Basin Research (1998) **10,** 393–408

Tectonic and sedimentological evolution of the Pliocene–Quaternary basins of Zakynthos island, Greece: case study of the transition from compressional to extensional tectonics

A. Zelilidis,* N. Kontopoulos,* P. Avramidis* and D. J. W. Piper†

**Department of Geology, University of Patras, 26500 Patras, Greece*

†Atlantic Geoscience Centre, Geological Survey of Canada (Atlantic), Bedford Institute of Oceanography, P.O. Box 1006, Dartmouth, NS, B2Y 4 A2, Canada

ABSTRACT

Pliocene–Quaternary basins of the Ionian islands evolved in a complex tectonic setting that evolved from a mid to late Cenozoic compressional zone of the northern external Hellenides to the rapidly extending Pliocene–Quaternary basins of the Peloponnese. The northern limit of the Hellenic Trench marks the junction of these two tectonic regimes. A foreland-propagating fold and thrust system in the northern external Hellenides segmented the former Miocene continental margin basin in Zakynthos and permitted diapiric intrusion of Triassic gypsum along thrust ramps. Further inboard, coeval extensional basins developed, with increasing rates of subsidence from the Pliocene to Quaternary, resulting in four principal types of sedimentation: (1) condensed shelf-sedimentation on the flanks of rising anticlines; (2) coarse-grained sedimentation in restricted basins adjacent to evaporitic diapirs rising along thrust ramps; (3) larger basins between fold zones were filled by extrabasinal, prodeltaic mud and sand from the proto-Acheloos river; (4) margins of subsiding Quaternary basins were supplied at sea-level highstands by distal deltaic muds and at lowstands by locally derived coarse clastic sediment.

INTRODUCTION AND GEOLOGICAL SETTING

Background and purpose

The Ionian islands of western Greece form part of the para-autochthonous Apulian foreland of the Hellenide orogen and include rocks of the Pre-Apulian (or Paxos) and Ionian isopic zones (=terranes) (Underhill, 1989). Overlying Apulian crystalline basement, Triassic halite and gypsum (only in the Ionian zone) are succeeded by thick Mesozoic and Palaeogene limestones in both zones. In the Oligocene, closure of the Pindos remnant ocean resulted in westward thrusting of the Pindos nappes and collision of Apulia with the Pelagonian microcontinent. A foreland basin developed in the Ionian isopic zone west of the Pindos nappes, within which several kilometres of lower Oligocene – lower Miocene flysch accumulated, passing westward into thin lower Miocene marls (Brooks *et al.*, 1988). The locus of thrusting propagated westward in the Ionian zone (Alexander *et al.*, 1990), so that by the late Oligocene several thrusts segmented the original flysch basin. Fold and thrust deformation in the northern Ionian island of Levkas (Fig. 1) resulted in a mid-Miocene unconformity (Clews, 1989), and similar deformation reached Zakynthos in the early Pliocene (Underhill, 1989). The thrusting propagated along Triassic evaporites (Kamberis *et al.*, 1996) and resulted in a series of thrust anticlines and intervening basins (Fig. 1). Thrusting continued through the Quaternary in Zakynthos, resulting in the progressive uplift of Pliocene and lower Pleistocene sediment and Quaternary marine terraces (Sorel, 1976). Triassic gypsum in the cores of the anticlines was mobilized into diapirs (BP Co Ltd, 1971; Brooks & Ferentinos, 1984; Underhill, 1988) (Fig. 1).

The regional tectonic evolution of the Hellenide orogen in the later Cenozoic is a consequence of the progressive collision of Eurasia and Africa in the eastern Mediterranean and Middle East. There are several competing regional hypotheses. Some authors emphasize the gravitational collapse of the Rhodope and Hellenide

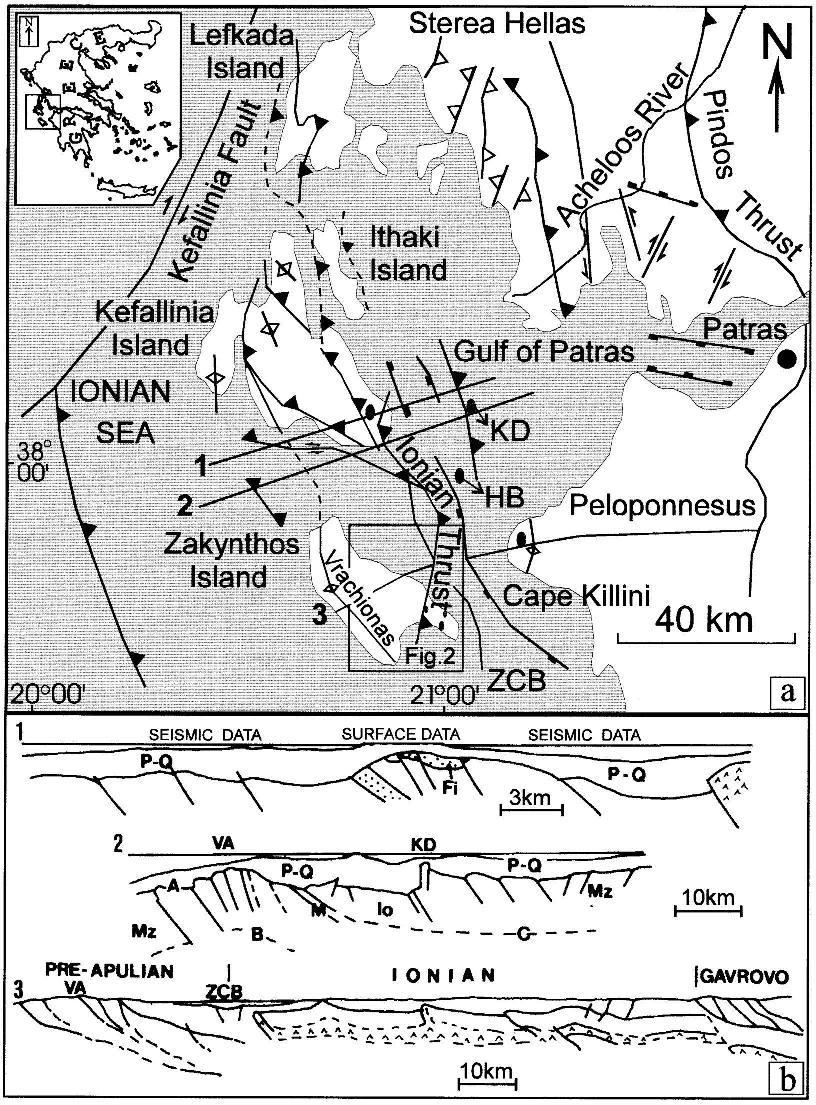

Fig. 1. (a) Regional map of the Ionian islands, based on Brooks & Ferentinos (1984), Underhill (1989) and Kamberis *et al.* (1996). Inserted box shows Ionian islands situated at the western part of Greece. (b) Structural cross-sections in the vicinity of Zakynthos (modified from Kamberis *et al.*, 1996) showing the relationship of basins to thrusts. ZCB = Zakynthos Channel Basin. Shaded ellipses are diapirs of Triassic evaporites, including HB = Hydra Bank; KD = Kefallinia Diapir; Fi = Flysch (Eocene–Oligocene); P-Q = Pliocene–Quaternary deposits; VA = Vrachionas anticline; Io = Ionian Zone; M = Miocene; A–C = main unconformities.

orogens (which show collisional thickening in the Cretaceous and Eocene respectively) (Doutsos *et al.*, 1993). Others have emphasized the importance of dextral strike-slip motion of Turkey relative to Europe, as a result of collision of Arabia and Eurasia (McKenzie, 1978). This motion is partly taken up by collision of north-western Greece with the Apulia–Adriatic platform of eastern Italy (Underhill, 1989) and partly by rollback of the South Aegean subduction zone (Royden, 1993), including the north-eastern segment of the subduction zone where the Ionian Sea is subducted beneath the Peloponnese. Subduction terminates at the northern strike-slip margin of Ionian oceanic crust just north of the Ionian islands (Fig. 1a). Although extensional basins formed in the Palaeogene in the northern Aegean region, only since the Miocene has extension been widespread in the Hellenide orogen (Mascle & Martin, 1989; Pe-Piper *et al.*, 1995). This progressive extension and collision (Taymaz *et al.*, 1991) resulted in clockwise rotation of the Apulian foreland by several tens of degrees between the late Oligocene and late Miocene (Kissel & Laj, 1988) and in a further 25° clockwise rotation of the Ionian islands since the early Pliocene (Kissel & Laj, 1988).

The Ionian islands are separated from mainland Greece by rapidly subsiding extensional basins of Pliocene–Quaternary age, notably the Zakynthos Channel basin (Brooks & Ferentinos, 1984) and the outer Gulf of Patras. Extension also occurred in the Gulfs of Patras and Corinth (Ferentinos *et al.*, 1985; Doutsos *et al.*, 1988; Zelilidis *et al.*, 1988; Leeder & Jackson, 1993), where basin-margin, marine, upper Pliocene sediments are widespread (Kontopoulos & Doutsos, 1985; Kontopoulos & Zelilidis, 1992). Rapid basin subsidence probably did not begin until the middle Quaternary (Piper *et al.*, 1990). The age of the oldest sediment in these extensional basins in western Greece is early Pliocene. Present rates of subsidence locally exceed 5 mm yr^{-1} (Chronis *et al.*, 1991) and the area is highly active seismically (Ambraseys & Jackson, 1990).

The late Cenozoic basins of the Ionian islands are thus located between a subduction zone and a region of rapid extension. They developed in a complex synsedimentary tectonic environment involving a foreland-propagating fold and thrust system, with diapiric intrusion of gypsum along thrust ramps and synchronous rapid extension further inboard. In this study, we examine the Pliocene–Quaternary basin facies evolution in the terrestrial sections that are best exposed, namely on the island of Zakynthos. We then apply our findings to the general evolution of the basins of this region.

Structural framework of Zakynthos

Three types of synsedimentary tectonic activity affected the Pliocene to Quaternary sediments of Zakynthos.

1 Compressional movements along the Ionian thrust are inferred to have taken place during the Pliocene and based on evidence from Kefallinia, probably continued in the Quaternary. Compressional structures deform upper Miocene sediments (Underhill, 1988) and lower Pliocene marls adjacent to the Ionian thrust (Bizon & Bizon, 1985). The Vrachionas anticline in western Zakynthos (Fig. 1a) developed in the Pliocene as part of the same foreland-propagating fold and thrust system (Underhill, 1989). The deeper distribution of these structures has been imaged on a deep-penetration seismic-reflection profile immediately north of Zakynthos (Kamberis *et al*., 1996).

2 Faulting and subsidence are widespread in Quaternary sediments of the Zakynthos Channel basin (Brooks & Ferentinos, 1984). The present Zakynthos Channel is bounded to the west by a master fault with a throw of at least 1 km (Brooks & Ferentinos, 1984, their fig. 11). The extensional Porto Zorou Fault in south-eastern Zakynthos (Fig. 2) is located in the footwall of the western bounding fault of the Zakynthos Channel basin and is parallel to this master fault (Underhill, 1988, 1989). It may have been the dominant fault in the basin in the early Pleistocene, with the present eastern bounding fault developing later and producing the mid to late Quaternary footwall uplift in eastern Zakynthos.

3 Triassic gypsum formed diapiric intrusions throughout the Pliocene–Quaternary (Underhill, 1988), rising through the shortened sedimentary cover from buried thrust ramps. The subsequent rise of the diapirs through Pliocene–Quaternary rocks, above decollement and thrust surfaces, appears to have been largely vertical, although reactivation of some thrust fault ramps may have influenced intrusion paths (Brooks & Ferentinos, 1984). Diapir intrusion initially caused extension and later complex rotation in the overlying Pliocene to Quaternary sediments (Underhill, 1988).

Previous stratigraphic studies on Zakynthos and adjacent areas

The lower Oligocene–lower Miocene Ionian foreland flysch basin of the Ionian islands was segmented by the late Oligocene by the westward-propagating thrust and fold system. In Zakynthos, upper Miocene flysch in the west passes eastward into shelf sediment in the east (Kontopoulos *et al*., 1997). Movement on the Ionian Thrust during the early Pliocene in the eastern part of Zakynthos (Sorel, 1976; Underhill, 1989) subdivided the pre-existing Miocene basin into two independent basins in the late Pliocene: the Alikanas basin in central Zakynthos and the Geraki basin in south-eastern Zakynthos (Fig. 2). Pleistocene sedimentation is largely restricted to the margin of the Zakynthos Channel basin.

The sedimentary succession of the Alikanas basin spans the entire Pliocene and extends to the early Pleistocene at Cape Krioneri, based on biostratigraphic data provided by Bizon & Mirkou (1969), Blanc-Vernet & Keraudren (1970), Bizon & Muller (1977), Dermitzakis *et al*. (1979), Tsapralis (1981), Frydas (1986), Underhill (1988) and Triandafyllou (1993), as summarized in figure 3. In the Geraki basin, a formal lithostratigraphic scheme has been defined by Dermitzakis *et al*. (1979) and Underhill (1988) (Fig. 3) and according to Dermitzakis *et al*. (1979), the Akra Davia Formation is late Pliocene, the Gerakas Formation spans the Pliocene–Pleistocene boundary and the Kalogeras, Porto Roma and Ag. Nikolaos formations are of early Pleistocene age. Fossil assemblages indicate that marine sediments are all of shallow-water origin (Heimann, 1977; Dermitzakis *et al*., 1979; Dermitzakis & Georgiades-Dikeoulia, 1987; Mirkou, 1987).

In the Patras region to the east of Zakynthos, the oldest postorogenic sediments are of Pliocene age and are principally of shallow marine or littoral origin (Frydas, 1987, 1989; Kontopoulos & Zelilidis, 1992). Sometime in the late Pliocene, marine sedimentation terminated and most lower to middle Pleistocene sediments in the Gulf of Patras are lacustrine, with the exception of a lower Pleistocene marine transgression. A similar evolution is seen in the Gulf of Corinth (Doutsos *et al*., 1988; Doutsos & Piper, 1990; Gawthorpe *et al*., 1994), where Pliocene marine sediments are followed by Quaternary lacustrine sedimentation. The first marine transgression into the Gulf of Patras took place during the marine highstand of isotopic stage 11 at about 400 ka (Stamatopoulos *et al*., 1994).

PLIO-PLEISTOCENE OF THE ALIKANAS BASIN

Western Alikanas basin

The strata of the Alikanas basin dip about 10°NE, exposing the oldest rocks in the west. Lower Pliocene deposits of the Agios Sostis (Fig. 2) are 10 m thick (Figs 3, 4A and 5A) and comprise mostly calcareous marls with an outer shelf fauna (Trubi Limestone of Heimann, 1977) with 10% interbedded fine to medium sandstones 4–20 cm thick, with only the lowest sandstone bed being 60 cm thick. The proportion of sandstone beds decreases higher in the section (Fig. 4A). Thicker sandstone beds have erosional bases, generally planar stratification in the lower part of the bed and low-angle

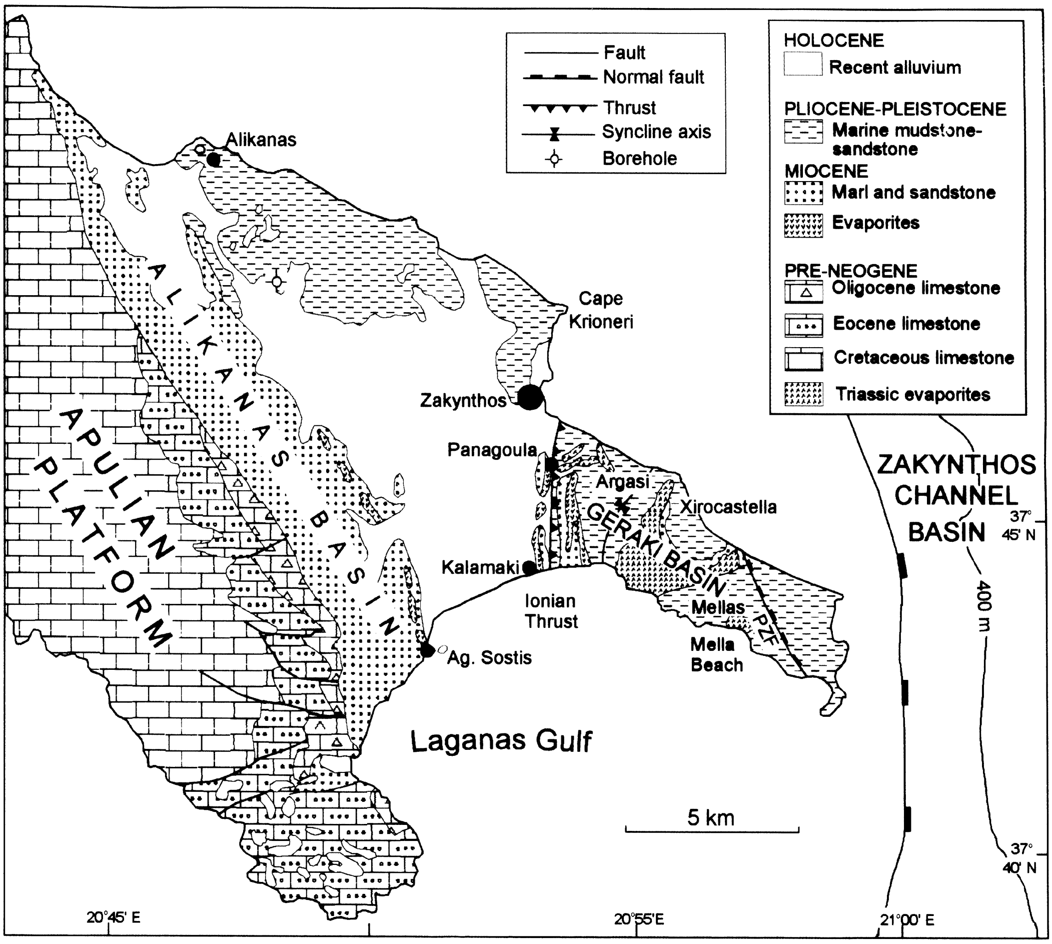

Fig. 2. Geological map of south-eastern Zakynthos island (modified from Underhill, 1988; Dermitzakis *et al.*, 1979). PZF = Porto Zorou Fault.

cross-bedding and rare ripples near the top of the bed. A few beds show trough cross-bedding. Upper and lower bedding planes appear highly bioturbated. The calcareous marls have internal erosional surfaces every 10 cm, with 1–5 cm of sand, marl clasts and rare rounded extrabasinal limestone pebbles (< 8 cm).

The exposed section at Alikanas (Figs 3 and 5B) of upper Pliocene comprises fossiliferous mudstones, overlain by alternating bioturbated silty sandstone and mudstone beds and at least one unconformity surface (Fig. 5B). Few primary sedimentary structures are preserved. Lower Pliocene sediments are known from the borehole located on Fig. 2 (Bizon & Mirkou, 1969).

Zakynthos town – Cape Krioneri section

Strata in this section are subhorizontal and gradually young northward, forming a 300-m-thick coarsening-upward sequence (Fig. 3). Three main facies are present. In the lower part of the section, marly claystone predominates (Fig. 3). The claystone contains abundant foraminifera and a sparse shelly fauna. The facies is weakly bioturbated, with *Chondrites* the most prominent trace fossil. Interbedded with the claystone are laminae of silt and < 5 cm beds of poorly sorted very fine sand (Fig. 5C). Higher in the stratigraphic section, sand beds reach 30 cm thickness (Fig. 5D). Sand beds commonly have an erosional base, rarely with a shell lag at the base or filling underlying burrows. Most beds appear slightly graded, passing gradually up into claystone. Most have horizontal lamination picked out by sandier and muddier laminae, some have low-angle cross-stratification and a few have rippled tops, with palaeocurrents indicating flows to the east. Rarely, synsedimentary convolute lamination is developed. In a few cases, woody detritus occurs in the sands.

In places in the middle part of the section, contorted and steeply dipping marly claystones are intercalated within the main claystone succession (Fig. 3). These slumped blocks of claystone are 5–20 m thick and commonly contain disrupted sandstone beds up to 50 cm thick. Overlying these slump blocks, the claystone contains up to 40% fine sand beds 10–30 cm thick, showing shelly lags, horizontal laminae, internal grading and sparse bioturbation similar to that in the thinner sands described above.

In the upper part of the section (Fig. 3) near Cape Krioneri, sand beds 5–20 m thick alternate with intervals

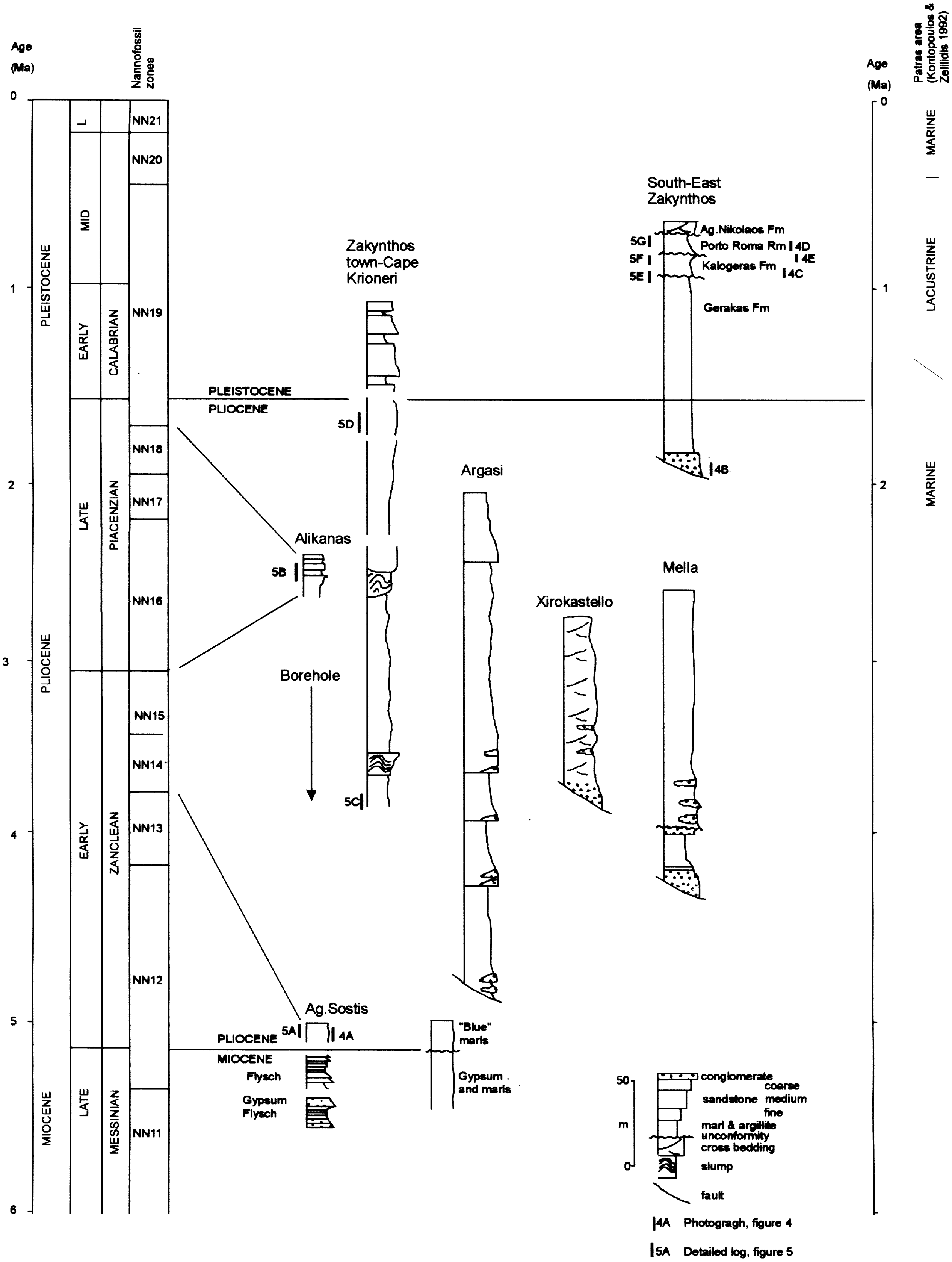

Fig. 3. Lithostratigraphy of the principal sections investigated in this study and their biostratigraphic correlation (for sources, see text). Stratigraphic distribution of calcareous nannofossil biozones after Frydas (1986) and Triandafyllou (1993).

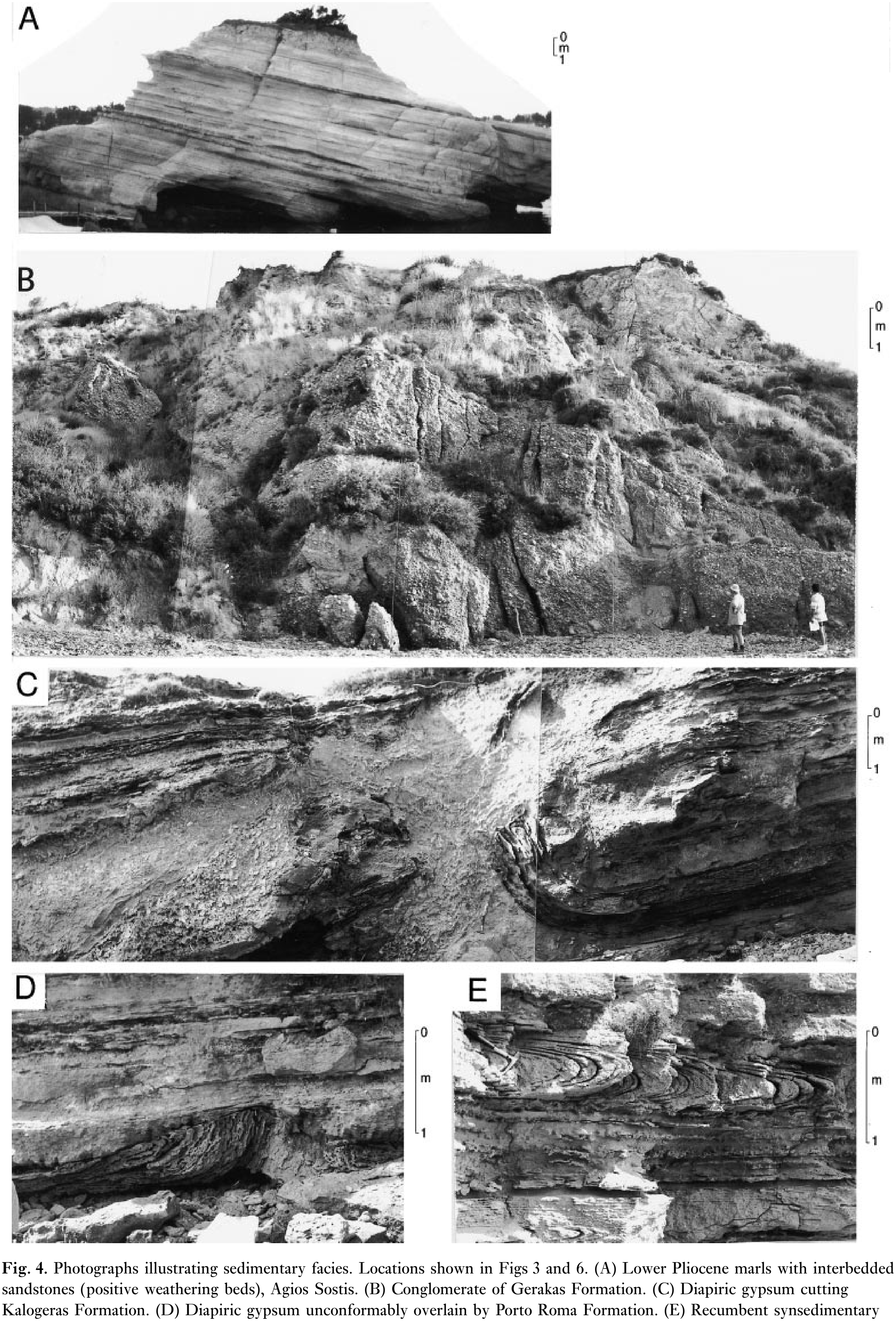

Fig. 4. Photographs illustrating sedimentary facies. Locations shown in Figs 3 and 6. (A) Lower Pliocene marls with interbedded sandstones (positive weathering beds), Agios Sostis. (B) Conglomerate of Gerakas Formation. (C) Diapiric gypsum cutting Kalogeras Formation. (D) Diapiric gypsum unconformably overlain by Porto Roma Formation. (E) Recumbent synsedimentary fold in Kalogeras Formation.

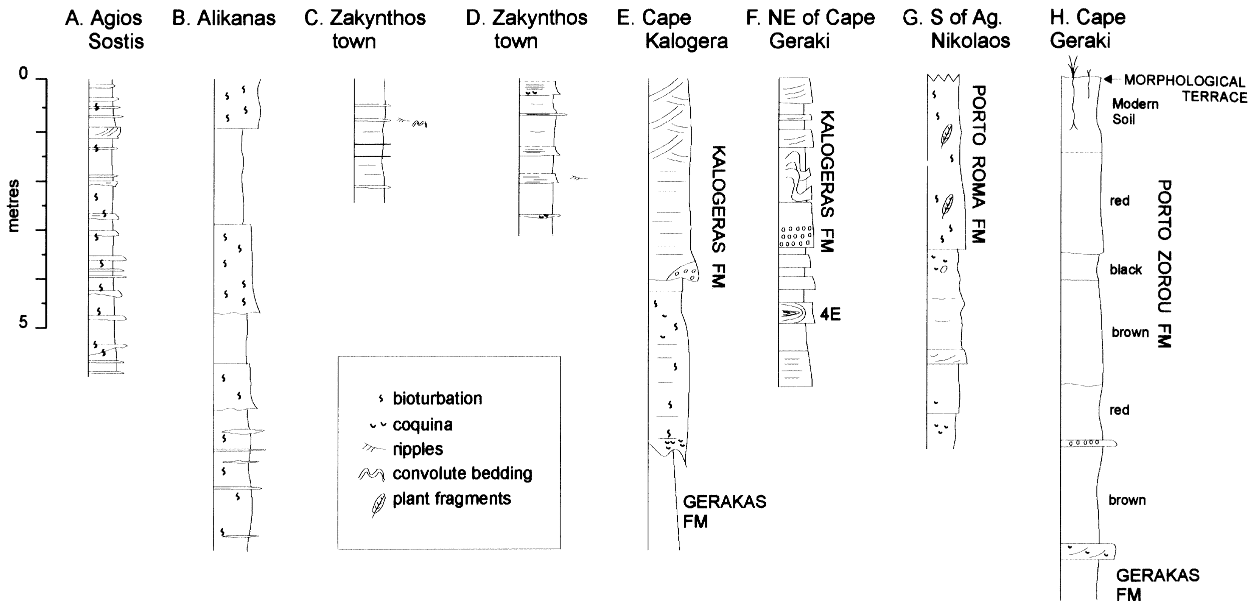

Fig. 5. Sedimentological logs of selected facies. Locations shown in Figs 3 and 6. (A) Part of lower Pliocene section at Agios Sostis. (B) Upper Pliocene section at Alikanas. (C) Part of upper Pliocene section at Zakynthos town. (D) Part of lower Pleistocene section at Zakynthos town. (E) Lower Pleistocene lower Kalogeras Fm at Cape Kalogera. (F) Lower Pleistocene upper Kalogeras Formation north-east of Cape Geraki. (G) Porto Roma Formation south of Cape Agios Nikolaos. (H) Type section of Porto Zorou Fm at Cape Geraki.

of marly claystone a few metres thick. The sand beds are of fine to medium sand and contain many shell fragments. They are intensely bioturbated, with cm-diameter burrows picked out by irregular cementation. Locally, lenticular bedding can be discerned. The base of the sand beds is sharp and planar. The marly claystone contains thin sand beds similar to those lower in the section.

Interpretation of the Alikanas basin

Terrigenous clastic sediment is most abundant in the eastern part of the Alikanas basin. In the lower Pliocene sections at Agios Sostis in the south-west and in the boreholes at Alikanas marls are most abundant, whereas in the upper Pliocene sections at Alikanas and Zakynthos town there is more sand. Precise biostratigraphic correlation (Triandafyllou, 1993) shows that the upper Pliocene section at Zakynthos town is also substantially thicker than that at Alikanas (Fig. 3). Fossil assemblages and sedimentary structures suggest that all the sediments accumulated in shelfal water depths.

At Agios Sostis, the lower Pliocene outer shelf marls are interbedded with sandstone beds that show many of the sedimentary structures associated with 'tempestites' (Walker & Plint, 1992) and the numerous erosion surfaces with rip-up clasts are probably also related to storms. The extrabasinal limestone pebbles provide evidence for uplift of the Vrachionas anticline, as part of the propagating fold and thrust system (Underhill, 1989) most clearly expressed in the Ionian thrust. This represents a major palaeogeographical change from the basin slope setting of the late Miocene (Kontopoulos *et al.*, 1997). Uplift of limestones in western Zakynthos is probably responsible for the westward thinning of the upper Pliocene clastic sequence from Zakynthos town to Alikanas.

The overall coarsening-up sequence in the Zakynthos section also shows an upward increasing amount of erosion at the base of beds, sand-bed thickness and the number of sandy beds with woody detritus. The abundance of woody detritus in the higher strata suggests the progradation of deltaic sediments across a clastic shelf and the sequence of marly claystone overlain by slumped claystone blocks may also be a consequence of delta progradation. The scale of the slump features, the sparsity of bioturbation and the lack of conglomerate (usually an indicator of local sources in a tectonically active environment) all suggest a large river as a source. The poorly bioturbated marly claystones and silt laminae were largely deposited in a prodelta environment from the discharge plume. The sand beds, like those of Agios Sostis, show many features of tempestites, including reworking of shell fragments and graded bedding. The trend towards thicker sand beds higher in the section probably represents greater proximity to sandy littoral facies. It is probable that the Zakynthos town section represents the episodic progradation of the proto-Acheloos river delta. Easterly palaeocurrents are consistent with modern climatic conditions, under which south-westerly storms predominate (Piper *et al.*, 1982a). The lower part of the Alikanas section appears transitional between the marly claystones and storm sandstones at Zakynthos town and the more bioturbated section at Agios Sostis. The upper part of the section shows abrupt changes from claystone-dominated intervals to silty sandstone-dominated intervals, similar to the section at Cape Krioneri.

There is no evidence that the subaerial delta ever prograded across eastern Zakynthos. The thick sand beds near Cape Krioneri, with lenticular bedding, shelly faunas and intense bioturbation, are typical of Neogene shelf sands found elsewhere in Greece (e.g. Piper *et al.*, 1976). They are probably analogous to delta-derived sands currently accumulating immediately south of the modern Acheloos delta in water depths of 20–40 m (Piper & Panagos, 1981; Piper *et al.*, 1987), transported by storm waves and tidal currents. The abrupt changes from claystone to thick sands probably reflect eustatic changes in sea level in this lower Pleistocene section.

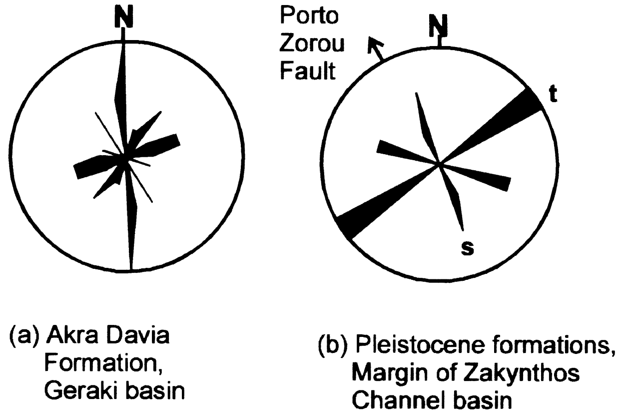

Fig. 7. Rose diagrams showing the orientation of synsedimentary faults for (a) the upper Pliocene Akra Davia Formation and (b) the Pleistocene formations of Geraki basin (Gerakas, Kalogeras, Porto Roma, Ag. Nikolaos Formations). s = synthetic normal faults; t = transfer faults.

UPPER PLIOCENE OF THE GERAKI BASIN

Description

The upper Pliocene Akra Davia Formation accumulated in three sub-basins (Mellas, Xirokastello and Argasi: Fig. 6), whose configurations were controlled by synsedimentary faulting and diapiric intrusions. The Formation unconformably overlies Miocene and lowermost Pliocene sediments (Bizon & Bizon, 1985; Underhill, 1989) and is locally juxtaposed against Triassic evaporites along extensional basin-bounding faults. In places, the lower part of the Formation has been strongly deformed by diapirs that appear related to N–S striking synsedimentary extension faults (Fig. 7a).

The Mellas subbasin has diapiric intrusions at its north-west margin and sediment thickness increases southward. The Akra Davia Formation is <150 m thick and comprises a coarsening-upward sequence (Fig. 3). At Mellas Beach, it shows a lateral contact with Triassic evaporites with several NW-striking extensional faults. In the hangingwall of these faults, poorly sorted limestone-clast conglomerate beds <5 m thick accumulated locally and pass upwards and laterally into reddish massive sandstone and olive-grey mudstone up to 10 m thick containing rare burrows, shell and plant fragments.

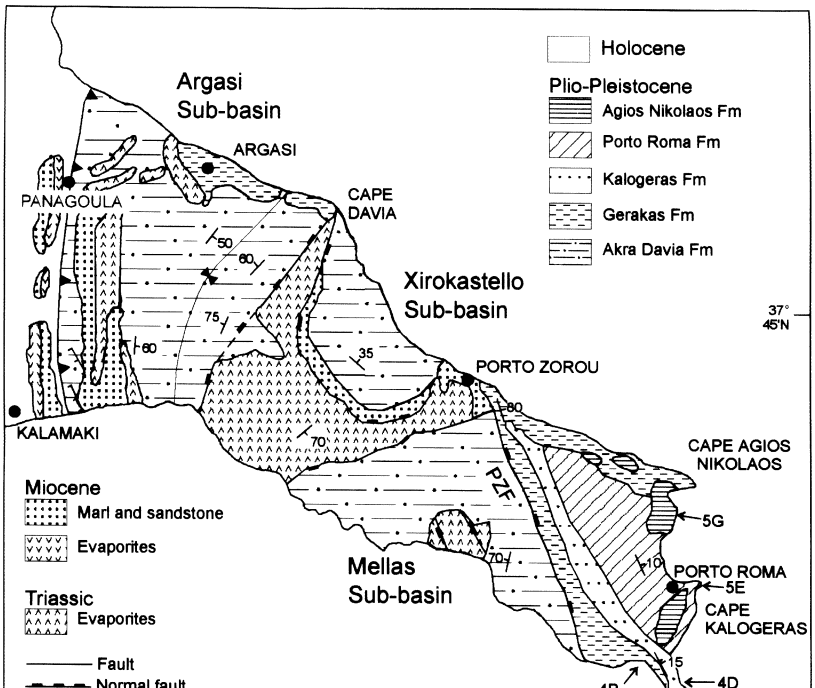

Fig. 6. Detailed map of the Geraki basin, showing relationship of diapirs to the bounding faults of sub-basins. PZF = Porto Zorou Fault. Numbers indicate locations of Figs 4 and 5.

These lower mudstones are overlain by < 10 m of brownish-grey massive mudstone, locally with shell fragments, with interbedded thin conglomerate beds with pebbles of gypsum (< 10 cm) and rare arenite. These observations show clearly that the diapirs were a sediment source to the lower Akra Davia Formation. The upper part of the formation forms asymmetric wedges, 1–1.5 km long and 80–120 m thick, that thin away from ENE-striking, synsedimentary faults (Fig. 7a), again providing evidence for synsedimentary diapiric activity. At the base are interbedded conglomerate and sandstone beds, forming a unit < 30 m thick that passes upwards into medium- to thick-bedded yellowish-orange sandstone.

The Xirokastello sub-basin has diapiric intrusions at its southern margin and Pliocene strata dip from 20° to 50°. Sediment thickness increases to the north-east. The oldest sediment is a mudstone unit, which appears to be overlain unconformably by a unit, < 100 m thick, of interbedded sandstone and mudstone beds. The yellowish-orange sandstones form trough cross-stratified beds (sets 1.5 m long and 50 cm thick). The greenish-grey mudstone occurs as lenses (2 m long and 80 cm thick) with common vertical (< 7 cm) and rare horizontal burrows. They contain common *Ammonia*, and rare *Elphidium* and ostracods, suggesting deposition in brackish water. Locally, the sandstone beds are massive and conglomerates are found near the basin margins in the hangingwall of bounding NNE-striking extensional faults.

The Argasi sub-basin has diapiric intrusions at its south-eastern margin and the sediment thickness increases to the north-west. The sub-basin sediments have been deformed into a NNE-trending syncline (Fig. 6). The main outcrop in the north-eastern part of the sub-basin contains a composite stratigraphic thickness of 350 m, located in a section about 500 m long in the hangingwall of a NNE-striking fault that separates Pliocene sediments from Triassic evaporites. Five fining-up cycles are recognized in this section (Section 4, Fig. 3). The base of each cycle consists of massive, yellowish orange sandstone beds, in places interbedded with minor conglomerate beds (generally lenticular, 1–4 m long, < 2 m thick, with 5 cm clasts). Higher in each cycle, sandstones interbed with and pass up into massive grey mudstone, with rare shelly fossils, burrows and wood fragments. In the western part of the sub-basin (near Panagoula and Kalamaki), the Akra Davia Formation rests unconformably on Miocene sediments. The basal sediments are lacustrine limestone and well-cemented blue-grey thinly bedded sandstone beds, overlain by < 15 m of mudstones with vertical burrows (< 10 cm) and rare body fossils (*Cardium*, *Cladocera cespitosa, Union*) that pass upwards into a unit (< 35 m thick) of thin-to thick-bedded sandstones. Basinwards, the upper part of the Akra Davia Formation consists of interbedded thick to very thick laminated sandstone beds (20–60 cm) with thinly laminated mudstone intercalations (5–10 cm) containing rare shelly fossils and burrows. This sandstone unit thickens towards the boundary faults from 30 m to 100 m.

Interpretation of the Akra Davia Formation

All three sub-basins show similar stratigraphic development. The basal part of the Akra Davia Formation consists of mudstone that rests unconformably on Miocene – lower Pliocene sediment. The assemblage of vertical burrows and body fossils suggests accumulation in a lagoonal environment, probably during the earliest phase of a marine transgression. The matrix-supported conglomerate units close to bounding faults represent small alluvial fans, passing rapidly into fossiliferous marginal marine sediments of the type described in detail by Piper *et al.* (1976). The overlying sandstone unit shows thickness variations related to bounding faults, suggesting that segmentation of the three sub-basins may have taken place during deposition of the Akra Davia Formation. These faults also controlled the locus of diapiric intrusions of Triassic evaporites that deformed the lower part of the formation much more than the upper part. The fining-upward cycles appear to represent fluvial deltaic sediment, with minor marginal marine intercalations (lagoonal mudstone). The thickness of the cycles (Fig. 3) suggests rapid subsidence. The occurrence of sandstone, limestone and gypsum pebbles in many of the sandstone sequences contrasts with the upper Pliocene sandstones of the Alikanas basin at Zakynthos town and suggests that a local source of sediment is likely.

UPPER PLIOCENE – LOWER PLEISTOCENE ON THE MARGIN OF THE ZAKYNTHOS CHANNEL BASIN

Description

Thick uppermost Pliocene and lower Pleistocene sediments unconformably overlie the Akra Davia Formation and outcrop only north-east of the extensional Porto Zorou Fault, in its hangingwall (Fig. 6). They thus appear to be marginal sediments of the Zakynthos Channel basin and correlate with upper Pliocene – lower Pleistocene sediments of the Gulf of Patras (Fig. 1a) that developed at the eastern margin of the Zakynthos Channel basin. Except for the lower Gerakas Formation, all appear to be marine sediment.

The base of the Gerakas Formation, near the hangingwall of the Porto Zorou Fault, consists of fining-upward massive conglomerates, up to 10 m thick (Fig. 4B). Conglomerates are clast supported with trough cross-stratification (2 m long, 1 m thick). Conglomerates at the base of the formation include < 1% boulders and ≃ 10% cobbles and pass upward into fine pebble (2–4 cm) conglomerate. Silty sand lenses (2 m long, 30 cm thick) are interbedded with the conglomerate, particularly in the upper part of the formation. Channels with NNE palaeocurrent direction are parallel to the Porto Zorou Fault. The conglomerates are overlain by, and pass laterally into, a unit of yellowish-white silty sand to sandy silt beds, up to 12 m thick, with rare (<1%) small conglomeratic lenses. The upper part of the Gerakas Formation consists of >15 m of massive blue-grey silty mudstone with fully marine fossils (Dermitzakis *et al.*, 1979; Triandafyllou, 1993) and bioturbated remnants of silt and sand beds.

The Kalogeras Formation, unconformably overlying the Gerakas Formation (Fig. 5E), is up to 10 m thick and consists of marine yellowish-orange medium sandstone and olive-grey massive mudstone beds, with *Chlamys*, *Ostrea* and *Cardium* present throughout the formation. One 50-cm-thick sandstone bed has synsedimentary recumbent folds (Fig. 4E). A contorted bed of interbedded sandstone and mudstone in the middle of the formation (Fig. 5F) increases in thickness towards the synsedimentary faults. The upper part of the formation, up to 1.5 m thick, consists of interbedded mudstone and medium sandstone beds with low-angle cross-stratification (Fig. 5F). The thickness of this formation was influenced by diapirism (Fig. 4C,D) and associated synsedimentary extensional faulting, which formed depressions 100–400 m long and <10 m deep. These extensional faults are either synthetic normal or transfer faults with respect to the bounding master extensional fault (Fig. 7b). Some synsedimentary diapirs that crop out in the south-eastern part of the Gerakas peninsula do not cut the overlying Porto Roma Formation. The unusual abundance of synsedimentary deformation compared with other formations may indicate high seismic activity associated with the emplacement of diapirs.

The Porto Roma Formation (Figs 3 and 5G), up to 12 m thick, consists of marine sandstone and siltstone deposited unconformably over the Kalogeras Formation. Outcrops are bounded to the south by a WNW-striking transfer fault. The formation appears to be broadly regressive, from outer shoreface to lagoonal. In the lower 3 m, bioturbated sandy silt with marine fossils interbeds with a thin coarse sandstone bed with low-relief erosional surfaces. Erosional channels trend NNW. The upper 8 m consists of bioturbated sandy silt with rare shelly fossils, common plant fragments and some root traces.

The Agios Nikolaos Formation rests unconformably on the Gerakas Formation mudstones and consists of sandstones with abraded *Ostrea*. It is generally about 6 m thick, but thickens to <20 m near NE-striking synsedimentary faults. It consists of strongly cemented horizontally bedded sandstones, locally with trough cross-bedding (sets <1.5 m, typically 0.8 m). Synsedimentary faulting created local unconformities (Fig. 3). Palaeocurrents determined from cross-bedding are bimodal N–S, but predominantly to the south, with small channels trending NNE–SSW.

Interpretation

Four facies associations are present in the Gerakas to Agios Nikolaos formations. Both fossil assemblages and sediments in the upper Gerakas Formation and overlying formations indicate deposition in shallow marine conditions, under varying water depths.

1 The conglomeratic lower Gerakas Formation lacks fossils. The presence of boulders at the base of the formation suggests a local source of coarse sediment and its distribution suggests that it developed along a fault-line scarp along the Porto Zorou Fault. The poorly sorted, clast-supported conglomerates with trough cross-stratification suggest deposition in the low-gradient part of an alluvial fan or on a subaerial fan-delta.

2 Thick mudstone sequences with fossils such as *Chlamys* and abundant planktonic foraminifers (reported by Dermitzakis & Georgiades-Dikeoulia, 1987) indicate fully marine conditions, rather than deposition in coastal lagoons. The mudstones appear thoroughly bioturbated and only thicker interbedded sands are preserved. Comparison with modern marine sediment deposition (e.g. Piper *et al.*, 1987) suggests deposition in water depths of several tens of metres.

3 Well-sorted cross-stratified sandstones with rare pebbles. These sandstones also contain fully marine faunas. They are commonly bimodally trough cross-stratified, particularly in the Agios Nikolaos Formation, suggesting the influence of tidal currents. In the Kalogeras Formation, interbedded mudstones are common. The dominant palaeocurrents towards the south are consistent with Coriolis forcing within Zakynthos Channel, which in this area is constricted by Cape Killini (Fig. 1a). The abundance of tidally influenced sandstones and the presence of fine pebble conglomerate lags indicates that tidal currents were stronger than at the present, suggesting a more constricted entrance to the Gulf of Patras than at present and probably shallower water in Zakynthos Channel. Sedimentation conditions may have been analogous to those in the modern Rion Straits at the eastern end of the Gulf of Patras (Piper *et al.*, 1987), in water depths of up to 50 m. Shallow erosional channels associated with these cross-bedded sandstones probably result from tidal scour.

4 Locally, in the Porto Roma Formation, lagoonal sediments are recognized on the basis of fossils and sediment facies criteria described by Piper *et al.* (1976).

The transitions from thick shelf mudstones to thick tidal sandstone sequences are abrupt and changes from thick shelf mudstones to lagoonal sequences imply tens of metres of relative sea-level change. This is an order of magnitude greater than subsidence associated with individual earthquakes. It therefore seems more probable that eustatic sea-level changes are the principal influence on the type of sediment facies developed, although these sea-level changes were superimposed on the effects of subsidence along the Porto Zorou Fault.

UPPER QUATERNARY MARINE TERRACES

In the early Pleistocene, progressive uplift of the eastern part of Zakynthos island resulted in the termination of marine sedimentation at Cape Krioneri and in the Agios

Nikolaos Formation. The upper Pleistocene Porto Zorou Formation (defined here) accumulated in a narrow zone of the hangingwall of the Porto Zorou normal fault (Fig. 6), unconformably overlying the Gerakas and Kalogeras formations, and is now about 20 m above sea level. At the type section at Cape Geraki (Fig. 5H), the base of the formation is a <1-m-thick sandstone with low-angle cross-stratification containing *Cardium* and *Pecten*. Reworked rod-shaped calcareous concretions are orientated NW–SE, parallel to the inferred coastline near the Porto Zorou Fault. The marine sandstone is overlain by 7 m of terrigenous sediment (red, brown and locally black mudstone and thin conglomerates) that wedges out rapidly eastward. At least two distinct morphological terraces are cut in the upper part of the formation. At Porto Zorou, the terrestrial mudstones are represented by a 3-m bed rich in plant debris that may represent a coastal marsh. Middle to late Palaeolithic tools from the Porto Zorou terrace suggest a Tyrrhenian age (isotopic stage 5) for the terrace (Kourtessi-Filippakis, 1996). Another prominent marine terrace, dipping NNE, unconformably overlies Pliocene–Pleistocene sediments from Zakynthos town to Cape Krioneri. In places, the terrace is represented only by reddening of underlying strata; elsewhere, thin sands and marls including *Cladocera* overlie the terrace.

DETRITAL MINERALOGY

Sediment provenance is inferred from the petrology of conglomerate clasts, determined in the field, and from the clay mineral composition of representative samples of upper Miocene to Quaternary sediments (Fig. 8). The basal conglomerates of the Akra Davia Formation are quite variable. In the Mellas sub-basin, they consist principally of limestones and minor chert, with stratigraphically higher conglomerates comprising gypsum and rare sandstone clasts. In the Argasi sub-basin, clasts comprise 90% sandstone and 10% limestone. Our observations concur with previous workers (Sorel, 1976; Underhill, 1988) who suggested that these clasts were derived from rocks immediately above the diapirs. The Akra Davia Formation has a clay mineral assemblage (Fig. 8c) with considerable amounts of smectite and kaolinite + chlorite that is similar in composition to that of upper Miocene sediments at Agios Sostis (Fig. 8d), consistent with evidence for local derivation of sediment from diapirically uplifted Miocene rocks.

The Plio-Pleistocene sediments at Zakynthos town lack conglomerates. Their clay mineral assemblage is almost devoid of smectite and resembles the high-illite assemblage of the Holocene Acheloos River (Fig. 8b). The Gerakas to Agios Nikolaos formations have somewhat higher smectite, with a clay mineral assemblage intermediate between that of the Miocene and the Zakynthos town section (Fig. 8a). This is consistent with the facies evidence for local derivation of sediment and the location of these formations on the edge of the Zakynthos Channel basin that was fed by the Acheloos River (Brooks & Ferentinos, 1984). Conglomerates at the base of the Gerakas Formation comprise 50% limestones, 40% chert and 10% sandstones. The composition of these clasts suggests that many are reworked from Akra Davia Formation conglomerates. High smectite content of the Porto Zorou Formation (Fig. 8a) probably reflects soil diagenesis under dry alkaline conditions during Pleistocene sea-level lowstand (cf. Piper *et al.*, 1982b).

DISCUSSION

The tectonic environment on Zakynthos

Movement on the Ionian Thrust during the early Pliocene separated the earlier Miocene basin on Zakynthos island

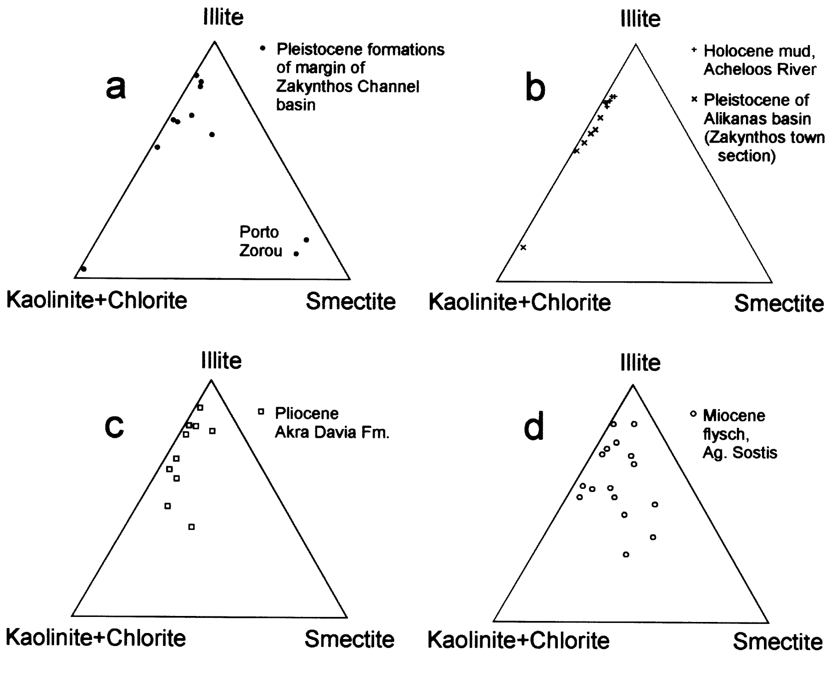

Fig. 8. Ternary diagrams showing variations in clay mineral abundance in the Miocene–Quaternary sediments of south-eastern Zakynthos. Clay minerals in the <2-mm fraction normalized to 100% using the weighting factors of Biscaye (1965).

into the Alikanas and Geraki basins (Fig. 9). During the Pliocene, extensional basins developed east of Zakynthos, including the Zakynthos Channel basin and the Gulfs of Patras and Corinth. Late Pliocene to Quaternary sedimentation in south-eastern Zakynthos is dominated by movement on the faults marking the western margin of the Zakynthos Channel basin.

In the Alikanas basin, lower Pliocene sediments were deposited over upper Miocene flysch with a local unconformity that in part reflects the post-Messinian transgression (Kontopoulos *et al.*, 1997). Limestone pebbles in these sediments provide evidence that uplift of the Vrachionas anticline had begun. This uplift was the result of a series of linked thrusts and anticlines propagating through the Apulian foreland (Underhill, 1989). The progressive uplift of the Vrachionas anticline is reflected in the youngest sediments in the Alikanas basin being found in the north-east part of the basin (Cape Krioneri) (Fig. 10). The propagation of the fold and thrust system to Zakynthos appears to have been later than further north: there is no evidence for the Vrachionas anticline in the upper Miocene of Zakynthos (Kontopoulos *et al.*, 1997), whereas in Levkas, there is an important mid-Miocene unconformity (Cushing, 1985; Clews, 1989). The first movement on the Ionian thrust in Zakynthos was in the earliest Pliocene, whereas to the north, the Kalamitsi thrust (marking the western margin of the Ionian zone) was active in the late Oligocene (Wilpshaar, 1995). This diachronous deformation probably results from the presence of the Hellenic Trench west of Zakynthos and the lack of a continental block west of the Pre-Apulian zone.

Diapirism of Triassic evaporites behind the Ionian Thrust was promoted by loading by the advancing thrust sheet and developed at thrust ramps. Precise diapir locations appear to have been controlled by basin-bounding extensional faults (Fig. 6). The lower strata in the Akra Davia Formation are highly deformed compared with the upper part of the formation, and overall the sediments show a coarsening-up sequence, suggesting progressive uplift of the source area. It is unclear to what extent unconformities within the Gerakas and younger formations are a consequence of continuing diapirism, or of hanging-wall uplift on the Porto Zorou Fault, one of the faults at the western margin of the Zakynthos Channel basin.

Evidence for the subsidence of the Zakynthos Channel basin is first seen in the latest Pliocene, with the development of alluvial fan conglomerates along the Porto Zorou Fault (Figs 9b and 10). The Porto Zorou Fault may have

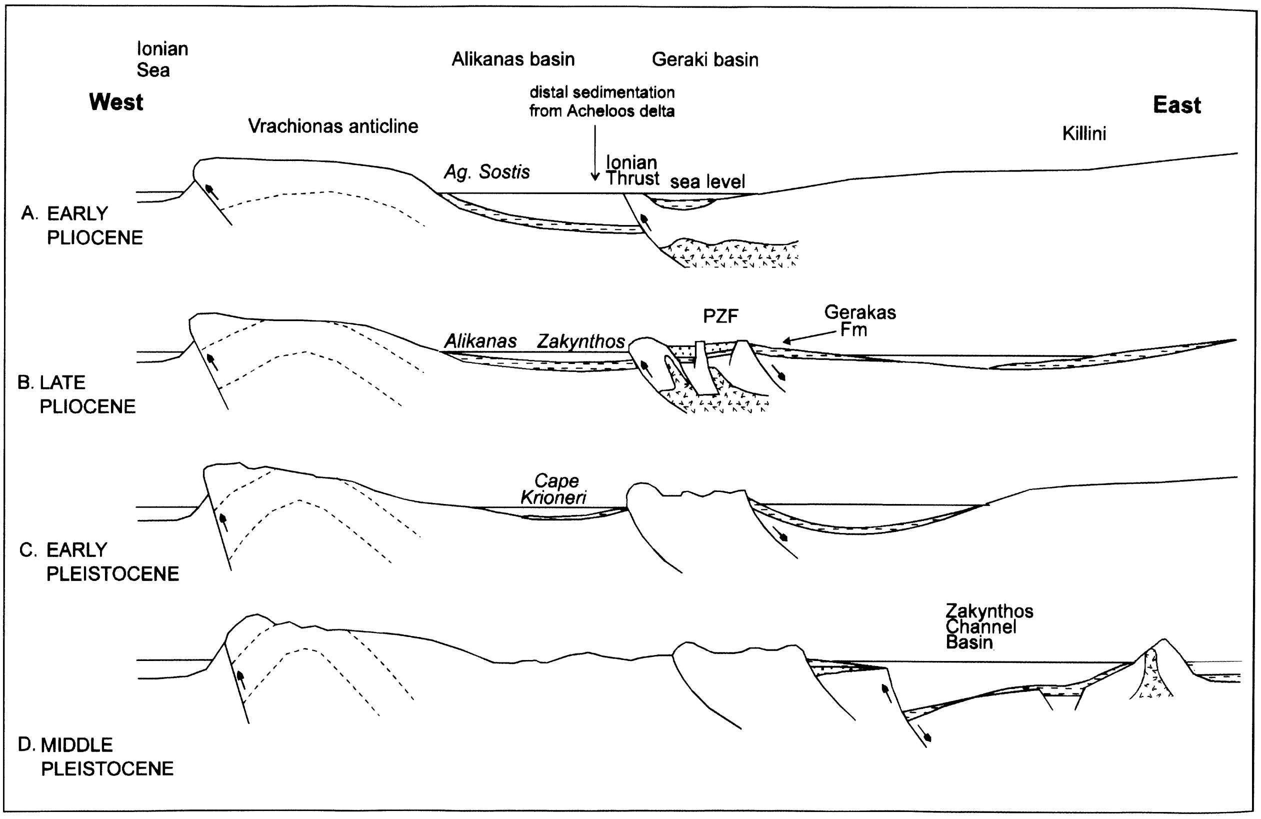

Fig. 9. Palaeogeographical cross-sections showing the Pliocene–Quaternary sedimentation in south-eastern Zakynthos and the role of surface faults. (A) Early Pliocene: segmentation into the Alikanas and Geraki basins and restriction of the Alikanas basin to the west by the rising Vrachionas anticline. (B) Late Pliocene: formation of local paralic sub-basins in the Geraki basin as Triassic diapirs continue to rise. Beginning of subsidence on the Porto Zorou Fault (PZF) and deposition of the Gerakas Formation. (C) Continued uplift of western Zakynthos and subsidence of the Zakynthos Channel basin. (D) Sedimentation restricted to the rapidly subsiding Zakynthos Channel basin.

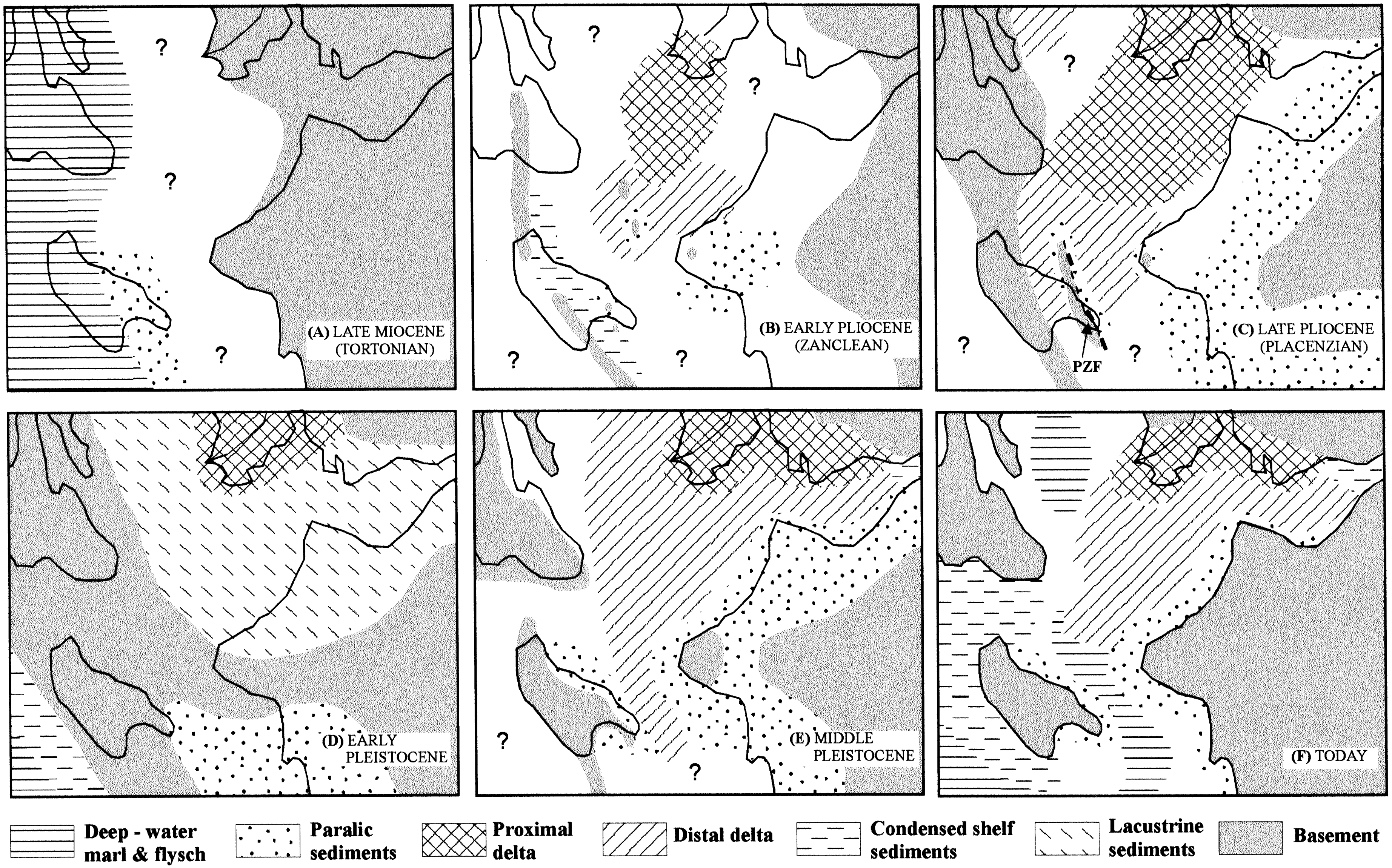

Fig. 10. Maps showing palaeogeographical evolution of the Ionian islands, based on our studies, Brooks & Ferentinos (1984) and Underhill (1989). (A) Widespread sedimentation in late Miocene (Tortonian); (B) Early Pliocene: segmentation into the Alikanas and Geraki basins; (C) Late Pliocene: formation of local paralic sub-basins in the Geraki basin as Triassic diapirs continue to rise. Beginning of subsidence on the Porto Zorou Fault (PZF) and deposition of the Gerakas Formation; (D) Early Pleistocene: continued uplift of western Zakynthos and subsidence of the Zakynthos Channel basin; (E) widespread sedimentation in middle Pleistocene; (F) present sedimentation.

marked the basin margin, but the thin sequence of lower Pleistocene sediment and numerous unconformities suggest that overall subsidence was slow. Many of the sediments show the influence of tidal currents, which were strong in the constriction between Cape Killini and Zakynthos, despite the low tidal range. Abrupt changes in inferred palaeo-water depth suggest that sediment accumulation was influenced by eustatic changes in sea level. Spectacular examples of synsedimentary tectonism, however, suggest periodic movement on secondary faults, analogous to changes of sea level of >1 m recorded after major earthquakes in the Gulf of Corinth (Vita-Finzi & King, 1985).

Sediments of the Gerakas Formation found in the hangingwall of the Porto Zorou Fault are of similar age to the upper Pliocene sediments in the Patras region (Kontopoulos & Zelilidis, 1992) (Fig. 1a), suggesting that latest Pliocene extension may have been regional. Seismic-reflection profiles show that the master bounding fault is now several kilometres to the east of the Porto Zorou Fault (Fig. 1b) and has a throw of at least 1 km (Brooks & Ferentinos, 1984, their fig. 10). The kinematic character of the Porto Zorou Fault and the pronounced fault-related subsidence in the Zakynthos Channel and Gulf of Patras argue for an extensional origin of Quaternary faulting in this area, rather than compression as proposed by Brooks & Ferentinos (1984). The age of the new master fault is unknown, but subsidence probably began in the middle Pleistocene (Piper *et al.*, 1990), eventually permitting marine transgression into the Gulf of Patras during isotopic stage 11 at about 400 ka (Stamatopoulos *et al.*, 1994). The footwall of this new master fault was uplifted and backtilted, terminating deposition of the marine sediments of the Gerakas to Agios Nikolaos formations (Figs 9 and 10). The only younger marine sediments are deposits of extreme marine highstands in the later Quaternary, that probably correlate with isotopic stages 5 and 7 (Stamatopoulos *et al.*, 1988; Piper *et al.*, 1990). If a similar age is inferred for Zakynthos, uplift rates are of the order of 0.1–0.3 mm yr^{-1}, an order of magnitude less than subsidence rates in the adjacent marine basins (Ferentinos *et al.*, 1985; Chronis *et al.*, 1991). A similar evolution is seen in the Gulf of Corinth (Doutsos *et al.*, 1988; Doutsos & Piper, 1990; Gawthorpe *et al.*, 1994), where a broad Pliocene graben was segmented by a middle Quaternary master fault on the south side of the Gulf, which resulted in rapid subsidence of the Gulf and footwall uplift on its south side, with the deposition of thin later Quaternary marine terraces.

Regional evolution of Pliocene–Quaternary basins

The evolution of Pliocene–Quaternary basins in the general region of the Ionian islands and the Gulf of Patras is a consequence of their position at the northern termination of the Hellenic trench. Oligocene–Miocene compression resulted in thrust and fold propagation in the Ionian–pre Apulian foreland in north-western Greece where the pre-Apulian zone was continuous with the Apulian plate of the Adriatic Sea and Italy. In contrast, from Zakynthos southward, the Ionian–pre Apulian continental basement was bounded south-westward by oceanic crust of the Ionian Sea and by the weak subduction boundary of the Hellenic Trench (Underhill, 1989).

The gulfs of Patras and Corinth mark an old crustal lineament that affected both Mesozoic facies (Robertson *et al.*, 1991) and Cenozoic flysch dispersal (Gonzalez-Bonorino, 1996). This lineament also marks the boundary between mid–late Cenozoic compression of the external Hellenides of northern Greece and late Cenozoic extension and concomitant uplift in the Peloponnese. As a result of the younger uplift in the Peloponnese, the river system is immature (Seger & Alexander, 1993). In contrast, rivers draining the northern external Hellenides, particularly the Acheloos and Arachthos, are large mature rivers that have built large deltas (Piper *et al.*, 1987). The late Cenozoic Acheloos delta lies at the boundary of the compressional area of north Greece and the Pliocene–Quaternary extensional zone of the Peloponnese. It has been progressively constricted eastward by the propagating thrust and fold system, but at the same time Pliocene–Quaternary subsidence in extensional basins has provided a sink for Acheloos deltaic sediment.

Progressive uplift on the thrust and fold system of the Ionian islands through the Pliocene culminated in the complete isolation of the Gulf of Patras from the open sea in the early Pleistocene.

This complex tectonic setting has nevertheless resulted in a predictable series of facies in the basins around Zakynthos.

1 The development of broad anticlines associated with thrusting led to deposition on their flanks of storm-dominated sediments with hiatuses and a limited supply of limestone clasts (e.g. Agios Sostis section).

2 More rapid uplift associated with gypsum diapirism resulted in small restricted basins, alluvial fans, cannibalism of Miocene clastic sediment to give thick fluvial and littoral sands with rapid lateral facies changes (e.g. upper Pliocene Geraki sub-basins). These sediments show overall cyclicity that may be climatically induced or may result from episodic tectonism.

3 The larger basins between fold zones were filled by extrabasinal sediment transported by the large proto-Acheloos river (e.g. Zakynthos town section). Progressive uplift of these basins and subsidence further inboard resulted in gradual eastward diversion of the Acheloos river sediment, so that the rapidly subsiding Quaternary basins were then filled by Acheloos sediment. More distal basins received fine-grained plume sediments, whereas bedload sediments accumulated in proximal basins, particularly at low stands of sea level.

4 The margins of these subsiding Quaternary basins were supplied at times of highstands of sea-level by distal deltaic muds and at lowstands of sea-level by locally derived coarse clastic sediment.

CONCLUSIONS

The Ionian island of Zakynthos is located near the northern terminus of the Hellenic Trench. Subduction roll-back resulted in less foreland-propagating thrust and fold tectonics than further north in the external Hellenides and has permitted rapid late Cenozoic extensional basin subsidence. Basins around Zakynthos show a systematic series of facies that result from this complex tectonic setting. Pliocene thrust and fold segmentation of the older late Miocene basin resulted in condensed shelf sedimentation on the flanks of low anticlines and rapid deposition of principally coarse-grained paralic sediments in basins between evaporitic diapirs that rose along thrust ramps. During the Pliocene, a coarsening-up sequence of alternating prodeltaic muds and sands from the Acheloos River accumulated in north-eastern Zakynthos. Progressive uplift of the Ionian islands throughout the Pliocene resulted in the eastward migration of the Acheloos River delta and culminated in the early Pleistocene with the isolation of the Gulf of Patras from the open sea, forming a Patras–Corinth lake. The evolving Quaternary basins of Zakynthos Channel and the Gulf of Patras eventually re-established a marine connection to the Gulf of Patras. These basins accumulated distal muds from the Acheloos delta. At their margins, particularly at lowstands of sea level, coarse locally derived sediments accumulated.

ACKNOWLEDGMENTS

We would like to thank Drs Philip A. Allen, Mike Brooks and John Underhill for their constructive reviews.

REFERENCES

Alexander, J., Nichols, G.J. & Leigh, S. (1990) The origins of marine conglomerates in the Pindus foreland basin, Greece. *Sediment. Geol.*, **66**, 243–254.

Ambraseys, N. & Jackson, J. (1990) Seismicity and associated strain in central Greece between 1890 and 1988. *Geophys. J. Intern.*, **101**, 663–708.

Biscaye, P.E. (1965) Mineralogy and sedimentation of recent deep-sea clays in the Atlantic Ocean and adjacent seas and oceans. *Geol. Soc. Am. Bull.*, **76**, 803–832.

Bizon, J.J. & Bizon, G. (1985) 37-d-1: Zakinthos, Gr. In: *Neogene of the Mediterranean Tethys and Paratethys: Stratigraphic Correlation Tables and Sediment Distribution Maps* (Ed. By F. F. Steininger, J. Senes, K. Kleeman, and F. Rogl), p. 100. Institute of Paleontology, University of Vienna.

Bizon, G. & Mirkou, R. (1969) Foraminifères planktoniques du Pliocène l'île de Zante (Grèce occidentale). *First Intern. Conf. Plankt. Microf., Geneva (1967), Proceedings*, 1, 285–297.

Bizon, G. & Muller. C. (1977) La limite Pliocene – Pleistocene dans l'ile de Zante. Lacoupe de la Citadelle. *C. R. Somm. Geol. Franc.*, **4**, 212–216.

Black-Vernet, L. & Keraudren, B. (1970) Sur la prèsence du calabrien à Hyalina balthica dans les îles de Zakynthos et de Kephallinia (Grèce). *Bull. La Museum l'Anthropologie Prehistorique Du Monaco*, **15**, 91–106.

Bp Co. Ltd. (1971) *The Geological Results of Petroleum Exploration in Western Greece*. IGME.

Brooks, M., Clews, J.E., Melis, S. & Underhill, J.R. (1988) Structural development of Neogene basins in western Greece. *Basin Res.*, **1**, 129–138.

Brooks, M. & Ferentinos, G. (1984) Tectonics and sedimentation in the Gulf of Corinth and the Zakynthos and Kefallinia Channels, western Greece. *Tectonophysics*, **101**, 25–54.

Chronis, G., Piper, D.J.W. & Anagnostou, C. (1991) Late Quaternary evolution of the Gulf of Patras. *Mar. Geol.*, **97**, 191–209.

Clews, J.E. (1989) Structural controls on basin evolution: Neogene to Quaternary of the Ionian zone, Western Greece. *J. Geol. Soc., London*, **146**, 447–457.

Cushing, E.M. (1985) *Evolution structurale de la marge nord ouest Hellenique dans l'ile de Levkas et ses environs (Grèce nord-occidentale)*. PhD thesis, Universitè de Paris-Sud (XI), Centre d'Orsay.

Dermitzakis, M.D. & Georgiades-Dikeoulia, E. (1987) Biostratigraphy and paleoenvironmental reconstruction of the SE part of Zakynthos island (W. Greece) during the Plio-Pleistocene. *Interim Colloquium, Evolution climatique dans le domaine mediterraneen au neogene, R.C.M.N.S., Montpellier and Barcelona* (1987), Collected Abstracts, p. 28.

Dermitzakis, M.D., Papanikolaou, D. & Karotsieris, Z. (1979) The marine Quaternary formations of SE Zakynthos island and their paleogeographic implications. *6th Coll. Geol. Aeg. Reg. (Athens* 1977), *Proc.*, 1, 407–415, *Inst. Geol. Min. Res., Athens.*

Doutsos, T., Kontopoulos, N. & Poulimenos, G. (1988) The Corinth-Patras rift as the initial stage of continental fragmentation behind an active Island arc (Greece). *Basin Res.*, **1**, 177–190.

Doutsos, T., Pe-Piper, G., Boronkay, K. & Koukouvelas, I. (1993) Kinematics of the central Hellenides. *Tectonics*, **12**, 936–953.

Doutsos, T. & Piper, D.J.W. (1990) Listric faulting, sedimentation and morphologic evolution of the eastern Corinth rift: first stages in continental rifting. *Bull. Geol. Soc. Am.*, **102**, 812–829.

Ferentinos, G., Brooks, M. & Doutsos, T. (1985) Quaternary tectonics in the Gulf of Patras, western Greece. *J. struct. Geol.*, **7**, 713–717.

Frydas, D. (1986) Calcareous nannofossils from Cenozoic of Paxos and Zakynthos islands. *Bull. Geol. Soc. Greece*, **18**, 143–159.

Frydas, D. (1987) Kalkiges Nannoplankton aus dem Neogen von NW-Peloponnes. *N. Jb.Geol.Paläont.Mon.*, **5**, 274–286.

Frydas, D. (1989) Biostratigraphische Untersuchungen aus dem Neogen der NW- und W-Peloponnes, Griechenland. *N. Jb. Geol. Paläont. Mon.*, **6**, 321–344.

Gawthorpe, R.L., Fraser, A.J. & Collier, R.E.LL. (1994) Sequence stratigraphy in active extensional basins: implications for the interpretation of ancient basin fills. *Mar. Petrol. Geol.*, **11**, 642–658.

Gonzalez-Bonorino, G.G. (1996) Foreland sedimentation and plate interaction during closure of the Tethys ocean (Tertiary; Hellenides; western continental Greece). *J. sedim. Res.*, **66**, 1148–1155.

Heimann, K.O. (1977) *Die Fazies des Messins und Untersten Pliozans auf den Ionischen Inseln (Zakynthos, Kephallinia, Korfou/ Griechenland) und auf Sizilien.* Dissertation, Technische Universität München.

Kamberis, E., Marnelis, F., Loucoyannakis, M., Maltezou, F., Hirn, A. & Streamers Group (1996) Structure and deformation of the External Hellenides based on seismic data from offshore Western Greece. In: *Oil and Gas in Alpidic Thrustbelts and Basins of Central and Eastern Europe* (Ed. by G. Wessely and W. Liebl), *Eage Sp. Publ.*, **5**, 207–214.

Kissel, C. & Laj, C. (1988) The Tertiary geodynamical evolution of the Aegean arc: a paleomagnetic reconstruction. *Tectonophysics*, **146**, 183–201.

Kontopoulos, N. & Doutsos, T. (1985) Sedimentology and tectonics of the Antirion area (Western Greece). *Bull. Geol. Soc. Italy*, **104**, 479–489.

Kontopoulos, N. & Zelilidis, A. (1992) Upper Pliocene lacustrine environments in the intramontane Rio graben basin, NW Peloponnesus, Greece. *N. Jb. Geol Paläont. Mon.*, **2**, 102–114.

Kontopoulos, N., Zelilidis, A., Piper, D.J.W. & Mudie, P.J. (1997) Messinian evaporites in Zakynthos, Greece. *Palaeogeogr. Palaeoclim. Palaeoecol.*, **129**, 361–367.

Kourtessi-Filippakis, G. (1996) New data of Paleolithic habitation on Zakynthos. *Archeol. Arts*, **60**, 77–80.

Leeder, M.R. & Jackson, J.A. (1993) The interaction between normal faulting and drainage in active extensional basins, with examples from the western United States and Greece. *Basin Res.*, **5**, 79–102.

Mascle, J. & Martin, L. (1989) Shallow structure and recent evolution of the Aegean Sea: a synthesis based on continuous reflection profiles. *Mar. Geol.*, **94**, 271–299.

McKenzie, D.P. (1978) Active tectonics of the Alpine Himalayan belt: the Aegean Sea and surrounding areas. *Geoph. J. R. Atr. Soc.*, **55**, 217–254.

Mirkou, M.R. (1987) Foraminifères benthiques plio-pleistocènes de Zanthe. Interpretation palèobathymetrique. *Ecl. Geol. Helv.*, **80**, 109–125.

Pe-Piper, G., Piper, D.J.W., Kotopouli, C.N. & Panagos, A.G. (1995) Neogene volcanoes of Chios, Greece: the relative importance of subduction and back-arc extension. *Geol. Soc. Spec. Publ.*, **81**, 213–232.

Piper, D.J.W., Kontopoulos, N. & Panagos, A.G. (1987) Deltaic sedimentation and stratigraphic sequences in post-orogenic basins, Western Greece. *Sediment. Geol.*, **55**, 283–294.

Piper, D.J.W. & Panagos, A.G. (1981) Growth patterns of the Acheloos and Evinos deltas, Greece. *Sediment. Geol.*, **28**, 111–132.

Piper, D.J.W., Panagos, A.G., Contopoulos, N. & Pe, G.G. (1976) Depositional environments of Pliocene littoral sediments, Gythion, southern Peloponnese. *Z. Dt. Geol. Ges.*, **127**, 435–444.

Piper, D.J.W., Panagos, A.G., Kontopoulos, N. & Spiliotopoulou, M. (1982a) Coastal processes and morphology, Gulf of Patras, Greece. *Z. Geom.*, **26**, 365–374.

Piper, D.J.W., Pe-Piper, G., Kontopoulos, N. & Panagos, A.G. (1982b) Plio-Pleistocene sedimentation in the western Lakonia graben, Greece. *N. Jb. Geol. Paläont. Mon.*, **11**, 679–691.

Piper, D.J.W., Stamatopoulos, L., Poulimenos, G., Doutsos, T. & Kontopoulos, N. (1990) Quaternary history of the Gulfs of Patras and Corinth, Greece. *Z. Geom.*, **34**, 451–458.

Robertson, A.H.F., Clift, P.D., Degnan, P. & Jones, G. (1991) Paleoceanography of the Eastern Mediterranean Neotethys. In: *Paleogeography and Paleooceanography of Tethys* (Ed. by J. Chanel, E. Winterer and L. F. Jansa), *Spec. Issue Paleogeogr. Paleoclim. Paleoecol.*, **87**, 289–343.

Royden, L.H. (1993) Evolution of retreating subduction boundaries formed during continental collision. *Tectonics*, **12**, 629–638.

Seger, M. & Alexander, J. (1993) Distribution of Plio – Pleistocene and Modern coarse-grained deltas south of the Gulf of Corinth, Greece. *Spec. Publ. Int. Ass. Sediment.*, **20**, 37–48.

Sorel, D. (1976) *Ètude nèotectonique dans l'arc ègèen externe occidental: les iles ioniennes de Kefallinia et Zakinthos et l'Elide.* Thèse 3[me] cycle, Université de Paris Sud, Orsay.

Stamatopoulos, L., Voltaggio, M. & Kontopoulos, N. (1994) ^{230}Th/^{238}U dating of corals from Tyrrhenian deposits and the palaeogeographic evolution of the western Peloponnesus (Greece). *Munst. Forsch. Geol. Paläont.*, **76**, 345–351.

Stamatopoulos, L., Voltaggio, M., Kontopoulos, N., Cingue, A. & La Rossa, S. (1988) ^{230}Th/^{238}U dating of corals from Tyrrhenian marine deposits nearby Varda (northwestern Peloponnesus) Greece. *Geogr. Fis. Dinam. Quatern.*, **11**, 99–103.

Taymaz, T., Jackson, J.A. & McKenzie, D.P. (1991) Active tectonics of the north and central Aegean. *Geophys. J. Intern.*, **106**, 433–490.

Triandafyllou, M.B. (1993) *Biostratigraphic and ecostratigraphic preservations based on calcareous nannofossils of the Plio-Pleistocene sediments of eastern Mediterranean*. PhD thesis, Athens University.

Tsapralis, B. (1981) *Contribution in the Pleistocene study of Zakynthos (Ostracodes – paleoenvironment)*. PhD thesis, University of Athens.

Underhill, J.R. (1988) Triassic evaporites and Plio-Quaternary diapirism in western Greece. *J. Geol. Soc. London*, **145**, 269–282.

Underhill, J.R. (1989) Late Cenozoic deformation of the Hellenide foreland, western Greece. *Geol. Soc. Am. Bull.*, **101**, 613–634.

Vita Finzi, C. & King, G.C.P. (1985) The seismicity, geomorphology and structural evolution of the Corinth area of Greece. *Philos. Trans. R. Soc. London*, **314**, 379–407.

Walker, R.G. & Plint, A.G. (1992) Wave- and storm-dominated shallow marine systems. In: *Facies Modes* (Ed. by R. G. Walker, and N. P. James), pp. 219–238. Geol. Ass. Canada.

Wilpshaar, M. (1995) Applicability of dinoflagellate cyst stratigraphy to the analysis of passive and active tectonic settings. *Geol. Ultraiectina*, **134**.

Zelilidis, A., Koukouvelas, I. & Doutsos, T. (1988) Neogene paleostress changes behind the forearc fold belt in the Patraikos Gulf area, western Greece. *N. Jb. Geol. Palaont. Mh.*, **5**, 311–325.

Received 23 July 1997; revision accepted 2 June 1998