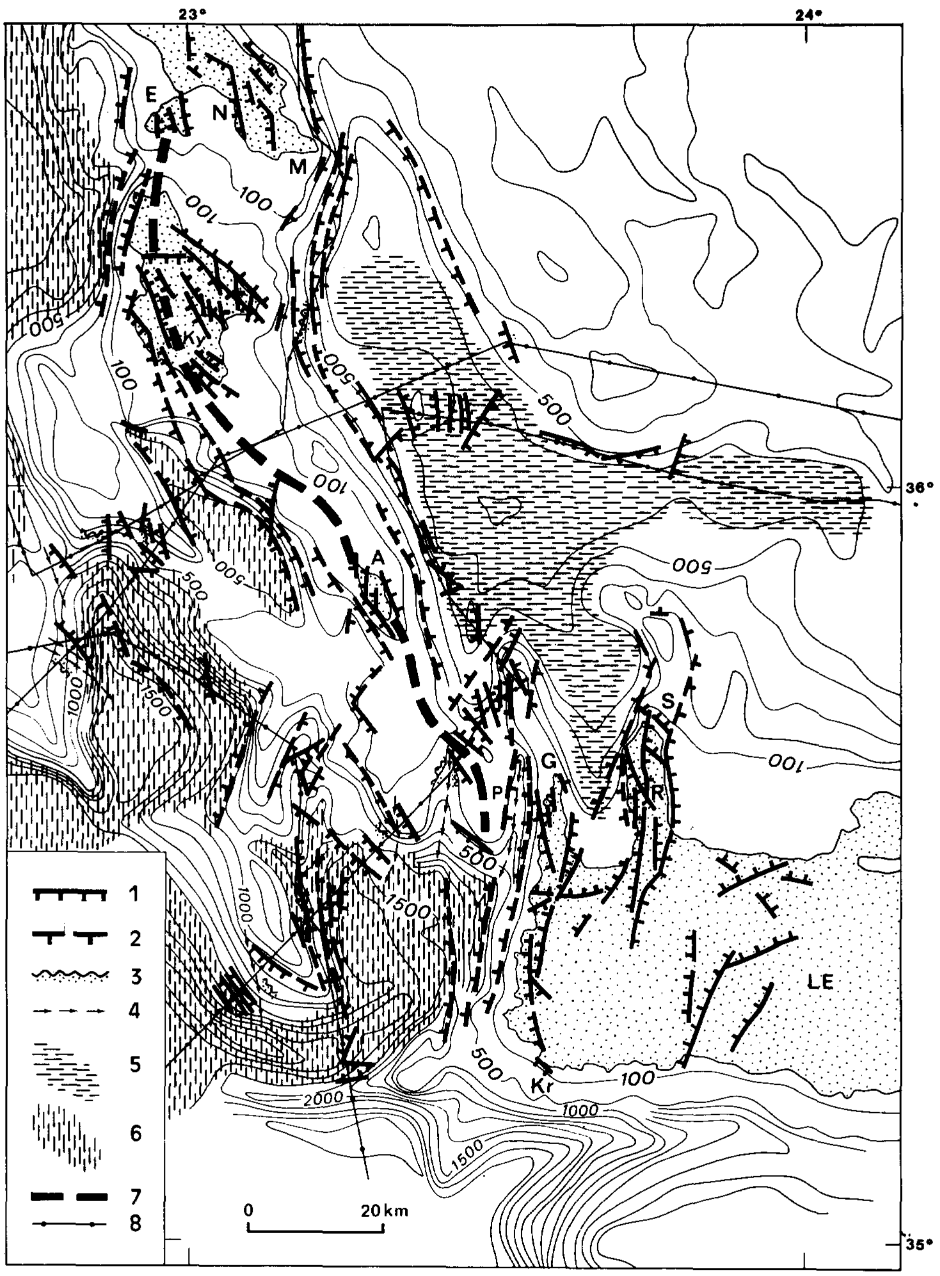


Fig. 2. Neotectonic structure of the Kythira strait area. E, Elafonissos; N, Neapolis; M, Cape Maleas; Ky, Kythira; A, Antikythira; P, Pontikonissi; G, Cape Gramvoussa; R, Rhodopou peninsula; S, Cape Spatha; Kr, Cape Krios; LE, Lefka Ori. 1, normal fault scarp; 2, probable normal fault scarp; 3, erosional morphologies; 4, submarine canyon; 5, deep Cretan Sea Basin; 6, deep sedimentary basins of the Hellenic margin; 7, morphological axis of the Kythira–Antikythira ridge; 8, ship-track lines. Submarine contours in fathoms (after Defense Mapping Agency, Hydrographic Center, Washington).

NNW–SSE (Fig. 3), and they also determine the regional shape of the ridge that crosses the Kythira strait (Fig. 2).

As the geological map (Fig. 3) and section (Fig. 4) suggest, the amplitudes and density of the Late Cenozoic normal faults have made stratigraphic reconstruction of the pre-Neogene succession difficult. Fault breccias are common. Two main formations of black and grey limestones have been identified. The lower formation of alternating limestones and dolomitic limestones contains rudists (*Radiolitidae* & *Hippuritidae*) and, hence, they are dated as Late Cretaceous. The upper formation of bedded limestones contains an abundant microfauna, with Nummulites (*Nummulites gr. globulus*); it is inferred to be of middle–late Eocene age; the upper layers containing Alveolines and Orbitolites (*Orbitolites gr. douvillei*). All sample analyses indicate a shelf-zone environment: neritic facies closely resembling those described by Thiébault (1973) from southern Peloponnisos. Thus we have established that the pre-Neogene formations of Antikythira belong to the Gavrovo–Tripolitza zone. Palaeogene flysch is absent as a consequence of pre-Tortonian erosion. On the islet of Pori, 8 km north of Antikythira, narrow zigzag folds affect a quite different formation of white–grey platy limestones with red intercalations. Landing and consequently sampling were impossible but the strata resemble the Upper Cretaceous platy limestone formation of the Pindos zone.

On Antikythira, the unconformable Neogene sedi-

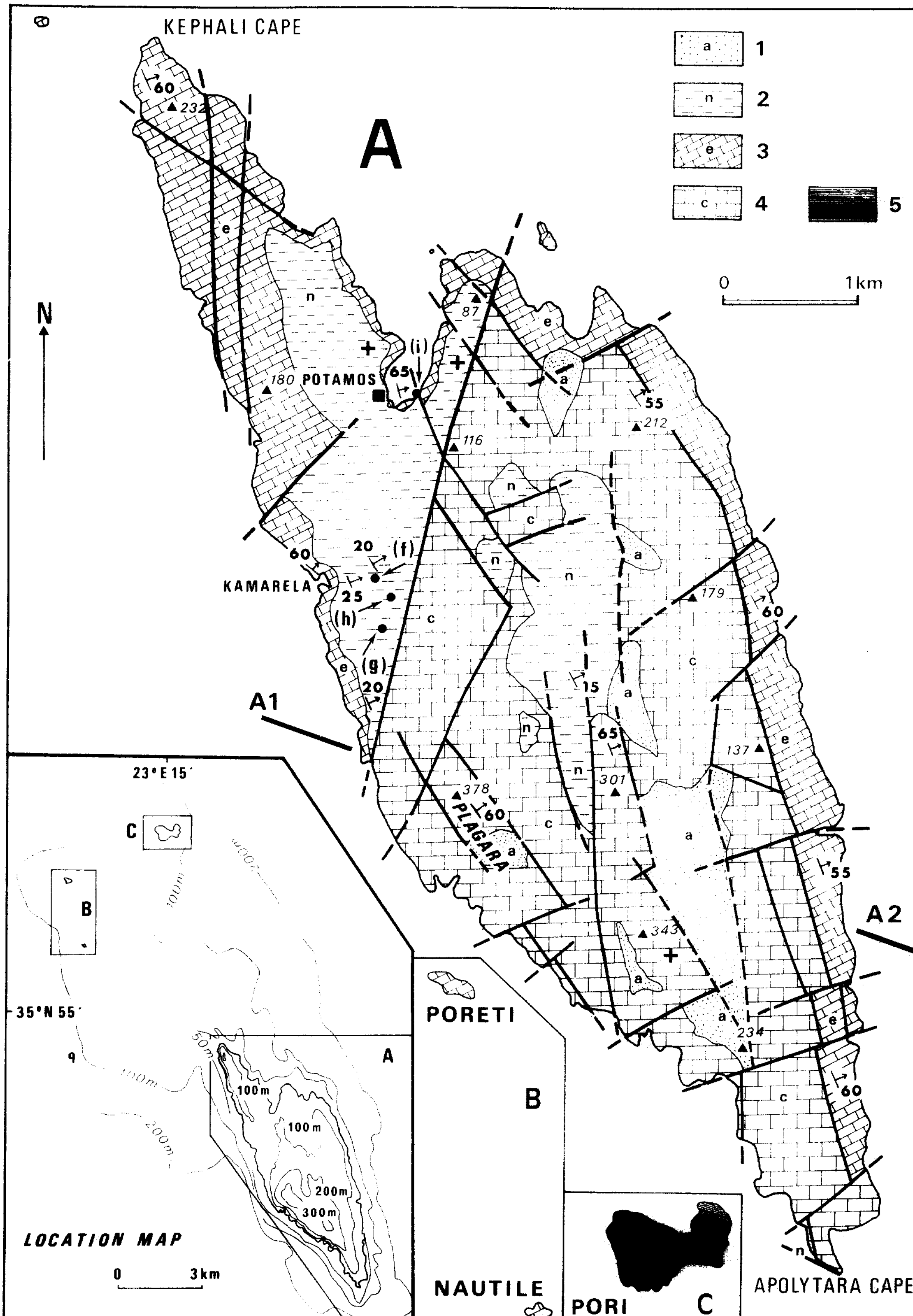


Fig. 3. Geological map of Antikythira island. 1, screes and alluvium (a); 2, Miocene sediments including sandy marls, marly clays and calcarenites (n); 3, Eocene limestones of the Gavrovo–Tripolitza zone (e); 4, Cretaceous limestones of the Gavrovo–Tripolitza zone (c); 5, platy limestones probably belonging to the Pindos zone (Pori islet). Small arrows (f, g, h and i) refer to fault analyses (see Fig. 5). A1–A2, trace of the section in Fig. 4.

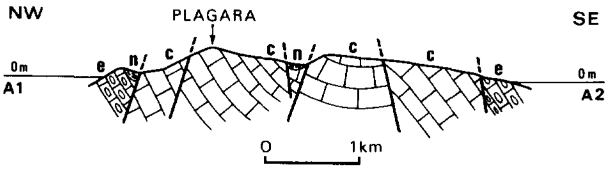


Fig. 4. Schematic geological section across Antikythira island. For location of section line and legend see Fig. 3.

ments exposed in two small faulted basins overlie both the Upper Cretaceous and Eocene limestones (east and west of the NNE–SSW Potamos fault, respectively); a small outcrop has also been found on Cape Apolytara (Fig. 3). Near Potamos and to the north, a yellowish calcarenite (dolomite limestones) is 4–10 m thick; this Potamos Formation of Middle–Late Miocene age was deposited in a coastal environment and contains *Borelis melo* (Fichtel & Moll 1798). The thicker Kamarela Formation is widely exposed south and southeast of Potamos; it is composed of yellowish marly clays and sandy marls, and is more than 20 m thick. The basal unconformity is marked by a ferruginous erosion surface. A reddish conglomeratic bed is present at the top. The lower marls contain microfossils of Early Tortonian age (corresponding to Sarmatian *pro parte*, with *Schackoinella sarmatica,* Weinhandl 1958).

As Figs. 3 and 4 show, the Late Cenozoic structural pattern of the island is dominated by NNW–SSE trends for both the faults and the strata. Cretaceous–Eocene limestones and Late Miocene sediments commonly dip 20° or more to the ENE, while the largest normal faults dip to the W and WSW. This pattern of tilted blocks bounded by normal faults explains the asymmetry of the island, the orographic axis being close to the fault-controlled western coast. The same asymmetry, with the main normal fault scarps facing the Hellenic trench and most blocks dipping gently toward the Cretan Sea, has already been observed on other islands of the southern Aegean arc such as Gavdos which has a size comparable to Antikythira or Crete which is much larger (Angelier 1979a). Note, however, that in the axial blocks of Antikythira island the bedding is horizontal in Tortonian sediments as well as in pre-Neogene formations. Note also that although NNW–SSW normal faults prevail and clearly control the morphology of the island, ENE–WSW and N–S striking faults are also present. The structural arrangement of the Miocene faulted basins is influenced by the existence of the N–S to NNE–SSW Potamos fault that intersects the whole of the island and the N–S fault east of Plagara; each basin being tilted to the ENE and bounded by one of these normal faults to the east.

CHRONOLOGY OF FAULTING

The faults that we have studied in detail (Figs. 5 and 6, Table 1) affect sediments in the Neogene basins and less commonly they displace Pleistocene marine platforms. The chronology of faulting has been established by taking into account stratigraphic data (Table 2) and field evidence for successive fault motions as well as geometrical and mechanical compatibilities between fault patterns observed in various localities and formations.

It is difficult, however, to reconstruct in detail the neotectonic evolution of the Kythira strait, because the Late Cenozoic sediments are not easily differentiated. For instance, stratigraphic analysis and geological mapping on Antikythira island have only enabled us to distinguish movements that occurred before or after the Early Tortonian (Fig. 3). Fortunately, because in this area the main extensional mechanisms were similar during the Late Pliocene and the Quaternary and older mechanisms are rarely reconstructed as a consequence of faults having been reactivated, the chronological uncertainties do not alter the determination of the patterns of stress and deformation that probably have prevailed since the Late Pliocene.

The most precise information for dating faulting events has been obtained on Kythira island where both the Upper Miocene (Pontian *sensu lato*) and the Pliocene are represented (Christodoulou 1965, Freyberg 1967, Theodoropoulos 1973). In the main NW–SE graben of central Kythira, the marine Pliocene sediments overlie Pontian deposits which have been slightly tilted and these Pontian deposits cover the Pindos nappe. In contrast, on both sides of the graben, the same unconformable Pontian sediments overlie the Mesozoic–Palaeogene limestones and flysch of the Gavrovo–Tripolitza nappe (Theodoropoulos 1973). Thus, the NW–SE graben was created prior to the Pliocene transgression. Synsedimentary structures in the graben show that faulting continued during the Pliocene. Finally, numerous normal faults cut the whole of the Pliocene series in Kythira as well as in other regions of the Aegean arc (Angelier 1979a, Mercier *et al.* 1979). Normal faults also affect the Pleistocene marine deposits of Tyrrhenian age that are exposed on the coast between Neapolis and Cape Maleas (Theodoropoulos 1973).

Most of these Late Cenozoic normal faults strike approximately NNW–SSE and control the structure and the morphology of the Cape Maleas peninsula as well as those of Kythira and Antikythira islands (Fig. 2). Thus, taking into account both the geometry of the fault pattern and the chronology that has been established on Kythira, one must conclude that the development of the ridge between Peloponnisos and Crete had begun prior to the Pliocene transgression and has continued since (Angelier *et al.* 1976). Stratigraphic studies on Crete have shown that the disruption of the Southern Aegean landmass into a mosaic of uplifted and subsided fault blocks began during the Late Serravallian, at the close of the Middle Miocene (Drooger & Meulenkamp 1973). A similar age has been proposed for the dislocation of the western Hellenic margin on the basis of stratigraphic analysis on the island of Zakynthos (Dermitzakis 1977). It is, thus, reasonable to assume that the tectonic movements which resulted in the fault pattern shown in

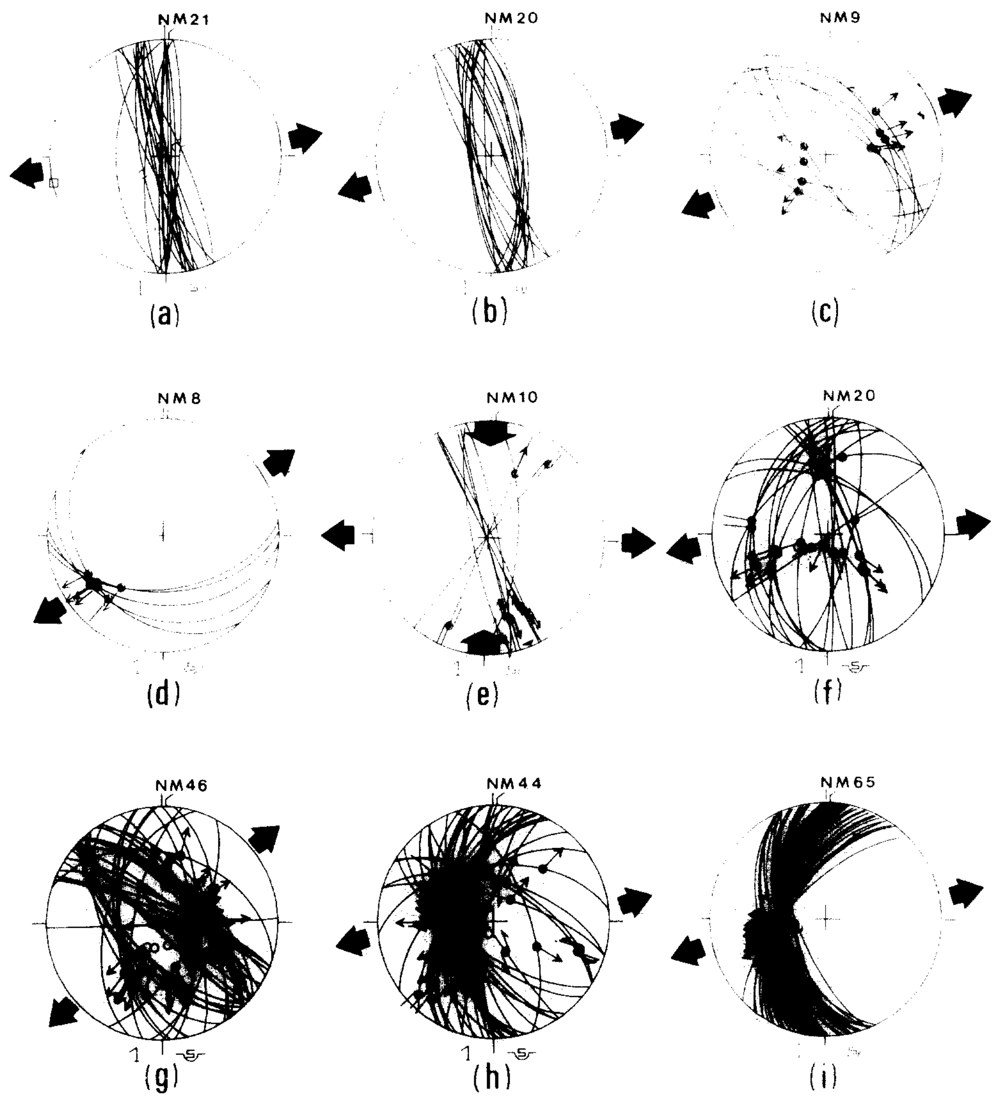


Fig. 5. Examples of fault populations analysed in the Kythira strait area. Corresponding sites are shown in Fig. 6 (also see Fig. 3 for Antikythira). Schmidt lower-hemisphere projections. Fault planes are shown as curves, slickenside lineations as dots with small centrifugal arrows (normal motion) or double arrows (strike–slip motion), and poles to bedding planes as small open circles. Large black arrows indicate the direction of extension (σ_3). Compression (σ_1) is subvertical except in diagram e. (a) 21 small joints and normal faults in Tyrrhenian marine deposits south of Neapolis (site NE2). The open square is a pole to a tension gash. (b) 20 normal faults and joints in Pliocene sediments between Mitata and Avlemonas (site CY2 of Kythira). (c) 9 measurements of large normal faults of the Kapsali graben (site CY3 of Kythira). (d) 6 measurements along a large E–W oblique-slip normal fault south of Agia Pelagia (site CY6 of Kythira). (e) 10 small conjugate strike–slip faults in Pliocene sediments northwest of Paleohorio (site CYB of Kythira). (f) 20 small normal faults in Miocene sediments of the Potamos basin (site AC2 on Antikythira). (g) 46 normal faults in Miocene sediments of Kamarella, Potamos basin (site AC3 of Antikythira). (h) 44 normal faults in Miocene sediments of the Potamos basin (site AC4 of Antikythira; note that some small strike–slip faults are also present). (i) 65 measurements along the large normal fault that limits the Potamos basin to the east (site AC5 of Antikythira; note that the fault strike ranges through 75° but the slickenside lineation provides a common axis).

Fig. 2 also began during this period, approximately 13–12 Ma ago. Numerous good fault mechanism determinations, however, are only available for the Late Pliocene and Quaternary, because older slickenside lineations have generally been erased by subsequent motions on fault planes.

FAULT MECHANISMS

Analyses of populations of fault planes and slickenside lineations from field measurements of strike, dip and pitch have enabled us to reconstruct the orientations (trends and plunges) of the principal stress axes that prevailed during the faulting event (maximum compressional stress σ_1, intermediate stress σ_2 and minimum stress σ_3). The methods used have already been presented and described in detail by Angelier (1979a, b). They include automatic plotting of data in diagrams such as Fig. 5, geometrical analysis of conjugate fault patterns, use of the method of right dihedra, and determination of stress tensors by mathematical means.

The main results that have been obtained in the area shown in Fig. 2 are summarized in Fig. 6 and Table 1. New data on Antikythira and Kythira islands as well as

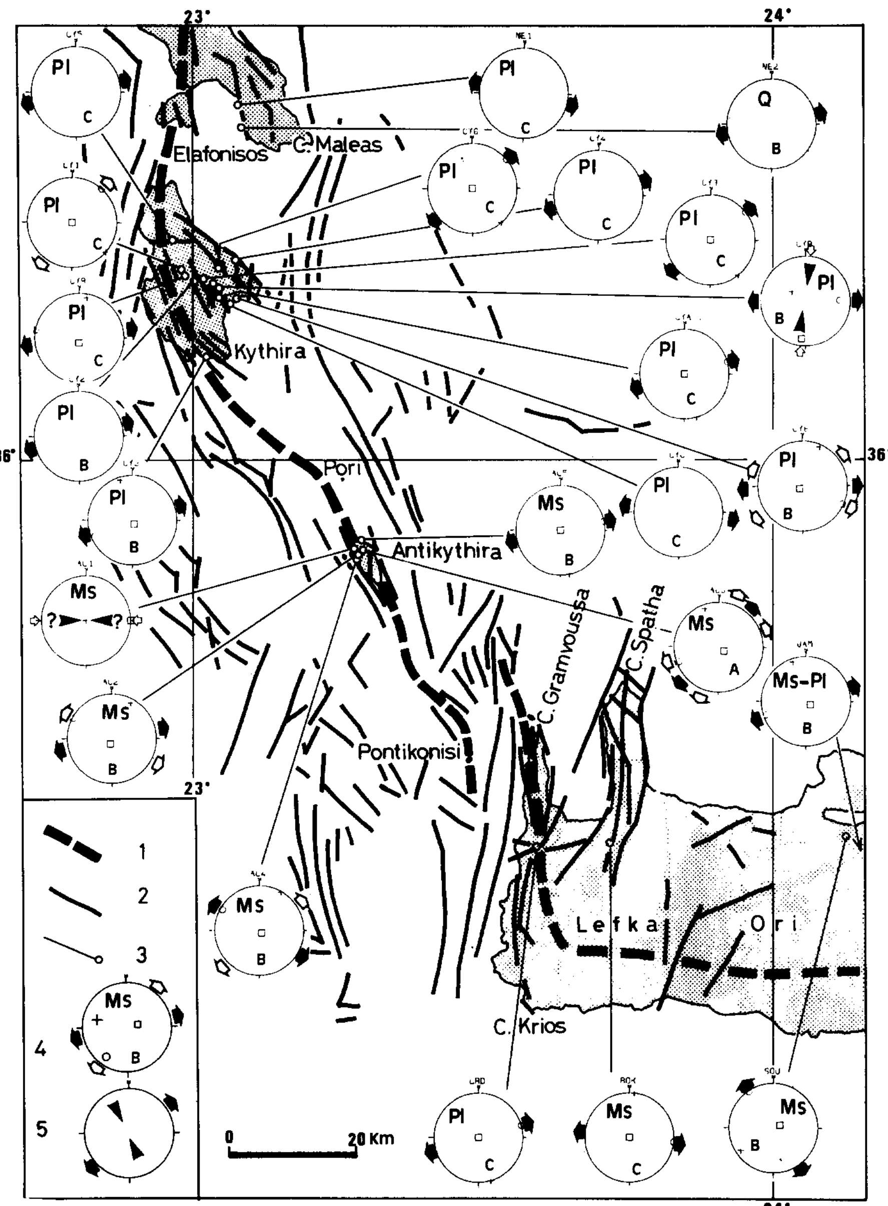


Fig. 6. Fault-population mechanisms reconstructed after field analysis in the Kythira strait area. 1, morphological axis of the arc; 2, main fault lines; 3, sites of fault population measurements; 4, main results obtained in one site. Directions of generalised extension given by black arrows, open arrows refer to directions of extension related to particular families of faults. Poles show the principal stress axes (σ_1 as squares, σ_2 as crosses, σ_3 as open circles). The quality of each determination is recorded as A (excellent), B (acceptable) or C (poor). The age of the youngest faulted sediments is indicated by Ms (Late Miocene), P1 (Pliocene) or Q (Quaternary). 5, conjugate strike–slip mechanism (direction of compression as centripetal arrows).

the results of previous neotectonic studies of the Kythira–Neapolis area (Angelier *et al.* 1976) and Crete (Angelier 1979a) have also been incorporated. The uniformity of the directions of extension is a striking feature of Table 1 and Fig. 6: all but a few computed azimuths of the subhorizontal minimum stress axis (σ_3) are in the range 076 ± 20°, that is close to ENE–WSW or E–W, whatever the age of faulting. The Late Cenozoic stress regime that resulted in the present pattern of normal faults in the Kythira strait was dominated by ENE–WSW to E–W extension.

However, approximately E–W normal faults are also present, especially on Kythira and on western Crete, where they affect Pliocene sediments. These E–W normal faults have been either reactivated as oblique–slip faults compatible with the main ENE–WSW extension or cut by large NW–SE or N–S normal faults. Thus, this E–W trending system of normal faults is generally older than the main fault systems (Fig. 6).

Small strike–slip faults also affect Tortonian (on Antikythira) and Pliocene (on Kythira) sediments. As Fig. 5(e) shows, these small strike–slip faults are consistent with a compression that trended approximately N–S. Senses of relative motion could not be determined on Antikythira. Such mechanisms closely resemble those described from Crete (Angelier 1979a) and can thus be attributed to a compressional event of Early Pleistocene age that has been identified in Greece (Mercier *et al.*

Table 1. Summary of results of tectonic analyses in the Kythira strait area, (also see Fig. 6) with some results omitted. N designates normal and D designates strike-slip mechanisms. The last column indicates the average azimuths of the direction of extension computed with weighting as function of the qualities of mechanisms (decreasing from A, excellent, to C, poor). The average azimuth for the whole area is 076°, as for Kythira

	Site	Azimuth of σ_3	Quality	Mechanism	Average azimuth of σ_3
Cape Maleas (Peloponnisos)	NE1	102	C	N	086
	NE2	078	B	N	
	CY1	040	C	N	
	CY2	076	B	N	
Kythira	CY3	073	B	N	
	CY4	078	C	N	
	CY5	078	C	N	
	CY6	050	C	N	
	CY7	050	C	N	076
	CY8	087	B	N	
	CY9	084–102	C	N	
	CYA	074	C	N	
	CYB	087	B	D	
	CYC	096	C	N	
	AC2	079–119	B	N	
Antikythira	AC3	051	A	N	072
	AC4	057–120	B	N	
	AC5	078	B	N	
Capes Gramvoussa–Spatha (Crete)	LRD	073	C	N	
	ROK	096	C	N	085

1979). This compressional event is quantitatively negligible relative to extensional tectonics (Table 2): strike-slip faults are rare and discontinuous, with throws generally less than one metre, whereas normal faults are very common and commonly have lengths of several kilometres and throws of several hundred metres. Note, however, that compressional tectonics played a significant role in the Ionian islands at the northwestern tip of the Hellenic arc (Mercier *et al.* 1979) and in the Strophades islands, west of Peloponnisos (Lyberis & Bizon 1981).

However, as discussed above, normal faulting clearly predominates in the Kythira–Antikythira area; two families of normal faults determining the structure of the islands. The first family strikes NW–SE to NNW–SSW, parallel to the axis of the submarine ridge, and controls the general morphology (Fig. 2). The other family strikes N–S to NNE–SSW, oblique to the axis. It determines the morphology of peninsulas in northwestern Crete (i.e. Capes Gramvoussa and Spatha) and it also plays a noticeable role in the structure of the Kythira–Neapolis area (Theodoropoulos 1973, Angelier *et al.* 1976) and on Antikythira island (Fig. 3). Although the N–S faults generally cut the NW–SE ones, both systems affect the Pliocene and were probably synchronous, resulting in a complicated relationship.

It is obvious, however, that as the largest portion of the studied area is under sea-level one cannot reconstruct the Late Cenozoic pattern of deformation without studying the submarine structures.

SUBMARINE FAULT PATTERNS

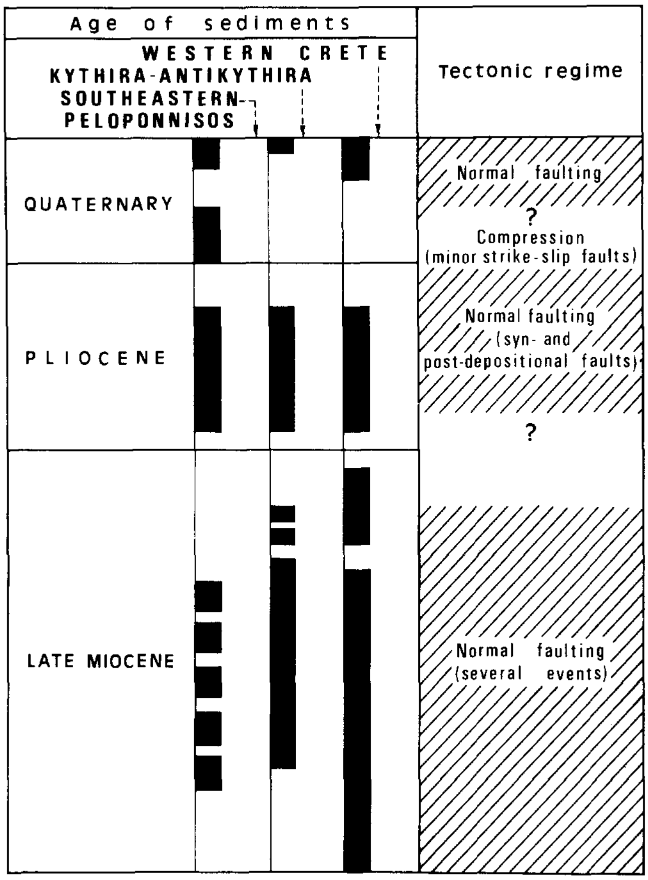
During the 1978 HEAT cruise of R/V Jean Charcot, detailed bathymetric mapping, using a multi-narrow beam echo-sounder Sea Beam (Renard & Allenou 1977), was carried out along the four ship-tracks shown in Fig. 2. The results obtained have already been presented for the Hellenic trench (Le Pichon *et al.* 1979a, Lyberis *et al.* 1980) and the Cretan Sea basin (Lyberis *et al.* 1981). In this paper, we focus on the data collected from the submarine ridge that connects Crete and Peloponnisos, also taking into account information given in available bathymetric charts (especially those published by the U.S. Hydrographic Center). The morphology of the ridge is that of a horst bounded by numerous fault scarps (Fig. 2). These scarps are of large extent and amplitude, usually being longer than 10 km and much higher than 100 m.

East of the ridge, the bottom of the western Cretan Sea basin is generally flat and approximately 1300 m deep. Our Sea-Beam records have enabled us to recognize several steep scarps, 60–300 m high, most of them striking either N–S to NNW–SSE or E–W. The scarps commonly separate secondary basins or plateaus; some corresponding to faults that were identified as normal faults by Jongsma (1975), on the basis of seismic-reflection studies.

The ridge itself strikes NNW–SSE (average azimuth 150°) and its width varies from 10 km near Antikythira to 25 km near Kythira. It is bounded by fault scarps, 200–1000 m high, that generally strike N–S to NNW–SSE on the eastern flank, from Cape Maleas in Peloponnisos to Cape Gramvoussa on Crete (Fig. 2). Sea-Beam records have shown, however, that a series of large scarps, up to 700 m high, strike NE–SW between Antikythira and Gramvoussa.

To the west, the ridge is bounded by a more complex system of fault scarps with dominantly NNW–SSE to

Table 2. Periods of fault movement. Black columns, sedimentary formations



NW–SE trends, N–S strikes being relatively rare. Near Kythira, it has been demonstrated that these faults cut Pliocene sediments (Blanc & Blanc-Vernet 1968). Numerous seismic-reflection profiles have shown that Late Cenozoic fault tectonics developed a pattern of successive basins and plateaus in the inner wall of the western Hellenic trench (Le Quellec 1979, Le Quellec *et al.* 1980), thus controlling the sedimentation of detrital material supplied by canyons (Vittori 1978). In detail, Sea-Beam records show complex erosional morphologies on the edges of these canyons (Fig. 2). Taking into account both the information from seismic-reflection studies and the comparison with field studies on land, it is quite reasonable to consider most faults of the western slope of the Kythira strait as normal faults; submersible studies of the inner wall of the Hellenic trench having effectively demonstrated the presence of normal faults (Le Pichon *et al.* 1979b). Figure 7 (D1 and D2) shows the distribution of fault-scarp strikes that have been recognized using Sea-Beam records on both sides of the submarine ridge. The main families strike NW–SE and NNW–SSE, as well as N–S on the eastern flank. On a larger scale, the bathymetry is effectively characterized by NNW–SSE to NW–SE and N–S to NNE–SSW scarp strikes (Fig. 2). These directions are also those of normal faults observed on land: as illustrated in Fig. 7 by diagrams B1–C1 (after field measurements of faults, Fig. 5) and B2–C2 (after geological maps, Fig. 3), for the islands of Kythira and Antikythira. Furthermore, the average direction of extension that one can determine for the whole of the strait by assuming that submarine faults are pure dip-slip normal faults is E–W to ENE–WSW, that is close to the direction that has been accurately determined using fault mechanism analyses on land (Fig. 6 and Table 1). We conclude that marine and land data on fault patterns and mechanisms are in good agreement in the area of the Kythira strait.

ROTATION OF PELOPONNISOS

A remarkable feature illustrated by Fig. 7 is the existence in southeast Peloponnisos of numerous NW–SE fault systems that are clearly oblique relative to the NNW–SSE fault systems that define the deep Plio-Quaternary grabens and most coastlines. The faults of southeast Peloponnisos shown in Fig. 7 were identified on Landsat orbital images, the distribution of strikes of all faults being summarized in diagram A4. Although it plays the major role in the recent deformation of the area, the NNW–SSE system is less visible on satellite images, because its faults are commonly hidden either under the sea (as along the eastern coast) or by slope detritus and recent deposits. In contrast, the NW–SE system is dense and its faults are clearly visible in uplifted pre-Neogene massifs. Figure 7 also shows the distribution of 905 Neogene–Quaternary faults that were measured in southeast Peloponnisos (diagrams A1–A3), all but a few of the faults being normal. NW–SE strikes are much less abundant than on orbital images (compare A1 and A4 in Fig. 7), because most of the NW–SE faults do not affect the upper Pliocene and Quaternary deposits in which the measurements were made. In addition, the structural analysis has shown that NW–SE faults are generally cut by NNW–SSE faults. As on Kythira and Antikythira, N–S to NNE–SSW striking faults are present. Diagrams A2 and A3 of Fig. 7 show that the dips of most faults average 60–65° (normal faults) or 85–90° (tension gashes) and that most pitches of slickenside lineations average 85–90° (pure normal faults).

Summarizing, the average azimuths of the main families of normal faults that affect the Late Cenozoic deposits of southeast Peloponnisos are 165–170° and 015–025°; the NW–SE strike (azimuth 130°) generally belonging to older faults. The 25–30° Late Cenozoic clockwise rotation of Peloponnisos was first proposed while studying the horizontal pattern of deformation of Aegea (Le Pichon & Angelier 1979) and has since been demonstrated using palaeomagnetic data (Laj *et al.* 1982). As stratigraphic and structural evidence suggests that most of the NW–SE faults of southeast Peloponnisos are older than the greatest part of this rotation, the initial trend of the NW–SE fault system was E–W to ESE–WNW (azimuth 100–110°), that is approximately parallel to the major E–W normal fault system that has been active in Crete since Serravallian–Tortonian times (Drooger & Meulenkamp 1973, Angelier 1979a): a con-

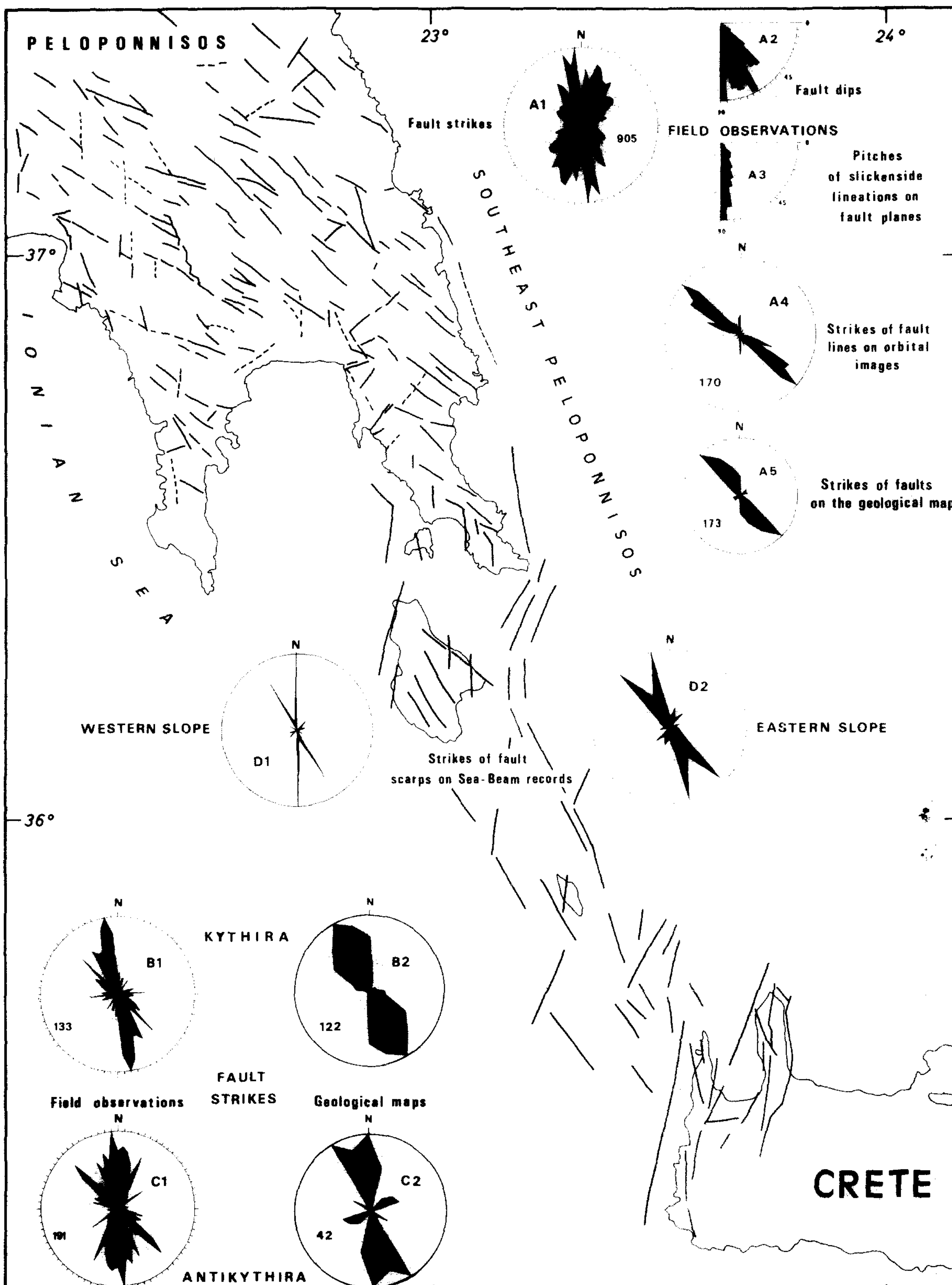


Fig. 7. Distribution of Late Cenozoic fault strikes in the Kythira strait area and in southeast Peloponnisos. Lines are major faults, from analysis of orbital images (land) and Sea-beam records (sea). (A) southeastPeloponnisos: distributions from field measurements of faults, from analysis of orbital imagery, from the geological map of Dufaure (1977). (B), (C) and (D) distribution of strikes on Kythira (B), Antikythira (C) and from the analysis of Sea-Beam records (D) (measured azimuths of scarp strikes counted with weighting as function of the vertical offset). Number of counted azimuths is indicated in each diagram.

clusion consistent with a 25–30° rotation. In addition, the initial strike of these old faults of south-east Peloponnisos fits very well with that of pre-Pliocene faults that have been identified by Jongsma (1975) in the western Sea of Crete. The N–S normal fault svstem, that has also been active in Crete since these times, progressively becomes NNE–SSW in southeast Peloponnisos and Kythira (Fig. 7); an observation also consistent with the rotation of Peloponnisos. On a broader scale, the implications of regional rotations in the whole of the southern Hellenic arc are discussed elsewhere (Angelier *et al.* 1982).

As Crete has not undergone significant rotation since the Late Miocene (Laj *et al.* 1982), the clockwise rotation of Peloponnisos may imply some equivalent counterclockwise rotation in the Kythira–Antikythira segment; no palaeomagnetic study, however, has been performed on the islands.

A MODEL OF DEFORMATION

The predominant role of extensional tectonics related to Late Cenozoic normal faulting is a major characteristic of the studied area as well as of other portions of the Hellenic arc. Notable peculiarities, however, are the submersion and the narrowness of this segment of arc. This contrast with other parts of the arc cannot be accounted for by variations of Alpine nappe structure,

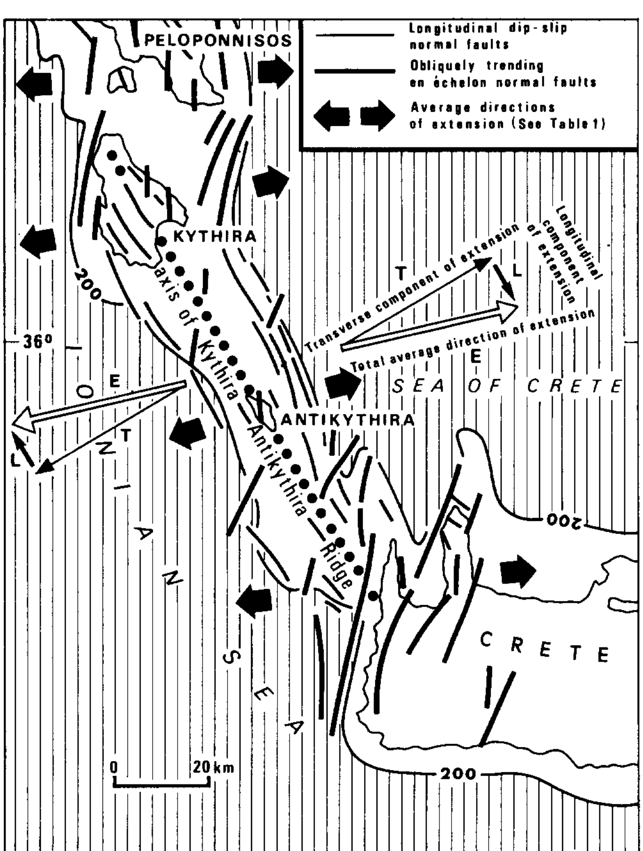


Fig. 8. Interpretation of the pattern of normal faulting in the Kythira strait area. Basins deeper than 200 fathoms are hatchured.

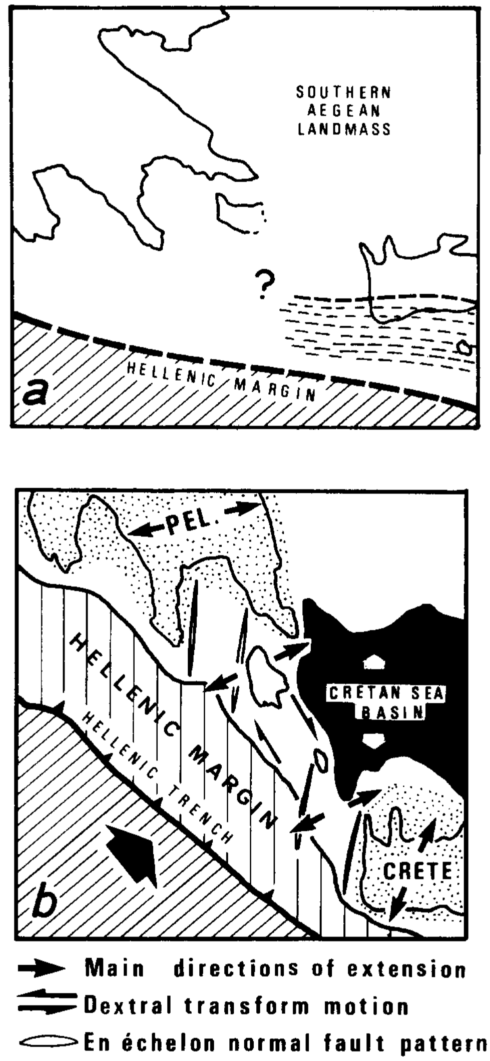


Fig. 9. Horizontal deformation of the southwestern Hellenic arc. (a) Beginning of Hellenic subduction associated with Aegean expansion, approximately 13–12 Ma ago (Late Serravallian–Early Tortonian). The dashed pattern designates the area where marine Late Serravallian deposits are identified. (b) Present day. Major blocks of the arc dotted. Wide black arrow: motion of Africa relative to Aegea. PEL, Peloponnisos.

because the orogenic belt of the Hellenides is continuous from Peloponnisos to Crete. We propose that it is a difference in the extensional neotectonic activity itself that explains the contrast between the wide uplifted segments of Peloponnisos and Crete, and the narrow low segment of Kythira–Antikythira. Directly evaluating extension rates by analysing the geometry of the faults as was done for Crete (Angelier 1979a) would be difficult in the strait where most faults are below sea-level. However, geological mapping (Theodoropoulos 1973 for Kythira and Fig. 3 for Antikythira) as well as bathymetric analysis (Fig. 2) suggest that the pattern of normal faults is especially dense in this area; no large, relatively rigid massif is present as on western Crete and southern Peloponnisos. Moreover, the Moho-depth map of Makris (1977) suggests that the crust is thinner beneath Kythira and Antikythira than beneath Crete and Peloponnisos. It is, thus, very likely that the whole of the Kythira strait has undergone more extension than adjacent segments of the arc (Peloponnisos and Crete). This, in turn, would easily explain its low altitude relative to other arc segments as extension induces subsidence.

Describing the evolution of faulting in southern Peloponnisos during the whole of the Late Cenozoic without taking into account the 25–30° clockwise rotation is impossible. The major normal fault system, that now strikes NW–SE, was trending WNW–ESE during the Late Miocene or Early Pliocene; the direction of extension probably being NNE–SSW. The system has been progressively rotated since and has become inactive, while a new major normal fault system, that now strikes NNW–SSE, has developed. Thus, the direction of extension was at first similar to that of Crete (N–S to NNE–SSW) and then became NE–SW to ENE–WSW during the Late Pliocene and Quaternary, while the Cretan Sea basin opened. This evolution is related to the progressive bending of the arc, which is demonstrated by the absence of rotation of Crete compared to the 25–30° rotation of Peloponnisos; the greatest part of the bending being absorbed in the Kythira–Antikythira strait. Unfortunately, the first stages of faulting are poorly understood, so that it is still difficult to reconstruct in detail the geometry of the deformation in the area prior to Late Pliocene–Quaternary times.

The pattern of the horizontal deformation related to the last stages of the Late Cenozoic normal fault tectonics is summarized in Fig. 8. The two major normal fault systems are NNW–SSE to NW–SE and N–S to NNE–SSW; analyses of fault mechanisms on land suggest that for both systems most faults are pure dip-slip faults, thus implying that corresponding directions of extension are ENE–WSW to NE–SW and E–W to ESE–WNW, respectively. This explains in detail, why the directions of extension that are accurately determined in Fig. 6 and Table 1 are close to ENE–WSW where NNW–SSE faults predominate, as near Antikythira, and close to E–W where N–S faults are abundant, as on both sides of the strait (Fig. 8). Finally, accurate analysis of the geometry and the mechanisms of faults exposed on the land, complemented by a study of submarine fault scarp patterns, suggests that the average azimuth of recent extension is 080 ± 5° in the 100 km long ridge between Peloponnisos and Crete. As the average strike of the ridge is 150°, the average angle between its axis and the direction of extension is 70°. Referring to Fig. 8, this means that the total recent extensional motion E includes a transverse extension T and a lateral motion L, relative to the axis of the ridge. The transverse motion T is principally related to movements on dip–slip normal faults that are longitudinal relative to the axis (i.e. NNW–SSE to NW–SE), while motion of obliquely trending en échelon normal faults (i.e. N–S to NNE–SSW) determine the lateral motion L that is equivalent to a dextral strike–slip displacement of both sides of the ridge (Fig. 8). However, the strike–slip component of recent extensional motion relative to the axis of the ridge is probably underestimated in Fig. 8, because most of the measurements that have enabled us to present a summary of the directions of extension in Table 1 were obtained in areas where the NNW–SSE to NW–SE fault system clearly predominates (see Fig. 6). On the contrary, areas where obliquely trending N–S to NNE–SSW normal faults of the en échelon system are well developed are either below sea-level (east and west of Kythira, north-west of Gramvoussa) or without abundant Late Cenozoic deposits that facilitate neotectonic analysis (e.g. the Gramvoussa and Spatha peninsulas, Fig. 2). Furthermore, most Sea-Beam records have been obtained in the central portion of the ridge, where the en échelon system is poorly represented. Consequently, the amplitude of the recent lateral motion L, shown in Fig. 8, is a minimum and could well be increased by a factor of two. In addition, we cannot exclude the possibility that dextral N–S strike–slip faults could be present, especially along the western coast of Crete, thus increasing the amplitude of this lateral motion. We could not find, however, any evidence that such system exists from our analyses of faults on land. The model depicted in Fig. 8 cannot, however, account for older deformations: their analysis will require accurate determinations of pre-Late Pliocene fault mechanisms and a knowledge of the local pattern of rotations that has prevailed in the area between Peloponnisos and Crete.

CONCLUSIONS AND DISCUSSION

Figure 9 summarizes the geometry of the deformation of the southwestern Hellenic arc. The general pattern of deformation of the Aegean region since 13–12 Ma has been described and discussed elsewhere on the basis of geological and geophysical observations (Le Pichon & Angelier 1979 & 1981, Angelier & Le Pichon 1980). This deformation resulted in both the clockwise rotation of Peloponnisos and the expansion of the Cretan Sea Basin (Fig. 9). Our results partly explain the particular role that the Kythira–Antikythira segment of the outer arc plays between the larger less deformed blocks of Peloponnisos and Crete. Faults that are parallel to the axis of the arc are well represented and correspond to mainly transverse extension; but, in addition, the presence of the en échelon normal fault system results in a dextral component of extensional motion parallel to the Kythira–Antikythira axis. Whether hypothetical large strike–slip faults are present or not, obviously modifies the amplitude of the dextral component of motion, but its existence seems beyond doubt. Our results principally deal with the last stages of the extensional deformation that began a long time before. However, the whole fault pattern of southeastern Peloponnisos is fairly consistent with its total 25–30° Late Cenozoic rotation; but the distribution of rotations between Peloponnisos and Crete still remains uncertain. In summary, the Kythira–Antikythira segment of the Hellenic arc acts as a complex zone of transform–extensional motion associated with rotations, at the western extremity of the N–S opening of the Cretan Sea basin.

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