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Structural development of Neogene basins in western Greece

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ABSTRACT

An account is given of the structural setting of the various Neogene sedimentary basins of western Greece. Compressional basins are attributable to foreland loading by the Alpine fold and thrust belt of the Outer Hellenides, and to active subduction in the adjacent western Hellenic arc. Late extensional basins are related to N–S crustal extension in the Aegean marginal basin and, in western Greece, are superimposed on the earlier compressional structures. The local seismicity provides evidence that the main E–W-trending basin-bounding faults of the extensional basins form a linked system that includes NW–SE- and NE–SW-trending transfer zones of transtension. The transfer zones are themselves the sites of small extensional basins.

INTRODUCTION

The Aegean region represents an ensialic marginal basin behind an active subduction zone located along the outer Hellenic arc. Although the age of initiation of subduction beneath the Aegean region is not known (e.g. Sorel & Mercier, in press), the present subduction regime has probably existed since at least mid-Miocene times (Le Pichon & Angelier 1979).

Focal mechanisms of shallow earthquakes (McKenzie 1972, 1978; Papazachos *et al.* 1984) show that the Aegean marginal basin is characterized by a large internal zone of crustal extension and a narrow peripheral zone in which crustal shortening dominates. The zone of extension occurs not only behind the active volcanic arc of the Cyclades but extends also into the arc–trench gap.

Towards the western and southwestern margins of the Aegean region, the extensional regime passes rather abruptly into a zone of compressional tectonics in which the maximum compression is roughly perpendicular to the margin (Mercier, Sorel & Simeakis, 1987). Structures resulting from this subduction-related compressional regime are, in western Greece, largely coaxial with earlier Alpine structures resulting from the collision of the Pre-Apulian microcontinent with Eurasia. Mercier *et al.* (1987) attributed the overall distribution of stress in the Aegean lithosphere to the combined effect of the push of the African plate on the Aegean boundary together with differential internal stresses resulting from lateral variations in crustal thickness and deep thermal structure.

Palaeomagnetic studies of Cenozoic rocks in the Aegean region (e.g. Laj *et al.* 1982; Kissel & Laj 1988) demonstrate that parts of western Greece have undergone substantial clockwise rotation during the Neogene simultaneous with compression.

The Alpine framework on which Neogene basins are developed consists of the three zones of isopics in the External Hellenides: the Gavrovo–Tripolitza, Ionian and Pre-Apulian zones (Aubouin 1959, 1965; Aubouin & Dercourt 1962; Temple 1968; Jenkins 1972; Smith & Moores 1974; Fig. 1). The Gavrovo–Tripolitza and Ionian zones are conventionally regarded as the two most external Alpine thrust sheets emplaced on to a stable Pre-Apulian autochthon as a late orogenic event. The stratigraphic differences on which the zones of isopics are defined do not extend into the Neogene, when subsidence in a foreland basin setting associated with an advancing thrust front affected the western Greek area (Underhill 1988, in press). A westward advance of the internal zone of crustal extension towards the western margin through the Neogene has led to the superposition of later extensional basins on the earlier compressional basins.

The main purpose of the present paper is to provide a systematic account of the various Neogene basins and to discuss the different structural controls on their development.

COMPRESSIONAL BASINS (Fig. 2)

Oligo–Miocene basin development and sedimentary patterns

The fold and thrust belt of the Outer Hellenides was initiated during the Alpine collision of the Apulian

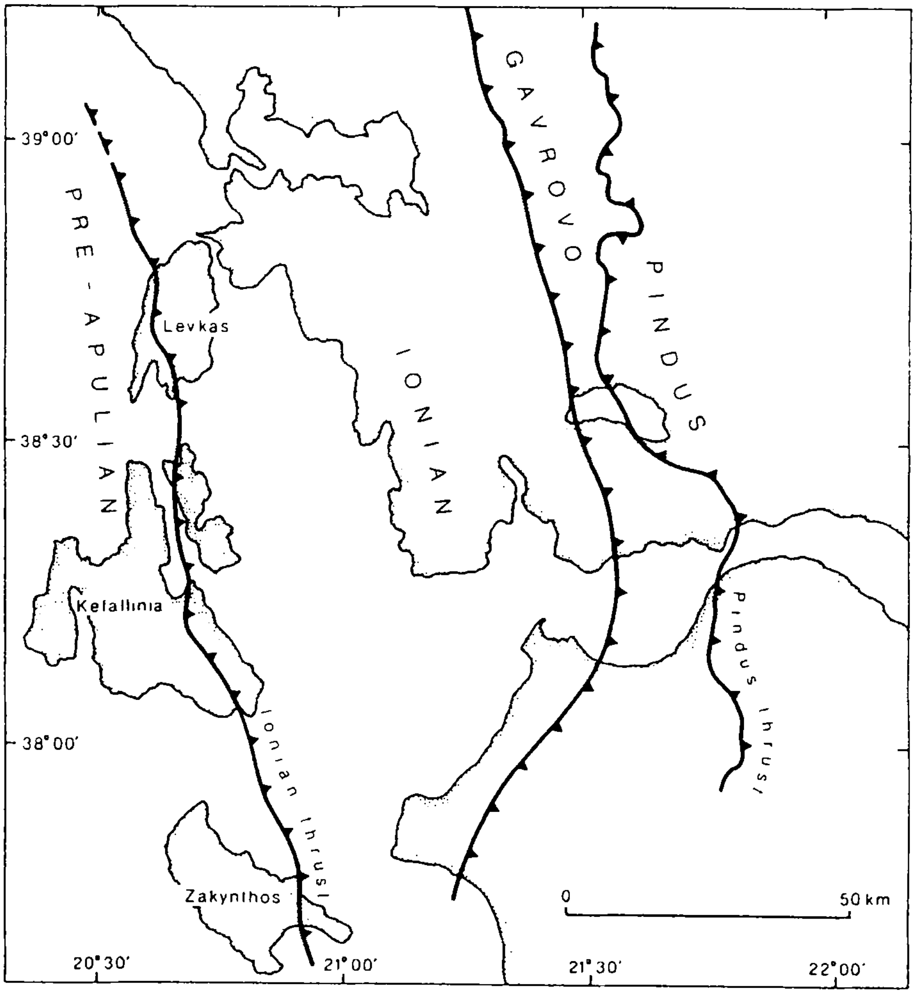

Fig. 1 Isopic zones of the Outer Hellenides of western Greece.

microcontinent with Eurasia, and deformation advanced progressively towards the foreland in a direction that is now westerly but was probably SSW before the clockwise rotational movements of western Greece during the Neogene (Laj *et al.* 1982; Kissel, Laj & Moller 1985; Kissel & Laj 1988). Displacement on the Pindus thrust had begun by mid-Eocene times, and this thrust remained active until the Oligocene (Fleury 1980). A thick Oligocene to early Miocene clastic wedge (the flysch of the Ionian and Gavrovo zones), derived from the Pindus mountains, was deposited on top of the Mesozoic to Eocene sequence, consisting mainly of carbonates, of the Outer Hellenides (Fig. 3a & b; Aubouin 1959; Fleury 1980).

At the end of the Oligocene, movement began on the Gavrovo thrust to the west of the Pindus thrust (Fig. 3c; Dercourt & Thiebault 1979) by progressive footwall collapse (Boyer & Elliott 1982; Butler 1987) and the Pindus thrust became inactive (Fleury 1980). Sedimentary facies and palaeocurrent patterns in the foreland basin indicate synsedimentary tectonic activity and this has been related to the advance of the thrust front through the basin (Clews, in press; Alexander & Nichols, pers. comm.).

A phase of compressional deformation in the Burdigalian to Early Langhian is recorded in the Ionian zone of Levkas (Cushing 1985). Two possible driving mechanisms for this deformation have been suggested: (1) continued foreland migration of the active front of the Hellenide fold and thrust belt (Clews, in press); (2) the onset of subduction in the western Hellenic arc. Underhill (in press) points out that these different causal mechanisms would be difficult to distinguish in the more external Hellenides. The beginning of the mid-Miocene is marked by a basin-wide hiatus and the Lower Miocene flysch of the Ionian and Gavrovo zones is unconformably overlain by clastic sediments of mid-Miocene to Quaternary age.

Flysch-like sediments are absent from the Lower Miocene of the Pre-Apulian zone and the carbonate-dominated sequences of Kefallinia and Zakynthos form part of a progressive westward thinning with associated facies change.

The close association between the thick turbidite-dominated clastic wedge and the advancing Hellenide nappe pile suggests an intimate tectonic control on Oligo–Miocene basin development and internal sedimentary patterns. Recent developments in our understanding of analogous settings elsewhere (e.g. Rocky Mnts, Apennines, Alps) have demonstrated that asymmetric foreland basins are developed in such positions as a result of lithospheric

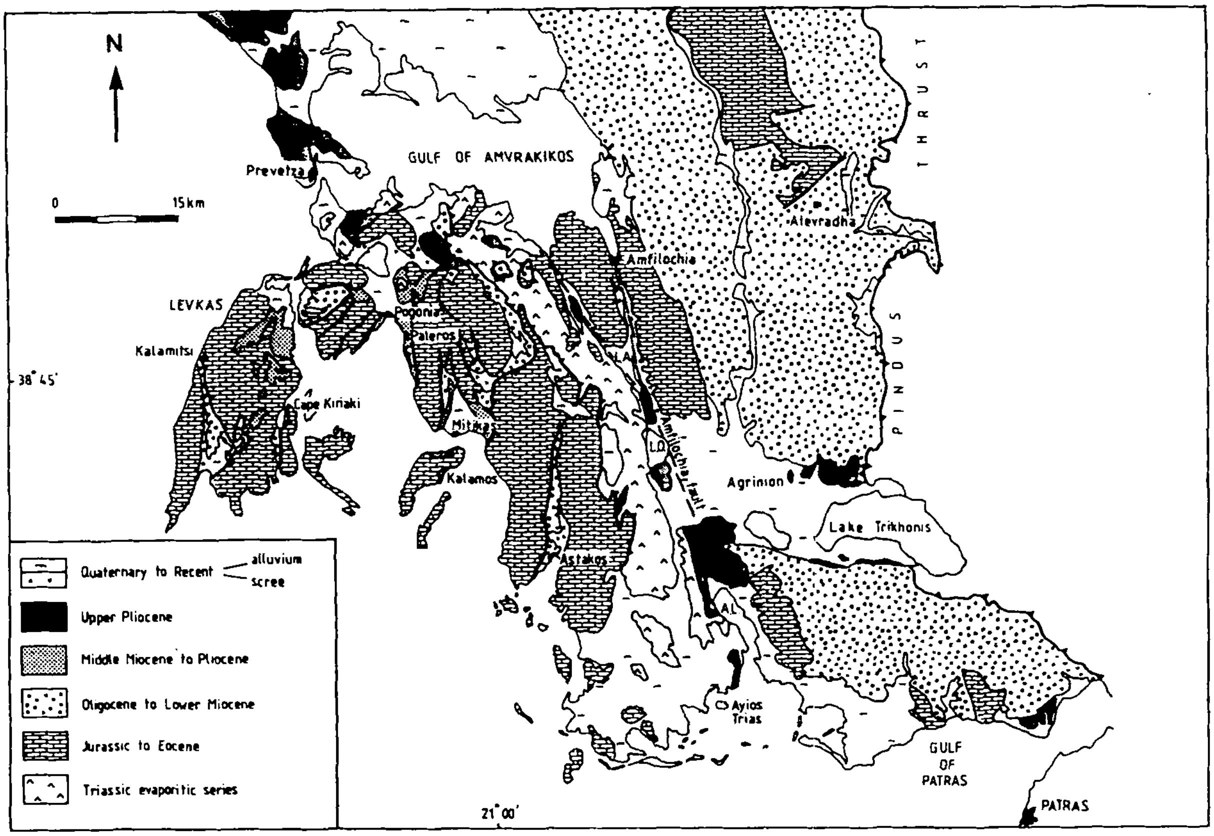

Fig. 2 Outline geological map of western Greece, showing the main Neogene units (after Clews, in press).

flexure ahead of the thrust load (see e.g. Allen, Homewood & Williams 1986, and references therein). The clastic wedge immediately west of the Pindus thrust is believed to represent an example of a foreland basin developed by the downwarping of the Ionian and Gavrovo zones. Sedimentary thicknesses and patterns change progressively westwards into the Pre-Apulian zone where little clastic input existed.

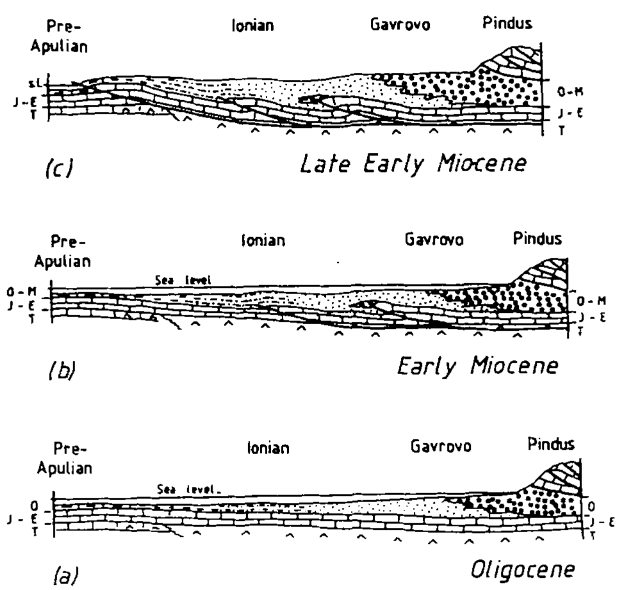

Fig. 3 Schematic restored cross-sections across the Outer Hellenides for (a) Oligocene, (b) Early Miocene, and (c) late Early Miocene times. T: Triassic; J–E: Jurassic to Eocene; O–M: Oligocene to Miocene.

Mio-Pliocene basin development and sedimentary patterns

Middle–Upper Miocene sediments on Levkas and Zakynthos occur close to the Ionian frontal thrust, locally known as the Kalamitsi thrust (BP Co Ltd 1971). Detailed descriptions of the sections on Levkas are given by Bornovas (1964) and Cushing (1985). The sequence on Levkas consists of breccias, conglomerates, sands and marls above a basal unconformity. A marine origin is indicated by rare fauna (Bornovas 1964). Diachroneity of basal beds above the unconformity was interpreted by Cushing (1985) as reflecting palaeorelief resulting from mid-Miocene tectonism. Coarse lithofacies predominate in the sequence, with clasts of limestone, chert, sandstone and shale. Limestone clasts are characteristic of pre-Oligocene Ionian zone carbonates and are readily distinguishable from the neritic limestones of the Pre-Apulian zone.

The lower part of the Upper Miocene sequence at Paleros on mainland Greece (Fig. 2) is conglomeratic with clasts predominantly of Mesozoic limestones and cherts. The upper part of the sequence consists of a series of fining-upward units of sands, silts and muds, 10 to 30 cm thick, with current ripples common in the sands and silts.

The silts and muds are finely laminated. Small-scale slumping is ubiquitous, and west-facing slump folds with approximately north–south-trending axes suggest a west-dipping palaeoslope (Clews, in press).

The proximity of the Kalamitsi thrust to the coarse detritus in Levkas, and the indication of an Ionian zone source area, suggest that the thrust sheet was emergent and was shedding sediment back towards the hinterland. The sedimentary sequence may be interpreted as a submarine apron with an easterly dipping palaeoslope (Clews, in press). The basin, forming behind the active thrust front and being carried forward by continued movement along the thrust plane, conforms to the definition of a thrust-sheet-top or piggy-back basin (Ori & Friend 1984). The fining-upward units of the Upper Miocene sequence at Paleros are interpreted as distal turbidites by Clews (in press) and the direction of palaeoslope implies an easterly source. A large thrust-sheet-top basin is therefore postulated, fed from both the Kalamitsi thrust front to the west and the older mountain belt to the east.

Although the Kalamitsi thrust was active in the early Miocene (Clews, in press), extensional growth faulting in the Pre-Apulian zone continued through to early Pliocene times (Sorel 1976) and acted as a major control on sedimentary patterns. Mid–Late Miocene deposition in the Pre-Apulian zone became increasingly dominated by clastic influences. A coarsening-up succession exists in which marls are replaced progressively by turbidites during the Serravallian–Tortonian. Thickness variations and sedimentary patterns in the Serravallian–Tortonian of Kefallinia, including the occurrence of olistoliths, indicate footwall uplift associated with at least one normal fault and suggest the presence of sub-basins at this time (Underhill, in press). Instability continued during the Messinian when the deposition of evaporitic sequences was disrupted by conglomeratic units. Deposition of these sequences in the Pre-Apulian zone was synchronous with uplift associated with the development and disruption of the thrust-sheet-top basin further east.

The extensional faulting in the Serravallian–Tortonian of Kefallinia may represent accommodation in response to lithospheric flexure on the external margin of the foreland basin (Underhill, in press). An angular unconformity of early Pliocene age in the Pre-Apulian zone of Kefallinia is associated with the reactivation of the earlier extensional faults as compressional structures (Sorel 1976) and suggests that the active thrust front had by then migrated westwards from the Ionian zone. Mid–Late Miocene compressional tectonics in Pogonia and soft sediment deformation in the Upper Miocene of Paleros suggest that the Ionian zone continued to be tectonically active at this time.

Pliocene deposits south of Pogonia comprise marine marls and muds with occasional calcarenites. The upward decrease in grain size from the coarse Miocene sediments suggests a cut-off in sediment supply or a lower energy of depositional environment. In Mitikas Valley, Middle and Upper Pliocene sediments (Doutsos, Kontopoulos & Frydas 1987) unconformably overlie folded and faulted Mesozoic to Lower Miocene strata, providing evidence of tectonism at some time during the mid-Miocene to Early Pliocene interval. BP Co Ltd (1971) recorded a thick sequence of conglomerates, sandstones and shales from the east of Mitikas Valley, with a freshwater fauna in the finer lithologies. This sequence has been interpreted by Clews (in press) as an apron of small, locally derived alluvial fans, possibly developed on the newly formed relief.

East of Mitikas, Pliocene sediments show a general decrease in grain size compared with their northern counterparts in Mitikas Valley. Thick oyster beds suggest euryhaline conditions, possibly indicating the local development of brackish marine conditions. In Levkas, apart from a single outcrop of Late Miocene–Early Pliocene marine muds in the north of the island, Pliocene sedimentation is exclusively continental (Cushing 1985).

The onset of continental conditions marks a regression that may be related to Pliocene global sea-level fall (Vail, Mitchum & Thompson 1977), or may reflect local tectonic activity along the Kalamitsi thrust. Indeed, uplift related to Mio–Pliocene deformation within the Ionian zone may have led to the progressive restriction of the thrust-sheet-top basin at both its eastern and western margins.

In Kefallinia and Zakynthos, a major unconformity separates Messinian sequences from those of the Middle Pliocene. Sedimentation persists in well-defined basins in the Pre-Apulian zone during the late Pliocene. Thick Plio–Cantabrian delta-fan conglomerates in western Kefallinia were shed westwards and southwestwards (Underhill 1985). Subsequent emplacement of thrusts in the Pre-Apulian zone to the west may have caused renewed uplift, thereby halting deposition in the Pre-Apulian zone except in local compressional basins.

In Quaternary times, the thrust-sheet-top basin sediments of the Ionian zone were deformed by approximately east–west compression in an episode of break-back thrusting (Clews, in press) and by local diapiric activity. Pliocene deposits in Mitikas Valley and Mio–Pliocene sediments in the Paleros and Pogonia areas were folded and overthrust by pre-Oligocene carbonates. A phase of compressional tectonics dated as early Quaternary has also been recognized in Levkas (Cushing 1985). Diapiric rise of Triassic evaporites from the decollement horizon beneath the Ionian thrust sheet caused the local development of peripheral sinks around diapirs, resulting in an extensional overprint on the more regional compressional tectonics (Underhill 1988). This process is most developed in the zone of active subsidence in the offshores zones of the Zakynthos and Kefallinia channels (Brooks & Ferentinos 1984).

LATE EXTENSIONAL BASINS (PLIO–QUATERNARY)

Late extensional basins (Fig. 4) in western onshore Greece are related to approximately north–south crustal extension. They represent the westernmost elements of an Aegean-

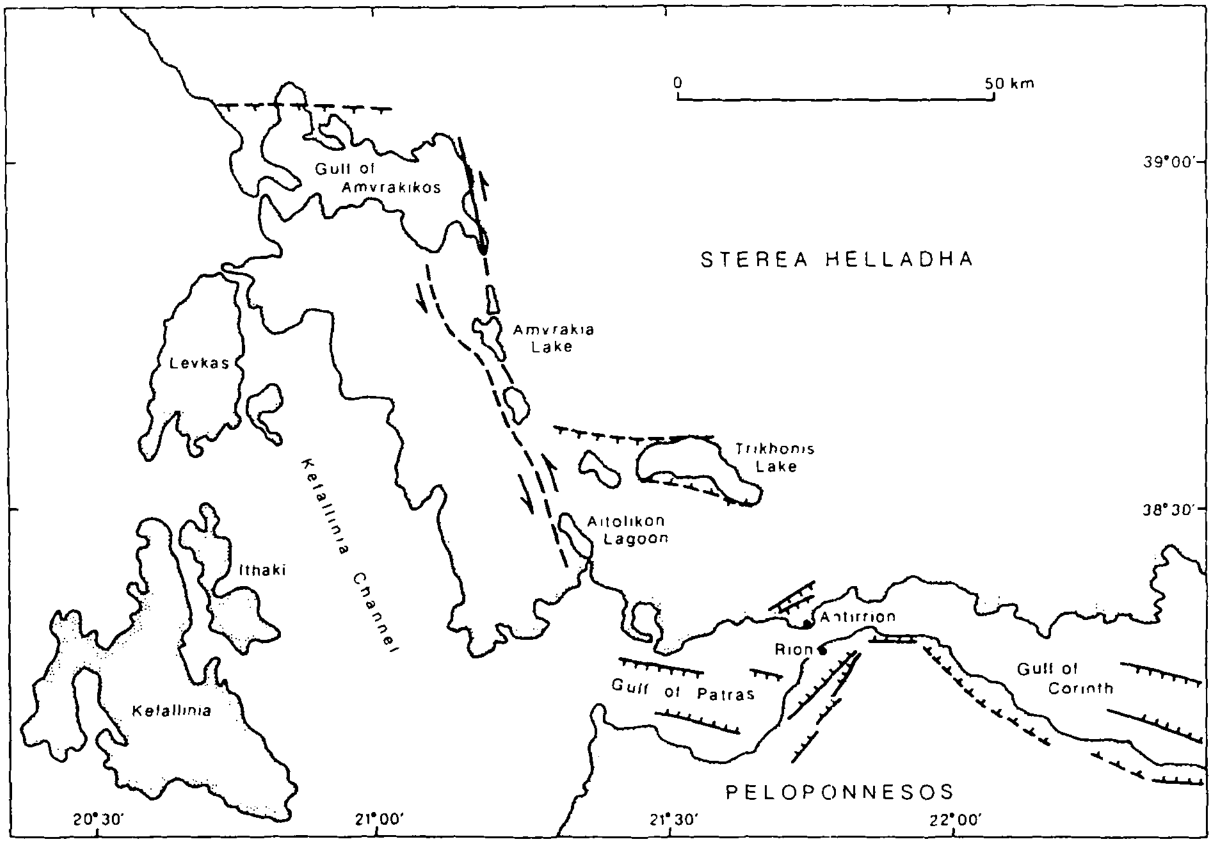

Fig. 4 Late extensional basins of western Greece.

wide marginal basin system causally related to active subduction in the western Hellenic arc. This system was probably initiated in the mid-Miocene (Le Pichon & Angelier 1979) and is still active. Throughout this period, extensional tectonics in the interior of the Aegean domain have coexisted with compressional tectonics in the western Hellenides. Associated with the westward migration of the locus of east–west compressional tectonics, as described in earlier sections of this paper, the limit of north–south crustal extension has advanced westwards. Thus, in western Greece, the late extensional basins are superimposed on earlier compressional structures. Typically, the later structures cross-cut the earlier structures, rather than reactivating them.

The main basins of western Greece belonging to this late extensional phase trend WNW–ESE and include the Gulf of Corinth, Gulf of Patras, Trikhonis Lake and the Gulf of Amvrakikos; related basins of different trend include the NE–SW waterway connecting the Gulfs of Patras and Corinth (including the Rion–Antirrion Channel), and the NW–SE depression containing Lake Amvrakia (Fig. 4).

Gulf of Corinth

The Gulf of Corinth is a complex asymmetric graben, which may be divided physiographically into a shelf, slope and floor (Heezen, Ewing & Johnson 1966). A large central area of the floor constitutes an abyssal plain with an almost constant water depth of 850 m, reflecting a state of nearly perfect dynamic balance between subsidence and sedimentation rates (Brooks & Ferentinos 1984).

The graben contains up to about 1 km of Quaternary sediments interpreted as being mainly turbidites with some intercalated debris flow deposits (Brooks & Ferentinos 1984). The abyssal plain has, during the development of the graben, been up to 4 km wider than at the present day but it has become restricted by recent differential movements along the northern margin. Higgs (this volume) discusses the structural evolution of the gulf and shows how the main depocentres have migrated in response to complex movement histories along intrabasinal and basin-bounding faults. There has been a progressive narrowing of the basin through time as fault-controlled subsidence has switched to antithetic faults concentrated towards the basin centre. Major footwall uplift during the Quaternary has raised thick Pliocene sandstone and marl sequences well above sea level immediately south of the gulf.

The main basin-bounding fault of the present-day Gulf of Corinth graben forms an impressive bathymetric feature along the southern margin of the gulf. Across this fault line there is about 3 km of relief, from the abyssal plain of the gulf to the peaks of flanking mountains in the northern Peloponnesos. The post-Pliocene displacement across the fault zone exceeds 4 km. The main fault scarp offshore is eroded by numerous canyons that feed sediment down on to deep sea fans. Some of the wedge-shaped deposits building out from the basin margins, as observed on high-resolution seismic profiling records (Brooks & Ferentinos 1984, fig. 4), may represent discrete fan units associated with individual canyon systems.

The gravity anomaly over the Gulf of Corinth and surrounding land areas is shown in Fig. 5. This map is a combination of Bouguer anomalies on land and free-air anomalies within the gulf, and is based on shipborne gravity surveys carried out during research cruises of RRS *Shackleton* (1/82) and RRS *Discovery* (137/83). A pronounced negative anomaly overlies the gulf, causing a local deflection in the pattern of north–south anomaly contours paralleling the regional strike of the Hellenides. The west–east rise in regional gravity is attributable to the effect of crustal thinning towards the interior of the Aegean domain (Makris 1977). The maximum amplitude of the local free-air anomaly is -65 mGal. This anomaly can be accounted for by a maximum thickness of about 1.2 km of sediments underlying the deep floor of the gulf (using a minimum reasonable density contrast of -0.45 Mg m^{-3}), an estimate in close agreement with seismic indications of sediment thickness (Brooks & Ferentinos 1984). There is thus no evidence for a substantial thickness of pre-Quaternary sediments under the floor of the gulf.

Gulf of Patras

By contrast with the Gulf of Corinth, water depths in the Gulf of Patras are everywhere shallow. Air gun seismic and 3.5-kHz profiling reveal several WNW–ESE growth faults defining, overall, a narrow asymmetric graben in the southern part of the Gulf (Ferentinos, Brooks & Doutos 1985). This localized graben is suggestive of a progressive narrowing of the zone of active subsidence within the Gulf of Patras basin, as with the Gulf of Corinth graben.

Although water depths in the gulf do not exceed about 130 m, the scale of structural relief in the basin is comparable to that in the Gulf of Corinth. Immediately to the north of the Gulf of Patras, mountains composed of Cretaceous–Eocene limestones rise to elevations of over 1000 m. Offshore drilling by the Public Petroleum Corporation of Greece in the outer part of the gulf proved that 1800 m of Neogene sediments overlie Triassic evaporites (Ferentinos *et al.* 1985). The maximum thickness of the Neogene is not known from drilling or seismics, but the Bouguer anomaly map (Fig. 6) shows a local negative anomaly, with an amplitude of about -40 mGal, largely restricted to the area of the gulf. Interpretation of this anomaly using a density contrast of -0.45 Mg m^{-3} for Neogene sediments produces a basin geometry consistent with the known Neogene thickness at the site of the offshore borehole and predicts a maximum Neogene thickness of about 2.5 km under the central part of the gulf. The Gulf of Patras exhibits the same structural trend as the Gulf of Corinth but the two basins are offset by about 25 km along a NE–SW offshore zone passing through the Rion–Antirrion Channel (Fig. 4).

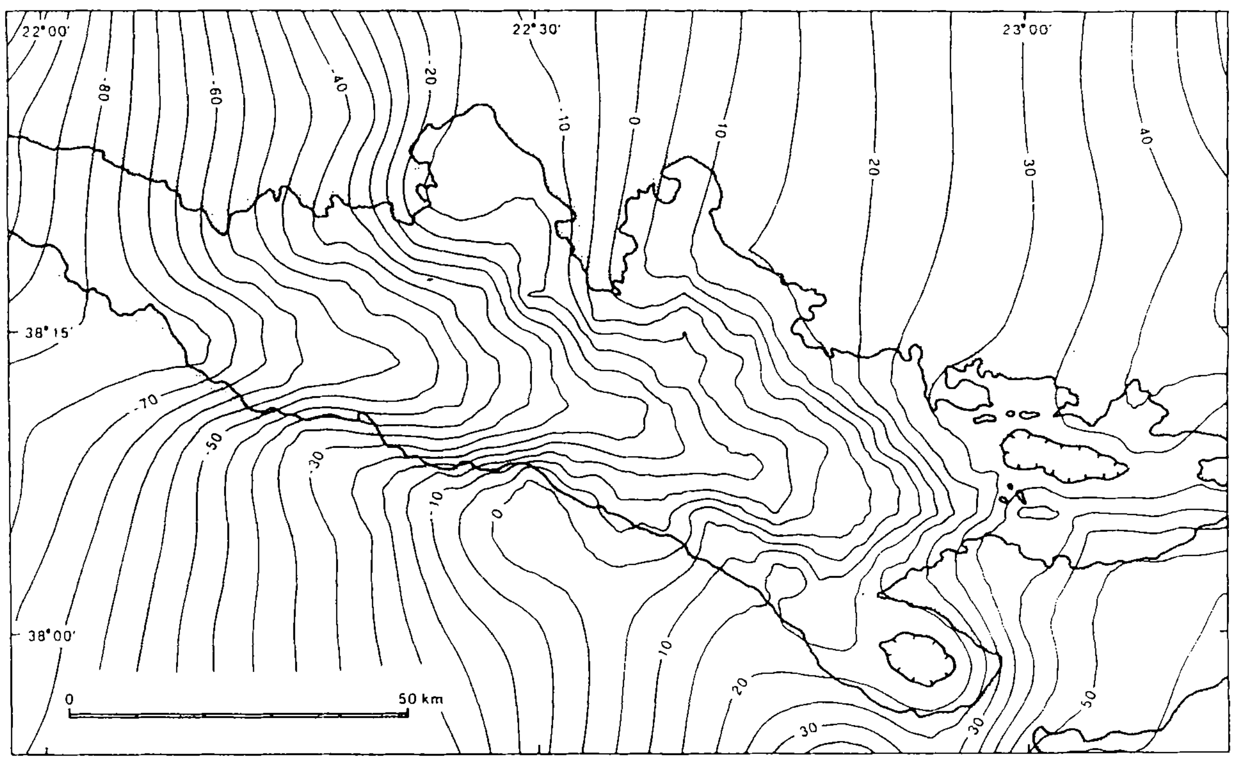

Fig. 5 Gravity map of the Gulf of Corinth area based on published Bouguer anomaly maps for the land area and unpublished free-air data for the gulf. Anomaly values on land have been re-adjusted to the gravity formula of 1967 (Kearey & Brooks 1984).

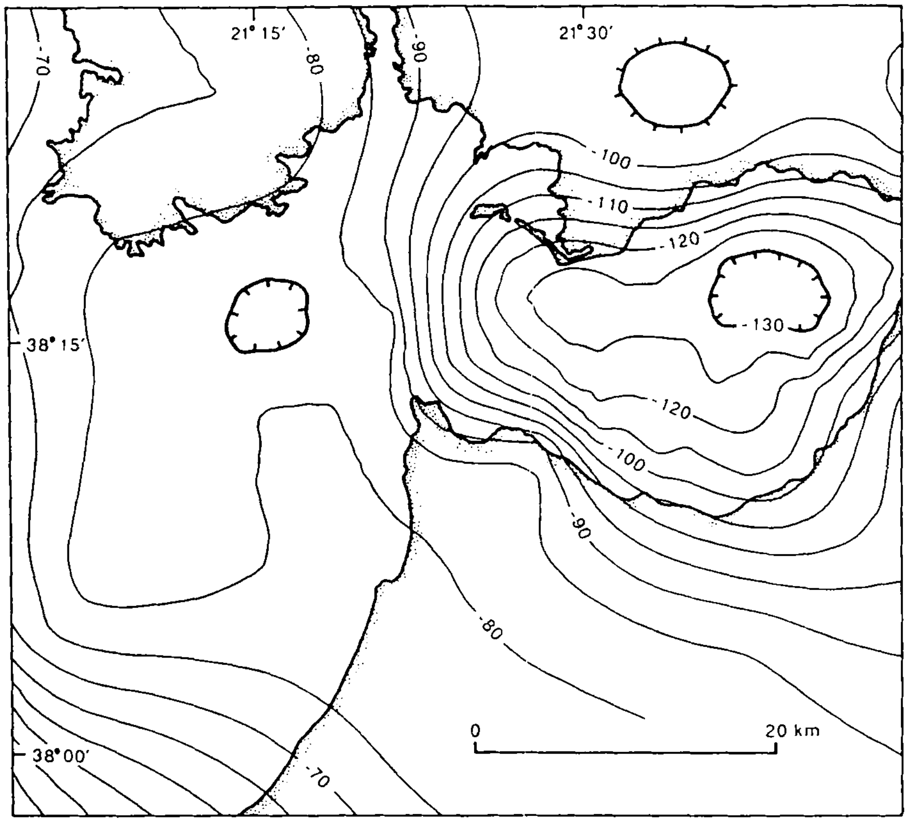

Fig. 6 Bouguer anomaly map of the Gulf of Patras area based on published maps for the land area and unpublished data for the gulf. Anomaly values on land have been re-adjusted to the gravity formula of 1967.

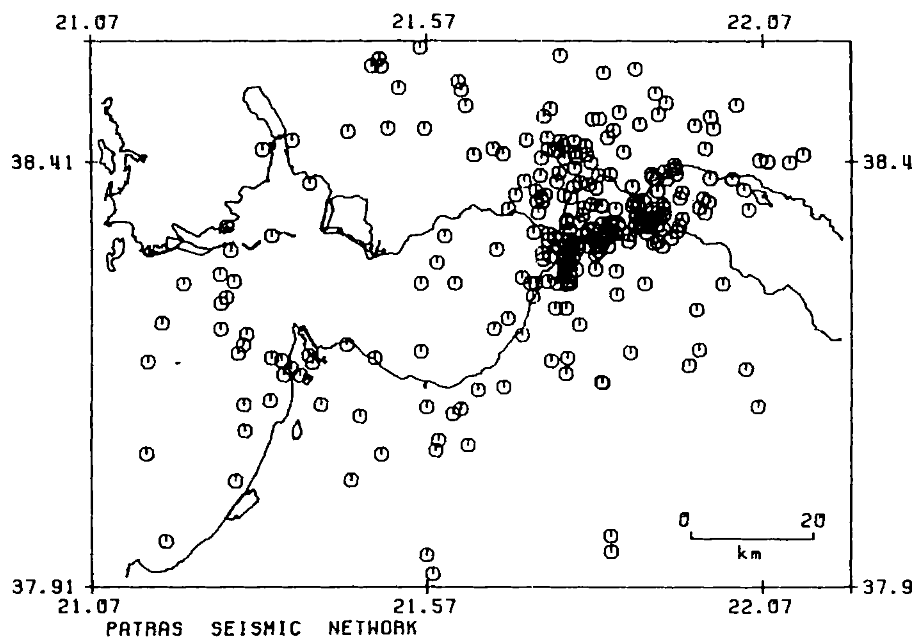

Fig. 7 Microseismicity of the Gulf of Patras area based on PATNET data (from Melis *et al.*, in press).

Trikhonis Lake

Very little sedimentological or structural work has been carried out in or around Trikhonis Lake, and the thickness of the lake sediments and the detailed geometry of the Trikhonis Lake graben are unknown. The main basin-bounding fault lies immediately to the south of the lake and is marked by a major escarpment.

Water depths in the lake reach 96 m. Sediment cores of up to 6 m in length were taken from the lake by Creer, Readman & Papamarinopoules (1981) as part of a study of geomagnetic variations over the past several thousand years recorded in Greek lake sediments. The oldest sediments cored yielded radiocarbon dates of 5–6000 yr BP, giving a recent sedimentation rate of about 1 mm yr^{-1}. Reconnaissance 3.5-kHz profiling in the lake in 1986 failed to reveal any faulting in the near-surface sediment layers (G. Ferentinos, pers. comm.).

The east–west depression containing Trikhonis Lake connects to the west with a low-lying, NNW–SSE-trending corridor extending from the Gulf of Amvrakikos to the Gulf of Patras and containing localized water-filled depressions, e.g. Lake Amvrakia and Aitolikon Lagoon (Fig. 8).

SEISMICITY AND TECTONICS

Seismicity of western Greece

The mainland area of Greece is the most seismically active part of the entire Alpine–Mediterranean belt, and comprehensive catalogues exist of seismic events that have occurred in Greece over the past two centuries (Galanopoulos 1960; Makropoulos & Burton 1981; Comninakis & Papazachos 1982).

For events in western Greece with magnitudes $M_s > 4.0$, the Gulf of Corinth and the Ionian islands coincide with zones of high seismicity whilst the intervening Gulf of Patras represents a well-defined seismic gap (Makropoulos & Burton 1981, 1984). A recent microearthquake study of the Gulf of Patras area, using data from the Patras Seismic Network (PATNET; Melis, Brooks & Pearce, in press), showed that the seismic gap in the Gulf of Patras exists for lower magnitude events also, and identified a localized zone of intense microseismicity centred in the Rion–Antirrion area (Fig. 7).

Microearthquake data recorded by PATNET over a six-month period, and by VOLNET (Volos Seismic Network) over a two-year period, show shallow events north of the Gulf of Patras, with focal depths of less than 10 km, and deeper events, with focal depths between 10 and 25 km, around Trikhonis Lake (Melis & Burton 1988). The VOLNET data especially, recorded over a longer period, demonstrate a continuity of seismicity between Trikhonis Lake and the Gulf of Corinth, including an area of high seismicity in the Rion–Antirrion area. The seismicity in this latter area is distributed in a diffuse zone dipping to the north–east and extending to a depth of about 25 km (Melis *et al.* in press, fig. 8). The network data demonstrate clearly that the hypocentres of local microearthquakes deepen northwards.

Focal mechanisms of shallow earthquakes in the Greek/Aegean region (McKenzie 1972, 1978; Papazachos *et al.* 1984) reveal a large internal zone of normal and strike-slip faulting associated with mainly north–south extension and a narrow peripheral zone of thrust faulting associated with radially directed compression around the outer Hellenic arc. In western Greece, the boundary between the internal zone of north–south extension and the radial zone of ENE–WSW compression passes through the outer part of the Gulf of Patras (McKenzie 1972, 1978; Pedotti 1988). In the area of the late extensional basins, focal mechanisms of local shallow earthquakes show normal faulting associated with north–south extension in the Gulf of Corinth area (Papazachos 1976) and a combination of normal and strike-slip faulting in the north–west Peloponnesos area adjacent to the Gulf of Patras (Pedotti 1988).

A seismotectonic model for the late extensional basins of western Greece

Melis *et al.* (in press) interpreted the pattern of microseismicity in the Gulf of Patras area in terms of a linked fault

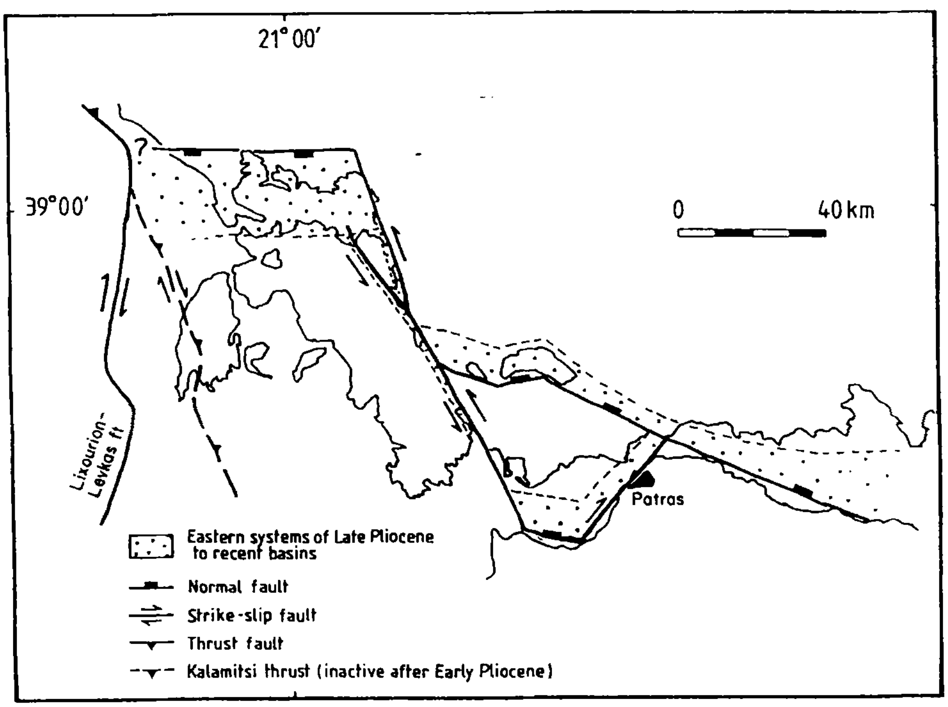

Fig. 8 Regional system of linked faults defining the late extensional basins of western Greece.

system connecting the individual late extensional basins of western Greece. They postulated that the active faults affecting the sea bed within the central part of the Gulf of Patras, which are apparently aseismic, are linked via a shallow decollement surface under the gulf to a deep WNW–ESE fault zone defining the southern margin of the Corinth graben and extending westwards to Trikhonis Lake (Melis *et al.* in press, fig. 9). The dipping seismic zone centred in the Rion–Antirrion area is assumed to be related to this deep fault zone. Aseismic faulting in the Gulf of Patras is attributed to the shallow surface of decollement into which the faults sole, probably at the level of the Triassic evaporites. A NE–SW transfer fault is postulated to account for the offset of the Corinth and Patras grabens, and local dextral strike-slip mechanisms (Pedotti 1988) may reflect movement along this transfer zone. A transfer zone is similarly postulated to account for the terminations of the Trikhonis Lake and Gulf of Amvrakikos graben structures along the NNW–SSE low-lying corridor containing Lake Amvrakia and Aitolikon Lagoon (Fig. 8). In the prevailing north–south extensional regime, both transfer fault zones are transtensional and, hence, provide sites for the development of localized extensional basins.

CONCLUSIONS

The main conclusions are enumerated below:

(1) The Neogene of western Greece records the unique development and disruption of sedimentary basins in association with a foreland-migrating thrust system and the initiation of a subduction-related marginal basin in an island arc setting.

(2) Oligo–Miocene basin development is consistent with the development of a major foreland basin ahead of the migrating Hellenide thrust load.

(3) Late Neogene tectonism dissected the foreland basin sequences by a combination of out-of-sequence thrusting, local diapirism and 'back-arc' extensional faulting. The latter is manifested by a major linked system of extensional basins.

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