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ORIGINAL PAPER

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Deformation of western Greece during Neogene clockwise rotation and collision with Apulia

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Abstract Following an Early Miocene phase of N–S extension affecting the entire Hellenides, 50° clockwise rotation affected western Greece. Modern GPS analyses show rapid southwestward motion in southwestern Greece over subducting oceanic lithosphere and no motion in the northwest, where Greece collided with Apulia. We aim to identify the deformation history of western Greece associated with the rotation and the collision with Apulia. The timing of the various phases of deformation is constrained via detailed analysis of vertical motions based on paleobathymetry evolution of sedimentary sequences overlying the evolving structures. The results show that accompanying the onset of rotation, compression was re-established in western Greece in the early Langhian, around 15 Ma. Subsequently, western Greece collided with the Apulian platform, leading in the Late Miocene to a right-lateral strike-slip system running from the Aliakmon Fault Zone in northern Greece via the Kastaniotikos Fault and the Thesprotiko Shear Zone to the Kefallonia Fault Zone, offshore western Greece. NE–SW compression and uplift of the Ionian Islands was accompanied by NE–SW extension in southwestern Greece, associated with faster southwestward motion in the south than in the north. This led in the middle Pliocene (around 3.5 Ma) to collision without further shortening in northwestern Greece. From then onward, NW–SE to N–S extension east of Apulia, and gradually increasing influence of E–W extension in the south accommodated motion of the Hellenides around the Apulian platform. As a result, curved extensional basin systems evolved, including the Gulf of Amvrakikos-Sperchios Basin–Gulf of Evia system and the Gulf of Corinth–Saronic Gulf system.

Keywords Vertical motions · Paleobathymetry · Aegean · Neogene · Biostratigraphy

Introduction

Following the Mesozoic Paleogene nappe stacking, the Aegean region became subjected to late-orogenic extension and formation of the Aegean arc since the Oligocene (Lister et al. 1984; Jacobshagen 1986; Gautier et al. 1999; Jolivet and Patriat 1999; Jolivet et al. 2003). In western Greece (Fig. 1), extension did occur in the Early Miocene (Van Hinsbergen et al. 2005b) but contrary to the central and southern Aegean, here extension ceased after the Early Miocene, not to return until the Late Pliocene, when numerous extensional basins—such as the Gulf of Corinth—started to develop (e.g. King et al. 1983; Doutsos et al. 1988; Goldsworthy et al. 2002). Instead, the area was dominantly subjected to compressional deformation (Mercier et al. 1972; Underhill 1989) and rotated approximately 50° clockwise since the Middle Miocene (Horner and Freeman 1983; Kissel and Laj 1988; Van Hinsbergen et al. 2005c). During some stage in the Neogene, northwestern Greece collided with the Apulian platform (Fig. 2), leading to the end of further convergence, whereas in southwestern Greece convergence between the lithosphere underlying the Ionian basin and the Aegean is still ongoing today (Kahle et al. 1993; McClusky et al. 2000; Sachpazi et al. 2000; Jiménez-Munt et al. 2003). In this paper, we aim to provide time constraints on the deformation phases that affected western Greece since the end of Early Miocene extension, contemporaneous with and following the

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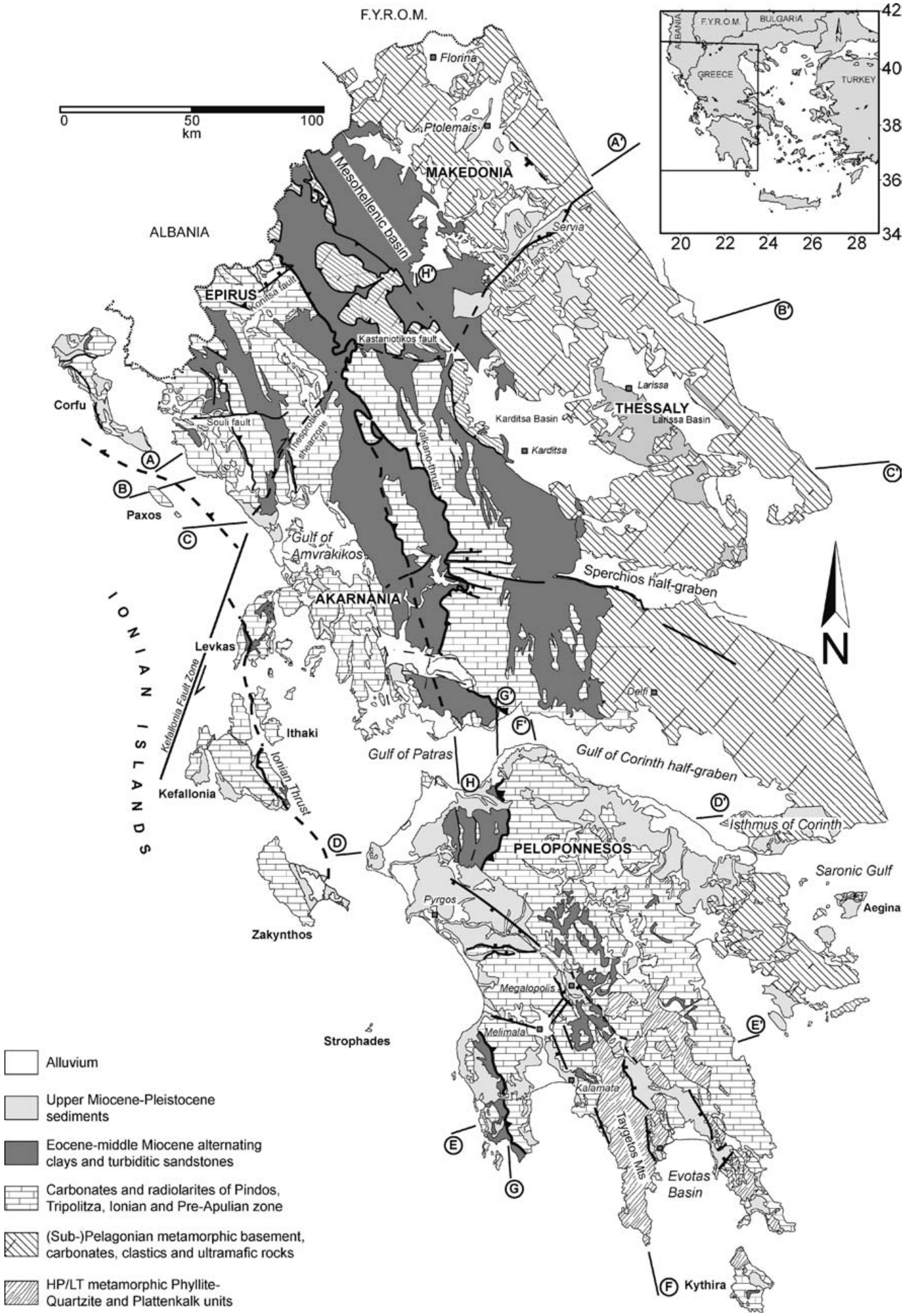


Fig. 1 Geological map of Western Greece, modified after Bornovas and Rontogianni-Tsiabaou (1983). Profiles A–A′ to H–H′ are shown in Fig. 4

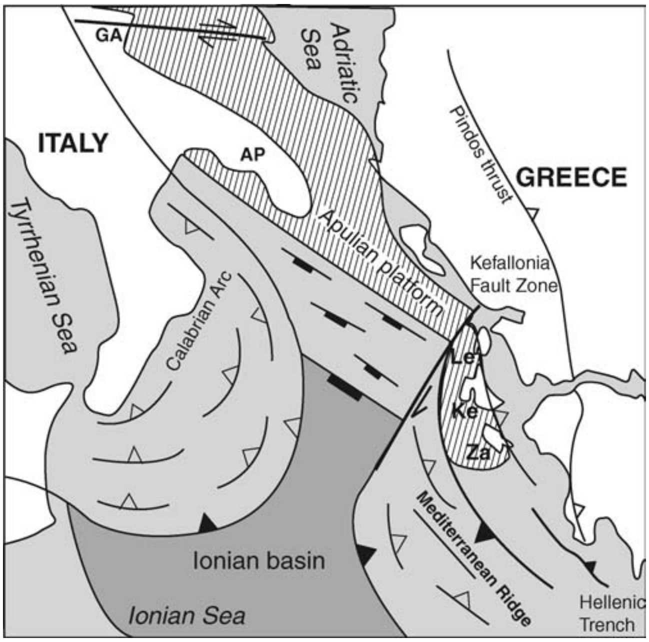


Fig. 2 Schematic map of the central Mediterranean area, modified after Finetti (1982), Underhill (1989) and Le Pichon et al. (2002), showing the configuration of the Apulian platform with respect to the Calabrian and Hellenic arcs. *Ap* Apulian Platform, *Ga* Gargano Peninsula, *Ke* Kefallonia, *Le* Levkas, *Za* Zakynthos

clockwise rotation of the area, to allow a better comparison between the apparently contradicting processes that affected western Greece and the central and southern Aegean during the Neogene. To this end, we have carried out detailed geohistory analyses of the Neogene sedimentary record of western Greece and combined these with the structural history of the area.

Geology of western Greece: a review

The nappes and nappe-bounding thrusts

Western Greece—running from the border with Albania to the southern Peloponnesos and from the Ionian Islands to eastern Thessaly and Makedonia (Fig. 1)—is characterized by a pile of NW–SE striking nappes separated from one another by major west-directed thrust faults. These nappes include the (Sub-) Pelagonian, the Pindos, the Gavrovo-Tripolitza, the Ionian and the Pre-Apulian units, each with its own distinct lithostratigraphy (Aubouin 1957; Jacobshagen 1986) (Figs. 3, 4). Each nappe consists of a Mesozoic to early Tertiary pelagic sediments, overlain by a clastic sequence of clays and turbiditic sandstones (flysch) that accumulated during nappe stacking in front of the active nappe-bounding thrust (Richter et al. 1978). The progressively younger ages of the flysch units from NE to SW thus reflect the southwestward propagation of nappe stacking.

North of the Kastaniotikos Fault (Fig. 1), only the Pindos flysch is exposed, but the pelagic sequence of the Pindos unit reappears in northern Albania (Meco and Aliaj 2000). The Valkano thrust (Figs. 1 and 4) emplaces Pindos rocks on top of Tripolitza flysch in the south, but on Pindos rocks in the north, indicating that it overprints an earlier (thrust-) contact between the Pindos and Tripolitza units (Fig. 4; compare profiles BB′ with CC′ in Fig. 5).

The Apulian platform and Ionian basin

The Apulian platform presently underlies the Adriatic Sea and the northern part of the Ionian Sea, and it crops out on the Gargano and Apulia peninsulas of southeastern Italy (Fig. 2). The Pre-Apulian zone is interpreted as the slope of the Apulian platform and it is exposed in southwestern Albania and on the Ionian Islands (Figs. 1, 2, 6, 7). The Apulian platform and its slope probably do not extend much further south than Zakynthos (Finetti 1982; Underhill 1989) (Fig. 2). The transition from carbonate to clastic sedimentation in the Pre-Apulian parts of Kefallonia and Levkas occurred in the Burdigalian (20.4–16.0 Ma: Lourens et al. 2004) or Langhian (16.0–13.6 Ma: Lourens et al. 2004). On Zakynthos, this transition was younger, around the Middle-Late Miocene transition (i.e. around 12–11 Ma) (Fig. 3). Marine sedimentation continued into the Pleistocene.

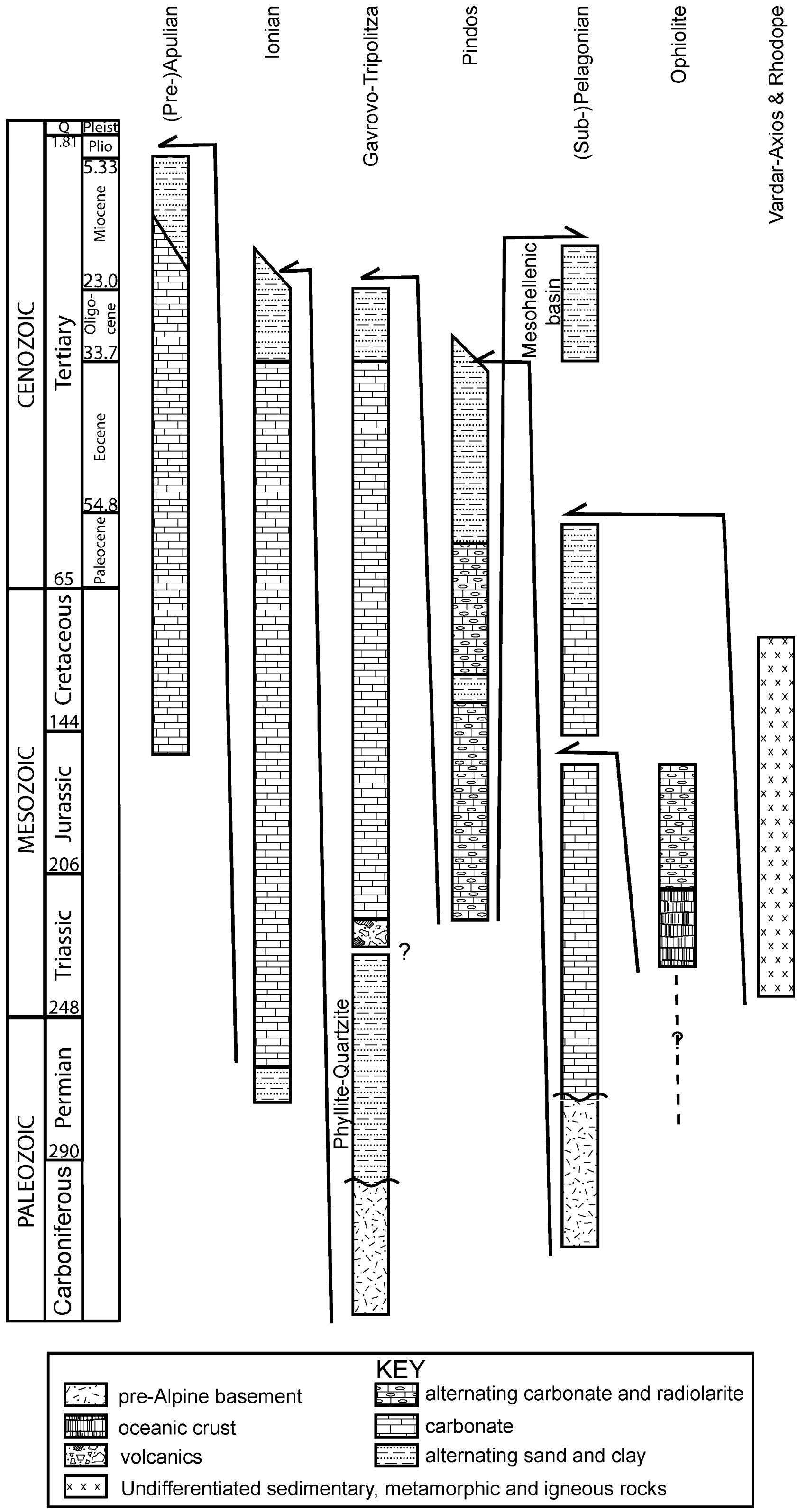
Offshore of southwestern Greece, however, lithosphere of the Ionian basin underthrusted below Greece. The sediments on top of the subducting plate have been folded in the Mediterranean Ridge (Finetti 1976; Le Pichon et al. 2002) (Fig. 2). The onset of deformation of the Mediterranean Ridge occurred between the base of planktonic foraminiferal zone N6 and the top of nannofossil zone NN5 (Kastens 1991), i.e. between 17.59 and 13.53 Ma (Lourens et al. 2004).

Post-nappe stacking deformation and sedimentation

Mesohellenic basin

The Pindos and (Sub-) Pelagonian units are unconformably overlain by sediments of the Mesohellenic basin (Fig. 1). This basin developed in front of east-directed thrusts since the Late Eocene (Fig. 1, Profile A–A′ in Fig. 5) (Zygojannis and Müller 1982; Barbieri 1992a, b; Doutsos et al. 1994; Kontopoulos et al. 1999; Meco and Aliaj 2000; Avramidis et al. 2002; Zelilidis et al. 2002). We consider these thrusts as ‘back-thrusts’, antithetic to the main Pindos Thrust and the Mesohellenic basin can be envisaged as ‘retroforedeep’, comparable in setting and size to the Aquitanian basin north of the Pyrenees described by Beaumont et al. (2000).

Fig. 3 Schematic lithostratigraphy of the nappes and timing of nappe emplacement, based on IGRS-IFP (1966), De Mulder (1975), Fleury (1975), Meulenkamp (1982, 1985), Pe-Piper (1982), Baumgartner (1985), Jacobshagen (1986), Kowalczyk and Dittmar (1991), Meulenkamp and Hilgen (1986), Thiébault et al. (1994) and Degnan and Robertson (1998)



The Klematia–Paramythia basin

The Ionian zone was subdivided into the internal, middle and external Ionian zone (IGRS-IFP 1966) (Fig. 4). After Late Oligocene to the earliest Miocene phase of emergence and erosion, a NE-dipping normal fault led to the Early Miocene foundering of the Klematia–Paramythia half-graben. Extension and subsidence lasted at least until 17 Ma (IGRS-IFP 1966; Avramidis et al. 2000; Van Hinsbergen et al. 2005b).

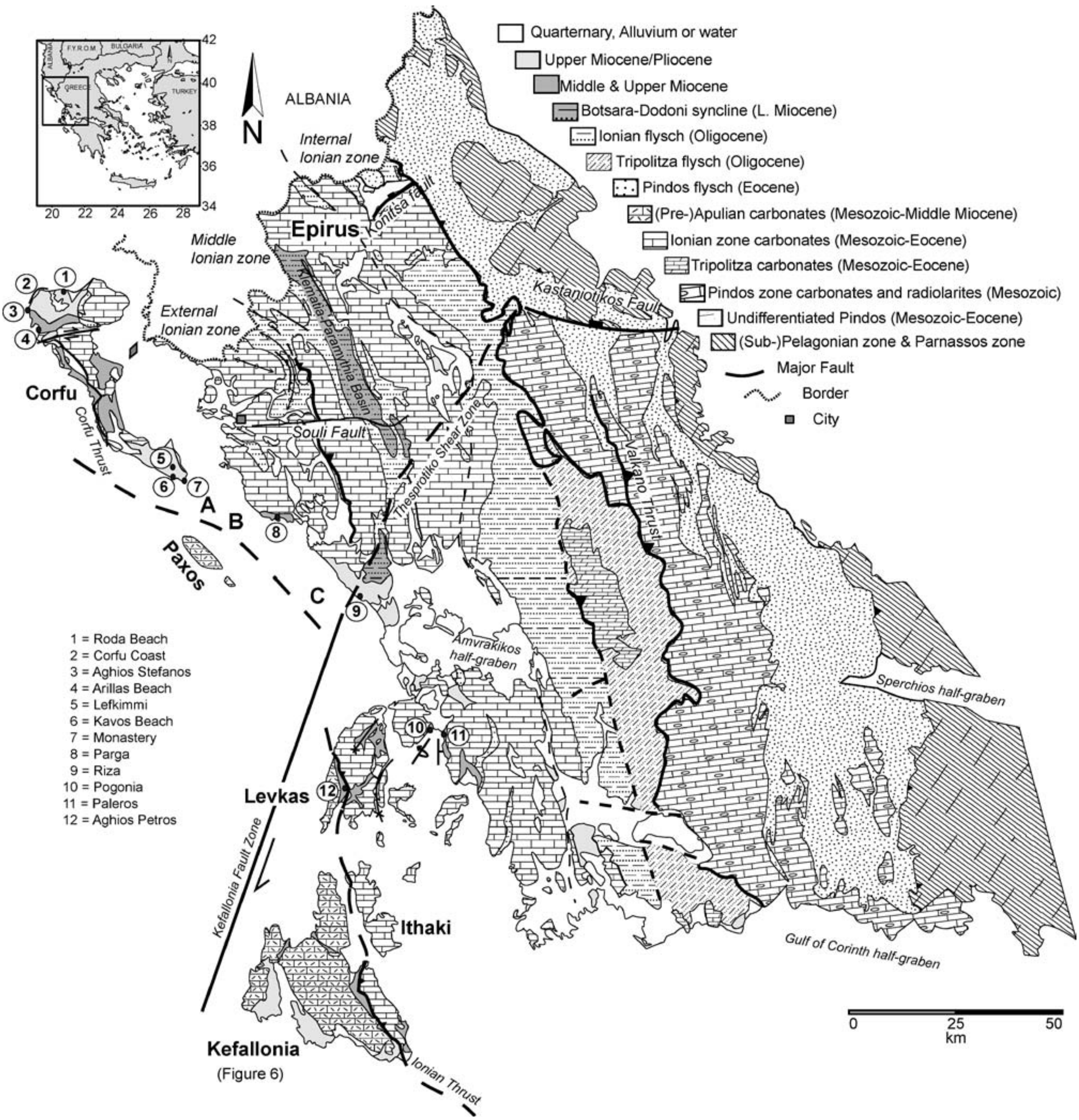


Fig. 4 Geological map of northwestern Greece, modified after Bornovas and Rontogianni-Tsiabaou (1983), with sample locations. *KPB* Klematia-Paramythia basin, *KF* Konitsa Fault AM. For key, see Fig. 1

The South Aegean core complex

The nappe stack on the Peloponnesos is cross-cut by the western part of the South Aegean core complex, which exposes two tectonostratigraphic units that are not exposed in the north: the high-pressure, low temperature (HP/LT) metamorphic *Phyllite Quartzite* and underlying *Plattenkalk* units (Fig. 1, 8). These units are separated from the Tripolitza and higher units by an extensional detachment (Jolivet et al. 2003) (Fig. 8, profiles DD′, EE′, and GG′ in Fig. 5). The Plattenkalk unit was interpreted as the underthrusted equivalent of the Ionian unit (Bizon and Thiébault 1974; Jacobshagen 1986; Dittmar and Kowalczyk 1991; Kowalczyk and Dittmar 1991; Kowalczyk and Zügel 1997) and the Phyllite Quartzite unit was interpreted to represent passive margin clastics, originally (un)conformably overlain by the Tripolitza unit (Van Hinsbergen et al. 2005b). The Phyllite Quartzite and Plattenkalk units underthrusted to a depth of tens of kilometres in the course of the Oligocene and exhumed in the course of the Early Miocene (Jolivet et al. 1996; Thomson et al. 1998; Ring

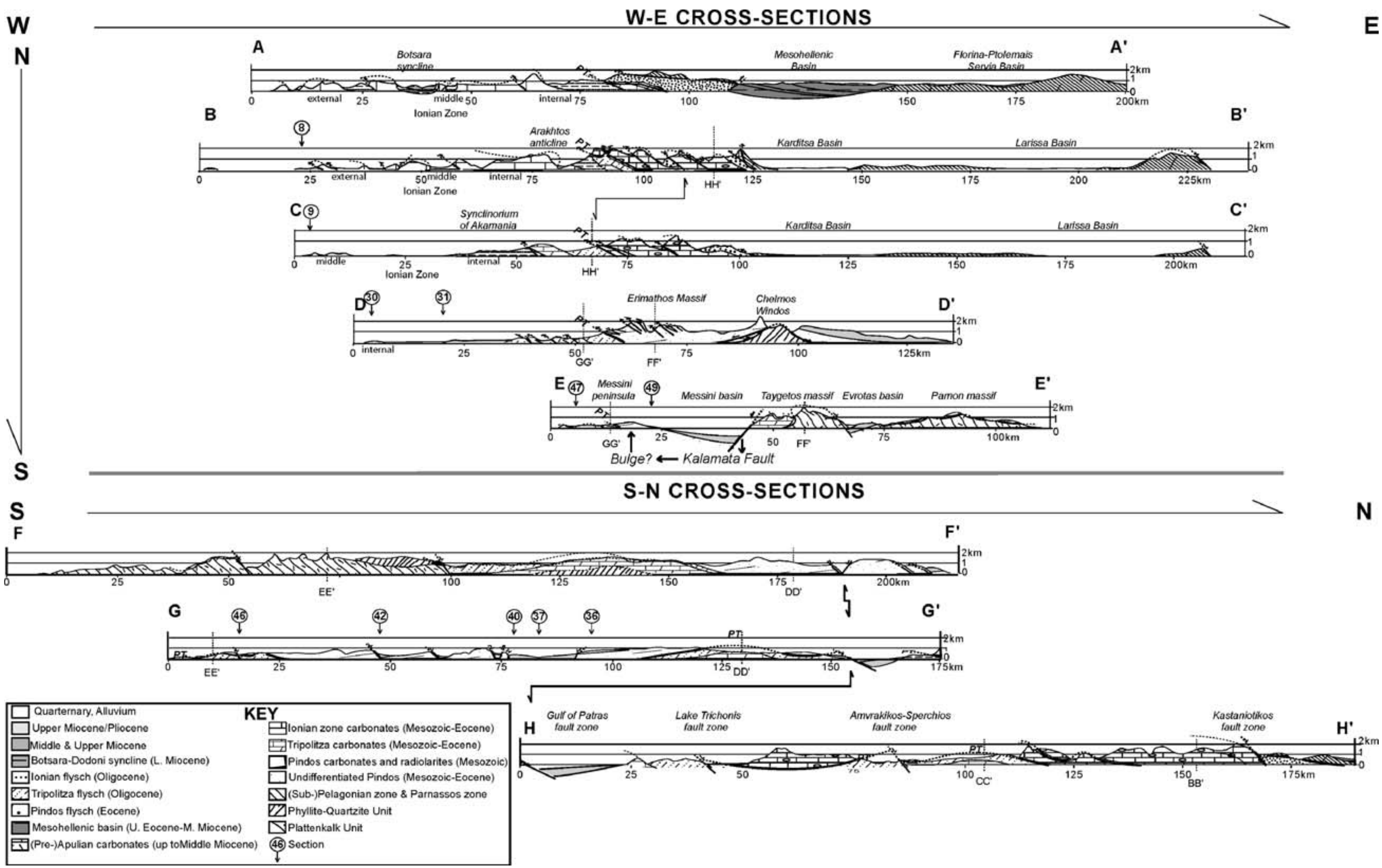


Fig. 5 Diagrammatic cross-sections constructed based on the geological map of Bornovas and Rontogianni-Tsiabaou (1983). The profiles are indicated on the map of Fig. 1. Profile E-E' is partly based on the seismic profiles of Papanikolaou et al. (1988)

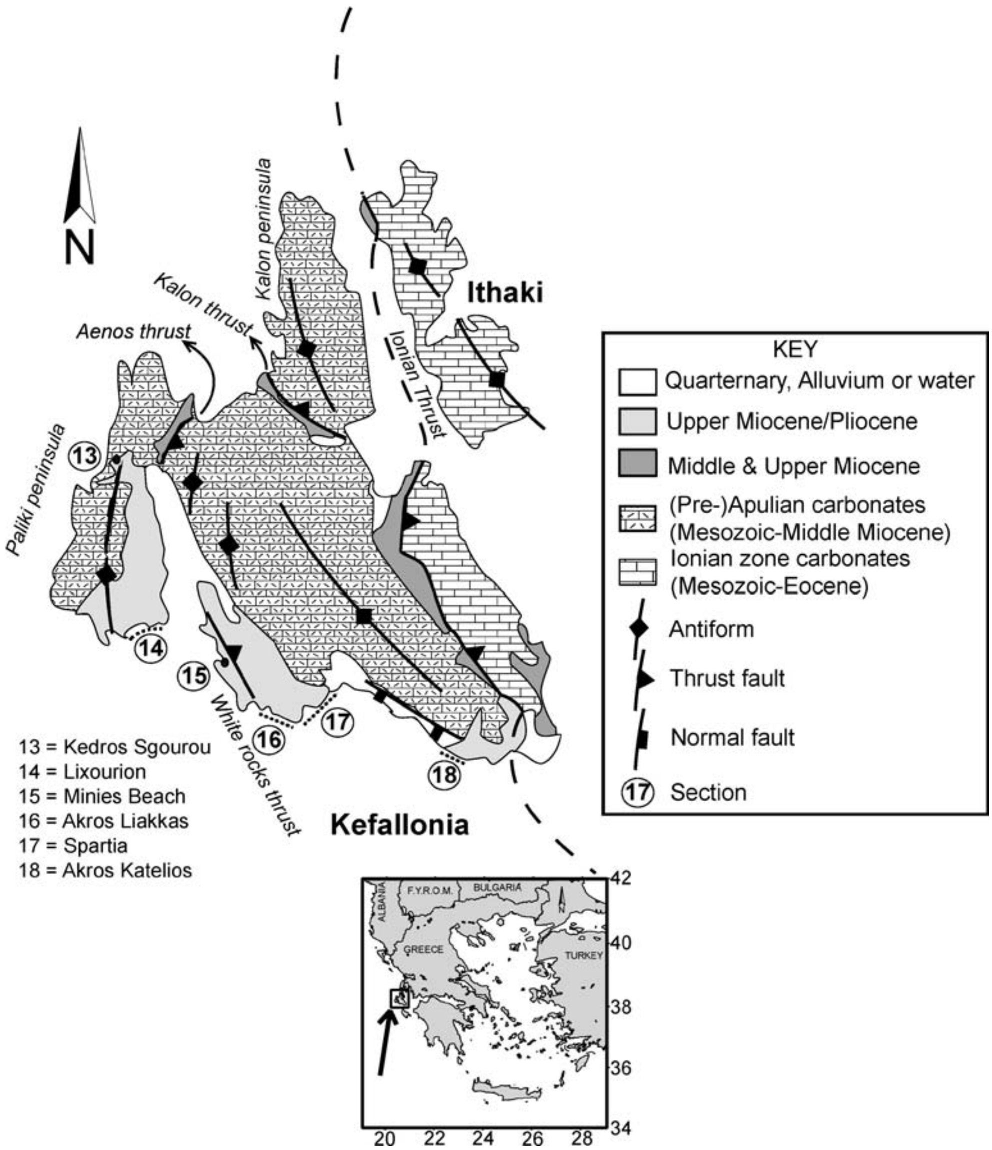


Fig. 6 Geological map of Kefallonia, modified after Bornovas and Rontogianni-Tsiabaou (1983) and Underhill (1989), with sample locations

et al. 2001) (Figs. 3 and 8). After the exhumation of the metamorphic units on the Peloponnesos, the detachment plane was deformed by west-verging folds (Profile DD′ and EE′, Fig. 5) (Doutsos et al. 2000; Xypolias and Doutsos 2000), with the fold axes plunging to the north (Fig. 8).

Transverse structures and late Neogene extensional basins

Thessaly and Makedonia In Thessaly and Makedonia, three extensional basin systems evolved in the course of the Late Miocene and Pliocene: the Florina–Ptolemais–Servia, the Larissa and the Karditsa basins (Fig. 1).

NE–SW extension formed the lacustrine Florina–Ptolemais–Servia basin since 8 Ma (Pavlides and Mountrakis 1987; Pavlides et al. 1998; Steenbrink 2001). The basin system is crosscut by the NE–SW trending Aliakmon Fault Zone (Doutsos and Koukouvelas 1998), which is in its latest motion a dip-slip normal fault. However, we observed slickensides on the Servia Fault (Fig. 1), which suggest that the dip-slip motion was preceded by a strike-slip activity. Lacustrine sediments of about 5 Ma old (Steenbrink et al. 2000) in the footwall of the Aliakmon Fault show no evidence for syn–sedimentary SE-ward tilting, but at present have a SE tilt. This provides a maximum age for significant dip-slip activity of the Aliakmon fault zone of ~5 Ma. South of the Aliakmon Fault Zone, the intramontane Larissa and Karditsa basins formed in response to Late Miocene and Pliocene NE–SW extension, followed by approximately N–S extension in the Late Pliocene and Pleistocene (Caputo and Pavlides 1993; Caputo et al. 1994).

Epirus and Akarnania The Aliakmon Fault Zone may connect in the west to the Kastaniotikos Fault that runs WNW–ESE across the Pindos unit (Figs. 1, 5) The latter fault zone links in the west with the NE–SW trending Thesprotiko Shear Zone (Jordan et al. 2005) that offsets the Lower Miocene of the Klematia–Paramythia basin.

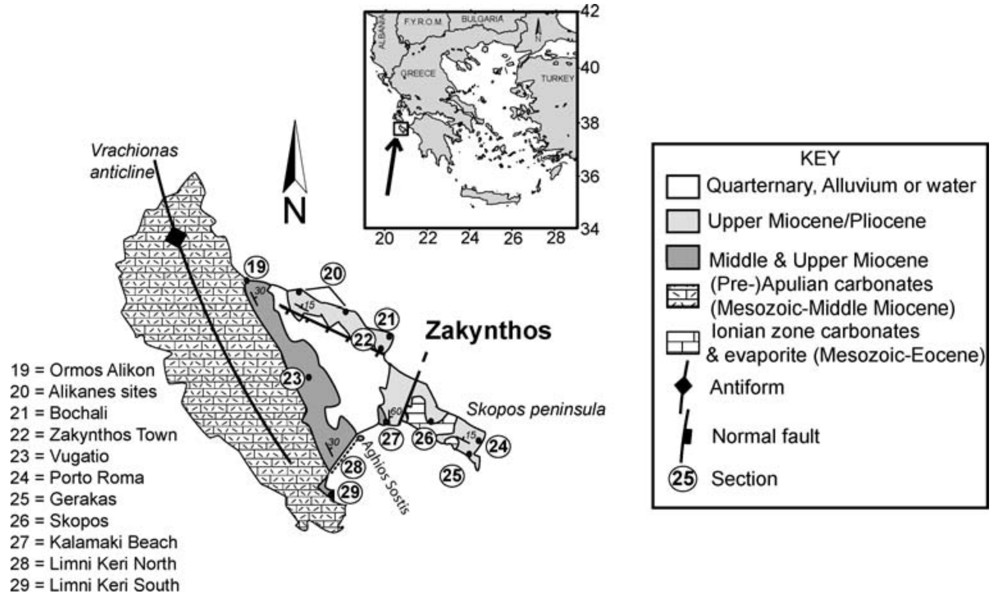


Fig. 7 Geological map of Zakynthos, modified after Bornovas and Rontogianni-Tsiabaou (1983), with sample locations

Contrary to the Thesprotiko Shearzone, its offshore lateral continuation, i.e. the Kefallonia Fault Zone, is seismically highly active (e.g. Scordilis et al. 1985; Kahle et al. 1993; Peter et al. 1998; Sachpazi et al. 2000). On Kefallonia, Levkas and in western Akarnania, the NNW–SSE striking west-directed thrusts and antiforms characteristic for the deformation all over Epirus are overprinted by drag folding along the Kefallonia Fault Zone (Cushing 1985; Underhill 1989; Louvari et al. 1999; Van Hinsbergen et al. 2005c) (Figs. 4, 6).

Finally, northwestern Greece contains a number of E–W and ENE–WSW striking normal faults that have recently been, or still are seismically active, such as the Souli and Konitsa faults (IGRS-IFP 1966; King et al. 1983; Doutsos et al. 1988; Boccaletti et al. 1997; Goldsworthy et al. 2002) (Fig. 1).

Curved extensional basin systems During the Late Plio–Pleistocene, two WNW–ESE striking, large-scale extensional fault zones crosscut mainland Greece which gradually curve eastward into a NNW–SSE direction (Fig. 9). The northern zone created the Gulf of Amvrakikos, connected through a fault zone crosscutting the Pindos mountains with the Sperchios half-graben and Gulf of Evvia (Mariolakos 1976; Clews 1989; Poulimenos and Doutsos 1996; Kranis and Papanikolaou 2001; Goldsworthy et al. 2002) (Fig. 1, profile HH′, Fig. 5). The southern one forms the Gulf of Patras–Gulf of Corinth system (Doutsos and Piper 1990; Armijo et al. 1996; Sorel 2000), which is connected in the east through the Isthmus of Corinth with the Saronic Gulf. The Isthmus of Corinth and the Saronic Gulf were already subjected to extension since the Early Pliocene (Collier and Dart 1991; Van Hinsbergen et al. 2004).

Extensional basins of the Peloponnesos Late Plio–Pleistocene extension resulted in the formation of, among others, the Gulf of Patras, Gulf of Corinth and the Pyrgos, Megalopolis, Evrotas, Kyparissia and Messini basins (Hageman 1977, 1979; Mariolakos et al. 1985; Zelilidis and Doutsos 1992; Van Andel et al. 1993; Poulimenos and Doutsos 1997; Doutsos et al. 2000; Sorel 2000) (Fig. 8). These are (half-) grabens that formed in response to both N–S (e.g. Gulf of Patras, Gulf of Corinth, Pyrgos basin, Molai graben, Kyparissia basin; Tavitian 1994; Stavrakakis 1996; Sorel 2000) and E–W extension (e.g. Messini and Evrotas basins: Papanikolaou et al. 1988; Poulimenos and Doutsos 1997). The profiles EE′, FF′ and GG′ (Fig. 5) illustrate the influence of the combined N–S and E–W extension directions. Additionally, a third curved system is defined by the Pyrgos basin–Megalopolis basin–Evrotas basin system (Figs. 8, 9). The Pyrgos basin formed as a result of N–S extension, whereas the Evrotas basin strikes NNW–SSW were formed as a result of an approximate E–W extension (Poulimenos and Doutsos 1997; Doutsos et al. 2000). The small graben of Molai (Tavitian 1994) and growth faults in the Evrotas basin fill, however, show that during the opening of the Evrotas basin N–S to NW–SE extension was also active. The Messini basin on the SW Peloponnesos (Figs. 5, 8) is bounded by the major west-dipping Kalamata normal fault at the western side of the Mani Peninsula (e.g. Papanikolaou et al. 1988; Mariolakos et al. 1997). The westernmost peninsula of the Peloponnesos is not fault bounded and seems to be a flexural bulge resulting from motion along the west-dipping Kalamata normal fault (profile EE′, Fig. 5).

The Kalamata Fault ends in the north of the plain of Melimala (Figs. 1, 8, 9), where it forms a triple junction with the WNW–ESE trending, N-dipping normal fault of the Kyparissia half-graben (see profile GG′ in Fig. 5) and a less well-developed normal fault zone striking

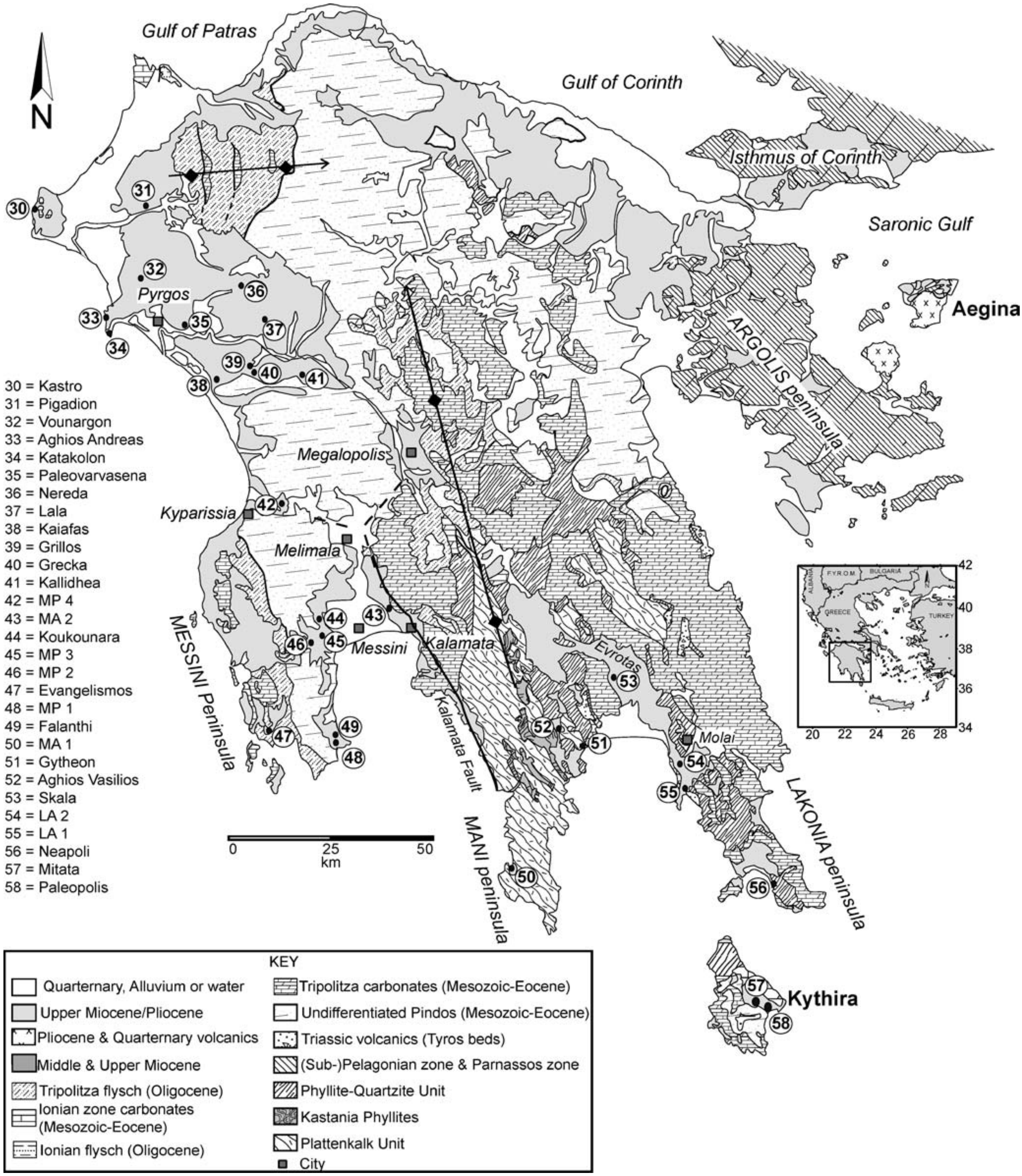


Fig. 8 Geological map of the Peloponnesos, modified after Bornovas and Rontogianni-Tsiabaou (1983), with sample locations

NE–SW (Bornovas and Rontogianni-Tsiabaou 1983). The Lakonia and Argolis peninsulas of the eastern Peloponnesos tilted southwestward in the course of the Late Plio–Pleistocene during NE–SW extension (Mariolakos 1976; Papanikolaou et al. 1988; Van Andel et al. 1993; Poulimenos and Doutsos 1997).

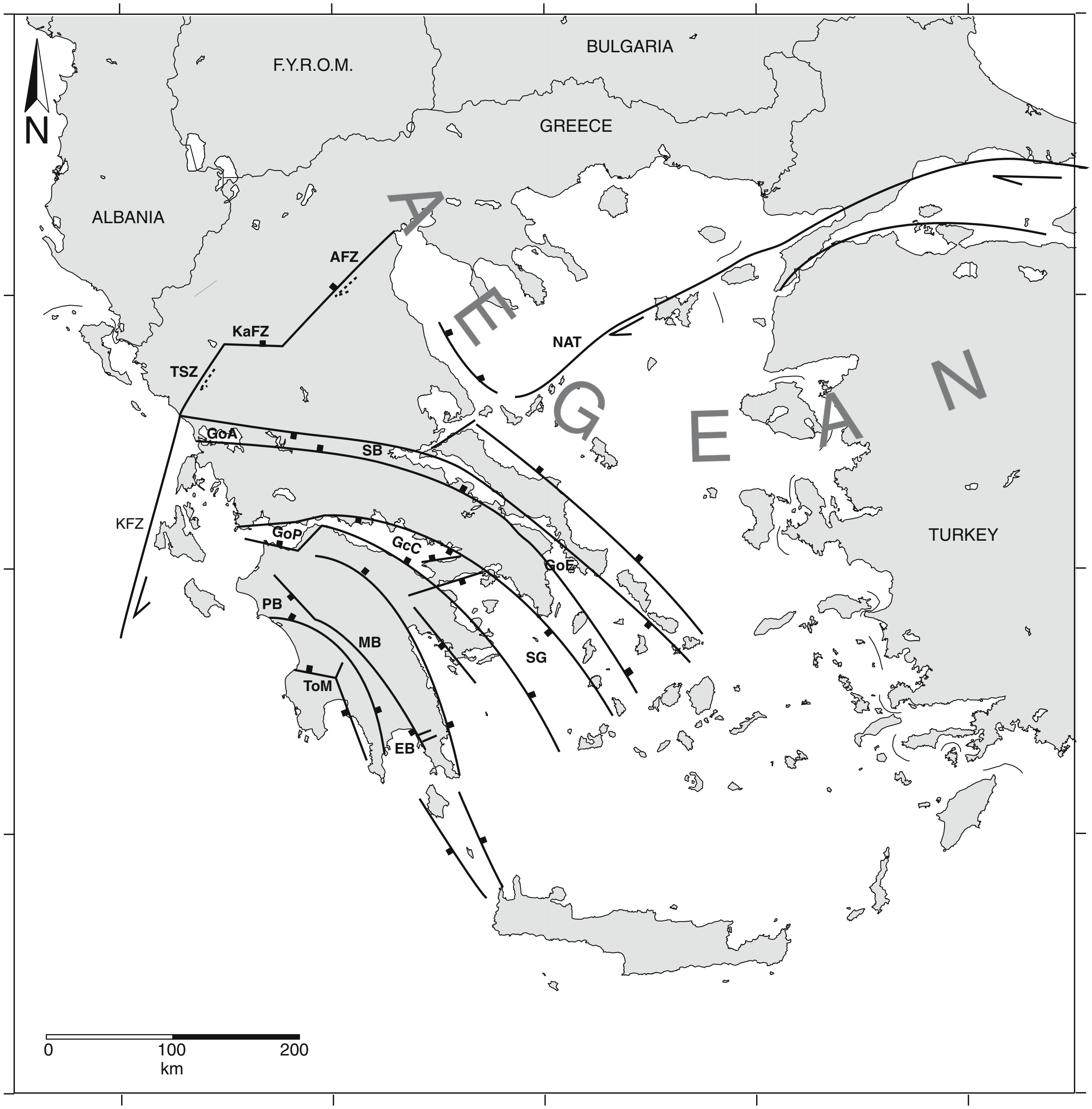


Fig. 9 Tentative interpretation of the curved basin systems of Western Greece. *AFZ* Aliakmon Fault Zone; *EB* Evrotas Basin; *GoA* Gulf of Amvrakikos; *GoC* Gulf of Corinthl; *GoE* Gulf of Evia; *GoP* Gulf of Patras; *KFZ* Kefallonia Fault Zone; *KaFZ* Kastaniotikos Fault Zone; *MB* Megalopolis Basin; *NAT* North Aegean Trough; *PB* Pyrgos; *SB* Sperchios Basin; *SG* Saronic Gulf; *ToM* Triple junction of Melimala; *TSZ* Thesprotiko Shear Zone

Almost all seismically active fault zones are associated with the Late Plio–Pleistocene basins. Therefore, the present-day motion pattern was probably generated in the course of the Pliocene. GPS measurements and earthquake analyses show that rapid southwestward motion occurs south of the Kefallonia Fault Zone nowadays, whereas no motion occurs to its north (King et al. 1983; Billiris et al. 1991; Kahle et al. 2000; McClusky et al. 2000; Jiménez-Munt et al. 2003).

Sampling, age calibration and paleobathymetry analysis

Paleobathymetry estimation and vertical motion analysis

To reconstruct the timing and amount of vertical motions associated with the tectonics of western Greece since the Middle Miocene, a total of 58 sedimentary sites and sections have been sampled (Figs. 4, 6, 7, 8). These were dated by means of bio- and magnetostratigraphy (Figs. 10a, b, c).

To estimate the depositional depth of the sediments, the general relationship between depth and the fraction of planktonic foraminifera with respect to the total foraminiferal population (%P) (Van der Zwaan et al. 1990) was used, following sample selection and counting procedures described in Van Hinsbergen et al. (2005a). If samples with evidence for downslope transport and carbonate dissolution are discarded, the relationship between %P and depth of Van der Zwaan et al. (1990) allows the estimation of the depositional depth of sediments with approximately 50 m of uncertainty at levels of a few hundreds of metres, to 150 m at depths around 1 km. The resolution generally increases with increasing numbers of counted samples. As a measure for the uncertainty, we have indicated the standard deviations of the depth estimates in Table 1.

The percentage of planktonic foraminifera also varies with oxygenation fluctuations (Van Hinsbergen et al. 2005a). Therefore, all samples with a fraction of oxygen stress markers amongst the benthic population (%S) exceeding 60% were discarded. Generally, %S-values were much lower and no large %S variations were found in the sections. This is in line with the geological setting of western Greece since the Middle Miocene, where sedimentation occurred in wide, deep and well-ventilated (foreland) basins. Some samples contain high fractions of quartz and rock fragments, which at deep marine levels probably result from downslope transport. These samples were discarded, or considered to give a minimum depth value.

Only sediments of Corfu Coast (2) show a significant bathymetry change during their deposition. Therefore, for all sections except Corfu Coast (2), depth estimates were averaged (Table 1). The depth was independently checked by means of the presence or absence of benthic depth markers (for discussion and list, see Van Hinsbergen et al. (2005a). The results are included in Table 1, and are generally in line with the calculated depth values.

The bathymetry is the resultant of sedimentary infill of the basin, eustatic sea level changes and tectonics. The lack of astronomically tuned ages has made detailed correction for eustatic sea level changes (of the order of tens of metres) impossible. The motion of a chosen stratigraphic level through time is obtained by adding the thickness of accumulated sediment to the paleobathymetry trend. This has been carried out for Corfu, Kefallonia and Zakynthos (Fig. 11).

Sampling, age dating and paleobathymetry estimation

Mainland Greece

Langhian to lower Tortonian (Tortonian: 11.6–7.2 Ma: Lourens et al. 2004) sediments around Parga (8), dated by IGRS-IFP (1966), overthrusted by Ionian carbonates (Fig. 4) show a paleobathymetry around 1,000 m (Table 1). In addition, Tortonian siltstones and turbiditic sandstones were sampled in Pogonia (10) and Paleros (11; Figs. 4, 10a). According to the regional dip, Paleros (11) is the younger continuation of Pogonia (10). The nature of the contact of the sequence with, and the age of the underlying sediments around Paleros and Pogonia is unknown. Both sections were deposited around 500–600 m depth (Table 1). Finally, Riza (9) contains Lower Pliocene alternating marls and sapropels (Figs. 4, 10a). Although no longer exposed, IGRS-IFP (1966) reported Messinian gypsum at the base of this section (Messinian: 7.2–5.3 Ma: Lourens et al. 2004). The depositional depth of Riza is 1,000 m or more (Table 1). IGRS-IFP (1966) reported Tortonian sediments around Riza with *Cibicides italicus*, indicative of a paleobathymetry around 1,000 m (Van Hinsbergen et al. 2005a). Deep marine conditions thus, already prevailed since the Late Miocene. Upper Plio–Pleistocene terrestrial conglomerates indicate Pliocene uplift.

Corfu

An Upper Miocene–Lower Pliocene clastic series was sampled along the northwestern coast of Corfu. The Tortonian of Arillas Beach (4) was deposited at a depth of 1,000 m or more (Table 1). Upper Miocene–Lower Pliocene sediments were sampled in Aghios Stefanos (3) and comprise fluvial and lacustrine deposits, correlated to the Lago Mare phase of the terminal Messinian (cf. Krijgsman et al. 1999). The overlying Lower Pliocene Trubi facies (Fig. 10a) was deposited at a depth of around 1,000 m (Table 1). It underlies the Lower Pliocene of Corfu Coast (2) (Fig. 10a) and the overlying section Roda Beach (1). Corfu Coast (2) and Roda Beach (1) reveal a strong shallowing trend from a depth of 1,000 to 200–300 m (Table 1). Adding the amount of accumulated sediment shows that the shallowing is initially accompanied by ongoing subsidence, and from approximately 4.5 Ma onward by uplift (Fig. 11).

The Lower Pliocene Trubi facies in the southwest of the island is deformed in the footwall of the west-directed Corfu Thrust (Fig. 4), a structure that was earlier described by Doutsos and Frydas (1994). In the south, samples were collected from Lower Pliocene clays and siltstones (Kavos Beach (6), Monastery (7) and Lefkimmi (5) Figs. 4 and 10a). The Lower Pliocene of Monastery (7) contains reworked Miocene—up to Tortonian—species of planktonic foraminifera. The paleobathymetry of Monastery (7) is estimated to be around 500–600 m, with large error bars because of reworking and downslope transport. Field observations suggest that section Kavos Beach (6) underlies Monastery (7). Its lithology is comparable to that of Monastery (7) and it is thus likely younger than the Lower Pliocene Trubi facies. The depositional depth of Kavos Beach (6) is estimated at 500–750 m.

Table 1 Calculated and estimated paleobathymetry values for all sites and sections

Code	Locallity	Samples (Gr code)	*N*	Depth (m)	SD (m)	Taxonomic estimate	Age
1	Roda Beach	11038–41	2	230	66	200–500	4.59–4.52 Ma
2	Corfu Coast	11381–481	56	Trend		Trend	5.09–4.59 Ma
3	Aghios Stefanos	11086–90c	7	1,064	44	500–900	~5.235–5.09 Ma
4	Arillas Beach	11072–85	8	1,011	80	> 1,000	~8.8–7.9 Ma
5	Lefkimmi	11001; 11034–37	5	393	233	200–300	3.98–3.596
6	Kavos Beach	11250–56	5	596	185	500–750	Early Pliocene
7	Monastery	11042–71	13	598	174	500–600	~4.4–4.2 Ma
8	Parga	11015–16	2	964	16		Middle Miocene
9	Riza	10941–60; 12001	21	905	85	> 1,000	~4.7–4.3 Ma
11	Paleros	12081–115	22	480	171	500–750	Late Miocene
10	Pogonia	12011–80	58	614	111	600–750	late Miocene
12	Aghios Petros (Levkas)	12951–12984	28	1,096	119	1,000 +	Early–late Miocene
13	Kedros-Sgourou	12475–77	2	1,146	129		~6.2–5.5 Ma
13	Kedros-Sgourou	12478–87	10	934	107	600–900	4.52–3.97 Ma
14	Lixourion	12326–70	36	426	232	300–500	2.6–1.6 Ma
15	Minies Beach	12488–95	7	78	17	100–300	< 4.52 Ma
16	Akros Liakkas	12371–88	10	767	237	500–1,000	~6.2–5.5 Ma
16	Akros Liakkas	12389–93	3	750	57	600–900	5.33–5.236 Ma
16	Akros Liakkas	12394–95	2	184	13	200–300	~4.52 Ma
17	Spartia	12396–422	26	76	45	150–250	4.4–3.5 Ma
18	Akros Katelios	12423–33	11	70	20	50–150	< 4.52 Ma
19	Ormos Alikon	11346–50	5	1,139	77	> 1,000	~7.45–7.25 Ma
20	Alikanes	10687–90	2	678	311	500–750	3.31–2.73 Ma
21	Bochali	11370–80	8	498	134	300–500	~1.3–0.075 Ma
22	Zakynthos Town	11091–113; 11351–70	24	817	190	500–900	1.9–1.6~Ma
23	Vugatio	11261–85	25	1,181	51	> 1,000	8.11–7.77 Ma
24	Porto Roma 1	11301–10a	11	377	73	300–500	0.99 Ma
24	Porto Roma 2	11298–300	3	422	148	300–500	0.88 Ma
24	Porto Roma 3	11294–97	4	82	19	200–300	0.70 Ma
25	Gerakas	11311–20	10	661	108	500–750	~1.8 Ma
26	Skopos	10680–81	2	1,054	9		5.05–4.12 Ma
27	Kalamaki Beach	10671; 11122–75	23	1,002	143	750–1,000	5.33–4.82 Ma
28	Limni Keri North	11321–45	11	1,176	46	900–1,000	7.24–6.78
29	Limni Keri South	11176–89	14	1,036	53	> 1,000	8.865–8.566 Ma
30	Kastro	10661–662	1	36			Late Plio–Pleistocene
31	Pigadion 1*	HH 703 A-M	11	37	1	0–50	Late Plio–Pleistocene
31	Pigadion 2*	HH 703 P-FF	6	38	2	0–50	Late Plio–Pleistocene
31	Pigadion 3*	HH 703 GG-OO	17	38	3	0–50	Late Plio–Pleistocene
32	Vounargon*	HH 700 A-U	14	98	11	50–100	Late Plio–Pleistocene
33	Aghios Andreas*	HH 320 A-G	7	51	6	0–50	Late Plio–Pleistocene
34	Katakolon*	HH 315 A-K	10	80	30	200–300	Late Plio–Pleistocene
35	Paleovarvasena*	HH 109 B-L	7	47	9	50–100	Late Plio–Pleistocene
36	Nereda*	HH 310 A-L	11	39	1	0–50	Late Plio–Pleistocene
37	Lala*	HH 2 A-N	13	44	12	50–100	Late Plio–Pleistocene
38	Kaiafas*	HH 313 A-G	7	43	4	0–100	late Plio–Pleistocene
39	Grillos*	HH 723 A-H	5	38	3	0–50	Late Plio–Pleistocene
40	Grecka*	HH 300 A-F	6	75	38	150–300	Late Plio–Pleistocene
40	Grecka II*	HH 305 A-F	5	46	12	50–150	Late Plio–Pleistocene
41	Kallidhea*	HH 301 A-V	20	53	22	50–100	Late Plio–Pleistocene
42	MP 4	10611	1	94	–	150–300	Late Plio–Pleistocene
43	Koukounara	10616	1	112	–		Late Plio–Pleistocene
44	MP 3	10622–22a	2	158	16	500–600	Late Plio–Pleistocene
45	MP 2	10619–21	3	340	256		NN18-NN19b
46	Evangelismos	10610	1	227	–		Late Plio–Pleistocene
47	MP 1	10603; 10607–08	3	372	131	500–600	NN14
48	Falanthi	10607–07a	1	283	77		NN14
49	MA 2	10625	2	116	34	0–50	NN19
50	MA 1	10655–56	2	116	34	300–500	NN15
51	Gytheon	10627–31	2	36	0		Late Plio–Pleistocene
52	Aghios Vasilios	10632–33	2	39	1		Late Plio–Pleistocene
53	Skala	10639	2	Sterile	–		Late Plio–Pleistocene
54	LA 2	10652	1	80	–	300–500	NN19b
55	LA 1	10647	1	47	–		Late Plio–Pleistocene
56	Neapoli	10641–42	1	36	–		Late Plio–Pleistocene
57	Mitata	Gr 2531–35	5	39	3	0–50	Tortonian
58	Paleopolis	Gr 2536–88	52	Trend			Middle–late Pliocene

Code refers to numbers in Figs. 5, 6, 7 and 8 *N* number of samples averages, *SD* standard deviation

The youngest Pliocene sediments of Corfu are exposed in small outcrops around Lefkimmi (5) and have an age between 3.98 and 3.60 Ma (Fig. 10a). Its depositional depth is significantly shallower than that of nearby Monastery (7), only about 200–300 m.

Taking sediment accumulation into account, the data suggest that on southern Corfu initial subsidence was followed by shallowing due to infill, without significant vertical motion (Fig. 11a).

Levkas

From Levkas, our analyses show that the depositional depth of the Lower to Upper Miocene section of Aghios Petros (12), dated by Bizon (1967) was 1,000 m or more (Table 1). Steininger et al. (1985a, b) show that clastic turbidite deposition occurred frequently in the section from the lower Tortonian onward. Meulenkamp (1982) reported strong, submarine folding (not slumping) on Levkas indicating a late Burdigalian or early Langhian (i.e. approximately 16 Ma) phase of compression.

Kefallonia

The stratigraphy of the Pre-Apulian unit of Kefallonia contains a thick Mesozoic to Paleogene series of carbonates, overlain by an Oligocene to earliest Tortonian series of folded, very deep-marine marls and interbedded turbiditic calcareous sandstones (Bizon 1967; B.P. Co. Ltd. 1971; De Mulder 1975). Along the southern coast of the island, Akros Liakkas (16) exposes a ~50°E dipping series of at least 100 m of Messinian clays, alternating with mass-transported conglomerate and sand beds, overlain by a 100-m thick series of Messinian evaporites and Lower Pliocene Trubi facies sediments and marls (Fig. 10b).

Akros Liakkas (16) is angularly unconformably overlain by a series of gently dipping Upper Pliocene silts, sands and calcarenites (Spartia (17), Fig. 10b). The cumulative thickness of the Spartia (17) is more than 300 m (Fig. 10b), but this thickness is largely the result of lateral sampling along a prograding sedimentary system. The paleobathymetry of both the pre-evaporite clastic sequences and the Lower Pliocene Trubi facies is deep marine (around 700–800 m), whereas the sediments of Spartia (17) were deposited at much shallower depth: 200–300 m above the first unconformity and between 0 and 100 m (Table 1) above the second one.

The shallow marine clastics of Akros Katelios (18) are, at least in part, time equivalent to Spartia (17) (Fig. 10b). Another series of probably Late Pliocene age was sampled in the section Minies Beach (15) (Figs. 7, 10b). These shallow marine sediments were strongly—probably synsedimentarily—tilted in the footwall syncline of the W-directed White Rocks Thrust (see also Underhill 1989) (Fig. 6). In summary, rapid Early Pliocene uplift affected southern Kefallonia and led to emergence and erosion, followed by the deposition of Upper Pliocene shallow marine deposits of Spartia and Akros Katelios (Fig. 11).

A remarkably different vertical motion history is reconstructed from the Paliki peninsula (Figs. 6, 10b). Kedros-Sgourou (13) is largely time-equivalent with Akros Liakkas (16). The Messinian basal part of the section is overlain by Lower Pliocene Trubi facies (Fig. 10b). Some small isolated outcrops of Messinian gypsum are exposed near Kedros-Sgourou. The paleobathymetry analysis shows that the Trubi facies of Kedros-Sgourou (13) was deposited around 900 m depth (Table 1).

The Upper Pliocene to Lower Pleistocene sections of Lixourion (14) were deposited between 300 and 500 m. The uplift of the Paliki peninsula, therefore, occurred much later (in the Pleistocene) than in the Akros Liakkas area, and there is no evidence for rapid Early Pliocene uplift. Instead, Trubi facies sedimentation continued throughout most of the Early Pliocene (Figs. 10b, 11).

Zakynthos

Most of our sections from Zakynthos (Figs. 6, 10c) were sampled by Duermeijer et al. (1999) from the Upper Miocene to Pleistocene terrigeneous clastic part of the Pre-Apulian unit and the Plio–Pleistocene on top of the Ionian unit. The stratigraphy and age calibration of the sections are shown in Fig. 10c.

The oldest sediments sampled have a Tortonian age (Fig. 10c) and contain alternating clays and sapropels [Vugatio (23), Ormos Alikon (19) and Limni Keri South (29)]. The paleobathymetry exceeds 1,000 m (Table 1). Limni Keri North (28) covers a lower Messinian sequence of alternating clays, turbiditic sandstones and large slumps (previously described as post-sedimentary chevron folds by Underhill 1989; Figs. 6, 10c). The paleobathymetry indicates depths of 900–1,000 m (Table 1). This clastic series is overlain by Messinian gypsum turbidites, presently exposed near the harbour of Aghios Sostis (see also Dermitzakis 1978). The sections in the Upper Miocene are all tilted to the east in the limb of the Vrachionas anticline (Fig. 7). Along Kalamaki Beach (27; Fig. 7), alternating Messinian gypsum and sterile fine clastics of the Lago Mare stage of the Messinian Salinity Crisis are overlain by Lower Pliocene Trubi facies and marls in section Kalamaki Beach (27) (Fig. 10c). [Halfway the poorly exposed interval between Aghios Sostis and Kalamaki Beach (27), we collected one sample from an exposure along the beach with a Messinian age.] The paleobathymetry of the Messinian part of Kalamaki Beach (27) is indeterminate. The Pliocene part was deposited at 750–1,000 m depth.

Kontopoulos et al. (1997) recovered Lower Pliocene nannofossils from marls and sandstones with hummocky cross-stratification near Aghios Sostis, with a significantly shallower tilt than the Messinian gypsum. We recovered a variety of Miocene planktonic foraminifera from these strata. These strata, therefore, have an age of Lower Pliocene or younger and result from erosion of the Miocene to possibly Lower Pliocene marine sediments. In the north of the island, 10–15° NE tilted, Upper Pliocene and Pleistocene alternating clays and bioturbated sands were sampled near Alikanes (20) and Bochali (21) and in Zakynthos town (22) (Figs. 6, 10c). The paleobathymetry reveals a shallowing trend: Alikanes (20) and Zakynthos Town (41) yield values of 700–800 m, whereas Bochali (21) was deposited around 500 m.

The Ionian unit on Zakynthos is exposed on the Skopos peninsula. Here, diapirs of Triassic gypsum, previously described by Underhill (1988) are overlain by strongly deformed Messinian gypsum and deep marine Lower Pliocene trubi facies (section Skopos, 26) and clays (see also Anapliotis 1963). In the southeasternmost part of the Skopos peninsula, coarse conglomerates are overlain by Plio–Pleistocene marls that contain thick beds of calcarenite in the top of the succession, previously described by Dermitzakis et al. (1977), Tsapralis (1981) and Triantaphyllou et al. (1997). The depositional environment of these conglomerates is uncertain. Possibly they represent a continental sequence, followed by rapid subsidence and the deposition of the Gerakas clays around 800 m of depth. A deepening trend, however, has not been found and was not reported before. Alternatively, these conglomerates could represent a submarine debris-flow inflicted by the uplift of the nearby diapirs. We analysed samples from Gerakas (25) and Porto Roma (24) (Figs. 7 and 12). The depositional depth of Gerakas (25) was approximately 500–750 m [which is in line with the estimates of Tsapralis (1981) based on ostracoda]. The sites of Porto Roma (24) show evidence for synsedimentary tilting and reveal a shallowing trend from 300–500 to (less than) 200–300 m (Table 1).

In conclusion, folding in the Vrachionas anticline probably occurred during the Late Miocene and Early Pliocene and led to the uplift and emergence of Tortonian and older deposits in western Zakynthos. The Skopos diapir formed in the course of the Pliocene, followed by deposition of the Upper Pliocene regressive sequences along the northern coast and the southeastern Skopos peninsula, representing Pleistocene uplift that affected the entire island (Fig. 11).

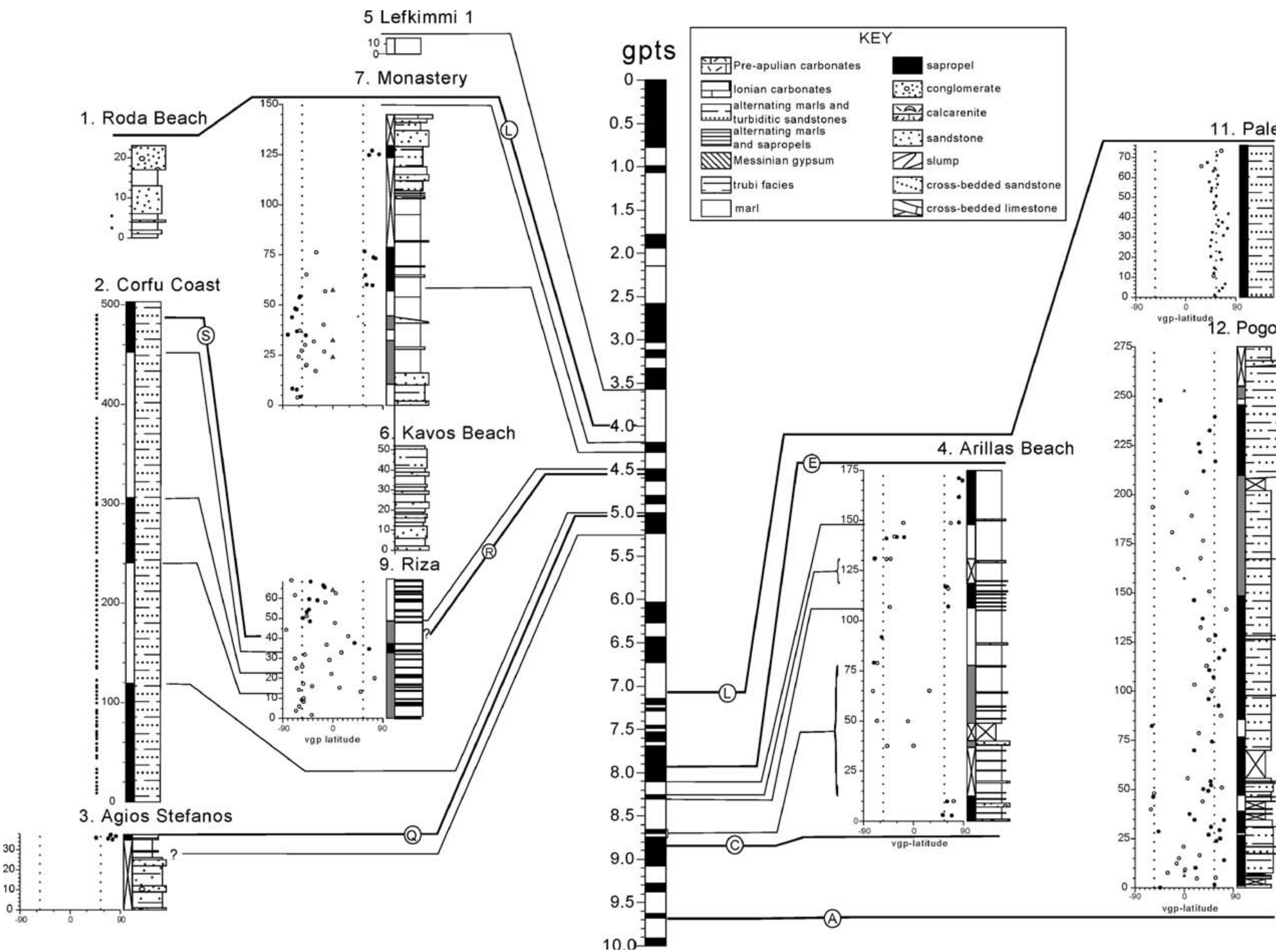
Fig. 10 **a** Detailed stratigraphic correlation of the sections of western Epirus and Corfu; **b** Kefallonia and **c** Zakynthos: Biostratigraphy of Spartia (17) was taken from Hug (1970): the logs, age determinations and magnetostratigraphy are based on Duermeijer et al. (1999). Kalamaki Beach (27) and Limni Keri South (29) were resampled and redated, and additional samples (*open circles*) were collected from Zakynthos Town (22). More detailed stratigraphic information of sections Roda Beach (1), Monastery (7), Riza (9), Pogonia (10), Paleros (11) is given in Appendix I. *Small dot* left of the logs represent sample levels. Paleomagnetic results are represented as virtual geomagnetic polar (*VGP*) latitude, with polarity zone interpretation. *Closed (open)* circles denote projections of (less) reliable ChRM directions. *Small dot* left of the logs represent sample levels. In the polarity column, *white* denotes reversed polarity; *grey* indicates less reliable and inconclusive polarity; *white plus cross* is used for intervals without data. Magnetostratigraphy of the Akros Liakkas, Spartia and Lixourion was reinterpreted from samples previously used for rotation analysis by Duermeijer et al. (2000). Magnetic polarity of three samples of Kedros-Sgourou (13) was reported earlier in Van Hinsbergen et al. (2005b). Numerical ages are taken from Rio et al. (1990), Hilgen et al. (1995, 2000), Backman and Raffi (1997), Lourens et al (1996, 1998, 2004), Sierro et al. (2001) and Raffi et al. (2003). *A* bottom regular sinistral *Neogloboquadrina acostaensis* (9.69 Ma); *B* bottom *Discoaster pentaradiatus* (9.367 Ma); *C* subtop *Catapsydrax parvulus* (8.865 Ma); *D* Cycle G50 of the Gibliscemi section (8.566 Ma); *E* low regular occurrence *Sphaeroidinellopsis seminulina* (7.918 Ma); *F* top *Helicosphaera stalis* (7.610 Ma); *G* subtop *Globorotalia menardii 4* (7.499 Ma); *H* bottom *Globorotalia menardii 5* (7.365 Ma); *J* bottom *Globorotalia conomiozea/miotumida* gr. (7.240 Ma); *K* bottom common *Amaurolithus delicatus* (7.22 Ma); *M* subtop *Globorotalia conomiozea/miotumida* gr. (6.504 Ma); *N* top *iscoaster quinquiramus* (5.537 Ma); *O Sphaeroidinellopsis acme* (5.30–5.21 Ma); *P* bottom *Globorotalia. margaritae* (5.08 Ma); *Q* bottom *Ceratolithus rugosus* (5.05 Ma); *R* bottom *Helicosphaera sellii* (4.52 Ma); *S* bottom *Globorotalia ruber* (4.52 Ma); *T* top *Globorotalia margaritae* (3.98 Ma); *U* top *Amaurolithus tricornicolatus* (3.97–4.12 Ma); *V* top *Globorotalia bononiensis* (2.39 Ma); *W* bottom *Globorotalia bononiensis* (3.31 Ma); *X* subtop *Discoaster tamalis* (2.80 Ma); *Y* bottom *Globorotalia inflata* (2.09 Ma); *Z* bottom *Globorotalia truncatulinoides* (1.88 Ma); *ZA* bottom common left-coiling *Neogloboquadrina acostaensis* (1.79 Ma); *ZB* bottom *Hyalinea balthica* (1.49 Ma); *ZC* disappearance of left-coiled *Neogloboquadrina* sp. (1.37 Ma; *ZD* reappearance of left-coiled *Neogloboquadrina* sp. (1.22 Ma); *ZE* top common left-coiled *Neogloboquadrina* sp. (0.61 Ma)

Strophades

It is not known which nappe underlies the islets of the Strophades south of Zakynthos (Fig. 1), because only Messinian gypsum and Lower Pliocene marls are exposed on the islet. These were sampled and analysed by Lyberis and Bizon (1981). They reported the presence of *Uvigerina* spp. and *Siphonina reticulata* from these marls, which are indicative of a depth exceeding 300 m (Van Hinsbergen et al. 2005a).

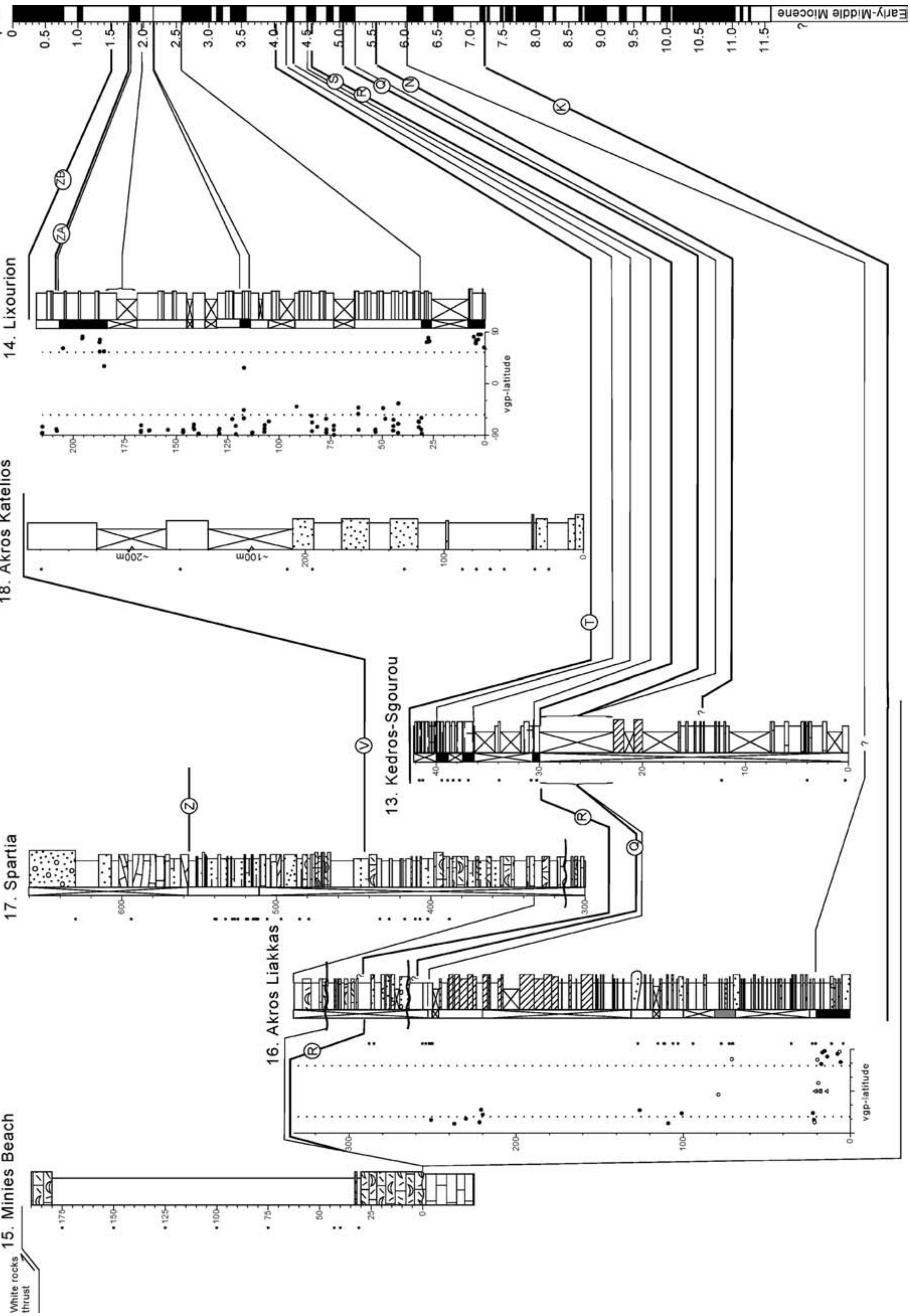
Peloponnesos

The Neogene sediments on the Peloponnesos have been dated earlier, mainly based on calcareous nannofossils by Hageman (1977; 1979), Piper et al. (1982), Frydas (1987, 1989, 1990, 1993) and Frydas and Bellas (1994). All marine sediments were assigned a Plio–Pleistocene age. Hageman (1979) reported large amounts of Middle Miocene nannofossils from the Upper Pliocene, indicating that a marine Middle Miocene cover has probably been present near this part of the Peloponnesos. The oldest deposits from which biostratigraphic information is available are found in the area around Falanthi (sites 45 and 46, Fig. 6). There, sediments overlying lignites contain a calcareous nannofossil assemblage indicating an age range between 5.28 and 4.12 Ma (Frydas 1990; Lourens et al. 2004). These Lower Pliocene sediments show evidence for a paleobathymetry as deep as 500–600 m. The lignites were described by Antoniadis et al. (1992), who assigned them to the Lower Pliocene, although conclusive evidence for this age is lacking. A



1. Roda Beach
2. Corfu Coast
3. Agios Stefanos
4. Arillas Beach
5 Lefkimmi 1
6. Kavos Beach
7. Monastery
9. Riza
11. Paleros
12. Pogonia
gpts
KEY
Pre-apulian carbonates
Ionian carbonates
alternating marls and turbiditic sandstones
alternating marls and sapropels
Messinian gypsum
trubi facies
marl
sapropel
conglomerate
calcarenite
sandstone
slump
cross-bedded sandstone
cross-bedded limestone
vgp latitude
vgp-latitude

Fig. 10a



apts
Early-Middle Miocene
14. Lixourion
18. Akros Katelios
17. Spartia
13. Kedros-Sgourou
16. Akros Liakkas
15. Minies Beach
White rocks thrust
vgp-latitude
~200m
~100m
ZB
ZA
S
R
Q
N
K
T
V
Z

Fig. 10b

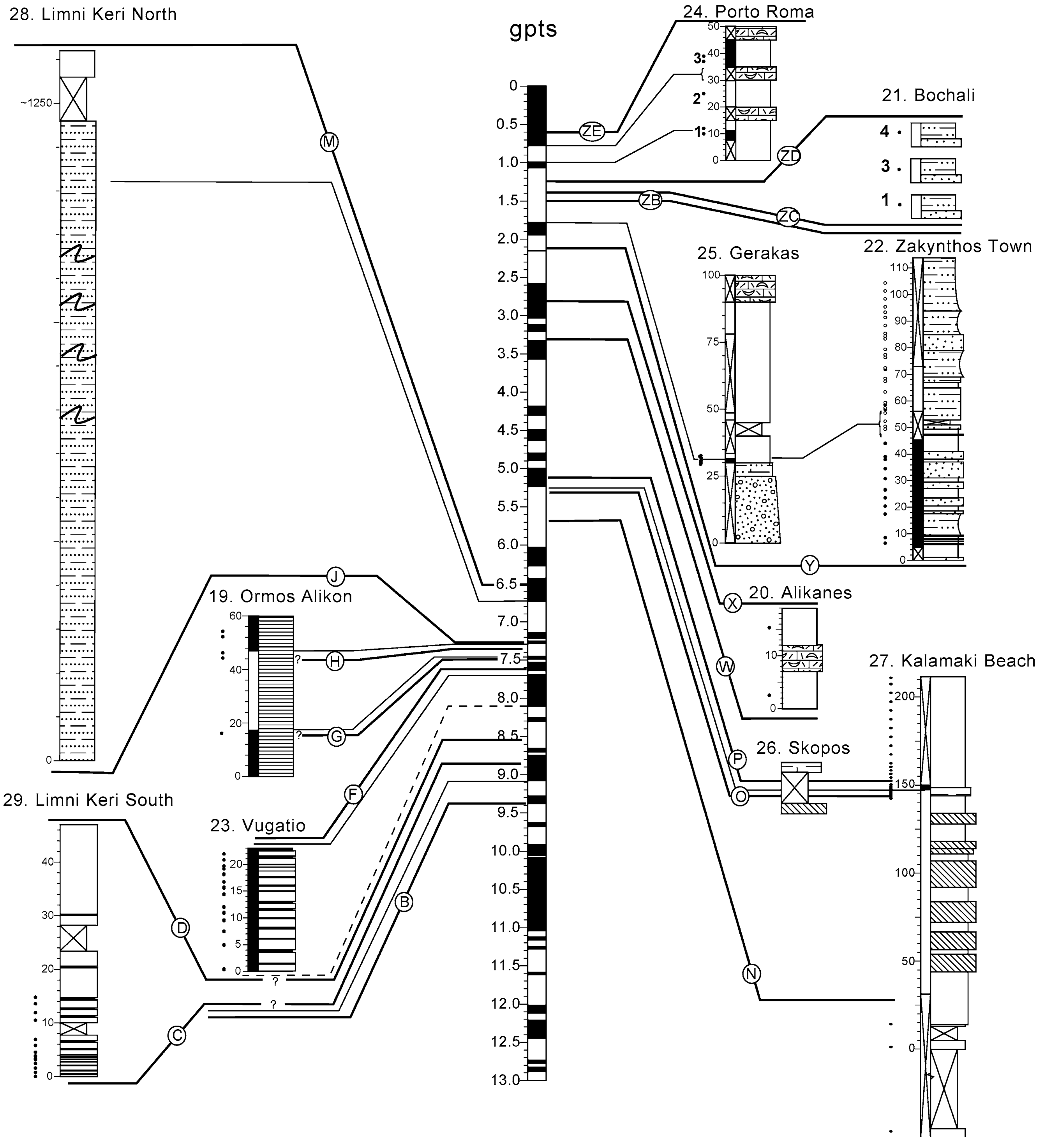


Fig. 10c

phase of at least 500–600 m of subsidence must, therefore, have affected the southern part of the Messini peninsula in the Early Pliocene or possibly during the latest Miocene–Early Pliocene transitional interval.

The Upper Plio–Pleistocene exposed on the Peloponnesos is invariably of shallow marine or terrestrial origin (Table 1). Contemporaneously, however, at least 800 m of subsidence occurred in the Gulf of Corinth since the Late Pliocene (Heezen et al. 1966). Brooks and Ferentinos (1984) estimated that as much as 2000 m of Late Plio–Pleistocene sediments were deposited in the Gulf of Patras. In the Saronic Gulf, several hundreds of metres of depth were already reached in the Early Pliocene (Van Hinsbergen et al. 2004).

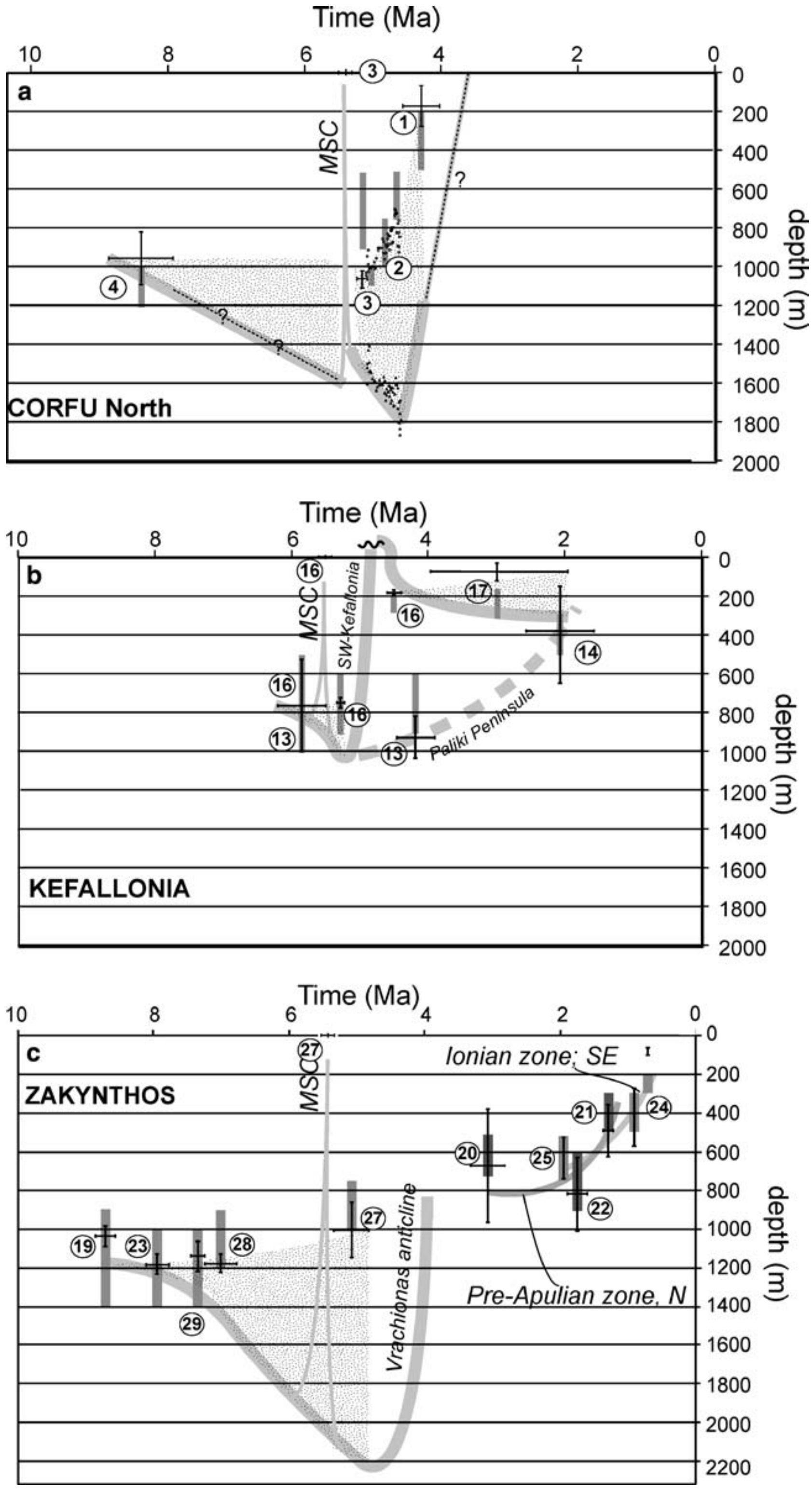


Fig. 11 Diagrams of Corfu, Kefallonia and Zakynthos, showing paleobathymetry trends and vertical motion of the base of the section. *Encircled numbers* correspond to sections of Figs. 5, 6, 7, 10a, b, c. See text for further explanation. *MSC* Messinian Salinity Crisis

Kythira

We analysed two sections from Kythira, sampled and described by Meulenkamp et al. (1977): Mitata (57) contains tilted fluvial sands and conglomerates and lacustrine to shallow marine (Table 1) silty clays and calcarenites. Marine Pliocene deposits of Paleopolis (58) unconformably overlie the basement and Tortonian deposits. Paleopolis (58) reveals a deepening from ~300 to ~750 m (Fig. 12, Table 1). The top of the section gives much shallower depths, although part of the shallow signal may be the result of downslope transport.

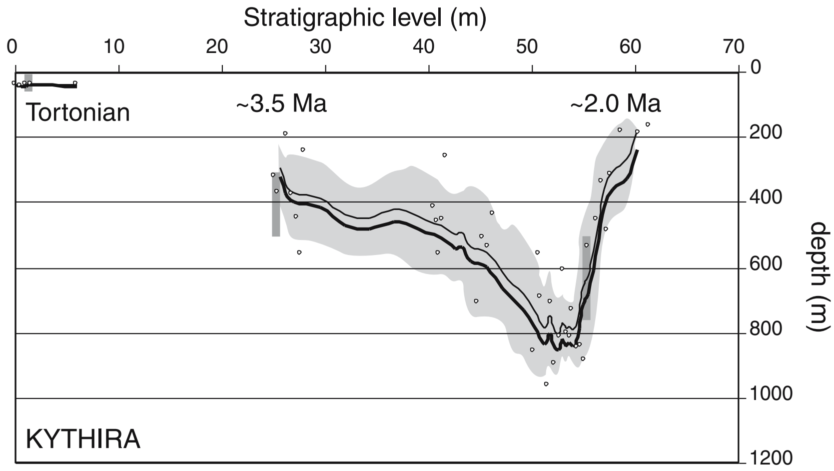


Fig. 12 Paleobathymetry and vertical motion curve obtained from the Tortonian (Mitata, 57) and Pliocene (Paleopolis, 58). For location see Fig. 8. Note that the *horizontal axis* represents stratigraphic level. The apparent uplift in the Late Pliocene may be an artefact of downslope transport. *Vertical dark grey bars* represent the depth estimates based on benthic depth markers, *shaded light grey area* represents the standard deviation on the depth estimate (*thin solid line*). *Thick solid line* represents the depth estimate plus the amount of accumulated sediment (i.e. the motion of the base of the stratigraphy with respect to sea level)

Age determination of the structures and deformation phases

This paper focuses on the deformation of western Greece since the onset of clockwise rotation starting around 15–13 Ma (Van Hinsbergen et al. 2005c). To unravel the geological history of western Greece, we will first discuss and summarize the various deformation stages and structures of western Greece and estimate their ages, based on the above literature review and the newly obtained paleobathymetry and vertical motion data and ages.

The folds of the Ionian zone

The age of the second phase of folding that inverted the Lower Miocene Klematia–Paramythia half-graben can be constrained to post-17 Ma, and prior to the formation of the Thesprotiko Shearzone, which offsets the Botsara syncline (i.e. the folded Klematia–Paramythia basin). The synsedimentary submarine folding reported by Meulenkamp (1982) forms Levkas in the upper Burdigalian to lower Langhian and may constrain the timing of the post-17 Ma folding of the Ionian zone. The overthrusted Middle to Late Miocene of Parga indicates that (possibly local) thrusting continued at least into the Late Miocene in the external Ionian zone.

Uplift of the Ionian Islands

The vertical motion reconstructions of Corfu, western Epirus (around Riza), Kefallonia and Zakynthos (Fig. 11; Table 1) reveal two distinct uplift phases: the first led to strong uplift in the Early Pliocene before 3.5 Ma and affected the entire western coast of Greece, possibly including the Strophades. In the external Ionian zone of Corfu and Riza, the structure(s) accommodating the uplift cannot be identified. However, on Kefallonia and Zakynthos evidence for the activity of thrusting and folding is found (White Rocks Thrust and Vrachionas anticline, respectively; Figs. 6, 7). These structures probably came into existence in the Late Miocene, as seen from the NE-transport directions in mass flows in the limbs of the Vrachionas anticline on Zakynthos (Limni Keri North, 28) and in the hanging wall of the White Rocks Thrust on Kefallonia (Akros Liakkas, 16). Moreover, the Messinian of Corfu is partly missing and the regional dip of the Tortonian and Pliocene of northern Corfu suggests an angular unconformity in between. Finally, the reworking of Tortonian foraminifera in the Lower Pliocene of southern Corfu indicates uplift and erosion in the hinterland during the Messinian. In northern Corfu, the base of Arillas Beach (4) was uplifted by approximately ~1,200 m since 4 Ma, whereas Roda Beach underwent a net uplift of only ~300 m. This differential uplift is in line with the presently observed 4°ENE-tilt of northern Corfu.

The second phase only affected the Paliki peninsula of Kefallonia and the island of Zakynthos and includes many hundreds of meters of uplift in the course of the Pleistocene. This occurred probably contemporaneous with clockwise rotation as reported by Duermeijer et al. (1999) and Van Hinsbergen et al. (2005c). The Pre-Apulian zone of Levkas (Aghios Petros, 12) confirms strong post-Late Miocene uplift (more than 1,000 m; Table 1), but does not allow further constraining of the timing of uplift.

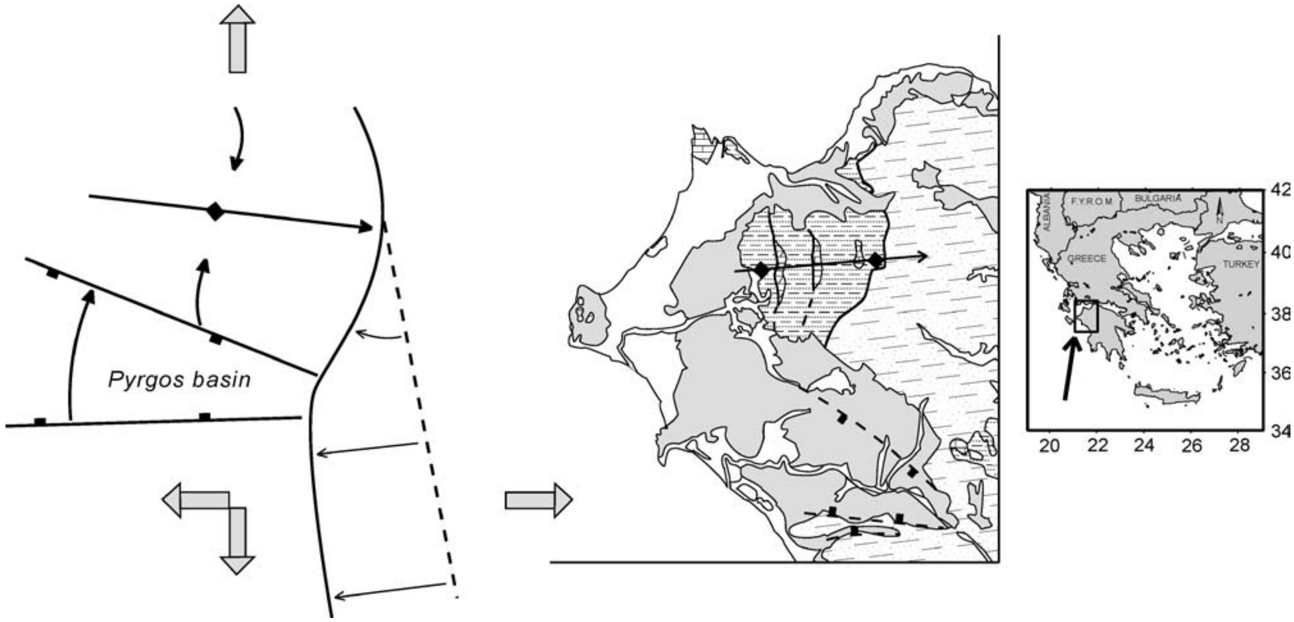


Fig. 13 Schematic drawing of the proposed Late Plio–Pleistocene development of the northwest Peloponnesos. A right-lateral component created by the increasing intensity of E–W extension from north to south superimposed on N–S extension created the opening of a triangular basin, accompanied by the formation or further development of a plunging antiform to its north

Kefallonia, Thesprotiko, Kastaniotikos and Aliakmon fault zones

As shown in Fig. 9, a fault zone can be traced from the offshore Kefallonia Fault Zone via the right-lateral Thesprotiko Shear Zone and the normal Kastaniotikos Fault to the Aliakmon Fault Zone. Presently, the Kefallonia Fault Zone is seismically highly active as a right-lateral transpressional strike slip zone, whereas the Thesprotiko and Kastaniotikos Faults are inactive. The modern Aliakmon Fault Zone accommodates dip-slip motion [e.g. the Servia Fault (Doutsos and Koukouvelas 1998); Fig. 1], but it is likely that these four fault zones once accommodated southwestward motion of the south with respect to the north (see below). This is in line with the analysis of the Karditsa and Larissa basins to the south of this fault zone, where Caputo and Pavlides (1993) and Caputo et al. (1994) reported NE–SW extension in the Late Miocene or Early Pliocene, followed by approximately NW–SE extension since the Late Pliocene and Pleistocene. To the north of the Aliakmon Fault Zone, NE–SW extension had only a minor influence since the Late Miocene (Pavlides and Mountrakis 1987). The NE–SW trending Aliakmon Fault Zone, therefore, most likely acted as a strike-slip fault during the Late Miocene and Early Pliocene, accommodating the difference in NE–SW stretching. In 'Transverse structures and late Neogene extensional basins', we estimated the maximum age of the transition from NE–SW to NW–SE extension at 5 Ma for the Aliakmon fault zone. The age of the Kastaniotikos Fault cannot be constrained further than syn- or post-activity of the Pindos Thrust.

The Thesprotiko Shear Zone accommodated approximately 20 km of right-lateral motion after the folding of the Botsara-syncline (which has a maximum age of 17 Ma, see above). Riza (9), which is positioned right on top of the Thesprotiko Shear Zone, underwent local, strong clockwise rotation after the Early Pliocene (Van Hinsbergen et al. 2005c), indicating post-Early Pliocene activity of the Thesprotiko Shear Zone.

Finally, the age of onset of formation of the Kefallonia Fault Zone can be estimated from the detailed investigation of the deformation of the Ionian Islands and western Akarnania. Cushing (1985) already suggested that the curvature of the fold-axial plains of Levkas from east-dipping, NW–SE striking in the south to west-dipping, NE–SW striking in the north was the result of further transpression along the Kefallonia Fault Zone in the north than in the south. In Akarnania, along the coast opposite to Levkas the Tortonian of Paleros (11) and Pogonia (10) is also folded along NE–SW striking axes, parallel to northern Levkas. Moreover, these sections have been rotated 35 and 50° clockwise, respectively (Van Hinsbergen et al. 2005c), probably as a result of dragfolding along the transpressional Kefallonia Fault Zone. Correcting for this post-Tortonian clockwise rotation yields strikes comparable to the general trend of fold axes in Epirus and Akarnania, indicating that the Kefallonia Fault Zone induced no drag folding prior to the Tortonian.

Further constraints on the timing of the Kefallonia Fault Zone can be derived from the island of Kefallonia, where the deformation of the Pre-Apulian zone since the Late Miocene was dominated by the interference between NE–SW compression and right-lateral strike-slip.

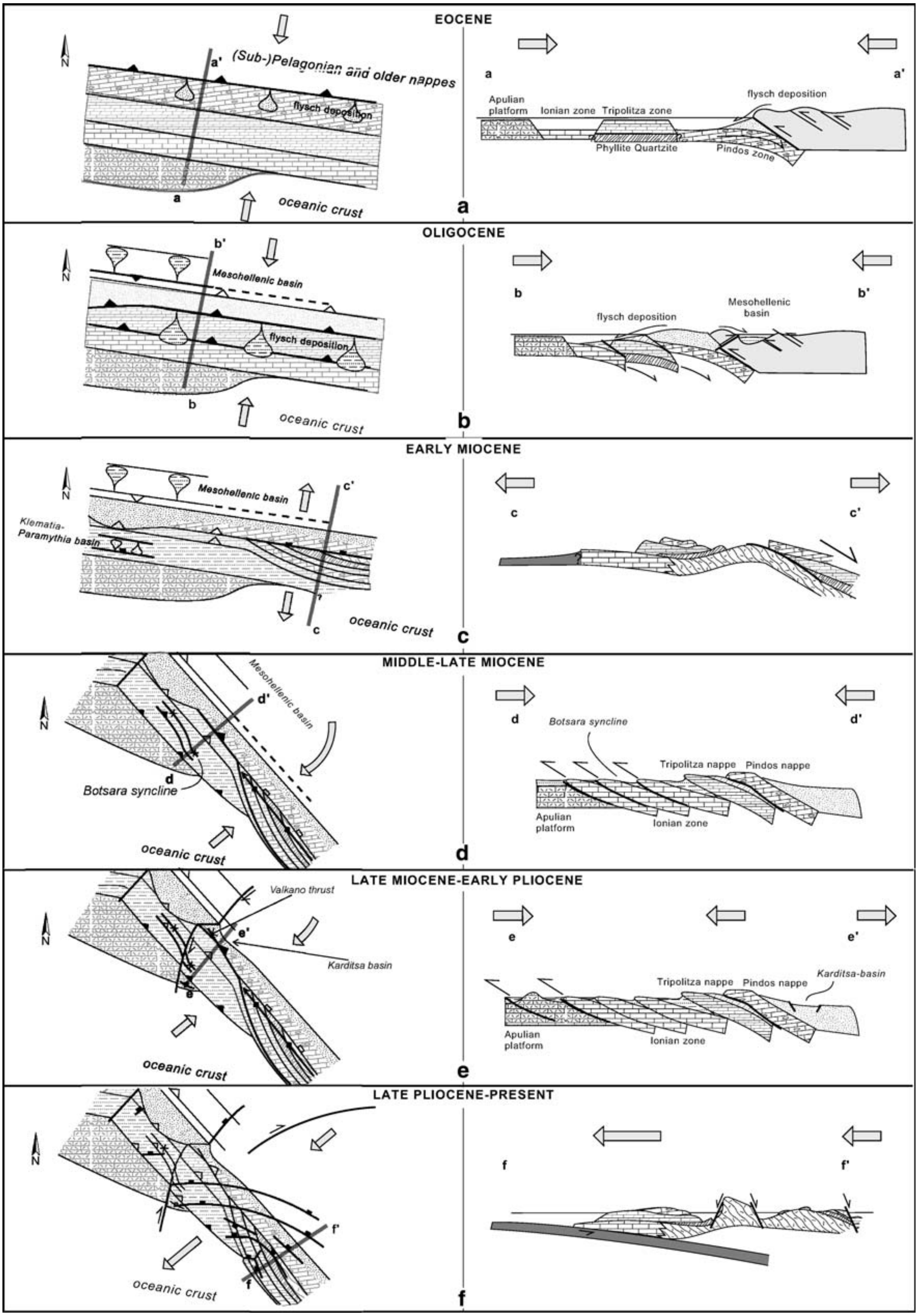


Fig. 14 Schematic paleogeographical and structural evolution of western Greece since the Late Mesozoic. For key, see Fig. 1

This can be inferred from a number of phenomena: the folds and thrusts on southeastern Kefallonia strike NW–SE, subparallel to the dominant trend of the Hellenides. This strike gradually changes to the NNE–SSW orientation of the Paliki peninsula antiform, i.e. subparallel to the Kefallonia Fault Zone (Fig. 6). In the northern part of the island, the Aenos Thrust and Kalon Thrust strike parallel and the orthogonal to the Kefallonia Fault Zone, respectively. No overprinting relationships can be inferred from these thrusts and they were probably created simultaneously sometime after the Early Miocene. The Aenos Thrust—striking subparallel to the Kefallonia Fault Zone—suggests right-lateral transpression.

Paleomagnetic data indicate a 90° clockwise rotation of the Paliki peninsula in the course of the Pliocene (Van Hinsbergen et al. 2005c), whereas hardly any rotation occurred since the Late Miocene in Akros Liakkas (16) and Spartia (17) (Van Hinsbergen et al. 2005c). The combination with the uplift history of both areas (Fig. 11) leads us to tentatively conclude that the deformation pattern of Kefallonia evolved entirely since the Messinian as a result of NE–SW compression and transpression along the NNE–SSW striking Kefallonia Fault Zone. The Kefallonia Fault Zone, therefore, most likely originated in the Messinian (Late Miocene).

The above analysis leads us to conclude that the Kefallonia Fault Zone, Thesprotiko Shearzone and Aliakmon Fault Zone formed a slightly curved, right-lateral strike-slip system—possibly including the Kastaniotikos normal fault—since the Late Miocene, which was active until the middle Pliocene. Afterward, the Aliakmon Fault Zone became dominantly a normal fault and the Thesprotiko Shear Zone became inactive, whereas the Kefallonia Fault Zone remained active as a right-lateral strike slip fault.

The smaller E–W (Souli-fault) or NE–SW (e.g. Konitsa Fault) striking, presently or recently seismically active normal fault zones of northern and northwestern Greece (Fig. 4), is probably related to the Late Pliocene and Pleistocene NW–SE extension.

Curved extensional basins of southwestern Greece

Most of the sedimentary basins of southwestern Greece were generated in the Late Pliocene and Pleistocene and are bounded by seismically still active faults (e.g. Gulf of Corinth, Gulf of Patras, Messini basin and Evrotas Basin). The prelude to the Late Pliocene basin formation occurred in the Early Pliocene, when subsidence affected the Isthmus of Corinth (Collier and Dart 1991) and Aegina in the Saronic Gulf (Van Hinsbergen et al. 2004). Additionally, rapid Early Pliocene subsidence must have affected the Messini peninsula of the Southwestern Peloponnesos, possibly already along the Kalamata Fault. Additionally, Kythira started to subside in the (late?) Early Pliocene (Fig. 12). The earliest signs of extension in the southeastern Aegean area are thus probably accommodated along NW–SE trending structures. The sedimentary basins of southwestern Greece contain a strong influence of N–S extension, probably comparable to the situation in northern Greece, with a gradually increasing influence of E–W extension from north(west) to south(east), superimposed on N–S extension. This is suggested by the curvature of the extensional basin systems of the Gulf of Amvrakikos-Sperchios Basin–Gulf of Evia, the Gulf of Patras–Gulf of Corinth–Saronic Gulf and the 15 basin-Megalopolis Basin–Evrotas Basin system, which all are characterized by approximately N–S extension in the northwest, grading into E–W extension in the southeast. Additional evidence is found on the northwestern Peloponnesos, where the triangular 15 basin opened first from south to north (Hageman 1979). To its north, a large, open eastward plunging antiform gives the Pindos thrust its curvature in mapview. We explain the eastward plunging antiform in combination with the triangular shape of the basin as the result further westward motion of the south with respect to the north, imposing a right-lateral component on the N–S extension that is responsible for the opening of the basin (Fig. 13). Finally, the triple junction of Melimala (Figs. 8, 9) may indicate the location, where N–S and E–W extensions were comparable in magnitude, leading to a multidirectional extension.

In conclusion, the sedimentary basins of southwestern Greece formed in response to N–S extension, with an increasing influence of E–W extension from north(west) to south(east). The prelude to the Late Pliocene phase of basin formation was formed in the Early Pliocene, probably by subsidence along NW-SE trending faults.

Reconstruction of the Tertiary tectono-sedimentary history of western Greece

The combination of the above summarized and dated structures of western Greece leads us to provide a scenario for the geological development of western Greece contemporaneous with the clockwise rotation of the area since the Middle Miocene and the collision with Apulia. First, the pre-Middle Miocene development will be summarized.

Mesozoic–Early Miocene

The Cretaceous to Oligocene history of western Greece is characterized by the formation of an accretionary wedge by an outward stepping of the thrust front and stacking of the (Sub-) Pelagonian, Pindos, Tripolitza and Ionian units (Figs. 3, 14) (Aubouin 1957; Jacobshagen 1986). In the latest Eocene to Early Oligocene time, i.e. approximately 35–30 Ma ago, a major rearrangement of the paleoenvironment took place in western Greece, associated with the onset of underthrusting of the Tripolitza and Ionian units below the Pindos nappe (Van Hinsbergen et al. 2005b). This forced the Pindos unit and the overlying ophiolitic rocks of the (Sub-)Pelagonian zone to thrust northward over the Sub-Pelagonian zone. In the front of this east-verging thrust, the Mesohellenic basin formed as a foreland basin (Figs. 5, 14) (Doutsos et al. 1994). In the west, this phase of uplift of the Pindos and (Sub-)Pelagonian nappes led to the transition from carbonate to flysch sedimentation all over the Tripolitza and Ionian zones (Figs. 3, 14) (IGRS-IFP 1966).

In the latest Oligocene or earliest Miocene, another rapid reconfiguration of the paleoenvironment occurred, associated with the end of underthrusting of the Tripolitza and Ionian units, and their internal shortening (Sotiropoulos et al. 2003; Van Hinsbergen et al. 2005b). After the latest Oligocene to earliest Miocene uplift of the Tripolitza, internal and middle Ionian units, NNW–SSE extension foundered the Klematia–Paramythia half-graben and was continuous until at least 17 Ma (Van Hinsbergen et al. 2005b) (Fig. 4). Simultaneously, the south-Aegean core complex formed a window in the nappe pile on the Peloponnesos. The amount of Early Miocene extension on the Peloponnesos [i.e. the width of the window exposing the metamorphic rocks prior to post-exhumation folding; >100 km: Fig. 8 and Van Hinsbergen et al. (2005b)] clearly exceeds the amount of extension associated with the formation of the Klematia–Paramythia basin in northwestern Greece (of the order of a few kilometers only). This difference in magnitude of horizontal extension may explain part of the present-day curvature of the Pindos unit (Fig. 14).

Middle-Late Miocene: compressional deformation during clockwise rotation

Between 15–13 and 8 Ma, western Greece and the Peloponnesos rotated 40° clockwise. The Ionian and higher units rotated, whereas probably no rotation occurred in (Pre-) Apulian zone (Van Hinsbergen et al. 2005c and references therein). This difference in rotation must have included a from north to south increasing amount of convergence between the Apulian and Ionian zones, with approximately 350 km of convergence at the latitude of Kefallonia. Several indications exist for the reestablishment of compression contemporaneous with the onset of rotation and the convergence between Apulia and the Ionian and higher nappes: a late Burdigalian to early Langhian phase of compression affected the external Ionian zone and the pre-Apulian zone of Levkas (Meulenkamp 1982). This phase of compression may correspond to the folding of the Lower Miocene Klematia–Paramythia basin, by inverting the basin-bounding normal fault to form the Middle Ionian Thrust (Van Hinsbergen et al. 2005b).

Additionally, folding and thrusting of the western Mediterranean Ridge was constrained to start between 17.5 and 13.5 Ma (see Geology of western Greece: a review). Extensional exhumation of the Phyllite–Quartzite unit continued until at least 15 Ma on western Crete (Thomson et al. 1998), providing a probable maximum age estimate for the folding on the Peloponnesos.

Re-establishment of compression in western Greece may have been slightly diachronous, becoming younger from north to south, but is largely contemporaneous with the (oldest possible) onset of clockwise rotation of western Greece around 15 Ma. We thus conclude that the rotation of western Greece was accompanied by internal shortening.

The more intense shortening of Epirus with respect to Akarnania includes the Lower Miocene and must have evolved during or after the Middle Miocene. We therefore suggest that it relates to the overriding of the continental (Pre-) Apulian zone by the Ionian zone of northern Epirus, as opposed to subduction of the oceanic Ionian lithosphere in the south (Fig. 2, 4). The deep-marine Tortonian sediments of Pogonia (10), Paleros (11), Arillas Beach (4) and in the area of Riza (IGRS-IFP 1966), with ages of 9–8 Ma and younger, were deposited during the last stages of, or immediately after the Middle to early Late Miocene rotation of western Greece. Probably, these sediments unconformably overlie the Lower Miocene (IGRS-IFP 1966). If true, this indicates that approximately 600–1,000 m of subsidence affected (part of) western Epirus, western Akarnania and Corfu prior to or in the early Tortonian. The reason for this subsidence remains open for speculation: no large extensional features have been identified in western Greece and the subsidence may reflect the flexural response to loading associated with the crustal thickening which is associated with the Middle to early Late Miocene rotation phase.

Late Miocene–Early Pliocene: onset of collision with Apulia

The vertical motion reconstruction of the Ionian Islands and the sedimentary basins in northern and southwestern Greece reveals that a reconfiguration of motion started to develop in the Late Miocene. The Pre-Apulian and external Ionian zones were shortened and uplifted along the western coast of Greece in response to NE–SW compression. Further west, the Karditsa and Larissa basins started to open in response to NE–SW extension and a right-lateral strike slip zone started to develop accommodating southwestward motion of the south with respect to the north. The subsidence of the southwestern Peloponnesos, Aegina and Milos in the the Saronic Gulf and its lateral continuation, and probably also Kythira, indicates the NE–SW extension also starts to play a role in southwestern Greece.

In our view, the Late Miocene to Early Pliocene situation shows strong evidence for an increasingly important role of the Apulian platform: the north collides with Apulia and the south moves over the subducting oceanic lithosphere of the Ionian basin. The increasing difference between the north and south leads was in the Late Miocene and Early Pliocene accommodated by the right-lateral fault system of the Kefallonia Fault Zone, the Thesprotiko Shear Zone and the Aliakmon Fault Zone.

The role of the Kastaniotikos Fault remains open for discussion. If both the Aliakmon Fault Zone and the Thesprotiko Shear Zone were active in the Late Miocene, it is reasonable to assume that the Kastaniotikos Fault formed a transtensional bend in the fault zone, which is in line with the observed sense of shear (Profile H–H′, Fig. 5). The Karditsa basin widens to the south and its opening most likely included rotation of the Pindos unit to its west, with a pole at the intersection between the Aliakmon Fault Zone and the Kastaniotikos Fault. The amount of inferred rotation is approximately equal to the angle between the folds axes on either side of the Valkano Thrust (Fig. 4). We suggest that the opening of the Karditsa basin was contemporaneous with either the formation or reactivation of the Valkano Thrust (Figs. 1, 4, 14). This rotation of the Pindos and motion of the Valkano Thrust is in line with faster and further southwestward motion in the south than in the north, concluded above.

The shortening of northwestern Greece came to an end in the course of the Pliocene, evidenced by the vertical motion histories derived from Corfu and the area around Riza (9). This was probably the result of the collision and end of further convergence between Apulia and northwestern Greece.

Late Plio–Pleistocene: motion of the Hellenides around Apulia

From 3.5 Ma onward, a no further contraction occurred between Apulia and northwestern Greece. The Thesprotiko Shear Zone and Kastaniotikos Fault were no longer active and mainland Greece became subjected to N–S to NW–SE extension, subparallel to Apulia.

Southwestward motion of southwestern Greece continued, but internal deformation increased, leading to the Late Plio–Pleistocene curved sedimentary basin systems and the basins. The dominant NW–SE to N–S extension and southward motion in northern Greece, gradually changing to E–W extension and westward motion in southwestern Greece indicates motion of Greece around the southeastern limit of the Apulian platform. This motion around Apulia is accommodated by the transpressional Kefallonia Fault Zone, shearing the Pre-Apulian slope off the Apulian platform.

The curved shape of the Late Pliocene and Pleistocene basins of southwestern Greece resembles the patterns predicted by analogue modeling of gravitational spreading around a ‘free boundary’ (Hatzfeld et al. 1997; Gautier et al. 1999), which in this case would be the southeastern margin of the Apulian platform. Note that the Gulf of Corinth and the Amvrakikos-Sperchios basins have been suggested to have also accommodated the westward motion of Turkey (e.g. Dewey and Sengör 1979; Reuther et al. 1993; Armijo et al. 1996; Goldsworthy et al. 2002).

The Late Pliocene change in stress regime of northwestern Greece and the formation of extensional sedimentary basins in southwestern Greece are thus the result of the collision with and subsequent motion around Apulia in the middle Pliocene, i.e. around 3.5 Ma.

Conclusions

To identify if, when and how deformation occurred in western Greece during the post-Early Miocene rotation and subsequent collision of the area with the Apulian platform, the geometry and inferred kinematics of the area have been combined with vertical motions derived from paleobathymetry and deposition trends in the marine sediments. After a phase of extension in the Early Miocene, basin inversion and compression accompanied the onset of rotation in the early Langhian, i.e. around 15 Ma, although the onset may become younger from northwest to southeast. Subsequently, western Greece collided with the Apulian platform, leading in the Late Miocene to a right-lateral strike-slip system running from the Aliakmon Fault Zone in northern Greece, via the Kastaniotikos Fault and the Thesprotiko Shear Zone to the Kefallonia Fault Zone, offshore western Greece. NE–SW compression and uplift of the Ionian Islands was accompanied by NE–SW extension in southwestern Greece, associated with faster southwestward motion in the south than in the north. This led in the middle Pliocene (around 3.5 Ma) to collision without further shortening in northwestern Greece. From then onward, NW–SE to N–S extension east of Apulia, and gradually increasing influence of E–W extension in the south accommodated motion of the Hellenides around the Apulian platform. As a result, curved extensional basin systems evolved, including the Gulf of Amvrakikos-Sperchios Basin–Gulf of Evia system and the Gulf of Corinth–Saronic Gulf system.

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