Paleogene deep-water sedimentation and paleogeography of foreland basins in the NW Peloponnese (Greece)

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Abstract: The NW Peloponnese (Greece) belongs to the west-verging Alpine thrust-fault belt. Deep-water sedimentation ensued, in the Gavrovo-Tripolitza and Ionian foreland basins, as relief was being built-up and the Tertiary compression migrated westwards. The deep-water sedimentation and the deep structure of the thrust-fault belt are hereby assessed on the basis of interpreted seismic profiles, borehole and field data. The sedimentation is controlled by sea-level changes and thrust activity. The highest sedimentation rates for the Gavrovo-Tripolitza and Ionian Zones are observed during the Early Oligocene. In the Late Eocene, within the Gavrovo-Tripolitza Basin, middle to outer fan associations prevailed (Drossia-Charavghi Formation) changing to a channeled sea floor (Roupakia Formation) as the Pindos thrust front approached. A deceleration of the Pindos' advancement, combined with sea deepening, changed the environment to distal fan and hemi pelagic (lower Skouras Formation). On top of the Skouras Formation a regressive episode is marked. In Late Eocene, clastic sedimentation was installed in the Ionian Basin. First, distal fan facies overwhelmed the Ionian carbonate sedimentation (Mavri Miti Formation). In the Early Oligocene the Santameri Formation witnesses basin stability with distal characteristics in its lower parts. The lower Peta Formation, during the Late Oligocene is similar to the previous one. A rapid and important uplift of the Pindos hinterland is marked in Peta's upper members.

Key words: Greece, NW Peloponnese, Paleogene flysch, deep-sea fan evolution, foreland basins, fold and thrust belt, seismic-borehole-field sedimentary data analysis.

Introduction

The continuous deformation of the continental margin of the Adria microplate, during its collision–subduction with the Europa since the latest Cretaceous, has resulted in the formation of numerous foreland basins associated with fold-thrust belts. Beaumont (1981), Stockmal et al. (1986), and others, using a variety of mechanical descriptions for lithospheric flexure, have modeled foreland subsidence and the resulting thickness of the foreland basin fill. These geodynamic models showed that a lithospheric flexure caused by overthrust loading, coupled with the resulting topographical expression across the thrust belt, is an adequate mechanism for generating a foreland basin and the stratigraphy of the deposits within it. The eustatic sea-level changes and the thrusting activity control the evolution of the clastic facies of the submarine fan formations in the foreland basins. The combined action of these factors is recorded by the patterns of the clastic sediments as well as by their stratigraphic organization and depositional geometry.

After Brunn (1956) and Aubouin (1959) the External Hellenides s.l. in Western Greece (Pindos, Gavrovo-Tripolitza, Ionian and pre-Apulian Zones), consist of a series of sub-parallel, north–south trending tectono-stratigraphic zones including several east-dipping thrust sheets and west verging folds.

From the Late Triassic to Eocene, a deep basin (Pindos Ocean) developed between the Adrian and Eurasian plates, comprising deep-water carbonate, siliciclastic and siliceous rocks (Alexander et al. 1990; Degnan & Robertson 1998). The collision–subduction between these plates, mainly during the Tertiary, has progressively affected the External Hellenides by compressive tectonics giving rise to folding and thrusting of the previously mentioned zones. The deformation started in the Late Eocene within the Pindos Zone while the Gavrovo-Tripolitza, the Ionian and the pre-Apulian Zones were progressively involved during the Neogene–Pleistocene (Underhill 1989; Kamberis et al. 1996).

In this paper new data are presented concerning the submarine fan (flysch) facies of the Gavrovo-Tripolitza and the Ionian Zones in the NW Peloponnese during the Late Eocene to Late Oligocene (Fig. 1). Furthermore an attempt is made in order to interpret the depositional environments and their control mechanisms acting during the deposition of the flysch sediments.

The study of the deep sea sediments and their paleogeographical significance are mainly based on the turbidite mesoscale facies concept of Mutti (1974, 1977), Mutti & Ricci-Lucchi (1972, 1975) and Ricci-Lucchi (1975). In a smaller scale, the Bouma (1962) elementary sequence was used where applicable.

Furthermore the present structural geometry of the External Hellenides, based mainly on subsurface geology such as seismic information and borehole data, has been outlined.

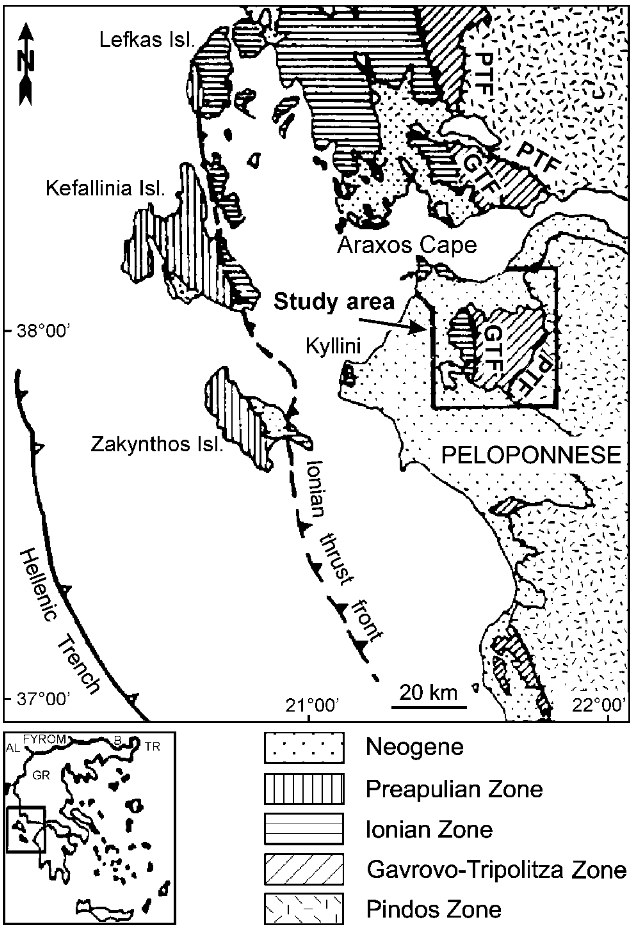

Fig. 1. Location map of the study area with the tectono-stratigraphic zones of the External Hellenides. **PTF** — Pindos thrust front, **GTF** — Gavrovo-Tripolitza thrust front.

Gavrovo-Tripolitza Zone

The Gavrovo-Tripolitza Zone represents a stable shallow carbonate dominated platform (Dercourt 1964). The Gavrovo-Tripolitza sediments include Middle Triassic to Upper Eocene/earliest Oligocene limestones and Eocene to Lower Oligocene flysch (Fleury 1980). The permanent shallow-water sediments (Late Jurassic to latest Cretaceous) are distinguished by successive regressive episodes (Bernier & Fleury 1980). Gavrovo-Tripolitza Zone is considered as part of the Adria microcontinent (Sorel 1976; Thiebault 1982; Underhill 1989). According to Alexander et al. (1990) this unit structurally underlies the Pindos thrust sheets.

Carbonate sequences

The carbonate sequences, exposed in Skolis Mt (Fig. 2), comprise Upper Cretaceous to Eocene thick-bedded carbonates starting with Senonian rudist limestones. According to previous studies the uppermost limit of the carbonate sediments, including the transitional beds, are of Early Oligocene age (Izart 1976; Fleury 1980). The transitional beds (3–10 m thick) are composed of a thin horizon of bioclastic carbonates with small-sized nummulites, marly limestones and marls. According to our field and micropaleontological analysis the transition between the carbonate and the clastic deep-water sedimentation was gradual without any interruption during the Eocene. This situation is typical in the northern flanks of Skolis Mt as well as further to the east, close to the Pindos thrust slices (Alepochori village, Fig. 2) but, due to the tectonic activity, it is not observed in the southern part of Skolis Mt. In the C–D section (Fig. 3) the contact between the limestones and the flysch of the Gavrovo-Tripolitza Zone is clearly marked by a strong well-organized reflector.

Turbiditic fan facies

The turbiditic fan facies, east of Skolis Mt, consist of three distinct formations. The prevailing typical facies types of these formations are presented in Fig. 4:

1 — The **Drossia-Charavghi Formation** is composed of Upper Eocene (Priabonian, biozone of *Turborotalia cerroazulensis* and *Areosphaeridium diktyoplokus)* to Lower Oligocene (biozone of *Globigerina ampliapertura)* pelitic facies. The lower horizons crop out close to the Pindos thrust front in three successive thrust slices near the Drossia and Alepochori villages. In these areas, the normal contact between the carbonates and the flysch sediments (Facies D_1, D_3 and G) is locally disturbed due to tectonic movements producing collapse structures.

Near the village of Charavghi, on the eastern flank of Skolis Mt, the uppermost horizons of the Drossia-Charavghi Formation are exposed. This part of the flysch consists of siltstones and thin-bedded alternations of very fine sandstones with small scale ripple marks (Facies D_1 and D_3). The sandstones include incomplete Bouma (1962) sequences (T_{c-e}, T_{d-e}). The above-mentioned flysch facies associations are interpreted as outer fan deposits, marking a sea transgression episode during the Early Oligocene.

2 — The **Roupakia Formation**, overlying the Drossia-Charavghi Formation, marks an important change in the sedimentation, characterized by the predominance of stacked channel sequences (Fig. 4). At the base of each sedimentary unit conglomerates (Facies A_2) and massive sandstone beds (Facies B) have been observed, while more pelitic intervals (Facies D) are progressively intercalated. The conglomerates have been deposited on the highly eroded upper surfaces of the pelitic beds of the underlying unit. Between these conglomerates are intercalated horizons presenting more chaotic structure expressed by slumps, olistostromes and olistoliths (Facies F) and rarely normal graded bedding corresponding to Facies A_1. In rare cases, imbricate pebbles indicate a roughly east-to-southwest current direction. Sandstones of Facies C succeed these conglomerates. Flute and groove casts show the same as above paleocurrent flow direction (Fig. 4). Mud clasts, fragments of lignite and vertical burrows, up to 60 cm in length, also exist suggesting an important input of sediments controlled by relatively strong turbidity currents during Early Oligocene times (biozone of *Globigerina ampliapertura* and *Wetzeliella gochtii).* From the base to the top of this formation, the lithofacies associations contain an increasing amount of

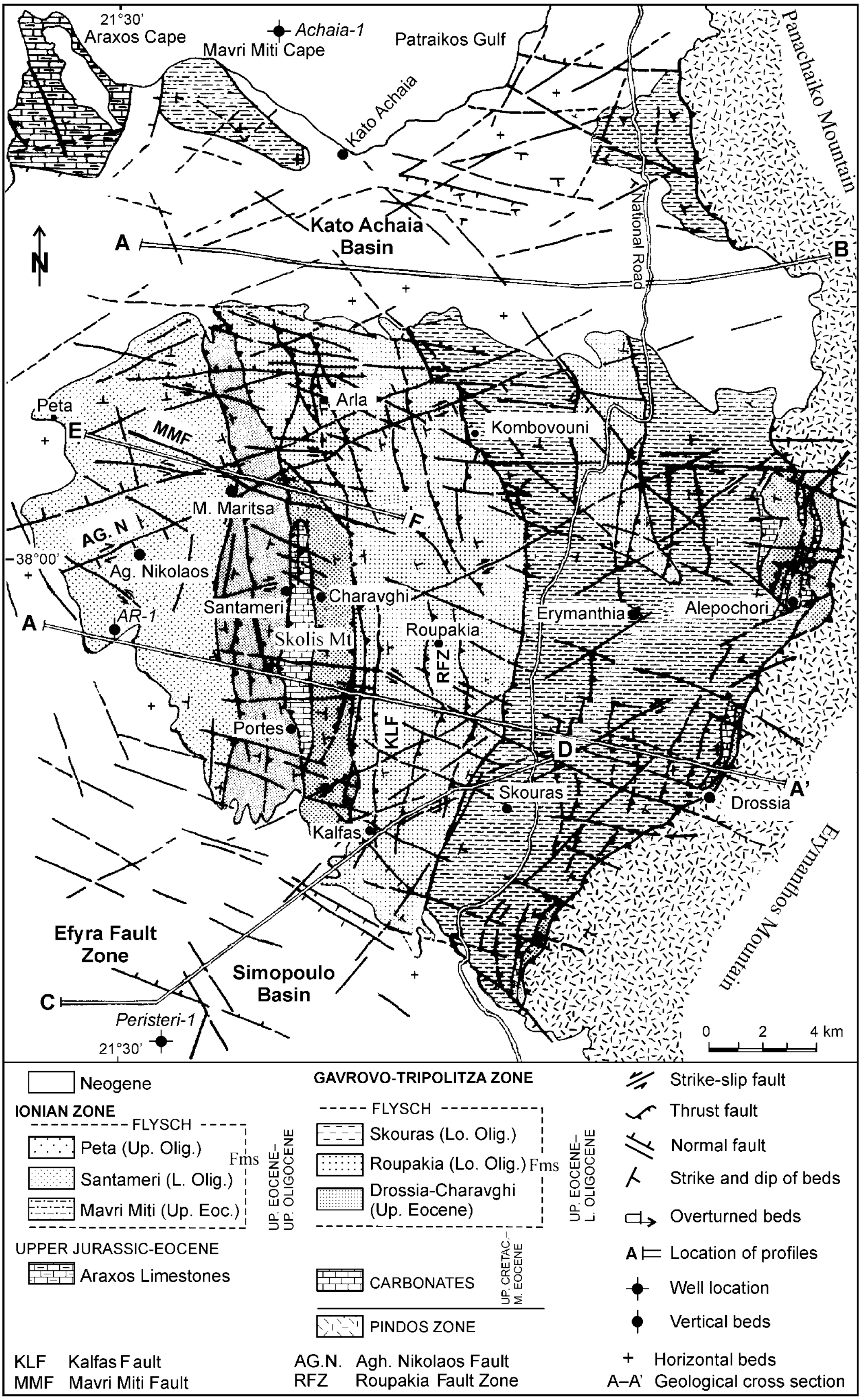

Fig. 2. Geological map of the NW Peloponnese.

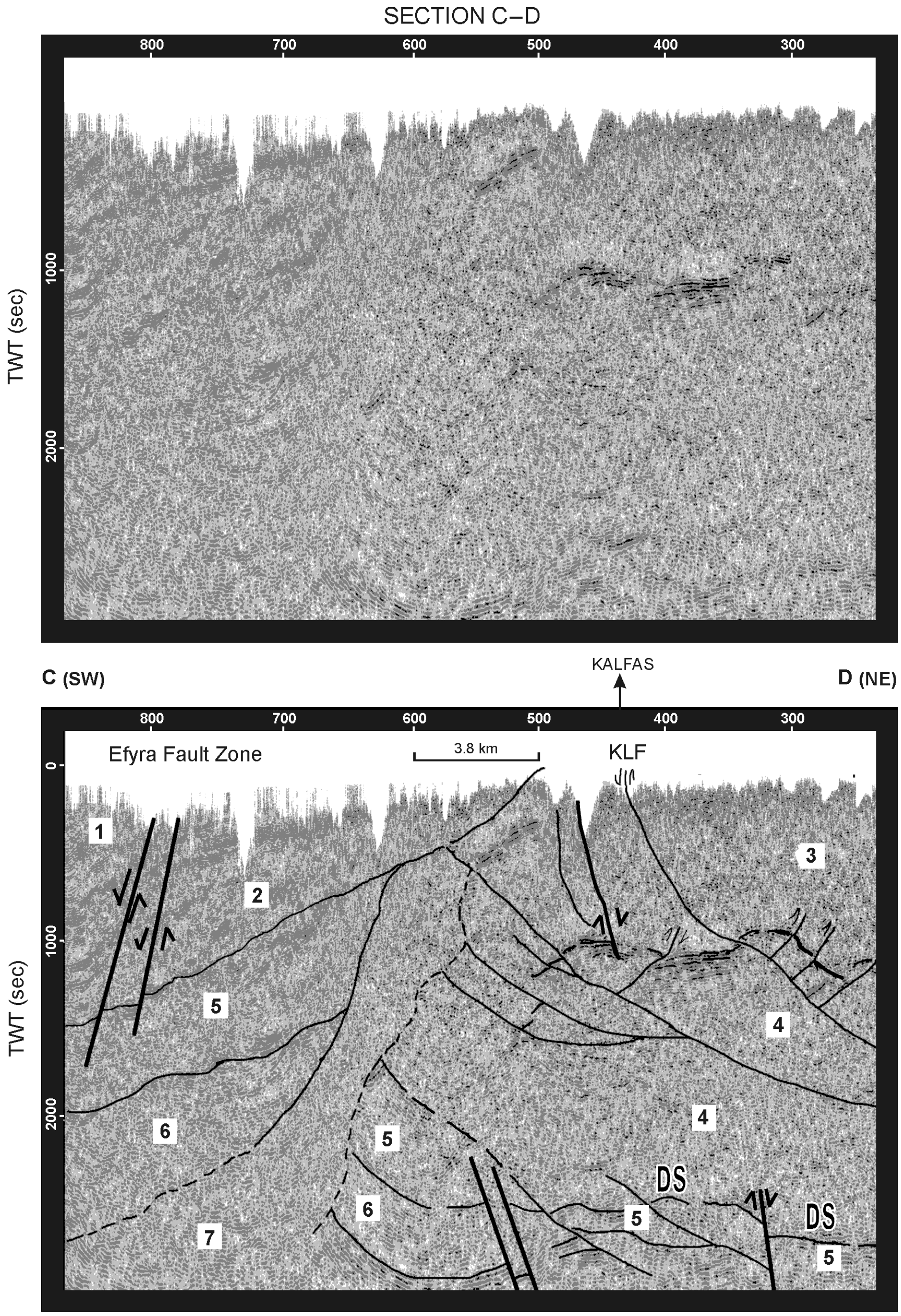

Fig. 3. Interpreted seismic profile C–D showing the faulted structure of the Gavrovo-Tripolitza Zone. **1** — Quaternary, **2** — Neogene (L. Miocene (?)–L. Pliocene, facies undivided), **3** — Flysch of Gavrovo-Tripolitza Zone, **4** — Gavrovo-Tripolitza limestones, **5** — Flysch of Ionian Zone, **6** — Ionian limestones, **7** — Ionian evaporites, **KLF** — Kalfas Fault, **DS** — Décollement surface. For location of the profile see Fig. 2.

pelitic material showing a decrease of the overall energy of the sedimentary gravity flows. The sandstones present an upwards fining and thinning sedimentary sequence with dispersed pebbles and small-scale synsedimentary structures such as slumps and ripple marks. Facies E are also presented. The above mentioned facies are interpreted as proximal and belonging to a levee environment.

3 — The **Skouras Formation** is mainly composed of pelitic facies overlying the Roupakia Formation (Fig. 4). The lowermost horizons are composed of alternations of coarse sand (C_2) and clays in positive sequences (middle fan). The intermediate parts of the Skouras Formation are dominated by thick pelitic intervals of facies D_3 and G, while upwards they alternate with coarser sandy facies of C_2 and D_1 type. These lithofacies are deposited during the Early Oligocene (biozone of *Globigerina ampliapertura* and *Wetzeliella gochtii*) in a deep marine environment (outer fan to basin plain). Abundant sedimentary structures, such as ripples, flute casts and slumps have been ob-

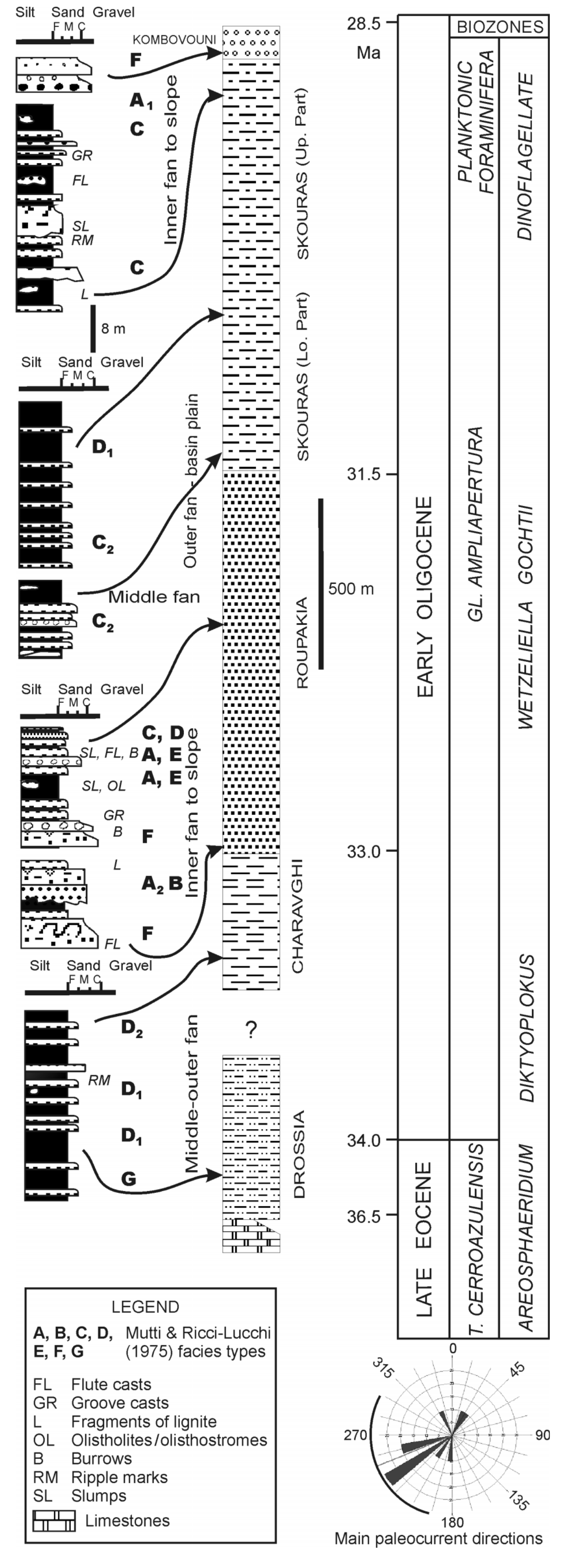

Fig. 4. Submarine fan facies of the Gavrovo-Tripolitza Zone with representative mesoscopic scale sequences.

served in the upper members of Skouras Formation sandstones suggesting a proximal marine environment. Finally, the uppermost part of the Skouras Formation is characterized by the presence of thick conglomerates, well exposed in the Kombovouni Mt (Facies F and A_1 mainly). These conglomerates include angular to subangular clasts derived from the Pindos Zone, indicating a relatively short distance of transport (Pavlopoulos & Tsagalides 1991). In the earliest Oligocene (Drossia-Charavghi Formation) the paleocurrent flow direction was towards the NW, while in a later stage (Early Oligocene) the direction was towards the SW (Roupakia and Skouras Formations). This remarkable change of the paleoflow direction may reflect a continuous change of the foreland basin topography (Pavlopoulos & Tsagalides 1991; Alexander et al. 1990).

The total thickness of the flysch formation in Gavrovo-Tripolitza Zone, deduced mainly by seismic and borehole data, is estimated about 3700 m.

Ionian Zone

The Gavrovo-Tripolitza carbonate platform passed westwards into an important intra-platform rift basin, the Ionian Zone. The oldest known strata of this zone are the evaporites including halite, anhydrite and gypsum accumulated in the Triassic basins and later mobilized. Although the evaporites crop out only in the Kyllini Peninsula, outside of the study area, they were drilled in several locations in the NW Peloponnese (Kamberis 1987; Kamberis et al. 2000).

Carbonate sequences

In continental Greece the lower parts of the sequence are composed of shallow water carbonates (Pantokrator Formation) corresponding to a pre-rift sequence of Early Lias (Aubouin 1959). Since the Late Lias a syn-rift period (Dercourt 1964; Karakitsios 1995) favours, in basinal areas, the deposition of pelagic sediments such as marls, carbonates, Posidonia shales and Ammonitico Rosso facies. Micritic cherty limestones (Vigla Formation) of the latest Malm to Early Cretaceous reflect the post-rift period. In the study area only the upper calcareous part of the Vigla Formation is exposed near the Cape Araxos, while the lower parts were identified in the well Sosti-1, located out of the studied area, close to the Peristeri-1 well (Fig. 2). The Upper Cretaceous to Lower Eocene deposits comprise alternations of microbrecciated and pelagic limestones. These carbonate facies, as well as in the Vigla Formation, include characteristic upwards thickening and coarsening carbonate sequences including ripple bedding levels indicating a deep-sea deposition environment.

During the Late Eocene meso- to large scale slumps and olistostromes are observed within the limestones expressing instability of the sea bottom during this period. This instability is also marked in the lower parts of the Ionian flysch sediments, as it is the case of the Mavri Miti Formation (s. next). A 21 m thick (1965–1986 m depth) carbonate-flysch transition zone has been found in the AR-1 well (Fig. 2) composed of limestones, marly limestones and subordinate sandstones, indicating an open marine to bathyal environment.

Turbiditic fan facies

The turbiditic fan facies, west of Skolis Mt, can be divided into three distinct formations, which have been recognized in field sections as well as in the AR-1 borehole lithology log (Figs. 5 and 6):

1 — The **lower Mavri Miti Formation** consists of uppermost Eocene turbidites (biozone of *Turborotalia cerroazulensis* and *Areosphaeridium diktyoplokus*), which are exposed in the Cape Araxos. In Klematia–Paramythia region, to the north in Epirus, a Middle Eocene age of the lowest parts of the Ionian flysch is reported (Avramidis et al. 2000). The lower parts of the Mavri Miti Formation are composed of Facies D_2 thin-bedded turbidites. These deposits are succeeded by Facies G including olistoliths of reddish Eocene limestone, with small-sized nummulites. These beds include chaotic deposits with olistoliths, sandstone olistostromes, slumps and flame structures.

Above these members, Facies C have been deposited separated from the previous ones by an erosional surface marked by the presence of mud clasts and fragments of lignite. The sandstone beds (Facies C) show a thickening and coarsening upwards trend and locally exhibit ball-and-pillow structures, in a lenticular sand body. Above the erosion surface the flysch sediments expose a fining-upward trend. The non-graded sandstones of Facies C_2 are bounded by even and parallel surfaces, including small scale ripple marks showing a weak current flow. The thickening upward trend of the pelitic intervals seems to be indicative of the deep marine conditions. Furthermore, in the upper part of this unit turbiditic sequences (Facies D_1) have also been observed.

In the upper members of the Mavri Miti Formation the flysch sedimentation is characterized by progressive coarsening-upward sequences. They are composed of Facies E, C_2 and D_1. Conglomeratic sandstones (Facies C) predominate at the top of this unit. Usually the top surfaces of the sandstone beds are in sharp contact with the overlying pelitic layers. The upper horizons of the Mavri Miti Formation are of Early Oligocene age (biozone of *Globigerina ampliapertura* and *Areosphaeridium diktyoplokus*), but contain an undeterminable fauna, while in previous works a post-Ypresian age has been reported (Tsoflias 1993).

In the section E–F (Fig. 7) the Mavri Miti Formation, as litho-seismic unit, is represented by the lower group of well organized reflectors. More sandy intervals correspond to the top of this unit and are composed of thin layered parallel reflectors.

2 — The Lower Oligocene flysch sediments (Fig. 5) are well exposed near the Santameri village, in the western side of the Skolis Mt. Pliocene-Quaternary sediments cover the transition between the Mavri Miti and the Santameri Formations. Thus, there is a lack of observation for about 700 m when compared to seismic sections.

The **Santameri Formation** consists mainly of very fine grained facies sediments (Facies D_2 and D_3) with sandstone intercalations (Facies C) especially in the lower part. A minor tectonic activity within this time interval (Early Oligocene) generated high energy turbidity currents capable of transporting coarser material. In this lower part the presence of *Helminthoidea* and *Paleodictyon* ichnofacies suggest a relatively

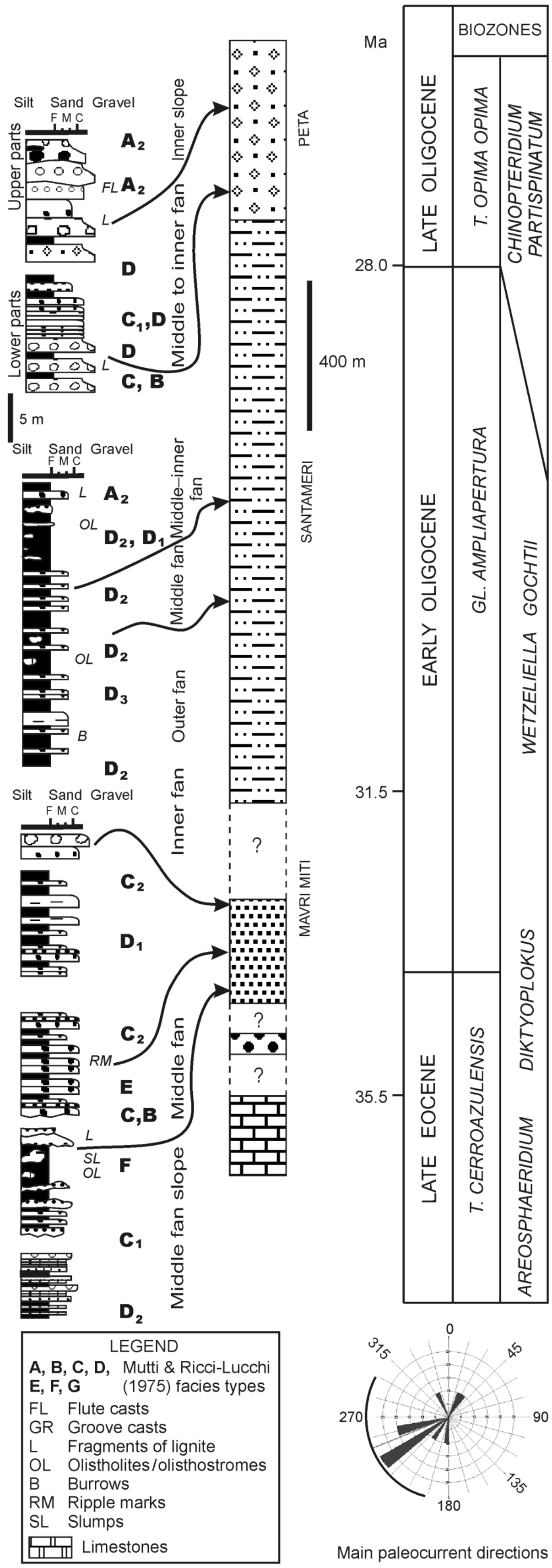

Fig. 5. Submarine fan facies of the Ionian Zone with representative mesoscopic scale sequences.

deeper marine environment (biozone of *Turborotalia opima opima* and *Wetzeliella gochtii*). The above mentioned lithofacies (D_2, D_3 and C) are mainly composed of fining upwards sequences. Within the uppermost horizons positive sequences are observed indicating the beginning of an important change of the depositional system (Mutti 1974). Furthermore, within these sequences, SW of Portes village conglomeratic lenses and limestone olistoliths are intercalated (Facies A_2?).

In the section of the Fig. 7 the middle to lower parts of the Santameri Formation are composed of fair to roughly marked reflectors. Towards the middle to upper parts of this formation the reflectors are relatively better portrayed marking the sandiest interval of this formation (see also Fig. 6, AR-1 well).

3 — In the Upper Oligocene **Peta Formation** (*Turborotalia opima opima* and *Chiropteridium partispinatum*) the dominant coarse sediments of Facies B and C, alternate with more pelitic intervals, mainly in the basal parts. In the uppermost strata (exposed north of Peta village) thick conglomerates and sandstones (Facies B and A_2) with high relief erosional surfaces mark a relatively shallow marine environment (inner to slope fan). These facies (B and A_2) are normally arranged in graded coarsening upwards beds. The pelites contain high amounts of sand and scattered pebbles. Field measurements of flute casts imply paleocurrents flowing to the southwest during the deposition of the Ionian flysch (Fig. 5). The Peta Formation corresponds to the uppermost part of the E–F section (Fig. 7) and it is represented by well marked thick parallel reflectors. The coarser material on the top of the Peta Formation corresponds to the well marked thickest reflectors, observed on the top of the E–F section (Fig. 7).

The total thickness of the Ionian flysch formation, as deduced from drillings (AR-1 well), seismic profiles and field observations is estimated at 2700 m, including the transitional beds.

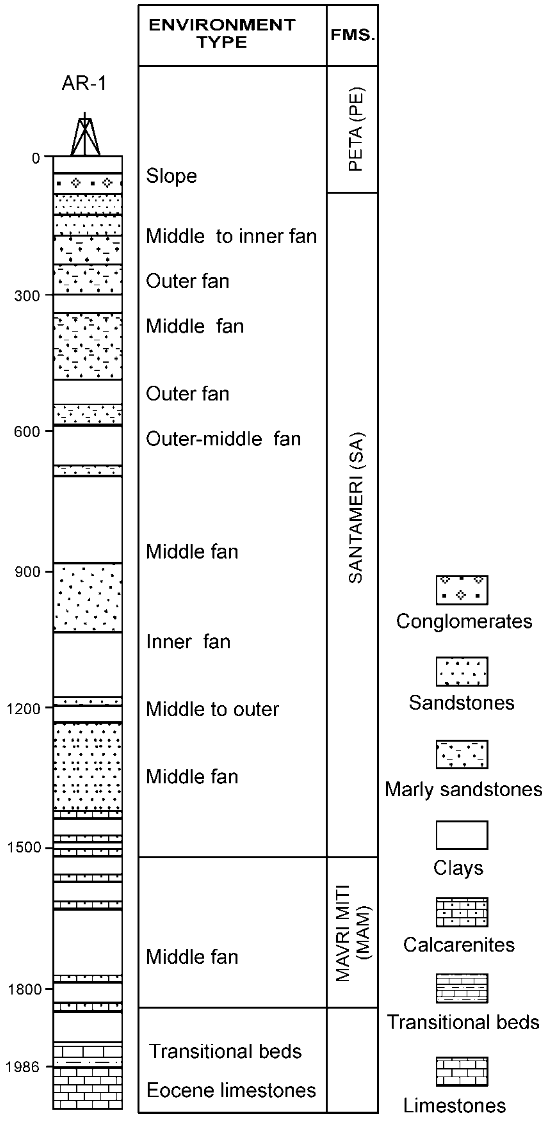

Fig. 6. Log of the Ionian deep-sea fan sequences (see also Fig. 5) issued from the AR-1 well. For location see Fig. 2.

Tectonic setting

On the basis of field observations and seismic profile interpretation, numerous compressional structures have been recognized, formed during westward thrust propagation.

The Pindos Zone is considered to be the main tectonic nappe in Western Greece. At the eastern part of the study area, the Pindos Zone is thrusted over the Upper Eocene-Lower Oligocene flysch successions of the Gavrovo-Tripolitza Zone inferring a post-Lower Oligocene westward movement. Likewise the Gavrovo-Tripolitza Zone has been affected by the westward migration of the orogenic front.

Between the Skolis Mt and the Pindos thrust front (PTF) several imbricated compressional structures are recognized (Fig. 2 and Fig. 9). The main structures correspond to intermediate-scale hanging wall anticlines and footwall synclines associated with high angle thrust and reverse faults, dipping to the east, as is the case near to Kalfas village (KLF), in Skouras (SKF) and Roupakia (RFZ) (Figs. 2 and 8). This results in the overall increase of the Gavrovo-Tripolitza flysch thickness (Fig. 8). In several cases, these faults seem to present a listric geometry passing downwards progressively into a low angle and finally subhorizontal surface situated at the top of the Ionian Zone flysch that is considered to be the décollement surface of the Gavrovo-Tripolitza Zone. This surface lies approximately at a depth of 4 to 5 km (0 to 2.5 sec T.W.T.) below the outcropping Gavrovo-Tripolitza flysch deposits, while it becomes deeper towards the Pindos thrust front (Fig. 9). In the C–D (Fig. 3) section, the décollement surface between the Gavrovo-Tripolitza limestone and the Ionian flysch is clearly outlined expressing a low angle thrust surface, dipping eastwards. The development of the hanging wall anticlines and imbricate fans has also been reported north of the study area, in Etolia-Acarnania region within the same foreland basin (Sotiropoulos et al. 2003).

This tectonic and structural style has been described in several foreland folds and thrust belts around the world such as in Southern Canadian Rocky Mountains (Bally et al. 1966), Southern Appalachians (Sachnic & More 1983) and Appeninnes (Kruse & Royden 1994).

The western limit of the décollement surface is a high angle east dipping surface corresponding to the Gavrovo-Tripolitza thrust front (GTF, Figs. 7 and 9) of Skolis Mt. As a result, the

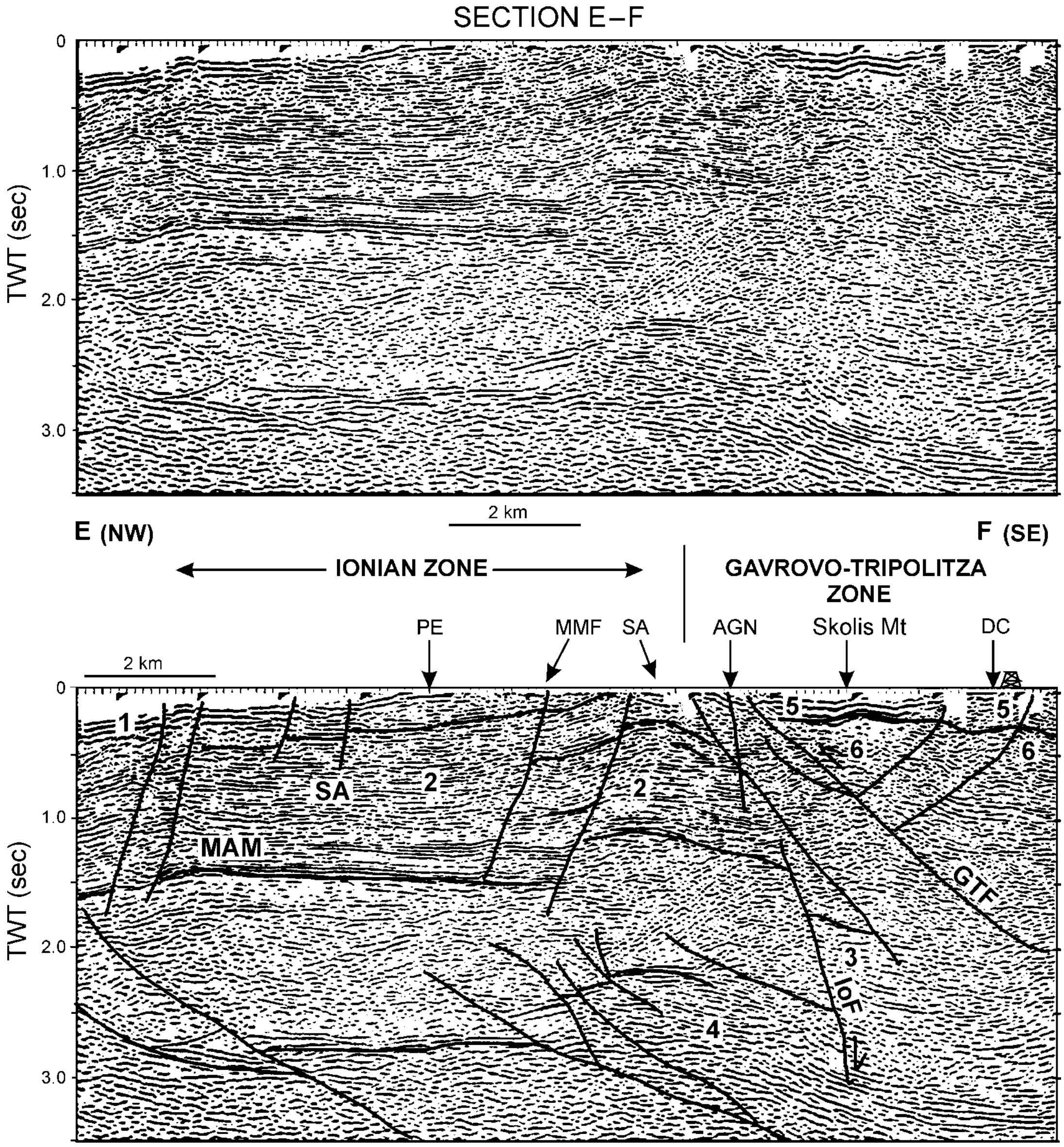

Fig. 7. Interpreted seismic profile E–F showing imbricate reverse faults in the Ionian Zone, close to the Skolis thrust (right uppermost part). Faults are marked in the footwall as well as in the basement of Triassic evaporites. The latter are involved in the shortening process in a relatively later time period. Note the normal fault IoF, underneath the Skolis thrust characterized by a significant throw (0.8 sec/TWT). **1** — Neogene-Quaternary, **2** — Flysch of Ionian Zone, **3** — Ionian limestones, **4** — Evaporites, **5** — Flysch of Gavrovo-Tripolitza Zone, **6** — Gavrovo-Tripolitza limestones, **IoF** — Ionian normal fault, **GTF** — Gavrovo-Tripolitza thrust front, **AGN** — Aghios Nikolaos fault, **MAM** — Mavri Miti Formation, **SA** — Santameri Formation, **PE** — Petas Formation, **MMF** — Moni Maritsas fault zone, **DC** — Drossia-Charavghi Formation. For location of the profile see Fig. 2.

Cretaceous-Eocene limestones of Skolis Mt may be considered as a characteristic narrow hanging wall half-anticline related to the thrust front. West of Skolis Mt the upper horizons of Ionian flysch successions are of Late Oligocene age (Peta Formation). Consequently it is considered that the Gavrovo-Tripolitza thrust activity started, at least since the earliest Miocene or earlier.

The Ionian Zone, successively, has been involved in the westward propagating thrust belt resulting in thrusts and large-scale folds. In contrast to the Gavrovo-Tripolitza Zone, the Ionian Zone is characterized by open anticlines (E–F section, Fig. 7) and synclines, reflecting a more ductile style deformation related to the presence of Triassic evaporites (Kamberis et al. 2000). It is assumed that these evaporites played an important role in the geodynamic evolution of the more External Hellenides.

The thrust faults, formed in the Ionian Zone, seem to sole downwards in the evaporitic layer considered as the detachment horizon of this zone (C–D section, Fig. 3). At the location of the above-mentioned compressional structures diapiric movements have been developed since the Pliocene, affecting the Pliocene-Quaternary sedimentation (Kamberis et al. 1996).

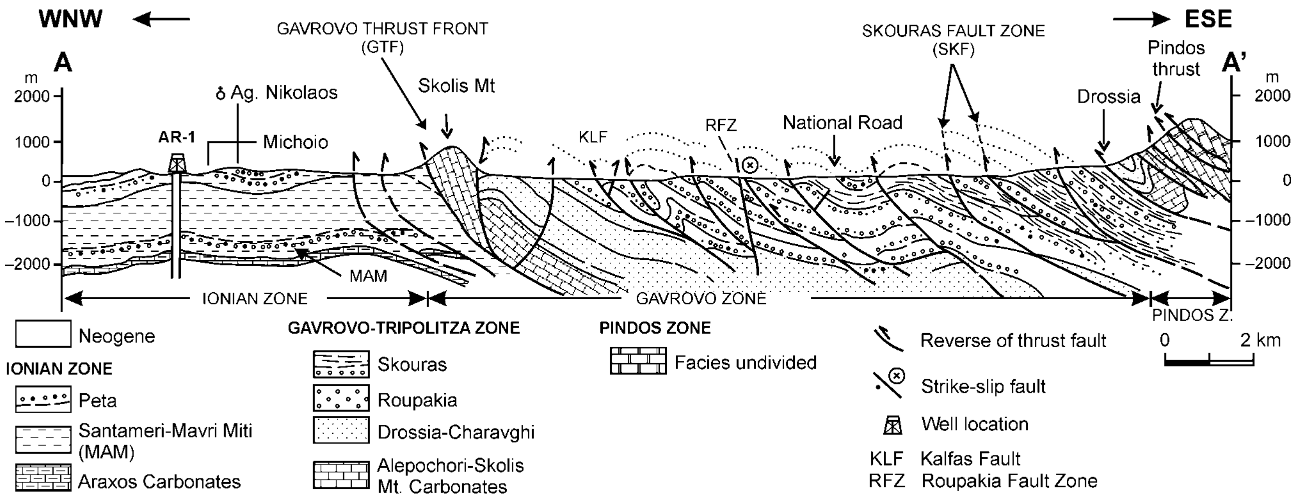

Fig. 8. Cross-section through the Skolis Mountain and the Pindos thrust (A–A′ in Fig. 2) showing the main contractional structures of the Gavrovo-Tripolitza and the Ionian Zones (modified from Kamberis et al. 2000).

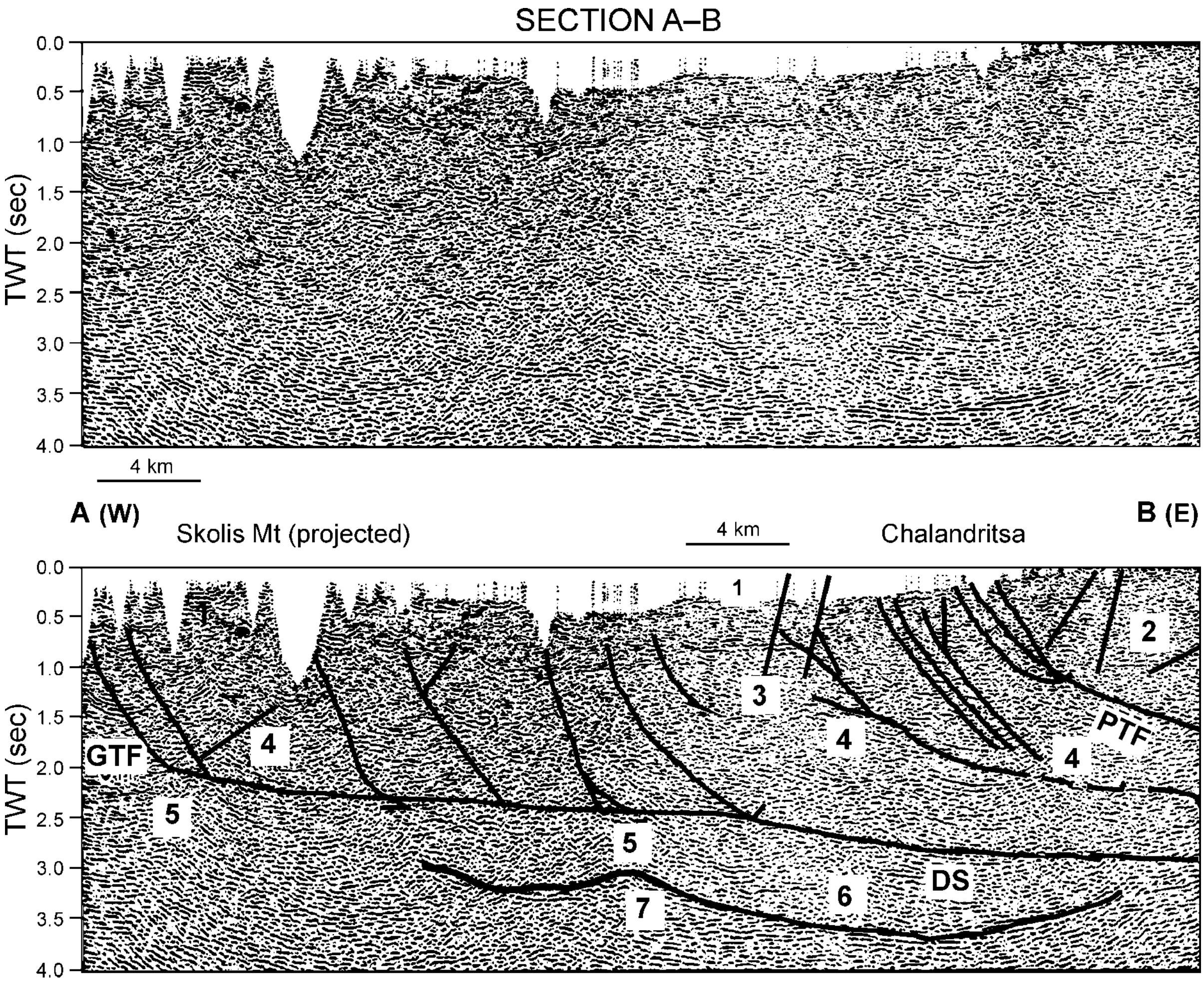

Fig. 9. Interpreted seismic profile A–B showing the internal structure of the Gavrovo-Tripolitza Zone in the Kato Achaia Basin. Notice the flexible deformation of the underlying Ionian Zone related to the presence of the Triassic evaporites. **1** — Neogene–Quaternary, **2** — Pindos Zone (facies undivided), **3** — Flysch of Gavrovo Zone, **4** — Gavrovo limestones, **5** — Flysch of Ionian Zone, **6** — Ionian limestone, **7** — Triassic evaporites, **GTF** — Gavrovo-Tripolitza thrust front, **DS** — Décollement surface, **PTF** — Pindos thrust front. For location of the profile see Fig. 2.

In the eastern part of the E–F seismic profile (Fig. 7) a shallow group of prominent seismic reflectors is distinguished and interpreted as the top of the Gavrovo-Tripolitza carbonate sequences. These limestones are covered by a few hundred meters thick of shaly flysch at Skolis area, west of the main Gavrovo-Tripolitza thrust. Thrust fault planes of lower order locally exist in the lower sandy flysch members of the Santameri Formation, as it is the case of Portes village and WNW of Arla village, near Maritsa Monastery. In these locations a typical ramp-flat geometry is observed in several outcrops. In the same seismic profile the lower group of reflectors correlates with the base of the flysch sequence. Ramp-flat thrust geometry is also recognized at the base of the Ionian carbonate series.

Furthermore, field observations in the Gavrovo-Tripolitza and Ionian Zones reveal the existence of several secondary WNW–ESE and ENE–WSW trending, sinistral and dextral respectively, strike slip faults (MMF and AGN, Fig. 2), which cut almost perpendicularly the previously mentioned thrust zones. These faults are considered as accompanying features, formed during the westwards thrust propagation in Miocene times (Kamberis et al. 2000). Some of these faults are reactivated, during the latest Cenozoic, as normal faults with quasi vertical angle of pitch. This is the case of MMF and AGN fault zones.

Moreover in the study area numerous mesoscale faults have been measured on Tertiary to Quaternary sediments. The WNW–ESE trending normal faults affect both the flysch and the Pliocene sediments, reflecting a more recent faulting process. This type of faulting has also been reported for an area adjacent to the southern coastline of the Patraikos Gulf (Fig. 2, Kamberis 1987; Zelilidis et al. 1988). In many cases, the WNW–ESE faults coexist with W–E and ENE–WSW trending normal faults in the previously mentioned areas. The formation of these faults is related to the late north-south extensional stage since the Late Pliocene.

Tectonosedimentary evolution of the foreland basins — Discussion

The examined deep-sea fan formations were deposited in foreland basins developed in front of an active thrust belt as the typical characteristics of this kind of basins are fulfilled:

— There is a horizontal compression generated by coupling at the contact between subducting and overriding plates (e.g. Kanamori 1986).

— There is a pre-existing thick pile of layered strata (e.g. Allmendiger et al. 1983) corresponding to the Pindos Zone for the Gavrovo-Tripolitza Basin and the Pindos and Gavrovo-Tripolitza Zones for the Ionian one.

— The foreland basins are a result of the deformation of a foreland fold-thrust belt (Fig. 10).

The long-term accumulation in a foreland basin requires tectonic subsidence. Nonetheless, surface processes impact such a strong influence on the stratigraphy that many times the pure tectonic signal is obscured. The main types of controls that are independent of the local tectonic activity are craton lithology, climate, eustatic sea-level and geological age.

In order to correlate the different types of the flysch sequences to the tectonic activity and/or other processes the following observations are made: the Gavrovo-Tripolitza submarine fan sequences, east of Skolis Mt, start in Late Eocene with the Drossia-Charavghi facies associations that correspond mainly to middle to outer fan deposits, indicated by weak turbidity currents with rare hemipelagic inter-

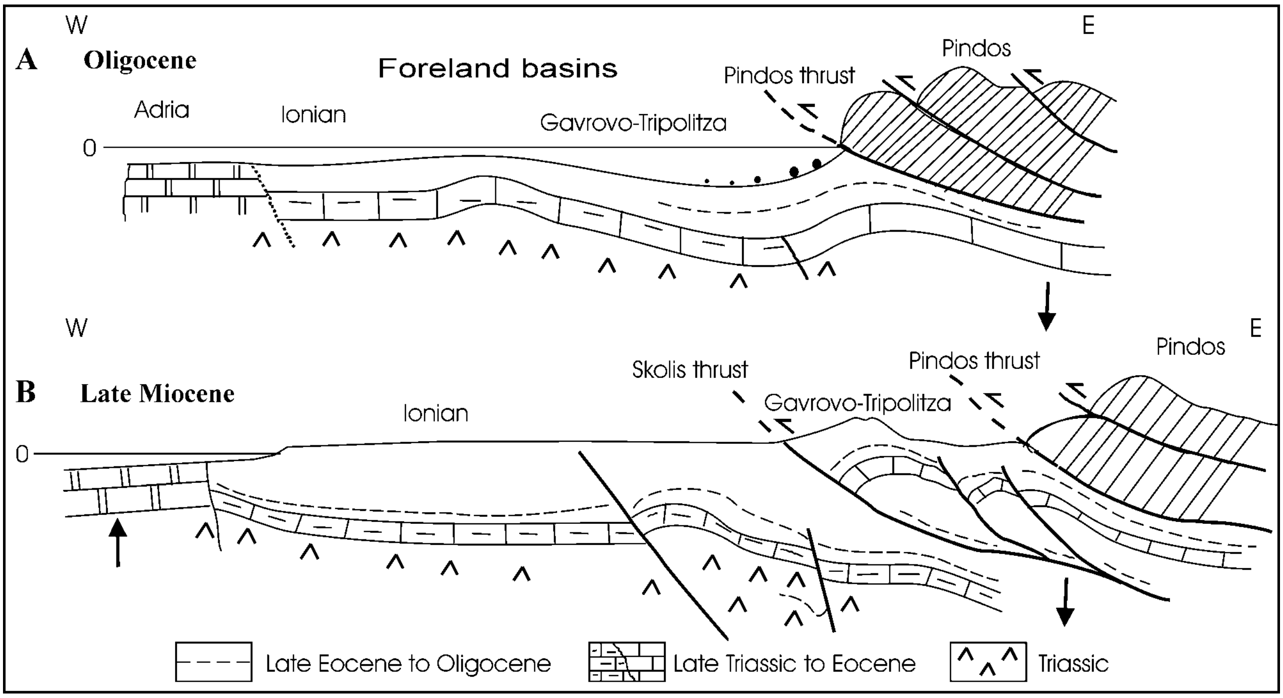

Fig. 10. Generalized situation of the foreland basins during Eocene–Miocene.

beds. The approach of the Pindos thrust pile, in the Early Oligocene, caused an important change of the sedimentary facies to inner fan/slope type deposits (Roupakia).

A calmer interval of the thrust advancement is clearly observed in the middle–upper members of the lower part of the Skouras Formation, which continue the up section evolution. Later on the environment changes to outer fan turbidites. The upper parts of the Skouras Formation indicate a new prograding phase of the thrust front, as the facies become coarser corresponding to an inner fan to slope/canyon environment.

From the above it is concluded that the flysch sediments east of Skolis Mt represent a deep-sea fan influenced by an important regressive sea level episode that took place in the Early Oligocene, during the deposition of Roupakia Formation. Progressively, a sea transgression followed, and resulted in the deposition of the Skouras Formation (Fig. 11). The sedimentological data of the Roupakia Formation imply a high-energy depositional environment. The high-energy conditions, which controlled the Skouras sedimentation, are not in accordance with the general trend of the eustatic curve (Haq et al. 1987; Fig. 11). This fact is related to the important tectonic activity along the Pindos thrust. The Kombovouni conglomerates, on the top of the Skouras Formation, mark the end of the sea transgression and correspond to the beginning of a new rapid regressional episode, which can be related to the combined action of tectonics and the rapid fall of the sea level.

West of Skolis Mt the Upper Eocene Mavri Miti Formation, in Araxos region, shows a prograding tendance of the compressional front. This is underlined by the proximal facies corresponding to an inner fan and slope environment. The overlying members show a more even area of deposition that favours the formation of an inner to middle fan environment with channel fill and overbank facies. Rapid deposition of sand produced flame and ball-and-pillow structures. The more or less important synchronous seismic activity has facilitated this procedure.

The lowermost part of the Mavri Miti Formation (Late Eocene) is mainly influenced by the tectonic activity to the east related to the Pindos thrust (Fig. 11). Progressively a more proximal environment (inner fan) was installed. The Lower Oligocene Santameri Formation corresponds mainly to a more distal environment. The presence of abundant trace fossils underlines a relative stability during the deposition of the Santameri Formation. The lithoseismic characters expressed by non-deformed thin layered, well-organized, continuous and parallel reflectors confirm this stability (Fig. 3). This event is in accordance to the corresponding eustatic high sea level. Towards the upper parts of the Santameri Formation a more proximal environment was installed.

The Upper Oligocene Peta Formation shows a rapid upward coarsening due to a more westwards progradation of the orogen as well as to the contemporaneous fall of the sea level (Fig. 10A).

The overall evolution of the Ionian turbiditic flysch deposits, considered together as a great scale unit, exhibit a prograding trend that becomes more important in the upper conglomeratic members of the Peta Formation.

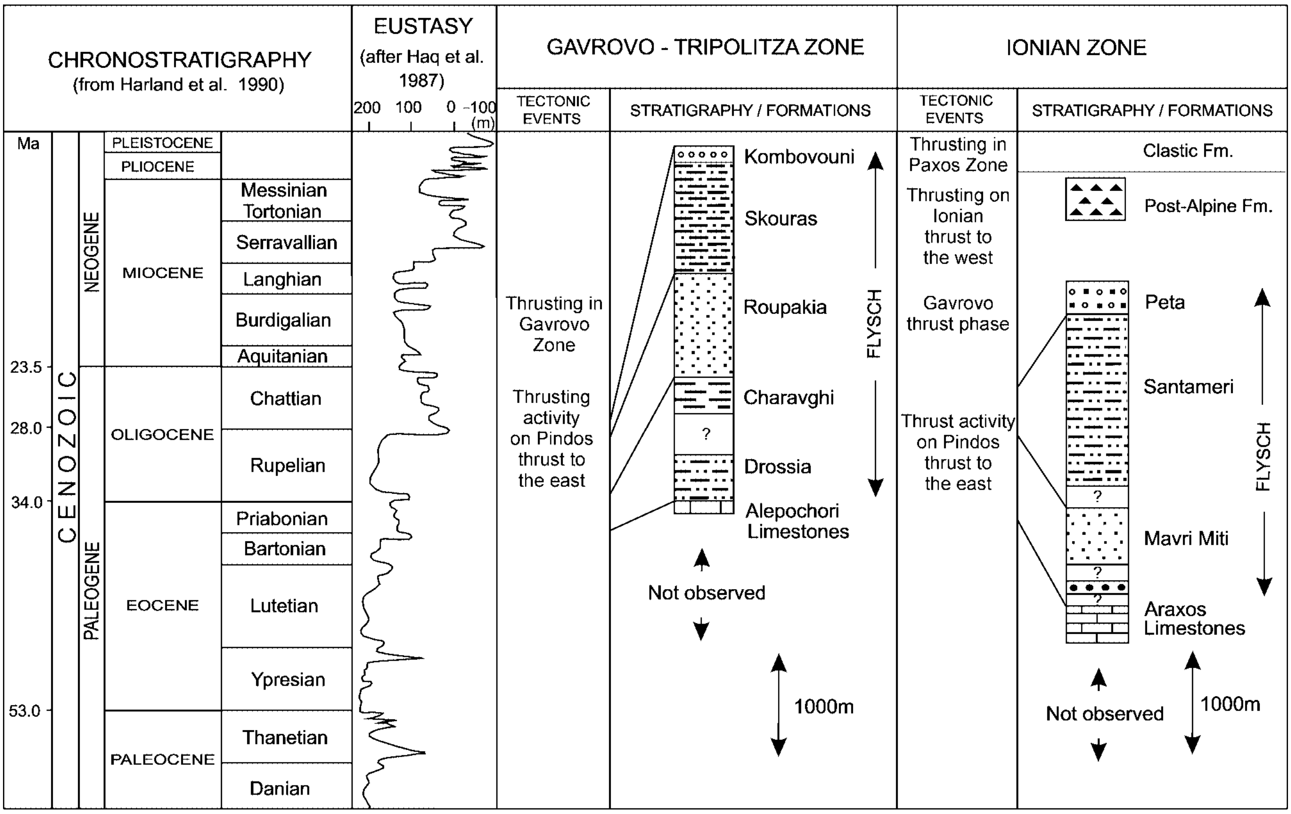

Fig. 11. Comprehensive diagram showing the tectonic events and the corresponding sedimentary formations in relation to the eustatic curves. The formations' thickness is based on well and seismic data.

The paleocurrents in both zones (Figs. 4 and 5) have a general W to SW flow direction with the exception of the Drossia-Charavghi Formation (Gavrovo-Tripolitza Zone) where the direction is NW. This is in agreement with the paleocurrent measurements reported from Piper et al. (1978) in the flysch of Etolia, north of the Patraikos Gulf and in Makrynoros area, west of Gavrovo Mt (Pavlopoulos 1983; Vakalas et al. 2001). In contrast they only partly correspond to the measurements of Gonzalez-Bonorino (1996) in the east of the Amvrakikos Gulf area.

In the studied area the progressive overthrusting to the west has resulted in the increase of the tectonic load and caused a lithospheric flexure towards the orogene (Fig. 8). Underhill (1989) has also reported the same assumption in the islands of Zakynthos and Kefallinia. Due to this overloading, a foreland basin was formed in front of the orogenic belt, where synorogenic sediments were deposited in deep water conditions. Furthermore, mountainous areas have been progressively formed reflecting the underlying footwall ramps as in Erymanthos and Skolis Mt. During this process, the more external parts have progressively formed a tectonic wedge migrating westwards.

The resulting overloading of the lithosphere, due to the Ionian thrust emplacement on the Preapulian Zone, since Early Pliocene, caused a further lithospheric flexure.

In order to calculate the sedimentation rates of the studied flysch formations, in both zones, the true thicknesses of these formations were calculated (Table 1). The calculation was a product of field observations, as on Fig. 8, combined with seismic information (Fig. 7) after conversion of the time scale to a depth one. The results are presented in Fig. 12 from where it is observed that the sedimentation rates in the Gavrovo-Tripolitza Zone are generally higher than those of the Ionian Zone. This is related to the combined action of the sea deepening during the Rupelian (Fig. 11) which provided more accommodation space, and the important thrust activity in the Pindos Zone to the east. During this period the sedimentation was extended to the most distal parts of the relatively deeper Ionian Basin. In the higher horizons of the Gavrovo-Tripolitza flysch (Skouras Formation) the sedimentation rate remains high compared to the partly synchronous Peta Formation of the Ionian Zone.

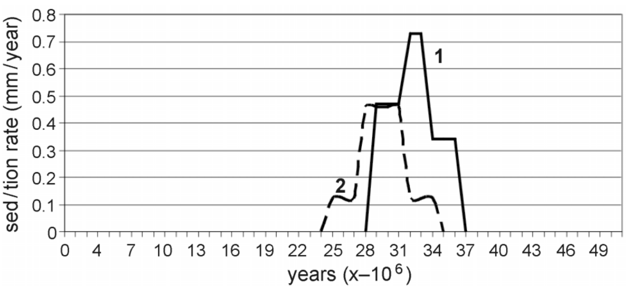

Fig. 12. Curves expressing the sedimentation rates during the flysch deposition in the Gavrovo-Tripolitza and Ionian Zones.

Conclusions

The stratigraphic and map pattern of lithofacies classification infer the depositional setting of ancient submarine fans within the foreland basins of the Gavrovo-Tripolitza and the Ionian Zones. This setting is related either to the eustatic sea-level changes and/or the thrust tectonic activity, which occurred during the Paleogene in the External Hellenides.

The Gavrovo-Tripolitza turbiditic fan sediments correspond, generally, to a more proximal depositional environment than that of the Ionian one. In addition, during distinct periods the deposition of the Gavrovo-Tripolitza flysch was more affected by the thrust activity along the Pindos thrust to the east than the Ionian flysch sedimentation. In the Gavrovo-Tripolitza sequences two major prograding cycles are distinguished:

a — The older one that ends with the Roupakia Formation is mostly related to a more intense tectonic activity. This event does not fit the general trend of the eustatic sea-level change.

b — The younger one includes the basal parts of the Skouras Formation. The Kombovouni Conglomerates, on the top of the Skouras Formation correspond to the beginning of an important sea level regressional episode.

The greater part of the Ionian turbiditic fan facies were deposited in a relatively deeper sea-fan environment than that of the Gavrovo-Tripolitza Zone. The lower part of the Ionian flysch (Mavri Miti Formation) is marked by two major overlapping fan prograding episodes. The end of the first one is marked by a well developed erosion surface and other typical structures of slope environment, while the second one is incomplete. The upper part of the Ionian flysch (Santameri and Peta Formations) reflects mainly a progradational sequence. The upper members of Peta Formation express an intense orogenic thrust activity, accompanied by a well marked important sea level regressive episode.

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Table 1: Thickness and sedimentation rates during the deep-water sedimentation in the Gavrovo-Tripolitza and Ionian zones.

Gavrovo-Tripolitza Zone					Ionian Zone				
Formation	Thickness (m)	Time span -10^6	Duration 10^6	Rate mm/years	Formation	Thickness (m)	Time span 10^6	Duration 10^6	Rate mm/years
Skouras (+Kombovouni)	1400	28.0–31.5	3.0	0.47	Peta	600	24.0–28.0	4.0	0.15
Roupakia	1100	31.5–33.0	1.5	0.73	Santameri	1600	28.0–31.5	3.5	0.46
Drossia-Charavghi	1200	33.0–36.5	3.5	0.34	Mavri Miti	500	31.5–35.5	4.0	0.12

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