Tectonic–sedimentary evolution of the western margin of the Mesozoic Vardar Ocean: evidence from the Pelagonian and Almopias zones, northern Greece

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Abstract: The Vardar Zone documents the Mesozoic–Early Cenozoic evolution of several small oceanic basins and a complex history of terrane assembly. Following a Hercynian phase of deformation and granitic intrusion within the Pelagonian Zone to the west, the Vardar Zone rifted in Permian–Triassic time, with the creation of an oceanic basin (Almopias Ocean) during the Late Triassic–Early Jurassic. During the Mid-Jurassic, this ocean subducted northeastwards beneath the Paikon Zone and the Serbo-Macedonian Zone, giving rise to arc volcanism and back-arc rifting. A second ocean basin, the Pindos Ocean, opened to the west of a Pelagonian microcontinent, also during Late Triassic–Early Jurassic time. During the Mid–Late Jurassic, ophiolites were emplaced northeastwards (in present co-ordinates) from the Pindos Ocean onto the Pelagonian microcontinent, forming the Pelagonian ophiolitic mélange within a flexural foredeep. This emplacement is dated at pre-Late Oxfordian–Early Kimmeridgian from the evidence of corals within neritic carbonates that depositionally overlie the emplaced ophiolitic rocks in several areas. Related greenschist- or amphibolite-facies metamorphism is attributed to deep burial following trench–margin collision and the attempted subduction of the Pelagonian continent. An inferred phase of NNW–SSE displacement, also of pre-latest Jurassic age, imparted a regionally persistent stretching lineation and related ductile fabric, apparently related to post-collisional strike-slip. The Pelagonian Zone and its emplaced ophiolitic rocks then underwent extensional exhumation during Late Jurassic–Early Cretaceous time. The western margin of the Vardar Zone experienced extensional (or transtensional) faulting, neritic carbonate and terrigenous clastic deposition, and intermediate–silicic magmatism during Late Jurassic–Early Cretaceous time. Oceanic crust (Meglenitsa Ophiolite) formed further east in the Vardar Zone during Late Jurassic–Early Cretaceous time, possibly above a subduction zone. A near-margin setting is suggested by the presence of a deep-water terrigenous cover, probably derived from the Paikon continental unit to the east. The Vardar Zone as a whole finally closed related to eastward subduction beneath Eurasia, culminating in collision with the Pelagonian microcontinent during latest Cretaceous–Eocene time, as recorded in foreland basin development, HP–LT metamorphism, ophiolite emplacement and large-scale westward thrusting. In contrast to models that suggest closure of the Vardar Ocean in the Mid–Late Jurassic, followed by reopening of a Cretaceous ocean, we believe that the Vardar Ocean remained partly open from Triassic to Late Cretaceous–Early Cenozoic time.

Compared with the westerly, more 'external' tectonic units of the Balkans and Hellenides, the Vardar Zone (Fig. 1) has remained poorly understood despite its critical bearing on the tectonic development of Tethys in the Eastern Mediterranean region (e.g. Smith 1993, 2006; Robertson *et al.* 1996). Also, its relation to the Pindos (Sub-Pelagonian) Zone to the west is still controversial. The Vardar Zone can be traced eastwards for > 500 km through Hungary, Croatia, Bosnia, Serbia, Macedonia (former Yugoslavia) and northern Greece until it runs into the northern Aegean Sea, with only fragments exposed on land further south (Fig. 1). The Vardar Zone of Northern Greece, also known as the Axios Zone, is located between the Pelagonian Zone to the west and the Serbo-Macedonian Zone to the east (Figs 1 and 2b). The Vardar Zone is traditionally divided into a series of tectonostratigraphic zones. From west to east these are the Almopias Zone, the Paikon Zone and the Peonais Zone (Mercier 1966; Fig. 2a). The Almopias Zone is further subdivided into the Western, Central and Eastern Almopias zones (Figs 2b and 3). The

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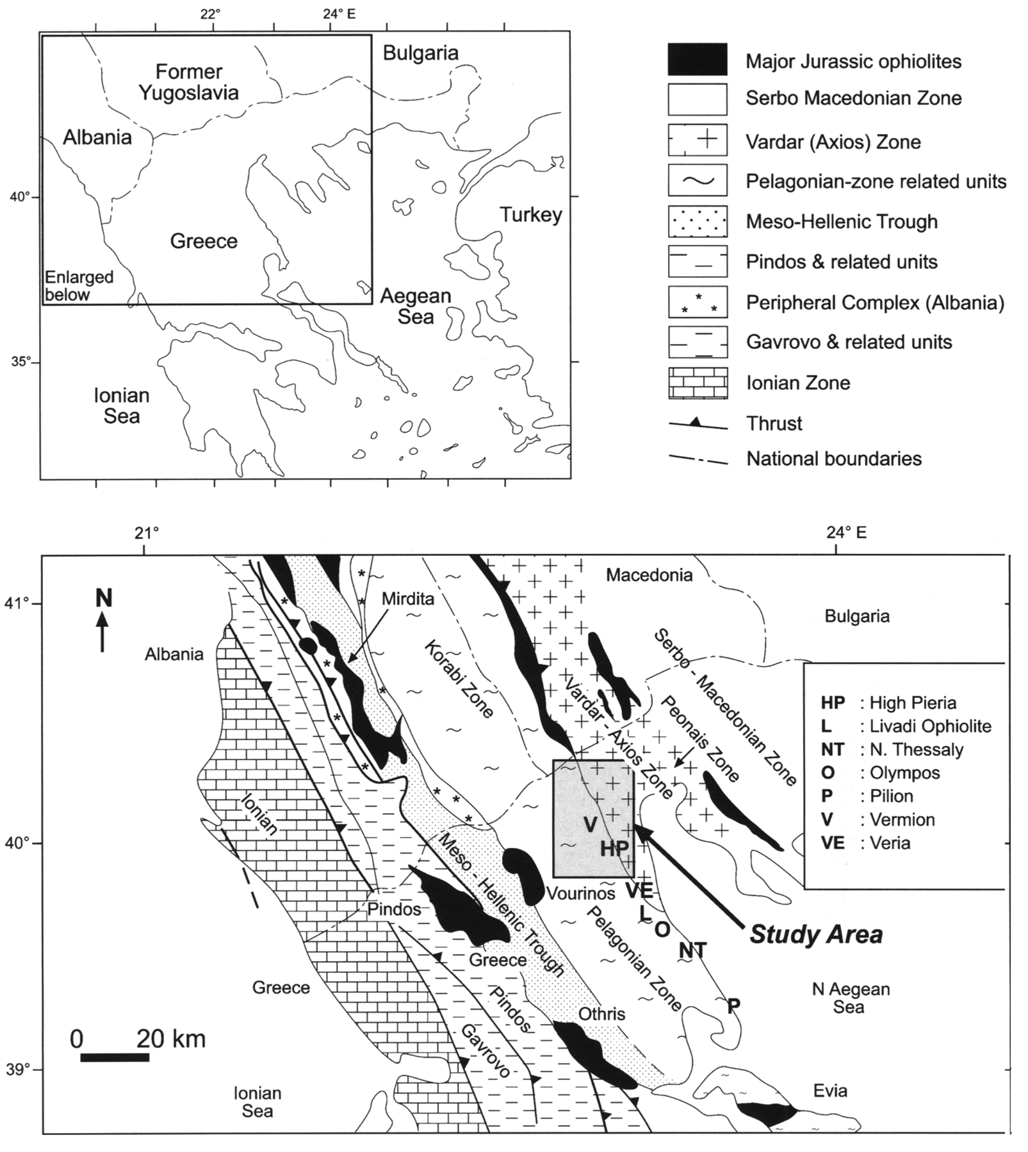


Fig. 1. Simplified outline tectonic maps of the Balkan region showing the study area in NE Greece (modified from Robertson & Shallo 2000).

Central Almopias Zone is the most diverse and is further subdivided into several tectonostratigraphic units, each with a distinctive stratigraphic sequence that can be correlated to give an overview of the tectonostratigraphic evolution through time (Fig. 4). Here, we will mainly discuss the traverse shown in Figure 2b, which exposes all of the units of the Almopias Zone and the eastern part of the Pelagonian Zone. We will also take account of correlative units exposed in the Voras Massif further north (Fig. 3) and relevant units exposed further south, especially in the Vermion Mountains (marked V in Fig. 1, see also Fig. 2a).

In general, the Pelagonian Zone and the Western and Central Almopias zones formed parts of the western margin of the Vardar Ocean during the Mesozoic, whereas the Eastern Almopias Zone comprises more oceanic lithologies derived from the Vardar Ocean and has a fundamentally

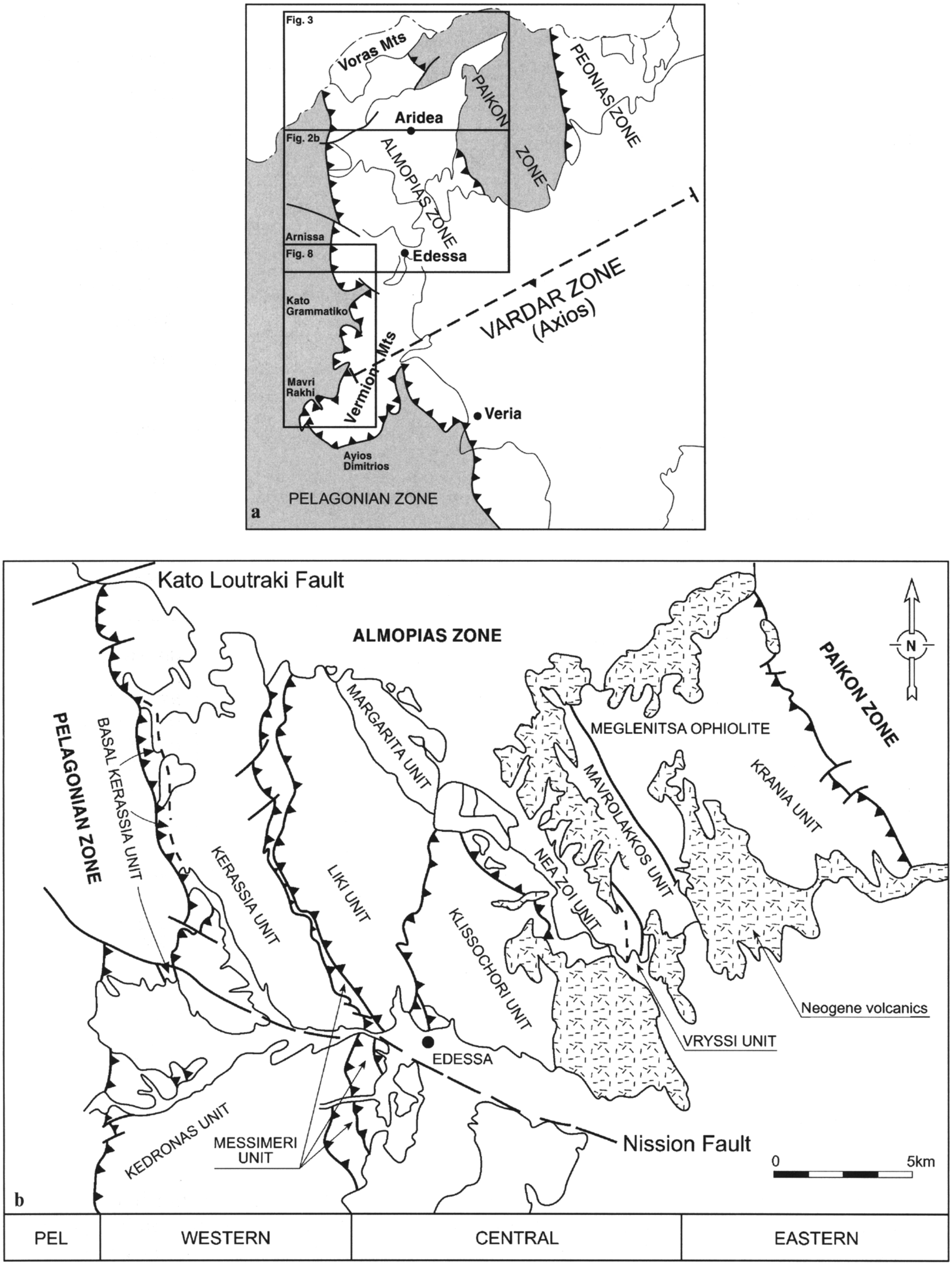


Fig. 2. (**a**) Outline tectonic map of the Vardar Zone; the small boxed area is shown in Figure 2b; the large boxed area is enlarged in Figure 3. (**b**) Tectonostratigraphic units of the Almopias Zone (modified from Mercier 1966).

different geological history until suturing during Early Cenozoic time. In particular, the Pelagonian Zone and the Western and Central Almopias zones experienced pre-Cretaceous deformation and metamorphism, which is not represented in the Eastern Almopias Zone. In this paper we use

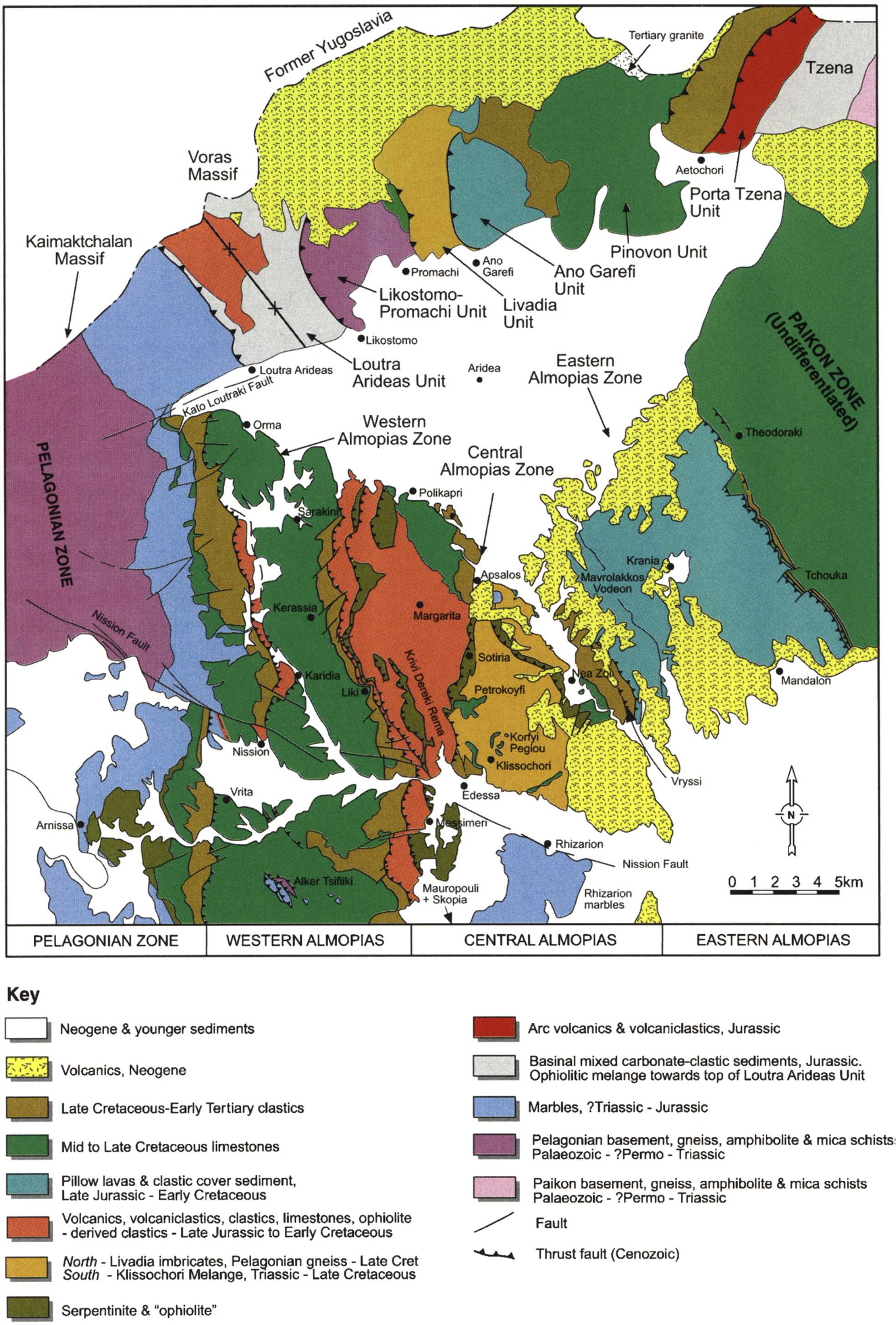


Fig. 3. Simplified geological map showing the main tectonostratigraphic units, lithologies and their ages in the Almopias Zone and the Voras Massif to the north, based mainly on mapping by Mercier (1966), Vergély (1984), Mercier & Vergély (1984*a*, *b*), Brown (1994) and Sharp (1994). Place names mentioned in the text are included.

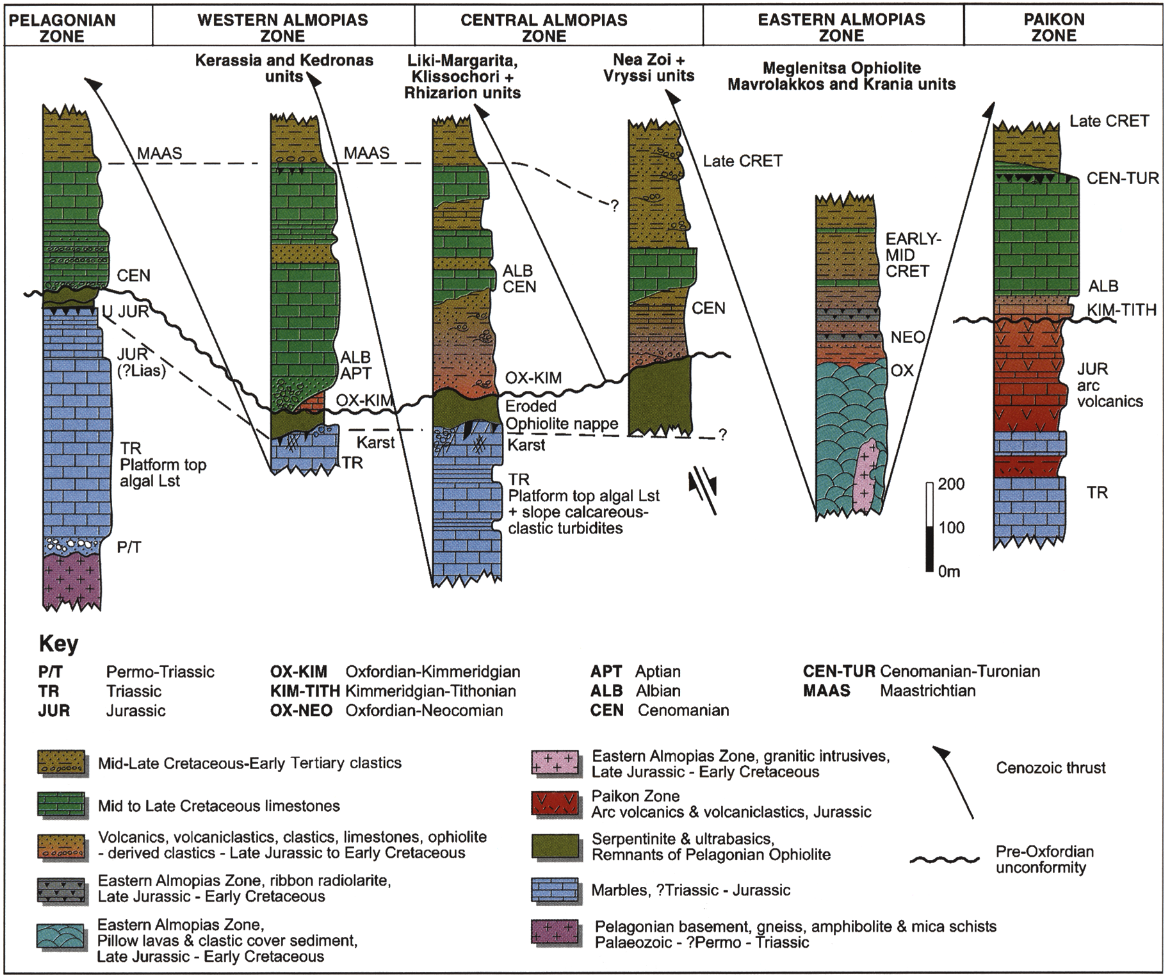


Fig. 4. Tectonostratigraphic units of the Pelagonian, Almopias and Paikon zones (modified from Mercier 1966).

the term Vardar Ocean for oceanic lithosphere that existed within the Vardar Zone, including the Peonais Zone in the east, whereas we use the term Almopias Ocean, more specifically, for oceanic crust that we interpret to have existed between the Pelagonian Zone and the Paikon Zone in the east (Fig. 2a).

Permian–Triassic rifting

Pre-rift 'basement' units are exposed within the Pelagonian Zone, where they mainly comprise metasedimentary and meta-igneous rocks (e.g. amphibolites), intruded by granitic rocks of Carboniferous age (Mountrakis 1984). Similar 'basement' rocks are exposed in the Paikon Zone within the Voras Massif, near the border between Greece and Macedonia (former Yugoslavia) (Mercier 1966; Brown & Robertson 2004; Fig. 3). A pre-Mesozoic tectonic fabric survives in the core of the Pelagonian Zone (e.g. in the Vernon Mountains; Mountrakis 1984), but elsewhere the foliation and deformation fabric mainly reflect Late Jurassic orogenesis, with only a minor imprint from the Palaeogene suturing of the Vardar Ocean.

The Pelagonian Zone shows evidence of Triassic rifting that could in principle be related to the opening of ocean basins to the west (Pindos Ocean) or to the east (Vardar Ocean), or both. The western Pelagonian Zone (e.g. in the Vernon Mountains) shows evidence of rifting, as indicated by metaclastic and metavolcanic units that unconformably overlie the metamorphic basement (Mountrakis 1984). In the eastern part of the Pelagonian Zone, as exposed in the Voras Massif in the north (e.g. Kaimaktchalan Unit; see Brown & Robertson 2004; Fig. 3) metasedimentary rocks and amphibolites apparently record rifting to form the Almopias Ocean to the east, although these units remain poorly dated. Within the Voras Massif, a rift-related clastic–volcanic succession overlies metamorphic basement and passes gradationally upwards into a thick Mesozoic carbonate succession further east (Likostomo–Livadia Unit and the basal part of the Loutra Arideas Unit; Fig. 3; Brown & Robertson 2004). Volcanic rocks (Promachi amphibolites) within this sequence exhibit a mid-ocean ridge basalt (MORB) chemical composition (Brown & Robertson 2004). Similar amphibolites of mainly within-plate basalt (WPB) (Fig. 5a) occur within the lower part of the Klissochori Unit of the Central Almopias Zone (Figs 2b, 3 and 4) and may also relate to Triassic rifting, although stratigraphic constraints are poor. The occurrence of rift-related volcanic rocks and sediments throughout the westerly Vardar units is consistent with the opening of an oceanic basin to the east, within the eastern Almopias Zone.

Two important transverse faults, the Kato Loutraki Fault in the north and the Nission Fault in the south (Figs 2b and 3) are believed to have been active during Triassic rifting. Sequences adjacent to these faults are dominated by clastic sediments and appear to separate areas of contrasting palaeogeography from Triassic time onwards (Sharp 1994). These faults trend at a moderate to a high angle relative to the regional NW–SE trend of the Pelagonian and Almopias zones, and are interpreted as transfer faults that subdivided the rifted margin into segments, with contrasting depositional and tectonic histories. Faults parallel to the rifted margin are more difficult to recognize, probably because they developed into thrusts during later compressional deformation.

During the Permian–Triassic a thick succession of clastic sediments, bimodal volcanic rocks and neritic carbonates developed along the western margin of the Serbo-Macedonian Zone (Dimitriadis & Asvesta 1993; Stais 1994; Fig. 1). The rift basalts range from WPB, to transitional, to MORB type (Dimitriadis & Asvesta 1993). The Paikon Zone is inferred to have rifted from the Serbo-Macedonian Zone to form a bordering microcontinent. The intervening deep-water basin is preserved within the Peonais Zone and is represented by the Svoula Flysch (Kaufmann *et al.* 1976; Kockel *et al.* 1977). A NE–SW facies change is observed, from continental facies within the Serbo-Macedonian continent, to marine units within the Peonais Zone during Triassic time (Stais 1994). The Serbo-Macedonian Zone was traditionally interpreted as a part of the Eurasian margin (Jacobshagen *et al.* 1978), but was recently reinterpreted as an exotic terrane that was amalgamated during the Jurassic deformation history (Himmerkus *et al.* 2006).

The eastern margin of the combined Pelagonian–Almopias Zone therefore records rifting to form a subsiding passive margin of Late Triassic–Early Jurassic age, bordering the Almopias (western Vardar) Ocean to the east. The Serbo-Macedonian Zone records coeval rifting. Three possible options for the setting of rifting are: (1) formation within a pre-existing Palaeotethyan Ocean located within the Vardar Zone (Mountrakis 1984; Robertson & Dixon 1984; Karamata & Vujnović 2000); (2) formation of a back-arc marginal basin related to subduction outwith the Vardar Zone, either southward subduction from a Palaeotethyan Ocean located

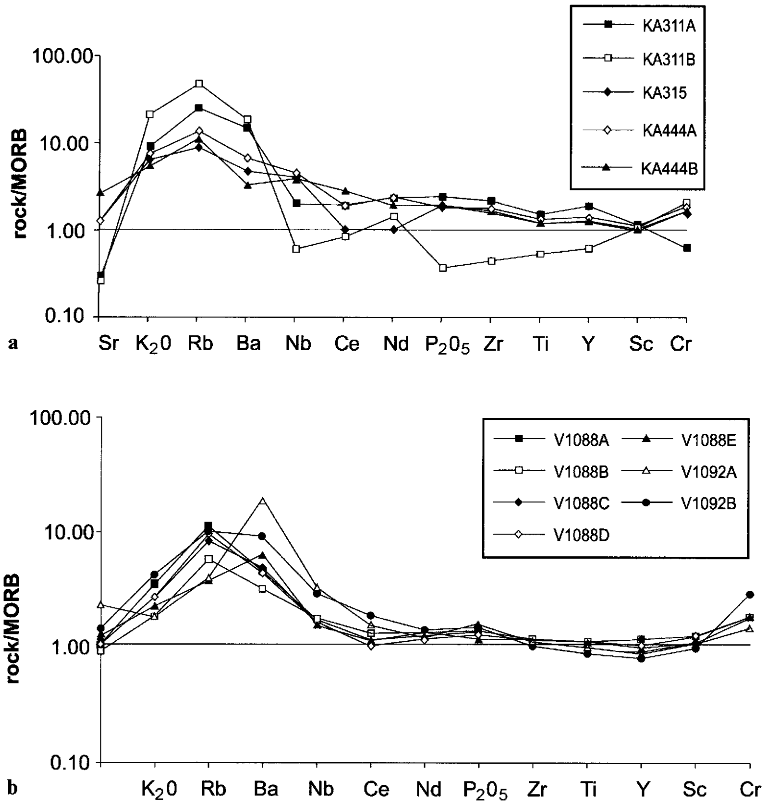


Fig. 5. (**a**) MORB-normalized 'spidergrams': (**a**) amphibolites from the Klissochori Unit (Central Almopias Zone); (**b**) basalts from the Late Triassic Vryssi Unit, easternmost Central Almopias Zone. Both the amphibolites and basalts are geochemically 'enriched' and are attributed to Triassic rifting of the Vardar Ocean. (See Table 1 for representative analyses.)

to the north (Şengör 1984), or northward subduction related to subduction in the South Aegean region (Stampfli *et al.* 2001, 2003; Stampfli & Borel 2002); (3) formation of a rifted small ocean bordering the Serbo-Macedonian Zone (Stais & Ferrière 1991; Dimitriadis & Asvesta 1993; Stais 1994; Brown & Robertson 2003, 2004).

Option (1) (Palaeotethys within Vardar) now seems unlikely, as there is no obvious evidence of any Palaeotethyan units actually within the Vardar Zone. Option (2), southward subduction from the north, has been tested based on studies of the Pontides in northern Turkey (Ustaömer & Robertson 1997; Ustaömer *et al.* 2005; see also Okay *et al.* 2001) and has been found to be invalid. Option (3), northward subduction from the south, has also now been tested based on studies of the South Aegean region and Crete, and is also now known to be invalid (Robertson 2006*b*; see also Smith 2006). The igneous and sedimentary evidence from the Vardar Zone as a whole is explicable in terms of the formation of a Triassic small ocean basin. In this interpretation the Pelagonian Zone formed part of Gondwana in Late Palaeozoic time, but later rifted away opening up a small ocean basin bordering both the western (Pindos) and eastern (Almopias–Vardar) margins of a Pelagonian microcontinent. This is comparable with the inferred Triassic rifting of the Tauride–Anatolide microcontinent from Gondwana further east in southern Turkey (Robertson *et al.* 2004). This interpretation is also consistent with the Serbo-Macedonian Zone being an exotic terrane that was amalgamated to Eurasia only during its Alpine history (Himmerkus *et al.* 2006). It is possible that a fundamental Palaeotethyan suture is located within units related to the Serbo-Macedonian and Rhodope zones and that all units to the south of this suture are Gondwana-derived.

Table 1. *Major and trace element XRF geochemical analyses of igneous lithologies. Major elements are in weight-percent oxide and trace elements in ppm.*

	Klissochori 1 KA311B	Klissochori 2 444A	Vryssi 3 Vio88A	Mélange 4 1142C	Mélange 5 12a	Mélange 6 11	Krania 7 K1042F	Krania 8 K944B
SiO_2	48.6	46.91	50.71	44.28	47.33	49.8	52.48	52.55
Al_2O_3	16.13	15.68	17.73	14.01	14.46	9.38	14.24	14.87
Fe_2O_3	9.09	12.08	9.01	9.44	12.43	8.01	11.97	10.66
MgO	9.57	6.52	6.75	9.06	9.82	13.11	6.62	3.35
CaO	7.87	9.84	2.14	17.94	7.91	16.76	5.53	5.87
Na_2O	1.57	2.51	5.51	0.22	2.61	0.55	4.48	2.39
K_20	3.22	1.16	0.52	0.95	1.29	0.25	0.14	0.55
$Ti0_2$	0.81	2.01	1.64	1.26	1.49	0.53	1.58	0.84
MnO	0.15	0.15	0.44	0.21	0.19	0.19	0.17	0.18
P_2O_5	0.04	0.22	0.16	0.16	0.18	0.05	0.18	0.07
LOI	2.52	2.39	4.84	3.23	2.34	1.98	3.4	9.06
Total	99.59	99.49	99.37	100.76	100.05	100.65	100.48	100.39
Ni	166	187	197	216	85	188	38	1
Cr	501	465	250	469	249	679	59	20
V	229	458	339	267	250	320	376	386
Sc	21	46	49	39	40	44	41	36
Cu	30	47	81	34	103	18	38	8
Zn	73	120	84	87	87	57	137	91
Sr	35	158	121	22	179	180	141	115
Rb	95	27	22	29	26	4	6	21
Zr	41	156	105	89	126	19	155	60
Nb	2	16	6	10	7	2	7	3
Ba	364	132	91	453	165	165	18	77
Pb	2.7	3	0	3	0.6	1	1	2
Th	b.d.	1	2	1	0.1	0	3	3
La	3	0.2	5	0	1.4	1	5	6
Ce	9	19	11	19	22	10	16	15
Nd	11	19	10	9	17	10	15	9
Y	18	41	32	25	31	14	52	18

LOI, loss on ignition; b.d., below detection limit.

Triassic–Jurassic: passive margin subsidence and ocean genesis

Mid-Triassic to Early Jurassic time was characterized by the formation of oceanic crust within the Vardar Zone. This oceanic crust has since been mainly subducted, but was located within the Eastern Almopias Zone (i.e. Vryssi Unit of Stais *et al.* 1990; Fig. 2b). Bordering continental units are represented by the Pelagonian Zone to the west and by the Serbo-Macedonian Zone plus the Paikon Zone to the east. However, it should be noted that significant strike-slip may have occurred such that these units did not necessarily face towards each other in the Early Mesozoic, as today. The Pelagonian Zone and the Western and Central Almopias zones were characterized by subsiding carbonate platforms during Late Triassic–Early Jurassic time. Thick supra-, inter- and sub-tidal limestone–dolomite loferite cycles accumulated within the Pelagonian Zone (e.g. Kaimakchalan Massif; Fig. 3) beginning in Mid–Late Triassic time. Despite deformation and regional greenschist-facies metamorphism, primary facies are locally well preserved. The supratidal facies typically are red marly limestones, whereas the intertidal facies are dolomite and limestone with well-developed laminar algal stromatolites (with fenestral fabrics) and sheet-prism shrinkage cracks, whereas the subtidal facies are typically dolomitic with Dasycladacean algae (e.g. *Griphoporella curvata*) and rare Megalodonts.

Evidence from the Klissochori Unit (e.g. the Rhizarion marbles) in the Central Almopias Zone (Figs 2b and 4) and from the Voras Massif further north (Loutra Arideas Unit; Migiros & Galeos 1990; Brown & Robertson 2004; Fig. 3) indicates the presence of an eastward-deepening, mixed carbonate–clastic slope sequence. Triassic?–Jurassic algal and fenestral (loferitic) marbles are also seen in the Livadia Unit of the Voras Massif. The Livadia Unit, together with the 'Rhizarion marbles' further south, formed an isolated area of platform sedimentation. This unit was apparently rifted from the main Pelagonian carbonate platform to the west to form a marginal fault block bounded by a rift basin, represented by the Loutra Arideas Unit of the Voras Massif (Fig. 3). Redeposited and hemipelagic deep-water basinal carbonates and subordinate clastic deposits are also present within the Pelagonian Zone, as represented by the undated Kato Grammatiko Formation of Sharp (1994), and can be interpreted as deposits within an intra-platform basin, located within the Pelagonian platform or along its eastern margin. Comparable intra-platform basins are recognized in the eastern Pelagonian Zone elsewhere in Greece, notably in the Argolis Peninsula (Clift & Robertson 1990*a*).

Triassic oceanic crust is preserved as tiny slices (< 5 m thick) of MOR-type pillow basalts within the Vryssi Unit in the most easterly part of the Central Almopias Zone, just beneath the basal thrust of the Eastern Almopias Zone (to the east of the Nea Zoi Unit; Figs 2 and 5b). The basalts are depositionally overlain by several metres of ribbon radiolarite of Late Triassic age (Stais *et al.* 1990).

Within the Paikon Zone to the east a metamorphosed Triassic?–Jurassic, mixed carbonate–clastic sequence (Gandatch Formation) is interpreted as a deep-water equivalent of clastic sediments within the Peonias Zone (Brown & Robertson 2003). Further east again, successions exposed along the SW margin of the Serbo-Macedonian Zone and the eastern adjacent Peonais Zone document a subsiding SW-facing carbonate platform during Anisian–Carnian time (Stais & Ferrière 1991; Dimitriadis & Asvesta 1993; Stais 1994). These platform carbonates are overlain by westward-deepening slope to basinal, mixed carbonate–clastic facies (Stais 1994). Associated volcanic rocks are of WPB to MORB type (Dimitriadis & Asvesta 1993). The sequence extends into the Early Jurassic as thick siliciclastic turbidites (Svoula Flysch; Kaufmann *et al.* 1976; Kockel *et al.* 1977). During Late Triassic–Early Jurassic time the floor of the Peonais basin between the Paikon and Serbo-Macedonian continental units is likely to have been represented by stretched continental crust.

In summary, the western margin of the Vardar Zone and the adjacent eastern Pelagonian Zone document Triassic rifting, then Jurassic passive margin subsidence related to the opening of a MOR-type oceanic basin to the east (i.e. the Almopias Ocean).

Early–Mid-Jurassic: eastward subduction of the Almopias Ocean

It is widely believed that the Almopias Ocean was subducted northeastwards beneath the Serbo-Macedonian margin during Early–Mid-Jurassic time (Vergély 1984; Bébien *et al.* 1986, 1987; Brown & Robertson 1994, 2003, 2004). The evidence for this is seen in the more easterly Vardar zones (i.e. the Paikon, Peonais and Serbo-Macedonian zones), outside the present study area (Fig. 1). This inferred subduction resulted in arc volcanism within the leading edge of the Serbo-Macedonian margin, represented by Paikon Zone (Bébien *et al.* 1980, 1994; De Wet *et al.* 1989; Brown & Robertson 1994, 2003, 2004). Contrary to recent suggestions of an oceanic arc origin (Stampfli *et al.* 2001), these volcanic rocks are seen to overlie continental crust within the Voras Massif (Mercier 1968; Brown & Robertson 2004; Figs 2a and 3). Back-arc extension is believed to have reactivated the inferred Peonais rift basin to form an intra-continental back-arc basin in which the Late Jurassic Guevgueli Ophiolite formed (Bébien *et al.* 1987; Mussalam 1991; Danelian *et al.* 1996; Brown & Robertson 2003, 2004). The Guevgueli Ophiolite retains primary intrusive contacts with adjacent metamorphic rocks, correlated with the Serbo-Macedonian Zone (De Wet *et al.* 1989; see Smith 1993). This suggests that this ophiolite is para-autochthonous with respect to the Serbo-Macedonian continental margin to the east.

Mid-Jurassic Pelagonian platform-margin collapse

Mid–Late Jurassic time (pre-Oxfordian–Early Kimmeridgian) was characterized by the collapse of the Pelagonian–Almopias carbonate platform and a transition to deeper-water hemipelagic sediments. The top of the Pelagonian platform succession (e.g. at Arnissa; Figs 3, 4 and 6) is gradationally overlain by a sequence (*c.* 20 m thick) of interbedded siliceous, micaceous and chloritic schists, ribbon cherts, cherty hemipelagic carbonates and siliciclastic sandstones (Arnissa Passage Beds Member of Sharp 1994). A less obvious transition is observed where basinal sediments previously existed (i.e. Kato Gramatiko Formation). Sedimentary structures in the upper part of the Arnissa Passage Beds Member are indicative of deposition by turbidity currents. Petrographic observations reveal an incoming of volcanic quartz, devitrified volcanic rocks and also of detrital chromite near the top of the sequence. X-ray diffraction

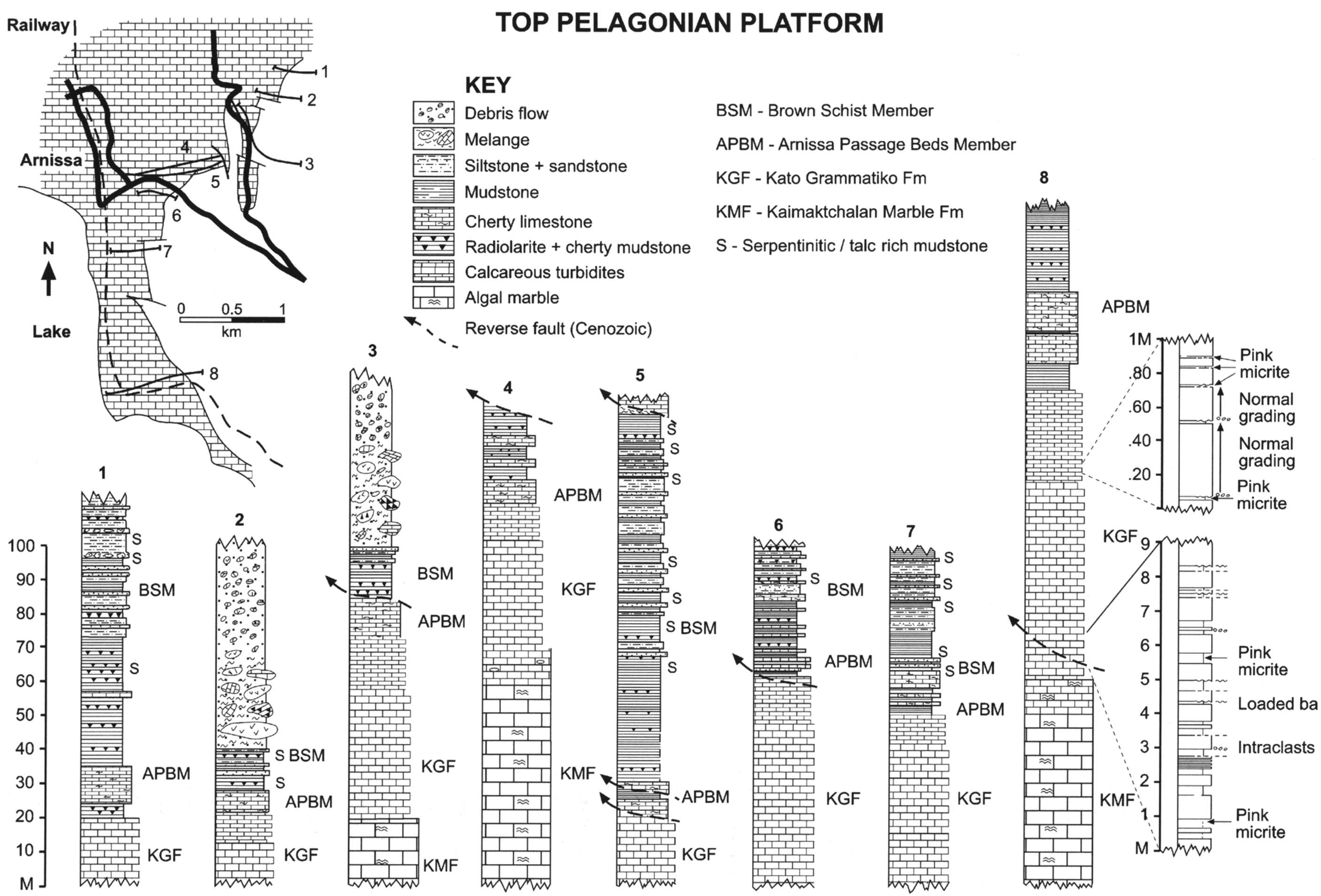


Fig. 6. Stratigraphic sections through the top of the Pelagonian platform in the Arnissa area, recording the collapse of the carbonate platform and the transition to a foredeep during ophiolite emplacement. The detailed logs in section 8 show calciturbidite facies (Kato Grammatiko Formation). (See Fig. 3 for the location of Arnissa.)

analysis of fine-grained sediments revealed an initial dominance of muscovite, chlorite (ripidolite and clinochore), quartz, albite, minor sphene and epidote, followed by an incoming of chromite and talc (serpentinite). This sequence is interpreted to record flexural subsidence and collapse of the Pelagonian platform ahead of emplacing ophiolitic units (Sharp *et al.* 1991). Similar sequences that can be related to platform collapse are known from the Pelagonian Zone elsewhere in Greece, including Vourinos (Zimmerman 1972; Naylor & Harle 1976; Vergély 1984), Othris (Smith *et al.* 1975; Ferrière 1976), Argolis (Baumgartner 1985; Clift 1992) and Evia (Euboea) (Robertson 1991).

Mid–Late Jurassic emplacement of ophiolitic debris flows and mélange

Within the eastern Pelagonian Zone (e.g. Arnissa area; Fig. 4), the platform-collapse sequence passes gradationally upwards into a largely ophiolite-derived, chaotic sedimentary sequence, the Brown Schist Member (Bijon 1982; Sharp 1994; Figs 6 and 7). This is overlain by ophiolitic mélange and by dismembered ophiolitic thrust sheets (Mercier & Vergély 1972; Bijon 1982). The lower part of the succession includes sandstones, mudstones and conglomerates of mainly epiclastic (detrital), basic igneous and subordinate siliciclastic origin. The higher levels are dominated by ophiolitic lithologies (e.g. serpentinite), mainly debris flows and turbidites within an argillaceous sequence, rich in talc, chromite and other ophiolite-derived material. Rare carbonate debris flows (e.g. Mavro Rema section; Fig. 8) associated with blocks of marble (commonly internally brecciated), are interpreted as talus shed from the collapsing Pelagonian carbonate platform. The debris flows as a whole are interpreted as the infill of a foredeep related to downflexure of Pelagonian continental crust ahead of an emplacing ophiolite. Similar collapse and related foredeep sequences are documented elsewhere in the Pelagonian Zone further south (e.g. in Othris, Evia and Argolis: e.g. Robertson *et al.* 1991). Similar features are also associated with Late Cretaceous Turkish ophiolites (Parlak & Robertson 2004), northern Syrian ophiolites (Al-Riyami & Robertson 2002), and with many other Tethyan settings, including Oman (Lippard *et al.* 1986; Robertson 1987, 2006*a*).

Upwards, the chaotic foredeep-type debris is covered by an ophiolitic mélange. This is made up of blocks or sheets of serpentinite ultramafic rocks (mainly harzburgite and dunite), with subordinate gabbro, basalt, diabase, amphibolite and recrystallized red ribbon chert, set within a strongly sheared, scaly serpentine-rich matrix. Good examples of this mélange are exposed in the Arnissa area and in the Vermion Mountains to the south (Figs 3 and 8). In places (e.g. Arnissa and Reis Tsifliki areas), the mélange is in direct tectonic contact with large masses of ultramafic rocks (hundreds of metres to several kilometres in size), although the strong Palaeogene deformation makes it difficult to determine original emplacement relationships.

The mélange is locally absent (e.g. Mavri Rakhi region of the Vermion Mountains; Fig. 8), where the Pelagonian carbonate platform is directly overlain by a large slice of serpentinized ultramafic rock, with amphibolite welded to its base. This amphibolite is interpreted as the remnant of a metamorphic sole.

Geochemical studies show that the blocks of extrusive rocks within the mélange range in composition from alkali basalts with WPB signatures (containing titanaugite), to subalkaline tholeiitic basalts with MORB affinities, to island-arc tholeüte (IAT)-type basalts; also, two samples exhibit boninitic affinities (Fig. 9a). This chemical variation suggests that the mélange basalts were derived from several different tectonic settings and were probably mixed within a subduction complex before being emplaced within the mélange. The IAT-type basalts and boninitic lavas are similar to those reported from the Vourinos Ophiolite (e.g. Bortolotti *et al.* 2004; Rassios & Moores 2006), whereas the alkali basalts might record either rift or accreted seamount basalts, and the MORB, remnants of Triassic–Early Jurassic oceanic crust. The predominance of harzburgite in the mélange blocks and thrust sheets is suggestive of a supra-subduction origin, in common with the Vourinos Ophiolite and many of the other Jurassic Balkan ophiolites (Pearce *et al.* 1984; Robertson *et al.* 1991; Rassios *et al.* 1994; Clift & Dixon 1998; Rassios & Smith 2000).

The mélange includes blocks and dismembered thrust sheets that can be closely compared with ophiolitic mélange exposed in the Voras Massif further NE (Fig. 3). A mixed carbonate–siliciclastic sequence, of Triassic–Jurassic age, is gradationally overlain by ophiolite-derived turbidites and debris-flow deposits (Loutra Aridea Unit; Migiros & Galeos 1990; Brown & Robertson 2004). In addition, ophiolitic rocks overlie metacarbonate platform rocks within the Western and Central Almopias zones (Fig. 4); these are mainly thin (<200 m) units of harzburgite, dunite, serpentinite and amphibolite (see below).

The lithologies within the ophiolitic mélange in the area studied are closely comparable with

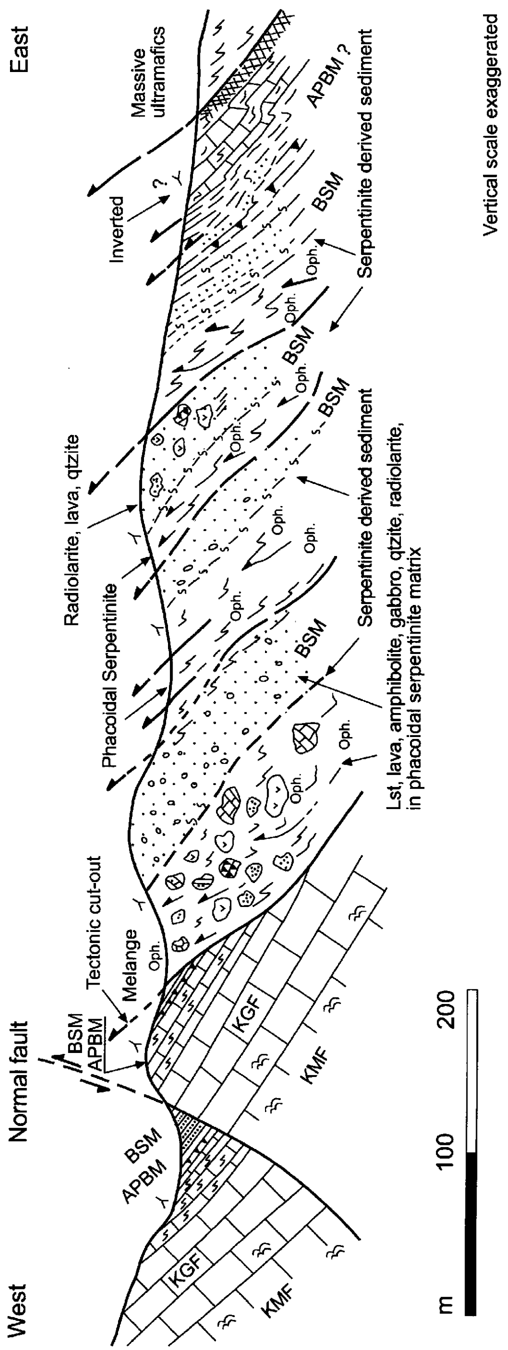


Fig. 7. Schematic cross-section through the uppermost levels of the Pelagonian carbonate platform in the Arnissa area showing the transition to ophiolitic mélange. The section was taken east of log 3 (Fig. 6). The thrust imbrication is mainly of Early Cenozoic age. Key as in Figure 6. (See text for explanation.)

the Vourinos Ophiolite and the accretionary Avdella Mélange beneath the Pindos Ophiolite further west (Kostopolous 1989; Jones & Robertson 1991; Fig. 1). The Vourinos Ophiolite and its locally preserved amphibolitic sole (Pichon & Brunn 1985; see Rassios *et al.* 1994 Rassios & Moores 2006) forms part of a regionally extensive thrust sheet above the Vourinos accretionary mélange. The mélange, as exposed in the eastern Pelagonian Zone (e.g. Arnissa area), is a mixture of chaotic mélange and ophiolitic thrust sheets. A very similar mixture is, for example, found within the mélange that lies beneath the Late Cretaceous Mersin ophiolite in southern Turkey (Parlak & Robertson 2004). Also, in northern Syria, the Late Cretaceous Baër–Bassit ophiolite is pervasively imbricated with accretionary material (Al-Riyami & Robertson 2002). In both of these areas this type of mélange and disrupted ophiolite thrust sheets developed in the frontal zones of an emplacing ophiolite, and a similar setting is inferred for the Pelagonian ophiolitic mélange. Palaeogene deformation of the Pelagonian and Almopias zones has resulted in thrust imbrication and further disruption of the mélange, with the serpentinite-rich intervals acting as décollement horizons (Figs 7 and 10).

Processes of ophiolite emplacement

One view is that all of the ophiolites were emplaced westwards from the Vardar Zone, (Mercier *et al.* 1975); another is that ophiolites were emplaced onto the Pelagonian Zone, towards the NE from a Pindos Ocean (Smith *et al.* 1975; Robertson *et al.* 1991; Doutsos *et al.* 1993), and a further option is that ophiolites were emplaced from both the Pindos Zone (e.g. Vourinos–Pindos) and from the Vardar Zone ('Vardar ophiolites') (Mountrakis *et al.* 1987). There are two main lines of evidence that could help to discriminate between these alternatives in the area studied: (1) structural data from the Pelagonian platform and the overlying ophiolitic mélange; (2) evidence from the age of the transgressive sedimentary cover. The effects of Early Cenozoic westward thrusting must also be considered.

During this work it was found that the entire Pelagonian platform and counterparts

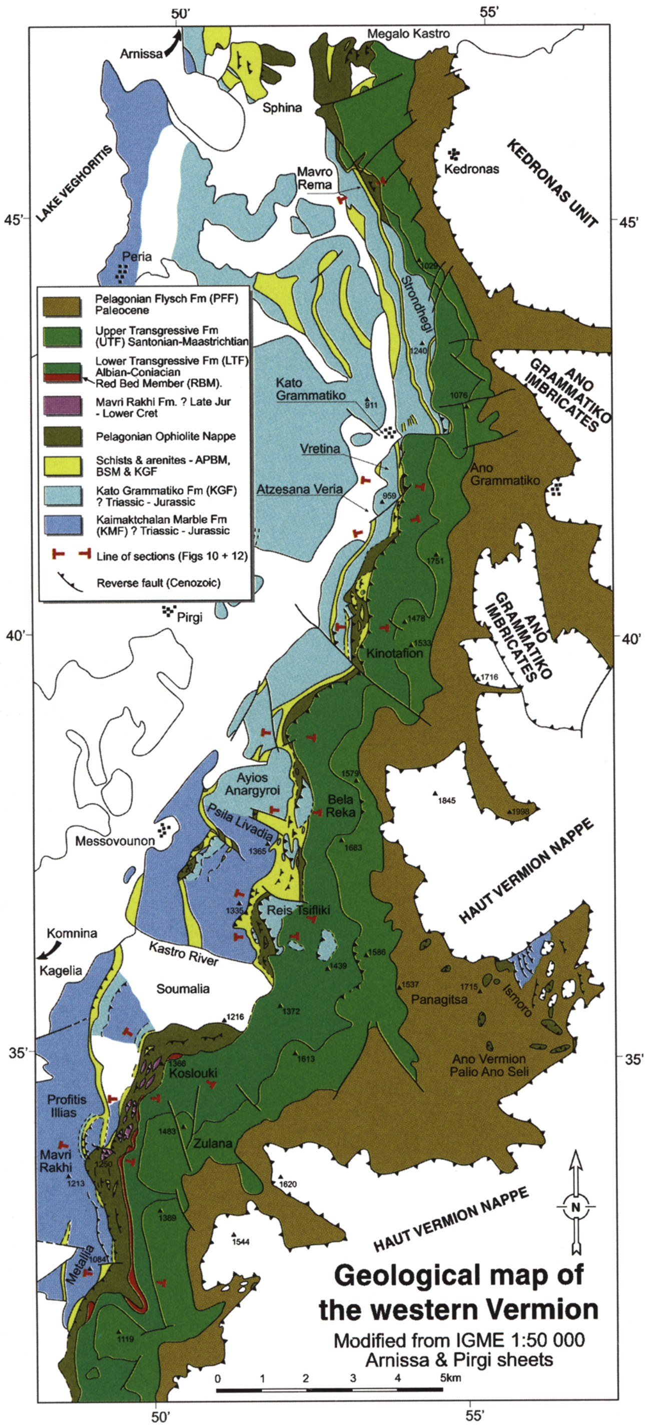


Fig. 8. Outline geological map of the Western Vermion Mountains. The locations of logged ophiolitic mélange in the Western Vermion Mountains (Fig. 10) and at Mavri Rakhi (Fig. 12) are shown. The overall structure relates to westward Palaeogene thrusting. The general location is shown in Figures 1 and 2a. Modified from IGME Arnissa and Pirgi sheets.

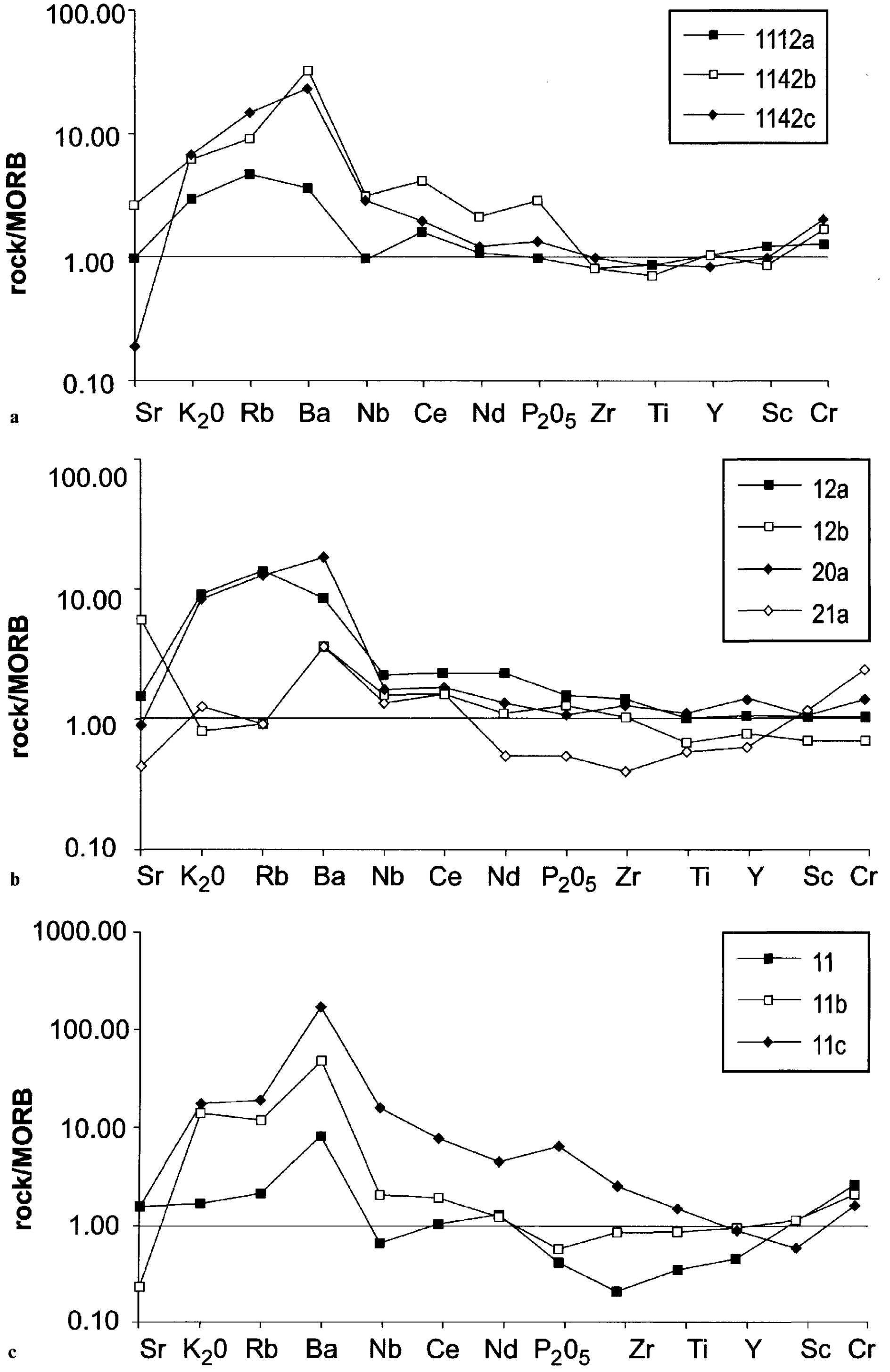


Fig. 9. MORB-normalized 'spidergrams' of lava and amphibolites from the Pelagonian ophiolitic mélange. (**a**) Basalts from the Arnissa area; (**b**) basalts from the Reis Tsifliki (Vermion) area; (**c**) amphibolites from Reis Tsifliki area. The 'enriched' composition should be noted. (See Table 1 for representative analyses.)

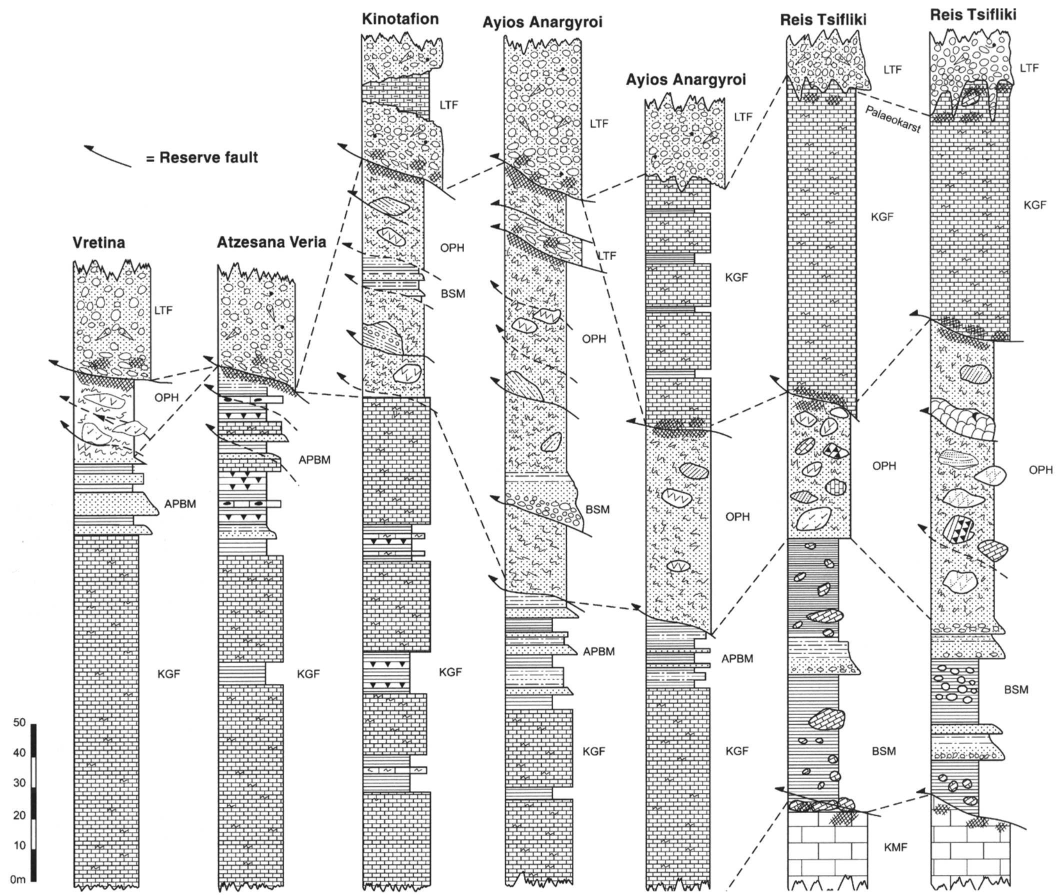


Fig. 10. Logged sections of the Pelagonian Zone in the Western Vermion Mountains. Location of logged sequences are indicated in Figure 8. The majority of reverse faults are of Palaeogene age. Faulting typically exploits serpentinite as a décollement horizon. Cretaceous cover sediments overthrust underlying units, resulting in locally intensive deformation of the basal Cretaceous limestones and red beds, and intense imbrication of the underlying ophiolitic lithologies. KMF, Kaimaktchalan Marble Fm; KGF, Kato Grammatiko Fm; APBM, Arnissa Passage Bed Member; BSM, Brown Schist Member; OPH, Pelagonian Ophiolite Nappe; LTS, Cretaceous Lower Transgressive Fm. Key as in Figure 6.

within the Western and Central Almopias zones, up to and including the ophiolitic mélange have experienced greenschist-facies, to possibly amphibolite-facies metamorphism, indicating that deep burial of the Pelagonian platform has taken place throughout this area. This was associated with ductile-style folding and the development of a penetrative deformation fabric (D_1) (Mountrakis *et al.* 1984; Kockel 1986; Mercier 1966). Flat-lying to slightly inclined isoclinal to similar folds are thickened in the hinge zones and thinned or sheared-out parallel to the foliation on the limbs. These folds are associated with the development of an axial planar cleavage, which is typically oriented parallel to bedding, and is also related to the formation of a pronounced stretching lineation that is oriented parallel to the B tectonic axis of folds. The trend of this lineation is remarkably constant throughout the study area with an average orientation of N150° (range N180°–N130°) (Vergély 1984; Sharp 1994). The vergence of the observed folds is variable, with both SW–WSW and NE–ENE fold vergence commonly being observed. In a small number of exposures in the Arnissa area C–S fabrics are locally indicative of top-to-the-SE motion (Sharp 1994).

Previous structural studies suggested that the easterly 'internal' zones of northern Greece as a whole, including the Pelagonian, Almopias and Paikon zones, experienced two major phases of deformation and metamorphism; i.e. JE-1 in the Late Jurassic (155–145 Ma) and JE-2 in the Early Cretaceous (130–110 Ma) (Mercier 1966, 1973; Vergély 1984). We confirm the presence (but not the vergence) of a JE-1-type event affecting the Pelagonian Zone and the Western and Central Almopias zones. Vergély (1984) used mainly fold vergence and cross-cutting relations to infer an initial ophiolite emplacement (JE-1) towards the west that was considered to be Late Jurassic in age; this was followed by a second ophiolite emplacement event, also verging towards to the west (JE-2) that was marked by cross-cutting folds and interpreted as being Early Cretaceous in age.

During this study we observed that the fold hinges of similar (ductile) folds are commonly oriented parallel to the stretching direction and thus cannot be used as kinematic indicators. Opposing vergence directions were observed more or less randomly even in local outcrops and cannot be interpreted as successive fold phases. In addition, some of the supposedly JE-2 folds deform Cretaceous facies including Albian–Cenomanian transgressive limestones and conglomerates, showing that, within the regional context, these folds must be post-Cretaceous in age (i.e. related to Cenozoic deformation). For example, many of the large SW-verging folds in the region (e.g. Messovounon and Ayios Dimitrios area; Figs 2 and 8) involve Late Jurassic and Cretaceous (Albian–Cenomanian)-aged sedimentary cover rocks that post-date ophiolite emplacement (Sharp 1994). This contrasts with earlier studies that attributed these particular structures to a regional Late Jurassic westward ophiolite emplacement (e.g. Braud *et al.* 1984; Vergély 1984).

Kilias (1991) described a similar NNW–SSE stretching lineation but without a preferential vergence from several localities in the Arnissa–Edessa area. Further west, near the Vourinos Ophiolite both SSW and NE (D_1) vergences were recorded. Kilias (1991) also reported data from quartz *c*-axis measurements. From the limited data available for this region the positions of the *c*-axis maxima on cross girdles (with reasonably well-defined outlines) are indicative of sinistral displacement, consistent with an easterly component of tectonic transport. In addition, structural studies of the Vourinos Ophiolite, adjacent small ophiolitic bodies and the underlying Pelagonian platform provide evidence of NE-directed displacement (Naylor & Harle 1976; Rassios *et al.* 1994; Rassios & Moores 2006). Rassios (pers. com.) noted that structures in the Pelagonian Zone south of the Vourinos Ophiolite (e.g. in the Triassic section south of the Aliakmon River and continuing into the Aliakmon area) exhibit kinematic indicators that indicate movement to the NE. NE-directed emplacement is also inferred for the Avdella Mélange beneath the Pindos Ophiolite further west (Jones & Robertson 1991; Rassios & Moores 2006). Structural evidence also supports generally east-directed emplacement of ophiolites onto the Pelagonian Zone, including Othris (Smith *et al.* 1979), Evia (Robertson 1991) and Argolis (Clift & Robertson 1990*a*, *b*), and also from Albania (e.g. Robertson & Shallo 2000) and former Yugoslavia (see Robertson & Karamata 1994; Karamata 2006).

The structural results for the region studied (e.g. Edessa–Arnissa) can also be compared with information from the region further south (south of the Aliakmon River) including the High Pieria, Olympos, Ossa and NW Thessaly area (Fig. 1). A pervasive stretching lineation with a variable trend (NW–SE or NNW–SSE) was reported from the Pelagonian Zone in NW Thessaly (Sfeikos *et al.* 1991; Sfeikos 1992), High Pieria (Yarwood & Aftalion 1976), Livadi (Nance 1981) and Olympos (Barton 1976; Kilias *et al.* 1990; Schermer *et al.* 1990; Doutsos *et al.* 1993; Schermer 1993; Fig. 1). Based on a kinematic study of the region as a whole, Wallcott (1996) and Wallcott & White (1998) observed that a SSE-directed stretching lineation (D_1) is well developed in NW Thessaly (as above) but becomes weaker and disappears southwards. Kilias *et al.* (1990) reported quartz *c*-axis data (i.e. maxima on crossed girdles) that indicate sinistral (eastward) transport for several localities in the NW Pieria mountain area, whereas the results from areas further south are indicative of dextral (west-directed) tectonic transport. In general, the metamorphic grade (M_1) ranges from amphibolite facies at high structural levels to amphibolite or upper greenschist facies at lower structural levels (Schermer *et al.* 1990; Walcott 1996).

Following the D_1 event, Walcott (1996) identified a weak east–west-trending fabric (D_2) in some areas of Thessaly that shows evidence of east-directed shearing; she related this either to continued thrusting or to an Early Cretaceous extensional event. The subsequent fabric (D_3) is indicative of top-to-the-SW tectonic transport, of inferred mid-Cretaceous to Early Cenozoic age according to most workers (e.g. Schermer 1993; Walcott 1996; Walcott & White 1998), punctuated by a phase of top-to-the-NE shearing (Lips *et al.* 1998). D_3 is associated with regional HP–LT metamorphism (M_2) (Schermer *et al.* 1990; Schermer 1993) and is attributed to generally eastward subduction beneath the Pelagonian continent.

Late Jurassic transgressive deposition

The deformed Eastern Pelagonian and Western Almopias zones are typically transgressed by cover sediments of mid-Cretaceous (Albian–Cenomanian) age (Mercier 1966; Sharp 1994) thus providing only limited constraints on the timing of the D_1 deformation and M_1 metamorphism. However, important fossil evidence from the Almopias Zone shows that transgression there began in the Late Jurassic. Specifically, in the Kerassia and Kedronas units of the Western Almopias Zone (e.g. Nission and Kedronas area; Figs 2b and 3), the basal sediments, which lie unconformably on ophiolitic units in some places, include reef limestone with corals of Late Oxfordian–Early Kimmeridgian age. Coralline fauna present (Fig. 11) include *Stylosmilia* cf. *michelini, Thecosmilia* cf. *langi, Cladocoropsis mirabilis, Dermosmilia* sp. and *Schizosmilia* cf. *rollieri*. Similarly, the ophiolitic mélange of the Loutra Arideas Unit further NE in the Voras Massif (Fig. 3) is unconformably overlain by shallow-water limestones, containing corals and *Cladocoropsis* sp. of Late Jurassic (Kimmeridgian?) age (Galeos *et al.* 1984; Brown & Robertson 2004). These relationships show that ophiolitic rocks were emplaced and eroded subaerially prior to, or during, Oxfordian time. Assuming that the underlying deformed ophiolites, represent part of the regionally emplaced Jurassic ophiolites these results constrain the emplacement of the Pelagonian ophiolitic mélange and the D_1 deformation as pre- to syn-Late Oxfordian–Early Kimmeridgian.

Within the Eastern Pelagonian Zone (near Mavri Rakhi; Fig. 8), the ophiolitic mélange is depositionally overlain by low-grade metamorphosed deep-water carbonates, known as the Mavri Rakhi Formation (Pichon 1976, 1977; Fig. 8). This unit begins with interbedded green arenites, cherts and mudstones (*c.* 20 m thick), passing gradationally upwards into siliceous carbonates (*c.* 30 m thick) (Fig. 12). Although lacking age-definitive fossils, calcispheres, aptychi, bivalves and possible calpionellids are present, suggesting a Late Jurassic–Early Cretaceous age. This unit is unconformably overlain, with a minor discordance, by Albian–Cenomanian limestones. The facies and stratigraphic position of the Mavri Rakhi Formation are very similar to the well-dated Late Jurassic–Early Cretaceous pelagic carbonates, which depositionally overlie the Vourinos Ophiolite (see Rassios & Moores 2006) and the Eastern Albanian ophiolites (e.g. Robertson & Shallo 2000). The Vourinos Ophiolite (e.g. Siatista, Krapa and Zygosti areas) exhibits a Late Jurassic–Early Cretaceous transgressive cover of slope to deep-water carbonate sediments that sit, with a minor angular discordance, on the ophiolitic extrusive rocks (Pichon & Lys 1976; Mavrides *et al.* 1979). A similar age and setting were inferred for the equivalent, Eastern-type Mirdita ophiolites in Albania (see Robertson & Shallo 2000).

Implications of structural and age information

The results from the area studied (Fig. 1) indicate that the pervasive D_1 fabric, characterized by the pervasive SSE–NNW stretching lineation, is of pre- to syn-Late Oxfordian–Early Kimmeridgian age. By contrast, in NW Thessaly, south of the Aliakmon River, which marks an important transverse discontinuity, radiometric dating of the early deformation fabric has yielded Early Cretaceous ages (130–100 Ma), i. e. 119 ± 15 Ma from granites in the Pieria region; 101 ± 13 Ma from augen schist from the Infrapierian unit (Yarwood & Dixon 1977); 124 ± 4 Ma and 123 ± 11 Ma from mylonitic Pierian granites

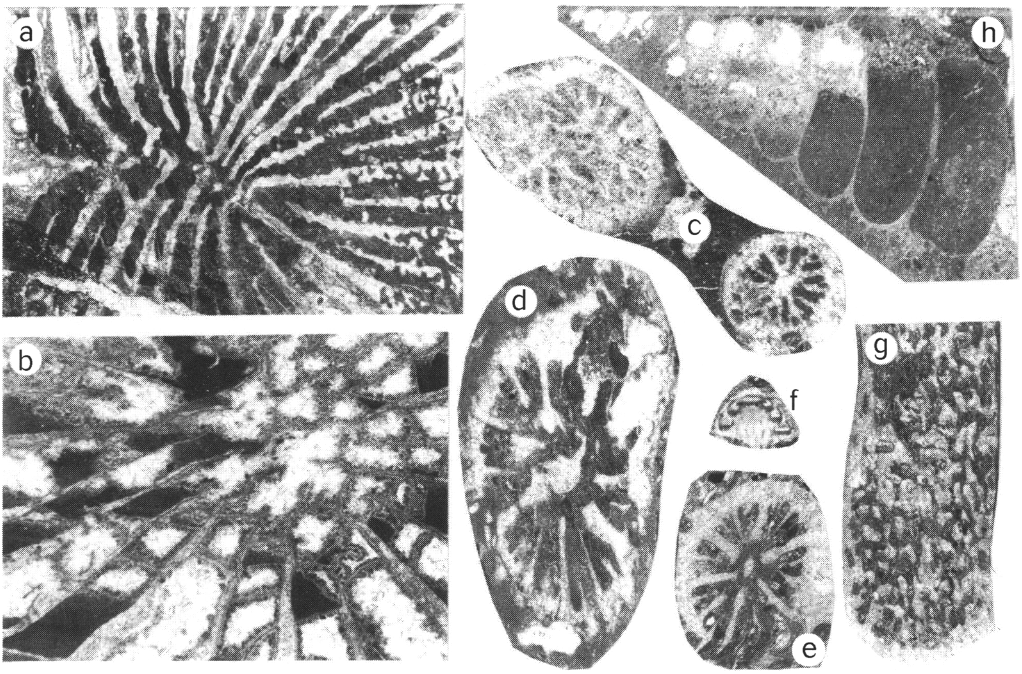


Fig. 11. Fauna from reefal limestone developed within the Basal Kerassia Unit, Western Almopias Zone, Nission area. **(a)** *Dermosmilia* sp. (Late Oxfordian–Early Kimmeridgian); **(b)** *Thecosmilia* sp. (Late Oxfordian–Early Kimmeridgian); **(c)** *Schizosmilia* cf. *rollieri* (Oxfordian–Kimmeridgian); **(d, e)** *Stylosmilia* cf. *michelini* (Late Oxfordian–Kimmeridgian); **(f)** *Neotrocholina* or *Trocholina* sp. (Late Oxfordian–Early Kimmeridgian); **(g)** *Cladocoropsis mirabilis* (lagoonal hydrozoan) (Oxfordian–Early Kimmeridgian); **(h)** spinose nerinid gastropod with geopetal structure infilled by faecal pellets and micrite at the base and sparite at the top. The samples were identified and dated by B. Rosen, British Museum of Natural History, London.

south of Olympos (Barton 1976) and 98 ± 2 Ma from the Intrapierian unit (Schermer *et al.* 1990). Assuming that the D_1 fabrics are contemporaneous north and south of the Aliakmon River, as is likely (but not proven), it is possible that the age of the D_1 deformation in NW Thessaly (e.g. nothern Pieria) was also pre- to syn-Late Oxfordian–Early Kimmeridgian (*c.* 155 Ma); if so, the Early Cretaceous radiometric dates could represent cooling ages. It is notable that the geological evidence from the Almopias Zone in the area studied (Western Vardar Zone margin) is indicative of an important Late Jurassic–Early Cretaceous extensional (or transtensional) phase (see below); this extension could have exhumed the Pelagonian basement and set the Early Cretaceous radiometric ages recorded from the Pelagonian Zone.

We are unable to confirm a 'JE-2' Early Cretaceous regional compressional event affecting the area. The Late Jurassic base of the transgressive cover of the Western Almopias Zone passes upwards into a Cretaceous succession without any intervening contractional or metamorphic event. Also, there is no evidence of the detritus expected if thick ophiolitic or other thrust sheets were emplaced westwards from the Vardar Zone onto the Pelagonian Zone during the Early Cretaceous. In addition, ophiolite-related units are absent from the Paikon Massif within the Vardar Zone. The Paikon Massif experienced ductile deformation and, according to Baroz *et al.* (1987), HP–LT metamorphism, associated with an early penetrative structural fabric (Vergély 1984; Brown & Robertson 1994). This fabric is sealed by Kimmeridgian limestones (Kromni Limestones; Brown & Robertson 2003), which pass upwards into Late Jurassic–Early Cretaceous clastic sediments (Ghrammos Formation) and then into platform carbonates (Cretaceous Transgressive Limestones) without any intervening Early Cretaceous JE-2 type event. Further east, the inferred back-arc basin, represented by the Guevgueli Ophiolite in the Peonais Zone, was uplifted and unconformably overlain by Late Jurassic–Early Cretaceous

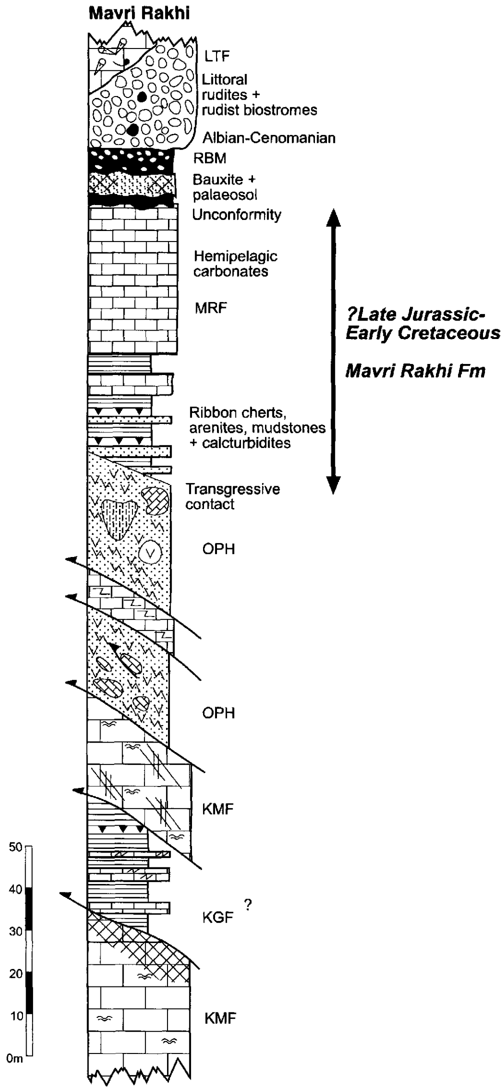


Fig. 12. Sketch section of the ? Late Jurassic–Early Cretaceous Mavri Rakhi Formation in the Western Vermion Mountains (see Fig. 8 for location). KMF, Kaimakchalan Marble Fm; KGF, Kato Gramatiko Fm; MRF, Mavri Rakhi Fm; OPH, ophiolite; RBM, Red Bed Member; LTF, Lower Transgressive Fm.

sediments (Mercier 1966; Bébien *et al.* 1987; Stais 1994) again without evidence of Early Cretaceous metamorphism.

A number of small meta-ophiolites are distributed throughout north–central Greece, including the High Pieria and Olympos areas (see Pe-Piper & Piper 2002). Schermer (1993) noted that with the exception of the Livadi ophiolitic complex these lack a well-defined D_1-type fabric. The ophiolites, as seen in the High Pieria area (Kilias *et al.* 1990), are dismembered and thrust over Pelagonian rocks. As they lack the well-defined D_1 deformation it is likely that they were emplaced, deformed and metamorphosed post-D_1. Early to mid-Cretaceous westward emplacement (Jacobshagen *et al.* 1978) is unlikely, as extension (or transtension) characterized the eastern Vardar Zone during this time (see below). Final emplacement of these ophiolites from the Vardar Zone during Early Cenozoic time is preferred.

The presence of a pelagic sediment cover, of probable Late Jurassic–Early Cretaceous age, in the western part of the Pelagonian Zone (Mavri Rakhi Formation) suggests that an implied thick overburden of the preserved meta-ophiolitic mélange (more than several kilometres of ophiolite?) was removed after its emplacement. As the transgressive sediments are relatively deep-marine it is unlikely that the structural overburden was removed simply by erosion, as there is no evidence of a non-marine basal conglomerate; it is instead more likely to have been removed by extensional (detachment) faulting, allowing exhumed ophiolitic mélange to be directly transgressed by relatively deep-marine carbonates. In addition, debris flows ('olistostromes') that are present within the Late Jurassic–Early Cretaceous aged pelagic carbonate succession overlying the Albanian (Mirdita) ophiolites may also relate to exhumation. These debris flows overlie the radiolarian chert cover of the Mirdita ophiolites. They include blocks of ophiolitic, continental margin and basement rocks. This unit is constrained as later than the Mid-Callovian to Early Oxfordian age of the radiolarian cover sediments but earlier than the Late Tithonian–Late Valanginian age of overlying calpionellid limestones (Bortolotti *et al.* 1996). As a result of the exhumation, material beneath the ophiolite could have been exposed on the seafloor and reworked oceanwards, giving rise to the observed multiple debris flows of Late Jurassic–Early Cretaceous age. This is an alternative to the genesis of these debris flows during the initial ophiolite emplacement onto the Korabi (Pelagonian) continental margin, as suggested by Robertson & Shallo (2000).

Mode of ophiolite emplacement, related deformation and exhumation

Taking account of the available evidence we propose the following hypothesis. The elongate Pelagonian micrcontinent collided with a west-dipping subduction zone within the Pindos Ocean, following Mid-Jurassic genesis of a supra-subduction-type ophiolite (e.g. Vourinos–Pindos;

Liati *et al.* 2004). The ophiolites in the study area were emplaced to the NE and the Pelagonian ophiolitic mélange was shed from the front of the advancing ophiolite (Fig. 13). This initial ophiolite emplacement did not by itself cause thick-skinned deformation or metamorphism of the underlying Pelagonian platform. With continuing convergence and the attempted subduction of Pelagonian continental crust a vast ophiolitic sheet was emplaced over the Pelagonian Zone. This is generally comparable with the latest Cretaceous attempted subduction of the Arabian continental margin, followed by rapid exhumation, as documented in the Oman Mountains south of the Semail Gap (e.g. Miller *et al.* 1998; Searle & Cox 2002). We then infer that a regional switch to strike-slip (transpression) took place during pre-Late Oxfordian–Early Kimmeridgian time (*c.* 155 Ma). This could relate to diachronous trench–margin collision, or to a change in microplate motion triggered by collision. A kilometres-thick competent ophiolitic slab was displaced subparallel to the relatively incompetent Pelagonian platform located beneath, and this was structurally thickened and deformed. This produced the pervasive NNW–SSE, D_1 stretching lineation and induced the amphibolite- or greenschist-facies metamorphism (M_1). We then infer a phase of extension-related exhumation during Late Jurassic–Early Cretaceous time (D_2). This evolutionary stage can be compared, for example, with the much younger (e.g. Mid-Cenozoic) exhumation of the basement of the Menderes Massif in western Turkey from beneath a thick pile of thrust sheets including ophiolites (Hetzel *et al.* 1995; Purvis & Robertson 2004).

A possible trigger for regional exhumation was a reversal in subduction polarity, from generally westwards (oceanwards) in the Mid–Late Jurassic, to generally eastwards (towards the continent). The subduction beneath the Pelagonian continent resulted in subsequent mid-Cretaceous to Early Cenozoic greenschist- to blueschist-facies metamorphism (M_2), associated with dominantly SW displacement (D_3) (Schermer *et al.* 1990).

Latest Jurassic–Early Cretaceous marine transgression

An unconformable contact between the underlying Pelagonian units (Triassic–Jurassic marbles and serpentinized ultramafic rocks) and the overlying cover rocks is exposed throughout the Almopias Zone (Fig. 4). Palaeo-karstic weathering is widely developed, with localized bauxite and laterite accumulations on marble and ophiolitic lithologies, respectively. Fissures in serpentinite were commonly infilled with ophicalcite, which is gradationally overlain by

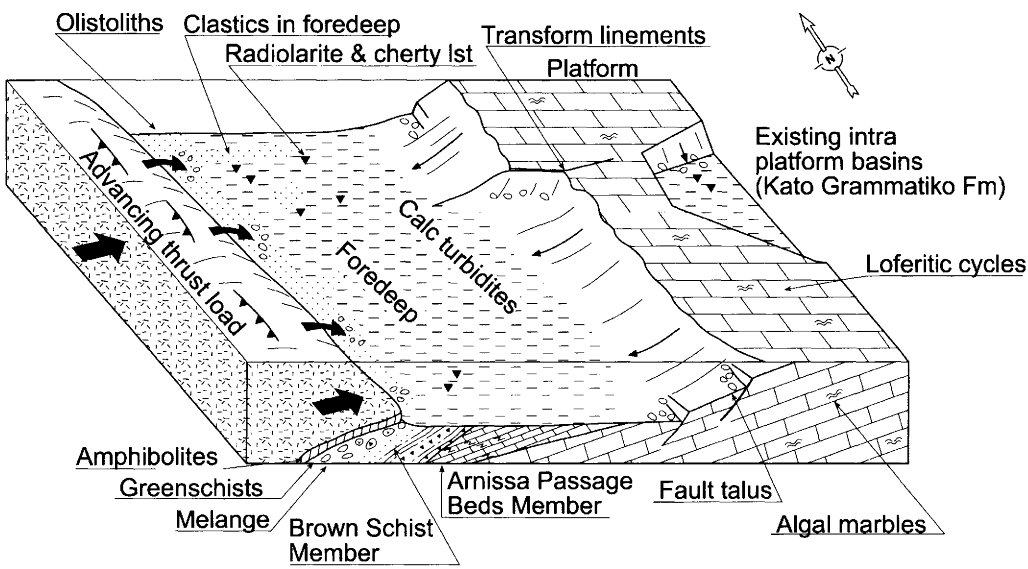


Fig. 13. Schematic block diagram representing the main depositional elements and interpreted depositional setting of the Eastern Pelagonian Zone from the Triassic to Mid–Late Jurassic. Noteworthy features are the northeastward ophiolite emplacement and the setting of the Pelagonian ophiolitic mélange within a foredeep above the collapsed Pelagonian platform. This interpretation applies also to the Vermion area further south.

serpentinite-derived clastic sediments. Higher in the succession clastic sediments are intercalated with intermediate–silicic volcanic rocks, which are at their thickest (200–250 m) in the Central Almopias Zone (Fig. 14). The intermediate-silicic volcanic flows and water-lain tuffs are seen to directly and unconformably overlie deformed ophiolitic lithologies in several sections (e.g. Mauropouli and Petrokorfi; Fig. 14). It is important to note that these volcanic rocks are not part of the underlying Pelagonian ophiolite that was previously emplaced in pre- to syn-Late Oxfordian to Early Kimmeridgian time.

Bijon (1982) previously reported the presence of IAT and depleted boninite-type rocks in the Klissochori and Nea Zoi units. During this study basic meta-igneous rocks, interpreted as detached blocks within the Klissochori Unit, were found to be mainly enriched within plate-type basalts, which we relate to Triassic rifting (see above). However, chemically depleted basalts with a marked negative Nb anomaly on MORB-normalized plots are also present (Fig. 5a).

In places in the Central Almopias Zone (e.g. in the Liki–Margarita and Klissochori units; Figs 2b and 3) the regional unconformity at the top of the deformed and metamorphosed Pelagonian units is dislocated by an important transverse fault zone that trends subparallel to the regionally important Nission Fault (Fig. 3). This area, termed the 'zone de broyage' by Vergély (1984), is characterized by a highly sheared and deformed unit, known as the Klissochori Mélange. This is a chaotic, mainly sedimentary unit that is strongly deformed in the lower part but much less deformed in the upper part. The lower part is clearly unconformable on the underlying Pelagonian platform (Fig. 15) and contains deformed and metamorphosed clasts of many of the rocks exposed in the Pelagonian platform, its basement and the ophiolite. The mélange is intersliced with lenticular sheared serpentinite in places. The mélange clasts are set in a matrix of sericitic and chloritic mudstone and are interpreted as multiple debris flows. The less deformed upper mélange unit contains similar clasts set in a little-deformed matrix, and grades upwards into the typical shallow-marine mixed terrigenous–clastic succession of mid-Cretaceous (Barremian–Aptian) age.

Previously the lower mélange unit was interpreted as being associated with the initial ophiolite emplacement, possibly as a foredeep sequence, whereas the upper mélange was interpreted as part of the Cretaceous cover (Vergély 1984). However, the presence of a definite unconformity between the Pelagonian platform and the ophiolitic mélange, and the overlying debris flows (mélange) indicates that the Klissochori Unit as a whole post-dates D_1 deformation and M_1 metamorphism of the Pelagonian platform and ophiolite. During this work we identified several local successions of the lower and upper mélange units but no overall intact succession and we were unable to confirm that a stratigraphical unconformity exists between two mélange units (see Vergély 1984).

The 'zone de broyage' associated with the Klissochori Mélange appears to correlate with an elongate highly deformed unit ('la bande broyé'), which extends NW–SE from the Western Almopias Zone (Liki–Margarita Unit) southwards through the area between Naoussa and Veria (Fig. 1); this comprises sheared, cataclastic and mylonitic serpentinite ('Veria Ophiolite'), together with blocks including diabase, schist and marble (Braud *et al.* 1984). These units are equivalent to the Western and Central Almopias zones further north. It is important to note that this zone includes reef limestones that were derived from the Late Jurassic cover of the ophiolite, which is locally preserved. This 'zone de broyage' is associated with intense westward thrusting and dextral strike-slip associated with the emplacement of the Vermion nappe over the Pelagonian Zone during Early Cenozoic time (Mercier 1966). The Klisschochori and Liki–Margarita units further north experienced similar intense Early Cenozoic deformation. The emplaced serpentinite at the contact between the Pelagonian platform and the overlying Late Jurassic–Cretaceous cover acted as a regional detachment (décollement) associated with the intense deformation of the lower Klissochori Mélange unit.

In this study we relate the Late Jurassic–Early Cretaceous units of the Central Almopias Zone to an important phase of extension (or transtension). Within extensional fault zones, footwall highs underwent subaerial erosion and karst development, whereas hanging-wall depocentres were infilled with coarse clastic sediments, volcanic rocks and minor intrusions (e.g. granophyre) (Fig. 14). Multiple dykes are interpreted as infills of transtensional fissures (e.g. Liki and Klissochori units). The existence of active hydrothermal systems is suggested by epidote mineralization. Some fault zones were apparently exploited by protrusions of ductile serpentinite that flowed onto the sea floor where they were covered by volcaniclastic and hemipelagic sediments, as seen within the lower mélange unit. Upwards, poorly sorted polymict arenites and rudites were deposited by debris flows and by high- to low-density turbidity currents (e.g. upper

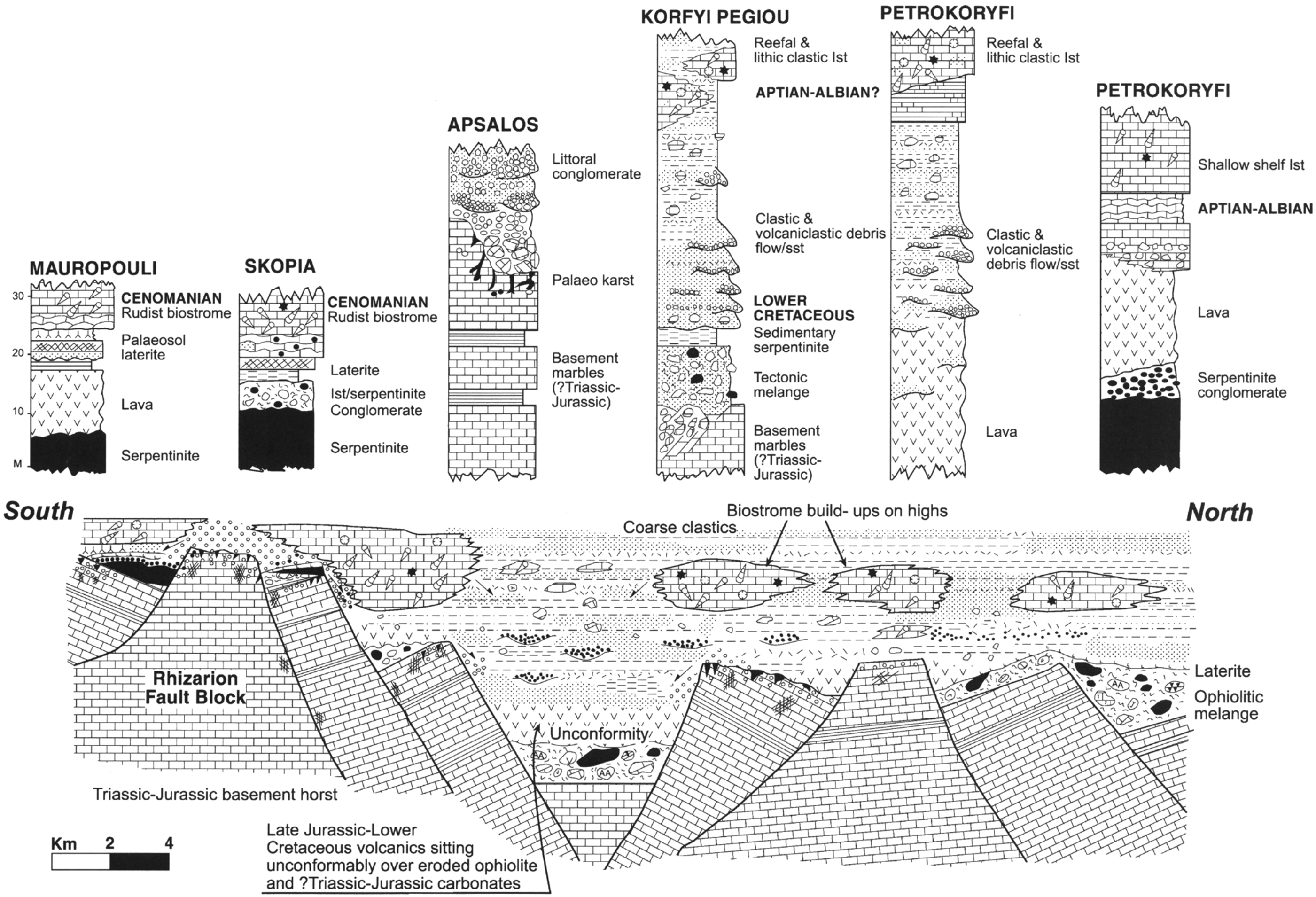


Fig. 14. Summary logs and schematic reconstruction of part of the Central Almopias Zone (Fig. 4) for Oxfordian–Kimmeridgian to Albian–Cenomanian time (see text for explanation.)

mélange unit at Petrokorfi; Fig. 14). Continuing volcanism is indicated by local intercalations of rhyolitic air-fall tuff. The clastic sediments include abundant metamorphic detritus, including augen-gneiss and garnet mica schist (Vergély 1984). The gneiss was presumably derived from the basement of the Pelagonian Zone or the Almopias Zone, which would require deep exhumation at least locally (e.g. along fault zones). The high-grade metamorphic detritus could, in principle, have been derived from an overriding crystalline thrust sheet, but there is no independent evidence that this existed. Facies trends and limited palaeocurrent data (Sharp 1994) suggest overall supply of the terrigenous sediments eastwards, from the Pelagonian Zone and the Central Almopias Zone, into the Eastern Almopias Zone.

The above relationship between the emplaced Pelagonian platform ophiolite nappe and sedimentary cover rocks could, in principle, be explained in two ways: (1) first related simply to a phase of extension (or transtension) along the western Vardar Zone margin; (2) related to more profound rifting to form a new Cretaceous oceanic basin within the Vardar Zone.

In option (1) the eastern margin of the Almopias Zone, including the Klissochori Mélange, acted as a zone of extension (or transtension) that was associated with intermediate–silicic volcanism, the emplacement of multiple debris flows, and with neritic carbonate and clastic sedimentation during Late Jurassic–Early Cretaceous time. The silicic composition of the volcanic rocks could reflect partial melting of thick underlying continental basement related to rifting.

In option (2) the intersheared serpentinites could be interpreted as emplaced fragments of the lower plate of an asymmetrically rifted continental margin that was associated with the exhumation of continental mantle onto the sea floor during Late Jurassic–Early Cretaceous time. The exhumed material in this model was represented by serpentinized peridotite that was hydrothermally altered to form ophicalcite in fissures, and then covered by silicic volcanic rocks, terrigenous sediments and, locally, by pillow lavas. For example, in the extreme SE of the Klissochori Mélange serpentinized dunites contain ophicalcite and lenses of pink micritic limestone in their upper part. The dunites are overlain by a thin horizon of serpentinitic or talc-rich mudstones and dunite-derived conglomerates (with reddened clast edges). These sediments are then covered, with a locally preserved primary contact, by little-deformed pillow lavas, lavas breccias and silicic extrusive rocks. This setting is, for example comparable with the exposure of continental mantle and the extrusion of overlying MOR-type extrusive rocks during the final stages of continental break-up to open the Late Jurassic Penninic Ocean in the Western Alps (Manatschal *et al.* 2003).

The main problem with model (2) is that in the Alps such exhumation took place in deep water, associated with radiolarite deposition, and there the volcanic rocks are of MOR type. By contrast, in the western Vardar Zone the associated sediments mainly accumulated in a shallow-water setting (e.g. reef limestones) and the volcanic rocks are mainly of intermediate–silicic composition. However, it is possible that various fragments of the Klissochori Mélange include a now-telescoped proximal to distal continent–ocean transition of Late Jurassic–Early Cretaceous age. Figure 16 shows a reconstruction of the edge of the Pelagonian Zone and the Central Almopias Zone between the relatively proximal Klissochori Unit and the more distal Nea Zoi Unit.

Further east, the Late Jurassic cover sequence of the Paikon Zone was represented by lagoonal facies, without volcanic rocks (Khromni limestones; Mercier 1966; Brown & Robertson 2003). Transgressive deposits of post-Late Kimmeridgian age in the Peonais Zone, further east again, were deposited in a marginal-marine to locally continental environment (Stais 1994). Early Cretaceous extension and exhumation were also inferred in the Paikon Zone (Brown & Robertson 1994) and elsewhere in the Pelagonian Zone (Doutsos *et al.* 1993).

The extensional (or transtensional) faulting and volcanism within the Western and Central Almopias zones effectively ended prior to Aptian–Albian time. Some areas, especially fault blocks, remained emergent, undergoing red-bed deposition and erosion of the metamorphic basement. Breaks in deposition occurred locally. For example, Late Jurassic reef build-ups are overlain by Early Cretaceous red beds and then by Mid–Late Cretaceous neritic carbonates in the Central Almopias Zone (e.g. Liki–Magarita Unit; Figs 2 and 16). The pre-Aptian–Albian time interval was thus marked by continuing tectonic instability along the Pelagonian–Almopias margin.

Late Jurassic–Early Cretaceous oceanic crust genesis

The Eastern Almopias Zone is dominated by two large exposures of ophiolite-related extrusive rocks, the Mavrolakkos Unit in the west and the Krania Unit in the east (Mercier 1966; Fig. 2b). Detailed field mapping has allowed the correlation of these two units as a single ophiolite

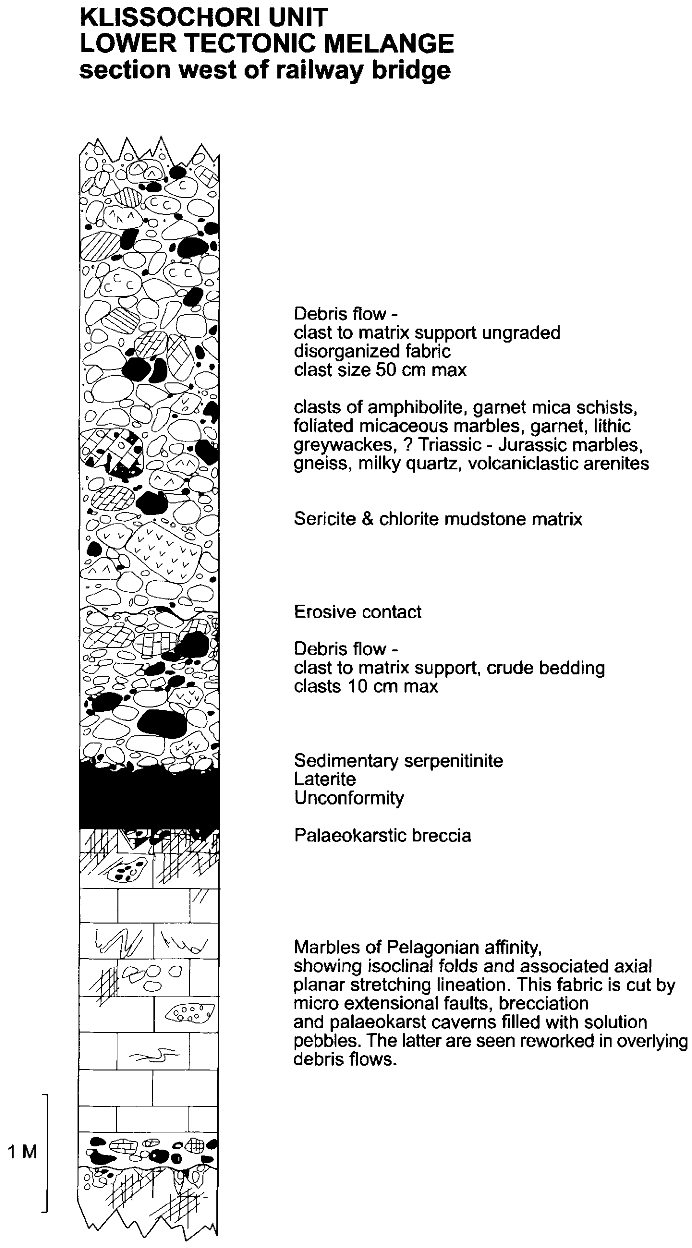


Fig. 15. Sedimentary log of the Klissochori Unit (lower mélange unit) in the type area (see Fig. 2b). The presence of coarse debris flows should be noted, overlying an erosional remnant of serpentinite with the Pelagonian platform carbonates beneath. These debris flows were emplaced adjacent to the Nission Fault, a major transverse structure.

thrust sheet, known as the Meglenitsa Ophiolite (Fig. 2b; Sharp 1994; Sharp & Robertson 1998).

The extrusive rocks of both units (c. >200 m thick) mainly comprise pillow lava and minor hyaloclastite. The lavas are overlain by a sequence (<50 m thick) of laminated black to green, ferruginous and micaceous mudstones, thin turbiditic sandstones and cherts with minor pillow lavas (Black Schist Member). Minor massive sulphides are present along the lava–sediment interface in the Krania Unit. A whole-rock K–Ar age of 110–134 Ma was obtained

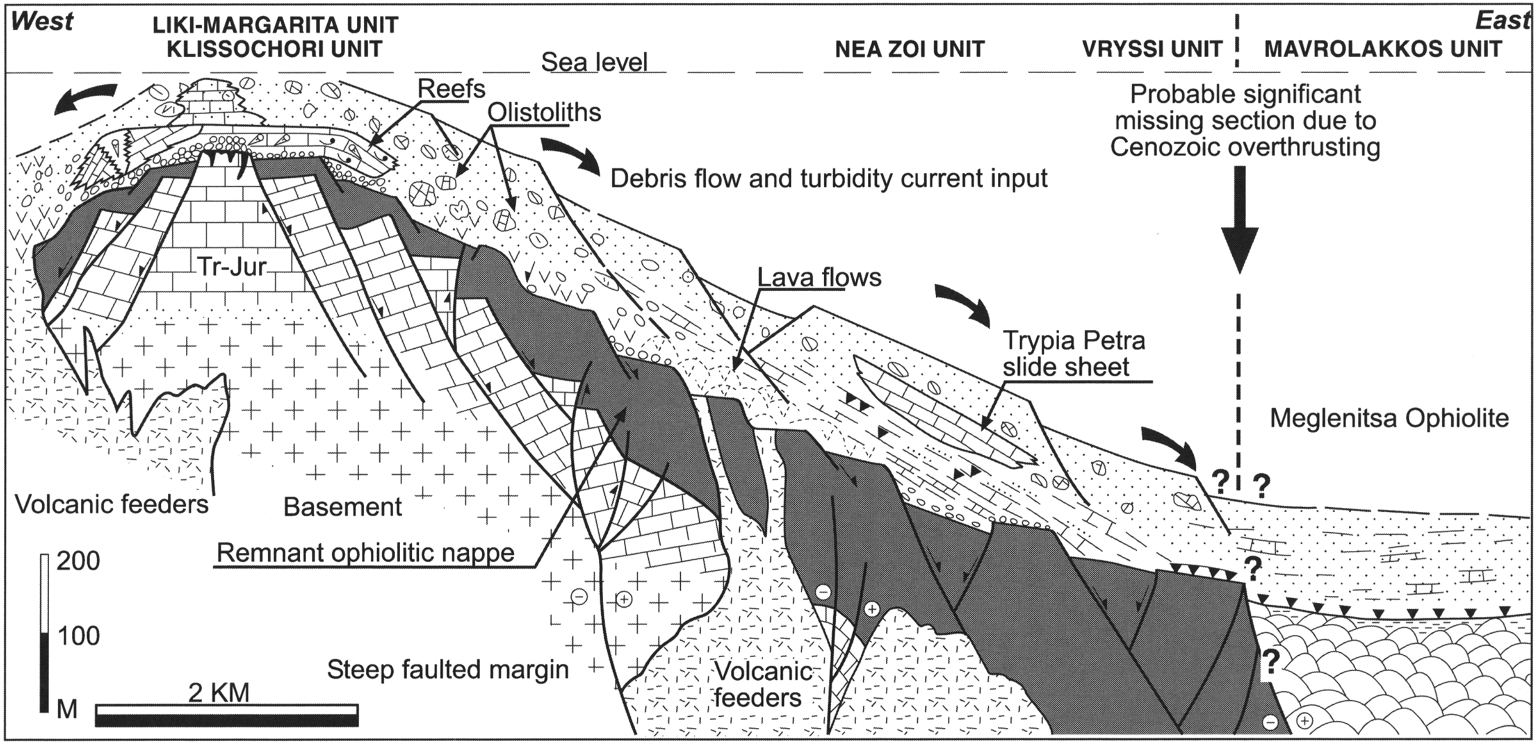


Fig. 16. Sketch reconstruction of the Mid–Late Cretaceous palaeogeography of the Central and Eastern Almopias zones as an east-facing rifted (transtensional?) margin of the Vardar Ocean. The nature of the contact between the Vryssi Unit (Central Almopias Zone) and the Mavrolakkos Unit (Eastern Almopias Zone) is uncertain as it is now an Early Cenozoic thrust (see Fig. 21 for alternative models).

from basic igneous rocks (basalt or diabase) (Bertrand *et al.* 1994), in keeping with radiolarian age data.

Above are radiolarian sediments, mainly ribbon radiolarites, radiolarian mudstones and siliceous mudstones, up to 25 m thick (Radiolarite Member). Pillow lava and thinly bedded sandstone are occasionally present at this level in the Mavrolakkos Unit. Radiolarian determinations (P. De Wever & H. YiLing; in Sharp & Robertson 1998) from the Mavrolakkos Unit indicate ages ranging from Late Jurassic (Callovian), to Early Cretaceous (Neocomian), possibly extending to Barremian–Aptian. Radiolarians from the Krania Unit yielded Late Jurassic (Mid-Oxfordian) to Early Cretaceous ages (Valanginian possibly extending to Berriasian–Turonian). Similar ages were reported by Stais (1994).

The radiolarian sediments pass gradationally upwards into a sequence (>200 m thick) of mudstones, siltstones and mixed sandstone–carbonate turbidites (Flysch Member). The Krania Unit in the east is coarser grained, mainly comprising terrigenous and calcareous turbiditic sandstones and debris-flow deposits, with clasts of basalt, neritic limestone, minor volcanic quartz, rhyolite, dolerite, granite, diorite and granophyre. There are also subordinate interbeds of arkosic sandstone and dacitic tuff. The turbidites and debris flows were thus mainly derived from an unmetamorphosed, mostly extrusive igneous terrane, together with smaller amounts of shallow-water carbonate and metamorphic material. The clastic sediments in the Krania Unit in the east are thicker and coarser grained than in the Mavrolakos Unit, suggesting derivation from the Paikon Zone (to the east) where similar lithologies are exposed (Grahmmos Formation; Brown & Robertson 2003). However, within the Mavrolakos Unit, limited palaeocurrent data indicate eastward to southward flow and slump folds are locally NE-vergent (Sharp 1994). Intraformational clasts of basalt and radiolarite were probably derived from subjacent oceanic crust. The upper age limit of the turbidites is constrained by the presence of unconformably overlying sediments of Late Cretaceous age along the western margin of the Mavrolakkos Unit.

The Krania Unit, including the Flysch Member, is cut by localized granite, granophyre and basaltic sills and also by north–south-trending dykes (Bébien *et al.* 1980; Sharp & Robertson 1998). Localized amphibole-bearing quartz diorite dykes cutting the clastic sediments were dated at 124 Ma (Bechon 1981), suggesting that the intrusive rocks are approximately contemporaneous with the underlying ophiolitic lavas.

Sharp & Robertson (1998) noted that the basalts of the Mavrolakkos Unit are of near-MORB composition but a few samples are relatively depleted (Fig. 17a–d). Basalts and occasional late-stage dykes from the northern part of the Krania Unit (Mavrolakkos Vodeon) show a spread from near-MORB to relatively enriched with a few relatively depleted samples, plus several samples that show a small but distinct negative niobium anomaly (Fig. 17a). Many samples from the Krania Unit as a whole (e.g. west of Krania village) again show a near-MORB to slightly enriched composition. However, several samples show a pronounced negative Nb anomaly (e.g. from the old Krania–Mandalon road-cut; Fig 17b) and are similar to some oceanic arc or back-arc basalts. These results, together with the presence of the local intrusions of granite and granophyre cutting the sedimentary cover, are suggestive of genesis in a subduction-influenced setting. On the other hand, the presence of the terrigenous sedimentary cover, albeit of deep-water origin, shows that this oceanic crust formed in a near-continental margin setting rather than an open-ocean setting.

The genesis of the Late Jurassic–Early Cretaceous Meglenitsa Ophiolite post-dates the emplacement and metamorphism of the ophiolitic rocks of the Pelagonian and Western–Central Almopias zones (pre-Late Oxfordian–Early Kimmeridgian). An Aptian–Albian unconformity is developed on the western margin of the Meglenitsa Ophiolite (Sharp & Robertson 1998). A counterpart of the Meglenitsa Ophiolite, the Ano Garefi Ophiolite in the Voras Massif further north (Brown & Robertson 2004), is also unconformably overlain by deep-water sediments of Aptian–Albian age (Mercier 1966; Brown & Robertson 2004). Alternative tectonic settings for the genesis of the Meglenitsa Ophiolite during Late Jurassic–Early Cretaceous time are considered in the Discussion and conclusions section.

Aptian–Cenomanian passive margin subsidence

During Aptian–Albian–Cenomanian time the western margin of the Vardar Zone experienced post-rift subsidence. Marine transgression of the combined Pelagonian and the Western and Central Almopias zones culminated in the development of an eastward deepening, mixed carbonate–clastic succession, characterized by fluvial–coastal plain to shelf environments.

Summary sedimentary logs of the Pelagonian Zone and the Western Almopias Zone applicable to this time are shown in Figures 4 and 18. Facies trends and palaeocurrent data are generally indicative of eastward sediment supply.

Within the Pelagonian Zone, a diachronous marine transgression progressively covered remaining exposed areas, with an upward transition from coastal plain–fluvial to marginal-marine settings. Overlying sequences are fully marine and accumulated mainly within middle to inner shelf settings. Inner shelf areas were characterized by littoral conglomerates, storm-influenced beds and rudist biostromes. Platy bedded, bioturbated carbonates and calcarenites rich in a mixed benthic–planktonic fauna were deposited in deeper-water, more offshore areas.

The Western Almopias Zone (i.e. Kerassia and Kedronas units; Fig. 18) represents an eastward continuation of the same east-facing margin sequence. Marked facies variations are evident, especially close to major transverse faults (e.g. Nission Fault) that are interpreted to have been still active. The facies of the Kerassia Unit exhibit an overall east- to SE-facing ramp geometry, from coastal plain to mid-outer ramp environments. Isolated subaerial highs persisted, for example in the Kedronas Unit, with karstic erosion and non-marine red-bed deposition.

Within the Central Almopias Zone, the Margarita Unit exhibits mainly neritic accumulation (Fig. 4). An isolated high, characterized by coral–rudist biostromes, developed in the Rhizarion region (e.g. Mavropouli and Skopia sections; Fig. 14), south of the Nission Fault, with contrasting coarse clastic sedimentation in the hanging wall of this fault-controlled basin directly to the north (e.g. Korfyi Pegiou section; Fig. 13). South of the Rhizarion fault block pelagic–hemipelagic deposition dates from Early Cretaceous time. North of the Kato Loutraki Fault (Fig. 3), Triassic–Jurassic marble and serpentinite are unconformably overlain by a mainly carbonate cover of Aptian–Albian age (Livadia Unit). The contact is locally marked by karstic weathering of Triassic–Jurassic marble, bauxite and red-beds (Brown & Robertson 2004). The easternmost part of the Central Almopias Zone, on the other hand, was characterized by deeper-water hemipelagic sediments of Aptian–Albian (and younger) age that unconformably overlie a local serpentinite basement (Nea Zoi Unit; Figs 3 and 4). By contrast, in the Eastern Almopias Zone, Aptian–Albian (or younger) sediments have only been recorded as exposures along the western edge of the Meglenitsa Ophiolite.

Regional comparisons show that a similar east-facing passive margin developed along the eastern Pelagonian Zone, both to the south (e.g. Sporades, Evia, Argolis; e.g. Clift & Robertson 1990*b*; Robertson 1990; Clift 1992) and further north in Macedonia and Serbia (see Karamata 2006).

Cenomanian–Turonian: depositional hiatus and relative sea-level rise

During Cenomanian–Turonian time, facies evidence points to a relative sea-level rise and deepening of the Pelagonian–Almopias carbonate margin. The west-to-east overall deepening trend persisted, with a few areas remaining subaerially exposed (e.g Arnissa region and Kaimatchalan Massif). Sedimentation during this time appears to have been influenced by a pulse of extension-related subsidence, coupled with eustatic sea-level rise (e.g. Sharland *et al.* 2001).

In the Pelagonian and Western Almopias zones, ramp-interior lagoonal and ramp-margin rudistic carbonates were terminated by intraformational conglomerates that were in turn abruptly overlain by deeper water outer-ramp and hemipelagic facies (Figs 18 and 19). Dramatic drowning of remaining subaerial highs is evident in parts of the Western Almopias Zone south of the Nission Fault, where coarse immature facies were deposited in a littoral setting (i.e. western and eastern sections of the Kedronas Unit; Fig. 18). In the Central Almopias Zone the Klissochori Unit experienced clastic sedimentation (Figs 4 and 14). The more distal Nea Zoi Unit further east (Figs 4 and 16) exhibits the emplacement of a large slide-sheet of mainly neritic carbonates ('Trypia Petra slide' derived from the Klissochori Unit?) and an upward short transition to coarse clastic sediments; some undated radiolarites are also present.

A similar relative sea-level rise is well documented further east, in the Paikon Zone, where a short period of subaerial exposure was followed by abrupt drowning of the carbonate platform and a transition to hemipelagic, then turbiditic, deposition (Sharp & Robertson 1993; Brown & Robertson 2003). The Paikon Zone remained an area of deep-water accumulation until Paleocene deformation.

Late Cenomanian–Turonian eustatic sea-level rise (Haq *et al.* 1987; Sharland *et al.* 2001) does not, by itself, explain features such as synsedimentary faulting, slumping and localized coarse clastic deposition in the area studied, and a tectonic trigger is likely. The Cenomanian–Turonian boundary was marked by uplift or

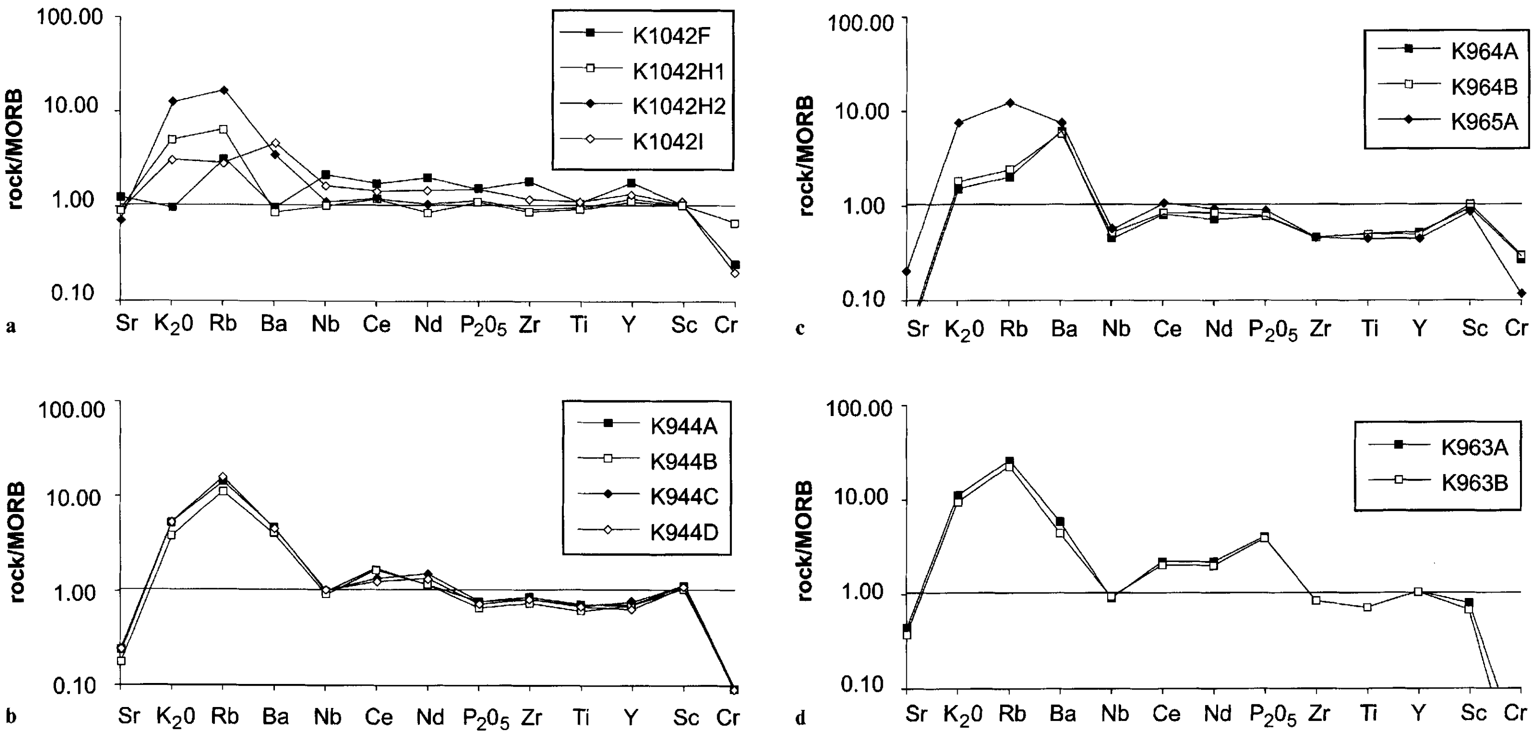


Fig. 17. MORB-normalized 'spidergrams' of tholeiitic basalts from the Krania (eastern) Unit of the Meglenitsa Ophiolite, Eastern Almopias Zone. (**a**) From the Lava Member, northern area, Mavrolakkos Vodenon; (**b**) from the overlying Black Schist Member, old Krania–Mandalon road-cut; (**c**) from the Flysch Unit, higher in the sequence, Litharia–Krania road-cut; (**d**) late-stage dykes cutting the sedimentary cover; Litharia–Krania road. The near-MORB composition should be noted.However, in particular the presence of a negative Nb anomaly is suggestive of a subduction influence. (See Table 1 for representative analyses.)

subsidence regionally; for example, the eastern margin of the Pelagonian Zone in Evia (Robertson 1990) and Argolis (Clift 1992) underwent abrupt subsidence at this time. In addition, the south Aegean region was affected by regional subsidence (Harbury & Hall 1988). Late Cretaceous ophiolites of supra-subduction-zone type, based on geochemical evidence, have been identified in the Argolis area (NE Peloponnese) (Clift & Robertson 1989), and in Crete and Karpathos (Koepke *et al.* 2002). It is possible that the Cenomanian–Turonian hiatus in the area studied was triggered by intra-oceanic subduction and supra-subduction spreading within the Vardar Ocean.

Campanian–Maastrichtian: resumed subsidence

Eastward deepening related to subsidence was re-established with the development of thick ramp–shelf margin to outer ramp–shelf carbonates during Late Santonian–Early Campanian time. The dominant control was eustatic sea-level rise (e.g. Hancock & Kauffman 1979). A north–south-striking, east-facing rudist-dominated ramp–shelf margin complex persisted in the Pelagonian, Western Almopias and parts of the Central Almopias zones during Campanian–Maastrichtian time. The northern Pelagonian Zone was finally transgressed, with the fault-bounded Kaimaktchalan Massif being the last area to be flooded. Eastwards, the Central and Eastern Almopias zones experienced coarse clastic sedimentation (Klissochori, Nea Zoi and Vryssi units), dominated by siliciclastic turbidites that are interpreted as having accumulated in a deep-water submarine fan or apron setting. In addition, fine-grained hemipelagic sediments accumulated to the south of the Nission Fault (e.g. in the Ano Grammatiko area; Fig. 8).

Late Maastrichtian: transition to a foreland basin

Late Maastrichtian time saw dramatic subsidence of remaining areas of the regional east-facing carbonate ramp–shelf (Figs 19 and 20). In the Pelagonian Zone, Western Almopias Zone and parts of the Central Almopias Zone (i.e. the Margarita & Klissochori Units; e.g. Petrokoryfi region, Fig. 14) there is an abrupt transition from neritic, rudist-bearing carbonates, accumulating within and along the margin of the carbonate platform, to hemipelagic micrites and deep-water terrigenous sediments of an inferred foredeep succession. In the Pelagonian Zone (Fig. 19), iron-phosphate encrusted and bored hardgrounds developed at the top of the platform carbonate succession, overlain by muddy ferruginous carbonates. An iron-stained subaerial emergence horizon developed in the Western Almopias Zone (Kerassia and Kedronas units), associated with fissuring of underlying limestones and the deposition of localized intraformational conglomerates. The base of overlying deeper-water calcareous mudstones includes well-rounded blocks of neritic limestone, derived from subjacent Maastrichtian rudistic carbonates. Late Maastrichtian–Paleocene siliciclastic sequences in the Pelagonian Zone are characterized by hemipelagic sediments and thin-bedded, fine-grained turbiditic sandstones that coarsen upwards, passing into high-density turbidity current and debris-flow deposits, with clasts including fresh basalt and metamorphic detritus. These successions appear to shallow upwards and may be fluvial near the top in places, before being overridden by SW-verging thrust sheets. A similar facies transition to siliciclastic turbidites is known elsewhere along the eastern margin of the Pelagonian Zone, including Evia (Robertson 1990) and Argolis (Clift 1992).

The profound facies change is interpreted to mark the transition to a foreland basin ahead of a westward-propagating thrust load (Sharp & Robertson 1993; Fig. 20). The more outboard sequences (i.e. Western Almopias Zone) underwent uplift and erosion, whereas the more inboard (westerly) sequences (i.e. Pelagonian Zone) became the sites of muddy, lagoonal deposition. Later, the entire margin collapsed, becoming a deep-water basin in which siliciclastic turbidites accumulated. This basin filled with coarser-grained, shallower-water sediments derived from the advancing thrust load.

Late Cretaceous eastward subduction

Late Cretaceous sediments near the contact between the Central and Eastern Almopias zones (Nea Zoi and Vryssi units; Fig. 4) are dominated by siliciclastic turbidites showing strong layer-parallel extension to form classic phacoidal fabrics. Such features could be related to the layer-parallel extension that characterizes some shear zones within thrust sheets but are also reminiscent of the phacoidal fabrics seen in many subduction complexes (e.g. Franciscan Complex, western USA: Cloos 1984). *In-situ* glaucophane was reported from Late Cretaceous turbidites in the Liki–Margarita Unit (Central Almopias Zone), particularly along the contact between the Liki–Magarita Unit and the Klissochori Unit

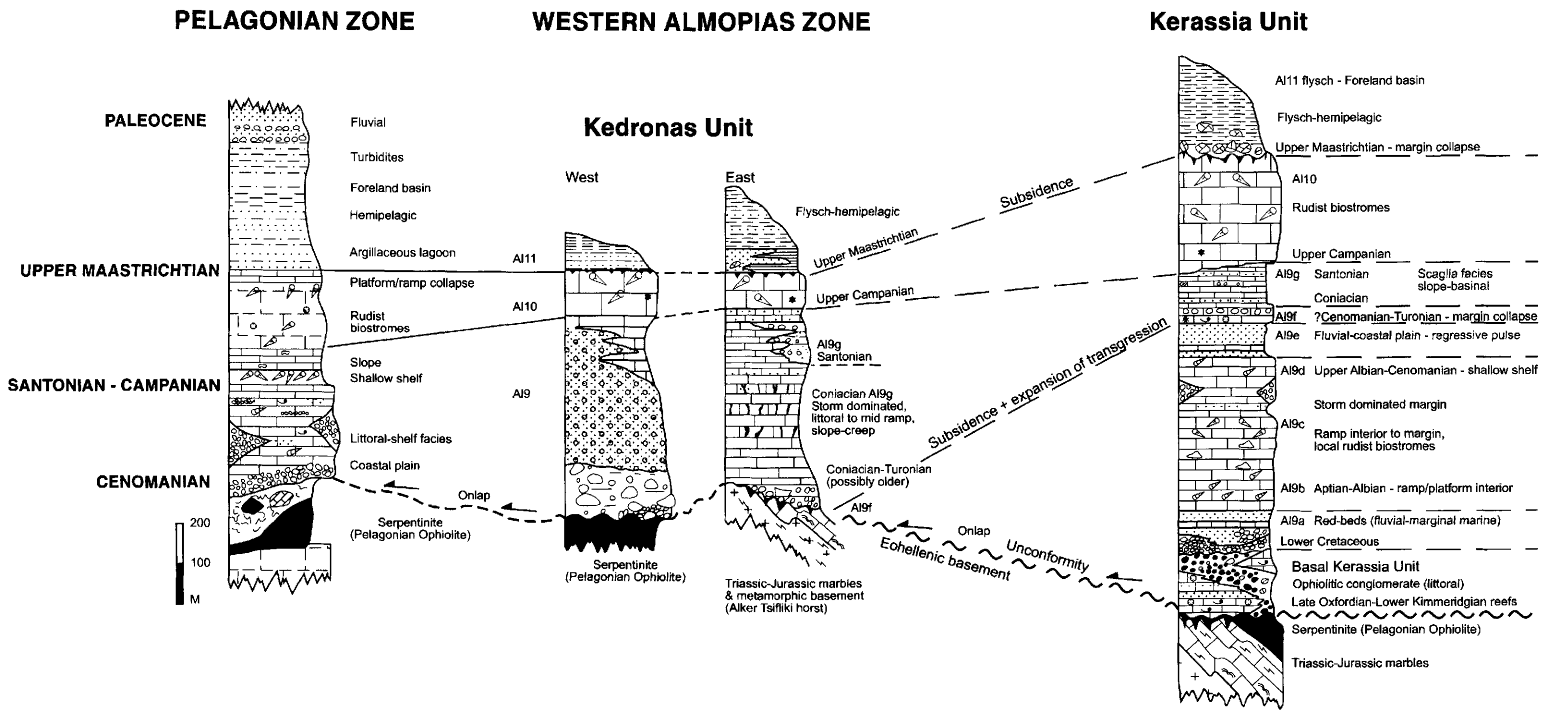


Fig. 18. Summary of the Cretaceous sequences of the Pelagonian and Western Almopias zones. Based on Campion (1966), Mercier (1966), Braud *et al.* (1984) and Sharp (1994). (See text for discussion of facies.)

(Vergély 1984; Sharp 1994). The HP–LT metamorphism is interpreted to relate to eastward subduction of the Vardar Ocean and the eastern edge of the Pelagonian–Almopias continental unit. The siliciclastic turbidites of the Nea Zoi and Vryssi units could thus represent part of an accretionary prism formed along the western margin of a Vardar Ocean rather than merely the distal part of a foreland basin.

Similar deformed siliciclastic turbidite facies occur in a comparable tectonic setting further south in Greece (e.g. Argolis Peninsula), where they are interpreted as a latest Cretaceous–Early Cenozoic accretionary prism related to eastward subduction and the related development of a foreland basin (Clift & Robertson 1989; Clift 1992). South of the study area in the Sporades Islands, an ophiolitic thrust sheet is present at the highest structural levels, and is interpreted to be the result of eastward subduction that culminated in westward ophiolite emplacement during latest Cretaceous–Early Tertiary time (Jacobshagen & Wallbrecher 1984). Comparable eastward subduction is inferred to have taken place along the eastern margin of the Pelagonian Zone in Greece and former Yugoslavia (e.g. see Karamata 2006). A number of the meta-ophiolites of the eastern part of the Pelagonian zone (e.g. High Pieria) that lack D_1 deformation are inferred to have been emplaced from the Vardar Zone during latest Cretaceous–Early Palaeogene time.

Palaeogene suturing

Regional-scale folding and thrusting took place with a southwesterly to westerly vergence, with the Almopias Zone overthrusting the Pelagonian Zone (Mercier 1966; Vergély 1984; Mountrakis *et al.* 1987; Fig. 2b). In the east, the Peonais Zone overthrust the eastern margin of the Paikon Zone, and the Serbo-Macedonian Zone generally overrode the Vardar Zone, as seen in the Voras Massif in the north (Brown & Robertson 2004; Fig. 3). However, the Eastern Almopias Zone (Meglenitsa Ophiolite) was clearly thrust both westwards and eastwards over the Central Almopias Zone and Paikon Zone, respectively (Sharp & Robertson 1993; Brown & Robertson 2003; Fig. 3). Indeed, the Meglenitsa Ophiolite was thrust northeastwards over the Paikon Massif to the highest structural level in the region, consistent with its low metamorphic grade (lower greenschist facies or less), and limited structural deformation (i.e. one phase of Tertiary folding; Sharp 1994; Vergély 1984; Sharp & Robertson 1998). However, north of the Kato Loutraki Fault the convergence direction was reversed and the equivalent Ano Garefi Ophiolite was thrust southwestwards, together with other units of the Voras Massif (Brown & Robertson 2004). This indicates that the thrust belt was segmented, with the Paikon segment experiencing local back-thrusting (Fig. 3).

The Pelagonian Zone was affected by large-scale imbrication and the development of regional-scale nappes (e.g. Haut Vermion Nappe), regional greenschist-facies metamorphism (Kockel 1986) and localized dynamic metamorphism, related to regional westward overthrusting (Braud *et al.* 1984; Fig. 8). The previously emplaced Pelagonian Ophiolite was detached from its Cretaceous cover in many areas, associated with re-thrusting and the development of large-scale SW-verging folds (e.g. Messovounon and Ayios Dimitrios areas; Sharp 1994; Figs 3 and 8). Similarly, in the Western Almopias Zone, the mainly calcareous cover was detached from underlying ophiolitic units, typically exploiting highly incompetent serpentinite. The underlying units were strongly imbricated, increasing the structural complexity of the ophiolitic mélange (e.g. in the Haut Vermion Nappe; Fig. 8).

Within the Pelagonian and Western and Central Almopias zones there is a general increase in the intensity of deformation from the west to the east. This is particularly evident in units of Mid–Late Cretaceous age. The probable explanation is that the downflexed passive margin in the east was located near the axis of the Vardar suture zone, whereas the Pelagonian Zone was located in a more westerly (external) position.

The Cretaceous cover of the Pelagonian Zone has generally undergone minimal penetrative deformation, whereas age-equivalent units in the Western and Central Almopias Zones (Kerassia, Margarita, Klissochori and Nea Zoi units) were more pervasively deformed. A similar eastward increase in deformation is seen in the Voras Massif, as far east as, and including, the Livadia Unit (Brown & Robertson 2004; Figs 2b and 3). In addition, as already mentioned, sediments of the Liki–Margarita Unit that post-date the Late Jurassic ophiolite emplacement contain evidence of HP–LT minerals of Late Cretaceous–Early Cenozoic age. Specifically, blue crossite crystals are developed parallel to the main S_1 and S_2 schistosity within deformed pelitic calcarenites that contain *Globotruncana* sp., proving a Turonian or younger age (e.g. exposures of Late Cretaceous Flysch exposed in railway sections

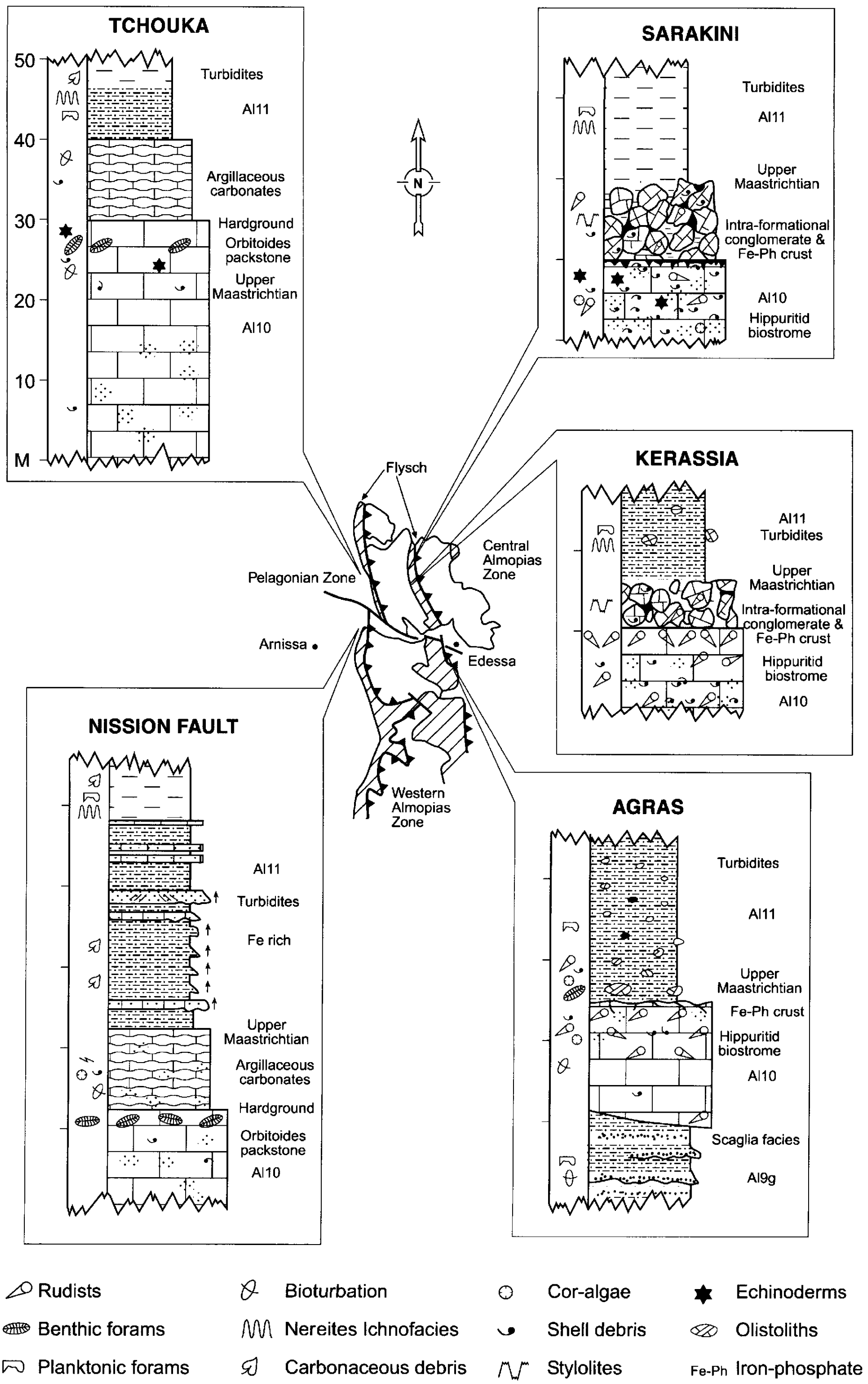


Fig. 19. Sedimentary logs of the top of the Cretaceous carbonate margin and the transition to flysch in the Pelagonian and Western Almopias zones. (See Fig. 3 for regional map.)

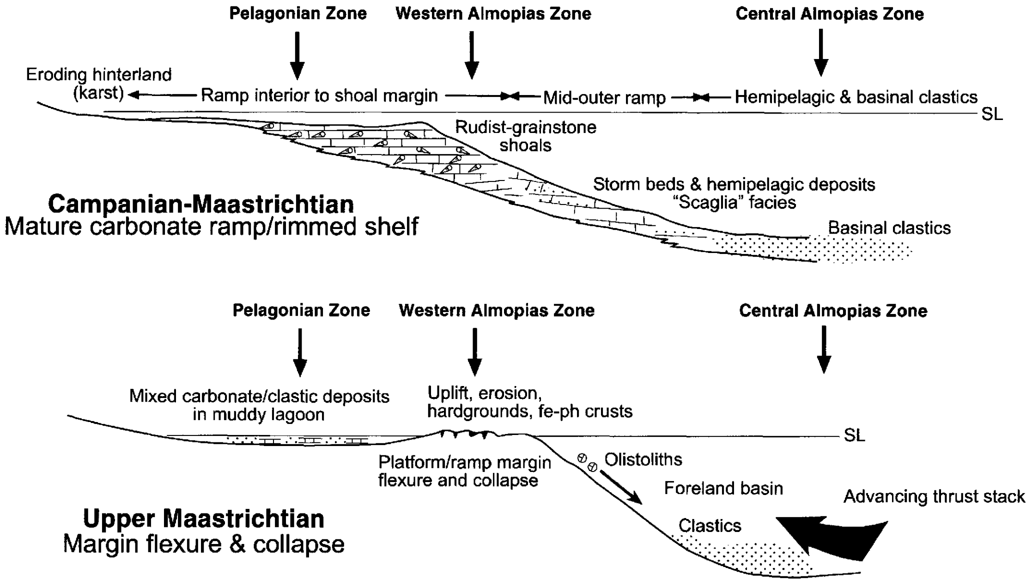


Fig. 20. Sketch interpretation of the Cretaceous carbonate margin and transition to flysch in the Pelagonian and Western Almopias zones.

at the southeastern end of Krivi Dereki Rema; Fig. 3).

In summary, HP–LT metamorphism, intense shearing, tight imbrication and westward nappe emplacement are together seen as the response to the attempted subduction of the Pelagonian microcontinent in a trench, dipping northeastwards beneath the Eurasian active margin and its amalgamated terranes (i.e. the Peonias and Paikon units).

Mid-Cenozoic–Recent: extensional collapse

An upper age limit for the regional Early Cenozoic deformation is provided by the presence of subhorizontal Eocene limestones and conglomerates in the Peonais Zone (Mercier 1966). During mid-Cenozoic time the thrust stack in the Vardar Zone underwent extensional collapse, probably related to 'roll-back' of a northward-dipping subduction zone in the south Aegean region (e.g. Kilias *et al.* 1999), probably beginning in the Oligocene (*c.* 25 Ma; Burchfiel *et al.* 2000). The post-Miocene neotectonic phase was also marked by widespread subaerial volcanism and continuing fault-controlled deformation (Mountrakis *et al.* 2006). This was related to subduction in the south Aegean region and the development of the Aegean arc behind a northward-dipping subduction zone (e.g. Le-Pichon & Angelier 1979).

Discussion and conclusions

The units of the eastern Pelagonian Zone and the Western and Central Almopias zones are interpreted as the western margin of the Vardar Ocean from Triassic to latest Cretaceous time. This area experienced rifting during the Triassic associated with alkaline magmatism and terrigenous sedimentation, followed by passive margin subsidence during Late Triassic–Early Jurassic time. During the Late Triassic–Early Jurassic the Pindos oceanic basin formed to the west of the Pelagonian microcontinent.

During Mid–Late Jurassic time, a vast ophiolite thrust sheet, including the combined Vourinos–Pindos ophiolite, was emplaced over the Pelagonian microcontinent, probably in response to the collision of the Pelagonian microcontinent with a SW-dipping intraoceanic subduction zone. Northeastward emplacement of this ophiolite from the Pindos Ocean to the west is favoured based on limited structural evidence from the area studied and additional evidence from the Jurassic ophiolites and underlying units throughout Greece, Albania and former Yugoslavia. The Pelagonian ophiolitic mélange in the area studied is interpreted as a subduction complex that was emplaced onto a regional foredeep following the flexural collapse of the Pelagonian passive margin. The ophiolite and the underlying Pelagonian platform experienced greenshist- to amphibolite-facies metamorphism (M_1), coupled with ductile deformation. The development of a pervasive NNW–SSE stretching fabric is inferred to relate to lateral (strike-slip) displacement of the emplaced ophiolite.

The ophiolite emplacement is dated as pre-Late Oxfordian–Kimmeridgian, from the age of depositionally overlying coralline limestones (assuming that the ophiolitic rocks of the Western Almopias Zone formed part of this regionally emplaced ophiolite). This has implications for the timing of thrusting (D_1) and metamorphism (M_1) further south in the Pelagonian Zone, which is traditionally believed to be no older than Early Cretaceous.

The Pelagonian ophiolitic mélange and the adjacent Vourinos and Albanian (Mirdita) ophiolites were overlain by deep-water carbonates along the western margin of the Pelagonian Zone (e.g. Mavri Rakhi Formation). The emplaced ophiolitic complex draped back into the relict Pindos Ocean to the west, where it was covered by deep-sea carbonate sediments.

The Almopias Zone was exhumed by Late Jurassic time; the Pelagonian Zone as a whole was exposed by mid-Cretaceous time and progressively covered by shallow-water carbonate sediments. The probable cause of the exhumation was extensional tectonics, coupled with erosion.

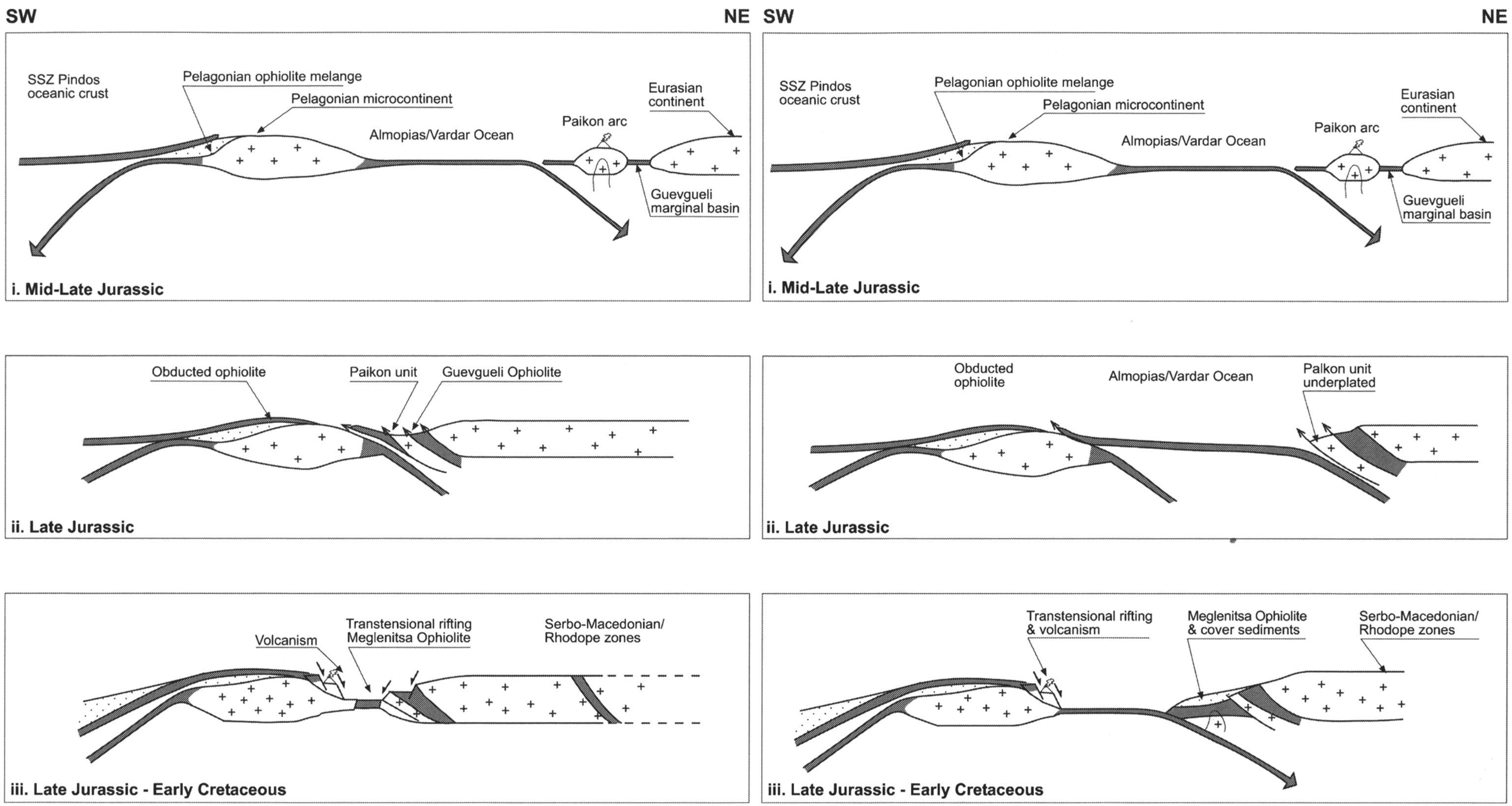
During Late Jurassic–Early Cretaceous time, following ophiolite emplacement, the Western and Central Almopias zones experienced extension, or transtension, associated with the emplacement of multiple debris flows ('mélange'), coarse clastic sedimentation and intermediate–silicic volcanism. This was followed by a resumption of passive margin subsidence along the eastern margin of the Pelagonian microcontinent, interrupted by a Cenomanian–Turonian hiatus that was at least in part tectonically triggered.

Further east, the Vardar Zone was the site of a Mesozoic ocean (Almopias Ocean) that opened in response to Triassic rifting. This ocean was subducted northeastwards beneath the Serbo-Macedonian Zone during Early to Mid-Jurassic time associated with opening of the Guevgueli marginal basin in the Peonias Zone. The Vardar Zone (Eastern Almopias Zone) includes large slices of basic extrusive rocks, associated deep-sea terrigenous and pelagic sediments (radiolarites), and minor intrusive rocks that are interpreted as the upper part of a dismembered Late Jurassic–Early Cretaceous ophiolite (Meglenitsa and Ano Garefi ophiolites). The Meglenitsa Ophiolite formed after the regional Mid–Late Jurassic ophiolite emplacement and was not involved in regional deformation (D_1) and metamorphism (M_1). This relatively young oceanic crust was later subducted eastwards beneath the Serbo-Macedonian Zone, by then amalgamated to Eurasia, during Late Cretaceous time and finally emplaced onto the Pelagonian–Almopias Zone, which by then had become a foreland basin; any other Vardar oceanic crust was subducted, leaving little trace. HP–LT metamorphism of probable Early Cenozoic age affected the eastern margin of the Pelagonian continent (Central Almopias Zone) related to collison with the subduction zone dipping eastwards beneath Eurasia and its previously accreted units.

An outstanding issue is the regional tectonic setting of the Late Jurassic–Early Cretaceous rifting along the western Vardar margin and the related intermediate-composition silicic volcanism and neritic clastic sedimentation, as seen within the Eastern and Central Almopias zones.

One option, shown in Figure 21a, is that the Almopias (Vardar) Ocean in northern Greece was completely closed during Mid–Late Jurassic time (pre-Kimmeridgian) (Papanikolaou 1996–1997), followed by the reopening of a new, Cretaceous oceanic basin (e.g. Jacobshagen & Wallbrecher 1984), possibly as a pull-apart basin (Sharp 1994). This model is consistent with the presence of terrigenous turbidites above the Meglenitsa Ophiolite. However, it is generally accepted that a wide Vardar Ocean still existed in the Cretaceous, sufficient to fuel extensive calc-alkaline magmatism along the Eurasian margin, as seen in the Serbo-Macedonian and Peonais zones (Dercourt *et al.* 1986, 1993, 2000; Ricou *et al.* 1988). If the Vardar Zone had completely closed in the Late Jurassic this would require a major spreading event during the Cretaceous.

A second option, shown in Figure 21b, is that the Meglenitsa Ophiolite formed by supra-subduction-zone spreading adjacent to the, by then, amalgamated Eurasian convergent margin to the NE. The ophiolite was finally emplaced when the Pelagonian microcontinent collided with the subduction trench during latest Cretaceous–Early Cenozoic time. This is consistent with the geochemical evidence of a subduction influence on the Meglenitsa Ophiolite, and the probable derivation of the deep-water clastic sediment cover of the ophiolite from the Paikon Zone to the east. The main problem is to explain the Late Jurassic–Early Cretaceous extensional or transtensional rifting along the western margin of the Vardar Ocean (i.e. within the Western and Central Almopias zones). This could possibly reflect a switch in subduction polarity within the Pindos Ocean to the west, from northeastwards in the Mid–Late



(a) "Closed ocean" model

(b) "Open ocean" model

Fig. 21. Alternative tectonic models. Both infer that the Jurassic ophiolite and the ophiolitic mélange were emplaced from the Pindos Ocean to the SW. (**a**) The Vardar Ocean closed in the Mid–Late Jurassic, followed by reopening of an oceanic basin, represented by the Meglenitsa Ophiolite, during Late Jurassic–Early Cretaceous time. (**b**) The Vardar Ocean remained open from the Late Triassic to the Late Cretaceous. A phase of extension–transtension affected the western margin of the Vardar Ocean, coupled with the genesis of new oceanic crust (Meglenitsa Ophiolite) during Late Jurassic–Early Cretaceous time; (**b**) is preferred but needs to be tested with additional evidence from former Yugoslavia.

Jurassic (resulting in compression–transpression) to southwestwards in the Cretaceous–Early Tertiary (resulting in extension–transtension).

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