Sedimentary evidence from the south Mediterranean region (Sicily, Crete, Peloponnese, Evia) used to test alternative models for the regional tectonic setting of Tethys during Late Palaeozoic–Early Mesozoic time

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Abstract: The south Mediterranean region, including western Sicily, Crete and mainland Greece (southern Peloponnese and Evia), is critical to an interpretation of the Late Palaeozoic–Early Mesozoic tectonic evolution of Tethys. Several contrasting tectonic models compete to explain the regional evolution. In a divergence-related hypothesis (Model 1) the south Aegean region experienced pulsed rifting along the northern margin of Gondwana that culminated in break-up to form the Pindos ocean in the region of Greece. In an alternative convergence-related hypothesis (Model 2) the south Aegean experienced Late Palaeozoic Early Mesozoic northward subduction, accretion and arc magmatism, culminating in 'Cimmerian' suturing of a Palaeotethyan ocean in latest Triassic time. In a third model, southward subduction of a Palaeotethyan ocean took place beneath the North Gondwana margin during Late Palaeozoic–Triassic time, giving rise to back-arc magmatism in an extensional setting. In addition, a more complex setting involving two opposing subduction zones (Andean-type and intra-oceanic) has also been suggested (Model 4), mainly based on lava geochemistry. To test these tectonic alternatives, mainly sedimentary studies were carried out in western Sicily, western and eastern Crete, the Peloponnese and Evia (eastern central Greece). Western Sicily was studied as a proxy for the unexposed deep Mediterranean south of Crete. Most of the available evidence supports the divergence-related (pulsed rift) hypothesis (Model 1). There is no clear evidence of sea-floor spreading (e.g. ophiolites) to the south of what became the Pindos ocean, or of plate convergence (e.g. magmatic arcs, subduction complexes), or collisional deformation in the south Aegean region that could be related to subduction or collision during the Mid-Carboniferous to Triassic, as in Model 2. Model 3 is not supported by evidence from the wider region (northern Greece, Turkey). Model 4 is not supported by evidence independent of igneous geochemistry. In the proposed interpretation, the northern margin of Gondwana initially rifted during Mid-Carboniferous to Early Permian time to form a wide deep-water basin. This was followed by further rifting, associated with volcanism during the Early Triassic; final continental break-up and spreading to form the Pindos ocean to the north during Late Triassic to Early Jurassic time then followed. Mid-Triassic uplift of part of the rift basin is explained as a flexural response to rifting as a precursor to opening of the Pindos ocean. Passive margin subsidence during the Early Mesozoic relates to opening of the Pindos ocean to the north. A subduction geochemical signature within some Triassic volcanic rocks, in this interpretation, is explained by melting of heterogeneous sub-crustal mantle, following an earlier, possibly Hercynian, subduction event.

The quest for 'Palaeotethys' of Late Palaeozoic to Early Mesozoic age in the Mediterranean region continues (Fig. 1). Most palaeomagnetic reconstructions suggest that a large westward-narrowing gulf of the super-ocean, Panthalassa ('Palaeotethys'), existed in the Eastern Mediterranean region by Late Permian time (e.g. Smith *et al.* 1981). What was the nature of this ocean? Where are its remnants? How does it relate to younger Mesozoic Neotethyan oceanic basins in the Eastern Mediterranean region? Deep-marine facies are known to have bordered the north margin of Gondwana, at least from Mid-Carboniferous time (Krahl *et al.* 1982; Kozur & Krahl 1984; Catalano *et al.* 1991; Kozur 1993, 1995), but their tectonic setting is controversial.

In a first, divergence-related Model 1 (Fig. 2), a Palaeotethyan ocean was subducted northwards beneath Eurasia, as indicated by evidence from the Pontides of northern Turkey and elsewhere along the southern margin of Eurasia. The Pelagonian Zone of Greece, eastern Crete and all of the units south of this continental fragment, known as the Pelagonian microcontinent, rifted from Gondwana during Early Mesozoic time

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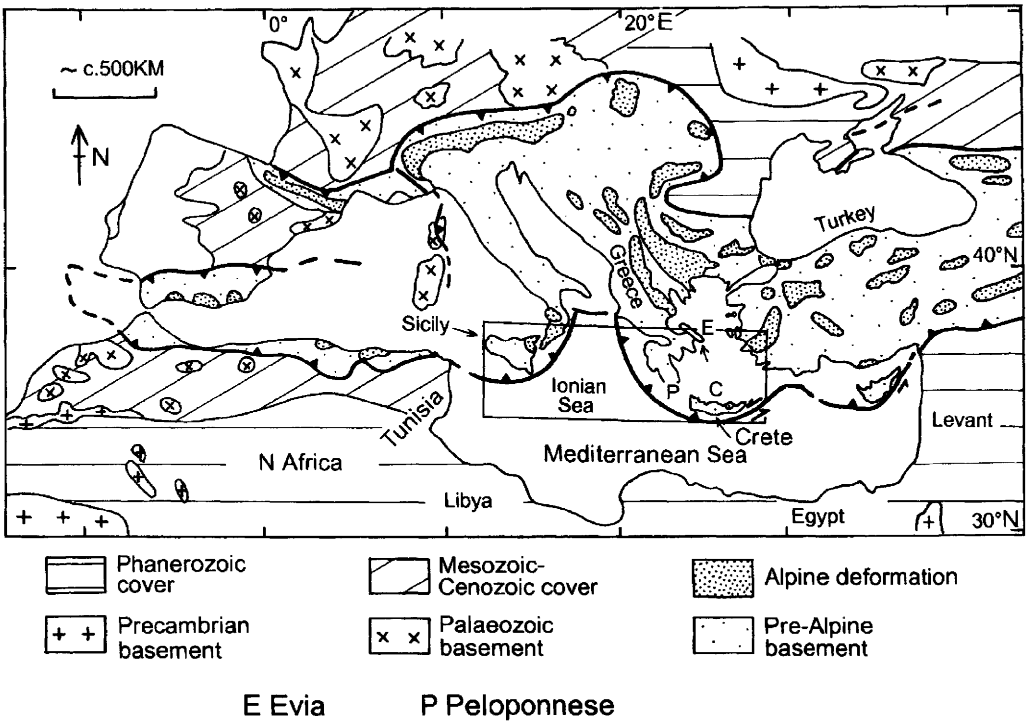


Fig. 1. Outline map of the Mediterranean region showing the major tectonic elements and the study area (within box). Modified after Papanikolaou & Ebner (1996–1997).

to create several Neotethyan oceanic basins. The principal oceanic realm between Eurasia and Gondwana in the Late Triassic lay to the north of the Pelagonian continent in this interpretation and thus no arc remnants or collisional suture existed further south. Model 1, in several variants, was favoured by workers such as Smith *et al.* (1975), Robertson & Dixon (1984), Dercourt *et al.* (1986, 1993, 2000), Robertson *et al.* (1991, 1996, 2004), Papanikolaou (1996–1997), Ricou 1996, Yılmaz *et al.* (1996) and Dornsiepen *et al.* (2001).

In an alternative convergence-related Model 2 (Fig. 2), a Palaeotethyan ocean was also subducted northwards beneath the southern margin of Eurasia during Late Palaeozoic–Early Mesozoic time; and again, the southern, Gondwana margin remained passive. However, a crucial difference is that the Palaeotethyan suture is inferred to be located much further south, within the south Aegean region, to the south of the Pelagonian continent, which is considered as part of Eurasia. In this interpretation an ocean opened along the northern margin of Gondwana during Late Ordovician–Early Silurian time and a continental fragment, termed the Hun terrane, was detached and drifted northwards until it was accreted to Eurasia, with the Palaeotethys opening in its wake along the northern margin of Gondwana. The northern Palaeotethys was in turn subducted beneath Eurasia during the Late Palaeozoic until the Hun terrane collided and was accreted during the 'Hercynian' orogeny. During this subduction a new ocean basin, termed Neotethys in this model, rifted along the Gondwana margin during Late Permian time detaching a Cimmerian microcontinent. The remaining Palaeotethys continued to subduct, opening several Triassic marginal basins (Vardar and Pindos) until it too sutured in the latest Triassic 'Cimmerian' orogeny. The remaining Neotethys survived in this interpretation until Early Cenozoic subduction and eventual suturing of the African and Eurasian plates in the Balkan region. Variants of this interpretation were proposed by several workers (i.e. Pe-Piper 1982; Stampfli *et al.* 1991, 1998, 2001; Stampfli & Borel 2002).

In a radically different Model 3 (Fig. 2), Şengör (1984) proposed that 'Palaeo-Tethys' was rooted in the north, adjacent to the southern margin of Eurasia (e.g. Pontides; Crimea), and that a Neo-Tethyan ocean rifted to the south of this, as one of several back-arc basins above a south-dipping subduction zone during the Triassic. This model, like the first, implies that south of the Pelagonian continent the Triassic setting was one of rifting, not subduction, collision or magmatism. This model has been tested and shown to be problematic based on studies in northern Turkey (e.g. Ustaömer & Robertson 1997), but has recently received renewed support from several researchers (e.g. Smith 1999, Karamata *et al.* 2006; Romano *et al.* 2006).

Finally, Pe-Piper & Piper (2002) have recently proposed an additional tectonic interpretation (Model 4; Fig. 2), based mainly on the geochemistry of Triassic volcanic rocks in Greece, which invokes double subduction (Fig. 2d). This infers Triassic northward subduction from a southerly Palaeotethys in the south Aegean region as in Model 2, but also the presence of an additional Triassic, southward-dipping intra-oceanic subduction zone located in the eastern part of a Triassic Pindos ocean.

Models 2 and 4 require the existence of a Late Palaeozoic–Early Mesozoic convergent margin and a collisional suture in the region of Crete and the Peloponnnese, whereas Models 1 and 3 locate the subduction zone of this age well to the north (albeit with opposite polarities) and imply a rift and passive margin evolution to have characterized the south Aegean region during the Triassic. The different models thus involve starkly contrasting inferences about the tectonic setting at this time, in this region.

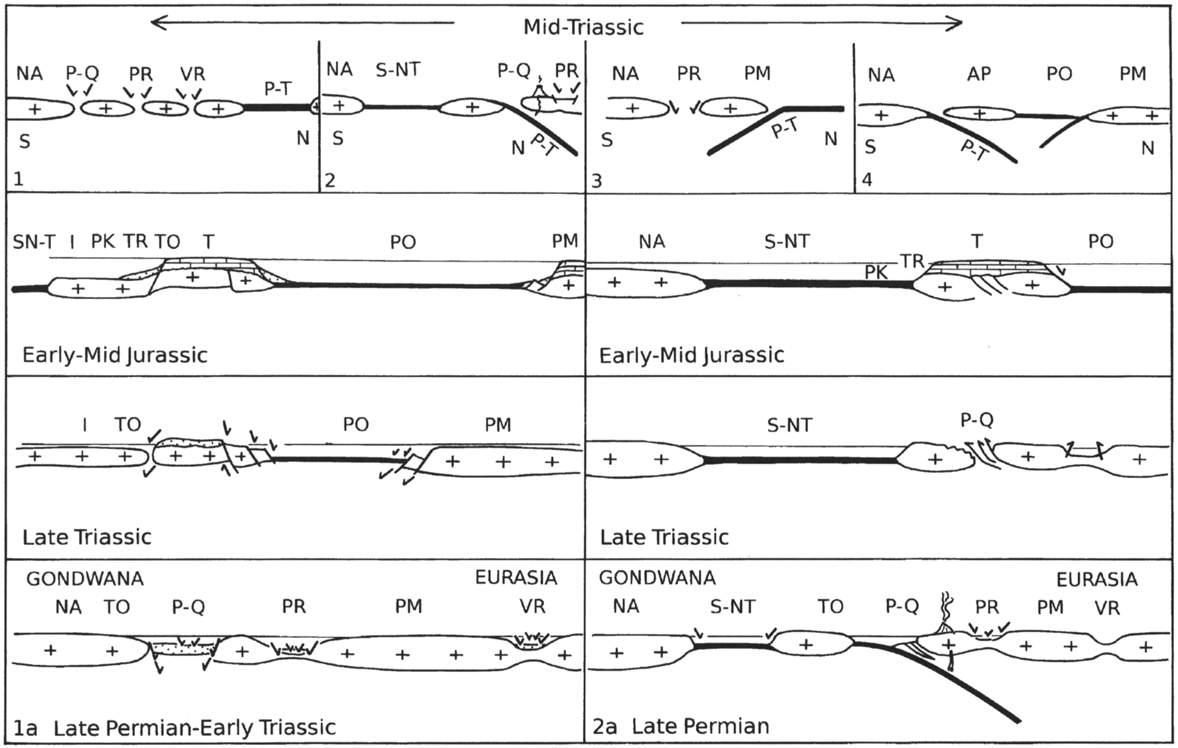
The primary aim of the paper is to present field-based sedimentary evidence from Sicily, Crete, the Peloponnese and Evia which will be used to test the above tectonic hypotheses in the light of the existing literature. The key requirement is to distinguish between generic models, which infer either divergence (Models 1 and 3), or convergence (Models 2 and 4) during pre-Jurassic time, rather than to test any one specific model, as variants of each of these models have been published and further alternatives may exist. The end-product will be a new tectonic model for the south Aegean region for Late Palaeozoic–Early Mesozoic time.

An immediate problem is that the evidence for the existence of any former oceanic crust located along the northern margin of Africa, south of Crete has been obscured by Cenozoic subduction and the present deep-marine basin. The timing and setting of Neotethyan continental break-up cannot be determined from the on-land record of North Africa alone (Guiraud *et al.* 2001). However, further west, in Sicily, Cenozoic northward subduction has already resulted in collision of a Tethyan accretionary prism with a promontory of Gondwana and, as a result, fragments of Late Palaeozoic–Early Mesozoic crust are exposed within a thrust belt in western Sicily (Catalano *et al.* 2000*a*, *b*). These units are critical to determine whether or not a 'Neotethyan' ocean existed in the South-Mediterranean during Late Palaeozoic–Early Mesozoic time. This area will be discussed first as a proxy for crust of this age south of Crete. The Upper Palaeozoic–Lower Mesozoic metasedimentary and metavolcanic units of Crete and the Peloponnese will then be considered. Evidence from the Pindos and Pelagonian zones further north in Greece is also important, particularly to determine if an early Mesozoic 'Cimmerian' collisional event affected these areas.

One persistent problem is that Tethyan nomenclature tends to be model-specific. Thus, for Şengör (1984) 'Palaeo-Tethys' is rooted in a relatively northerly location, whereas for Stampfli *et al.* (2001) their Palaeotethys is rooted further south, and, by definition, Neotethys even further south again (Fig. 2). When such a model-specific nomenclature is adopted, one is at once locked into hypothesis confirmation rather than hypothesis testing (see Robertson & Mountrakis 2006). For this reason, a looser, non model-dependent approach is used here. The term Palaeotethys as used here refers generally to older (i.e. pre-Mid-Jurassic) oceanic crust, and the term Neotethys to generally younger oceanic crust (i.e. Late Triassic–Early Cenozoic).

The writer was unable to discriminate between the alternative tectonic models from the literature alone, and so decided to embark on a field-based study of the critical areas that has taken several years (Fig. 1, inset). There is no simple shortcut to understanding the pre-Jurassic tectonic evolution of the south Aegean region other than in-depth studies of the lithological assemblages in each of these areas, followed by comparisons and synthesis, which also takes account of evidence from the wider region and modern tectonic settings. A substantial body of new information has became available during this work, mainly concerning the sedimentary facies and palaeotectonic setting of the Upper Palaeozoic–Lower Mesozoic units in the region. The main results of a 10-year study of comparable units in western Turkey were recently summarized elsewhere (Robertson *et al.* 2002) and will be drawn on in the discussion section.

The criteria for discriminating between tectonic settings are first outlined. The alternative



AP	Apulia	P-T	Palaeotethys	T	Tripolitza platform
I	Ionian basin	P	Pindos ocean	TO	Taleo Ori
NA	North Africa	PR	Pindos rift	TR	Tripali
PK	Plattenkalk	SN-T	South Neotethys	VR	Vardar rift
PM	Pelagonian microcontinent				

Fig. 2. Alternative tectonic models for the south Aegean region. Model 1, Northward subduction with the trench located in a southerly location; Model 2, Northward subduction with the trench located in a northerly location; Model 3, Southward subduction with the trench located in a northerly location; Model 4, Double subduction model. (See text for discussion and references). The four models are shown at the top of the figure, then Models 1 & 2 are shown in more detail below.

possible interpretations of each area are reviewed, and an indication of which model is favoured is given before moving on to the next area. Salient aspects of the wider regional setting, outside the area studied (e.g. Eurasian margin; central European Hercynian orogen) will be considered in the discussion section.

Many of the Upper Palaeozoic–Lower Mesozoic units of Crete and the SW Peloponnese, in contrast to western Sicily and Evia, have undergone HP–LT (blueschist-facies) metamorphism (Seidel 1978). For example, the extensive Phyllite–Quartzite unit in Crete was at least partially metamorphosed under high-grade conditions (8–19 kbar, 300–400 °C) during Late Oligocene–Early Miocene time (Seidel *et al.* 1982; Theye *et al.* 1992; Zulauf *et al.* 2002). However, primary sedimentary lithologies and sedimentary structures, including stratigraphical way-up evidence, are still commonly recognizable. For this reason, rock types will be generally referred to here in their pre-metamorphosed states, dropping the ubiquitous 'meta-'; i.e. psammitic schists were commonly sandstones, marbles were limestones or dolomites, and pelitic schists were mudstones, etc. An informal stratigraphical terminology is used, as different local names have often been used for similar units in different areas. The time scale is that of Gradstein *et al.* (2004). Coordinates given refer to the present day unless specified otherwise.

Criteria for recognizing tectonic settings

A combination of biostratigraphical, sedimentary, igneous and structural evidence (termed tectonic facies; Robertson 1994) allows different tectonic settings to be distinguished. Some of the main, relevant tectonic settings are as follows.

Divergence-related tectonic settings include rifts, failed rifts (aulacogens) and intra-platform basins. Sedimentary environments associated with passive margins and marginal platforms include siliciclastic shelves and carbonate platforms. Tectonic settings associated with spreading centres and oceanic basins include spreading ridges, abyssal plains, continental fragments, oceanic seamounts and oceanic plateaux. Conversely, tectonic settings associated with convergence-related settings include supra-subduction zone spreading centres (i.e. many ophiolites), oceanic arcs, subduction–accretion complexes, fore-arc basins, intra-oceanic back-arc basins and intra-continental back-arc basins. Tectonic settings associated with collisional tectonic settings include intra-oceanic collision zones, foreland basins and the sedimentary products of collision ('molasse'). Additional tectonic settings characterize strike-slip-related settings (e.g. pull-apart basin), which could also be relevant here. Recognition of such tectonic settings in the south Aegean region should allow the alternative tectonic models to be distinguished.

The recognition of such tectonic settings in metamorphic terranes as in the south Aegean region is obviously difficult, but still possible where the protoliths of the sedimentary and igneous rocks can be recognized and where the sediments are reasonably well dated. As a cautionary note, however, it should be noted that metamorphic rocks that have undergone HP–LT metamorphism, like those of the south Aegean region, have been exhumed from a subduction zone setting so that parts of the original record may have been lost. Also, some subduction settings involve net loss of material from the overriding plate (i.e. subduction erosion) such that some critical tectonic units (e.g. accretionary prisms) may be lost.

The main tectonic settings that would be expected to occur for each of the four main alternative tectonic settings of the south Aegean region are as follows. In a divergence (rift)-related model (Models 1 and 3) the tectonic facies would be those of rifts, passive margins and Atlantic-type ocean basins. In a convergence (subduction)-related model (Models 2 and 4) the expected tectonic settings for the Triassic would identify both active margin (e.g. subduction complexes; magmatic arcs) and collisional settings (e.g. foreland basins). Also in these models, additional divergence-related tectonic settings would characterize the Late Palaeozoic, inferred break-up and spreading of 'Neotethys' adjacent to Gondwana. Time relations are therefore clearly critical to distinguish the tectonic alternatives.

The southward subduction hypothesis (Model 3) should also be characterized by divergence-related tectonic settings, but coupled with igneous geochemical evidence of subduction. Finally, the model invoking both southward and northward subduction (Model 4) would imply the existence of two belts characterized by convergence-related tectonic settings and two belts of subduction-related magmatism, one intra-continental (Andean type) and the other intra-oceanic.

In summary, in the extension-related models (Models 1 and 3) only a limited number of tectonic settings would be represented, whereas many more would need to have existed for the convergence (subduction)-related models (Models 2 and 4).

Tectonic units of the south Mediterranean region

The entire south Aegean region and areas to the west, as exposed in the Italian region (e.g. Calabria and Sicily), comprise piles of thrust sheets that were mainly emplaced during Cenozoic time related to northward subduction and suturing of the Neotethyan ocean. In this paper, units will be discussed in turn, working structurally upwards on a regional basis, beginning with those at the structural base that restore closest to Gondwana and ending with those that restore furthest north.

As noted above, the most southerly unit, representing Neotethyan crust that formerly separated North Africa from the Cretan units, has been subducted or is located deep beneath the Sea of Crete and is not exposed and has not been sampled by drilling. The main evidence for the existence of this southerly oceanic basin is the record of subduction obtained from the Mediterranean Ridge accretionary complex south of Crete (Camerlenghi *et al.* 1995; Chaumillon & Mascle 1997), and the record of Cenozoic HP–LT metamorphism within the Cretan nappes (Seidel 1978). A history of rifting is documented by wells and exposures in North Africa to the south (Guiraud *et al.* 2001) but it is not possible to determine from this when spreading of an adjacent southerly Neotethyan ocean began; possibilities include Late Permian, Mid–Late Triassic or Late Jurassic–Early Cretaceous. For this reason it was decided to study the Late Palaeozoic–Early Mesozoic of Sicily as a proxy for the oceanic basin between North Africa and Crete. This is reasonable, as the entire North African margin from the Nile to Morocco shows evidence of a comparable history of rifting during Late Palaeozoic–Early Mesozoic time and, indeed, some workers restore the Sicanian basin (Sicily) of this age to a location south of Crete or even further east (e.g. Garfunkel 2004). In Sicily, the Cenozoic thrust belt exposes units that formed in a southerly, Sicanian basin bordering Gondwana from the inception of this basin, during Late Palaeozoic time. There is thus an opportunity to determine the tectonic setting of the North African margin in this region during this period. In particular, does the lithology present record a rift setting as in Models 1 and 3 or a spreading-related setting as in Models 2 and 4?

In the discussion below relevant aspects of the geology of western Sicily will be considered and it will be shown that the available evidence supports Model 1 and that there is no firm evidence in support of Models 2, 3 or 4.

Evidence from the Permo-Triassic of eastern Sicily

Relevant outcrops occur in two main areas: Lercara–Roccapalumba and in the Sosio Valley (Fig. 3). These units are unmetamorphosed, in contrast to many of those of the south Aegean, which will be discussed later. The Permo-Triassic units of western Sicily were finally emplaced as a result of Late Neogene (Miocene–Pliocene) southward thrusting over the North African continental margin (Catalano *et al.* 2000*a*, *b*). This took place during closure of the Mesozoic Piedmont–Ligurian ocean to the north. Field evidence is augmented by results from local shallow drilling (e.g. in the Sosio Valley) and from hydrocarbon exploration drilling (e.g. Roccapalumba-1 well) (Catalano *et al.* 1991). It will be argued that the sedimentary sequences are most consistent with a progressively deepening rift basin that nevertheless continued to be supplied periodically with shallow-water carbonate debris and terrigenous clastic deposits.

Lower Permian clastic sediments and basic igneous rocks

Exposures are few and far between (e.g. in railway cuttings) in an area of rolling hills and farmland. The main lithologies are terrigenous and carbonate turbidites, siltstones and shales, with subordinate detached blocks and coarse carbonate debris flows of mainly shallow-water-derived material. The subsurface thickness exceeds 1000 m (e.g. in Roccapalumba-1 well), although stratal repetition is possible. Exposures are commonly highly deformed, steeply dipping, or inverted. Typically, thinner bedded units are strongly sheared, whereas thicker-bedded, more competent units have survived as undeformed beds, cemented by calcite spar. The mudstones within the turbiditic sequence are dated as Early Permian (Artinskian) by conodonts (Catalano *et al.* 1991). Occasional alkaline basic sills have been reported (Di Stefano & Gullo 1997).

The Early Permian clastic sediments were examined in small exposures, SSW of a railway station, SE of Roccapalumba (*c.* 300 m west of the River Torto). Mudstones are greyish to reddish in colour and include deep-water trace-fossils of the *Nereites* ichnofacies (Kozur *et al.* 1996). The mudstones are intercalated with thin-bedded (5–10 cm) siliciclastic turbidites, exhibiting well-developed, partial Bouma sequences. Individual sandstone turbidites reach 1.6 m in thickness in this area. Thin-section study shows that carbonate grains are typically more

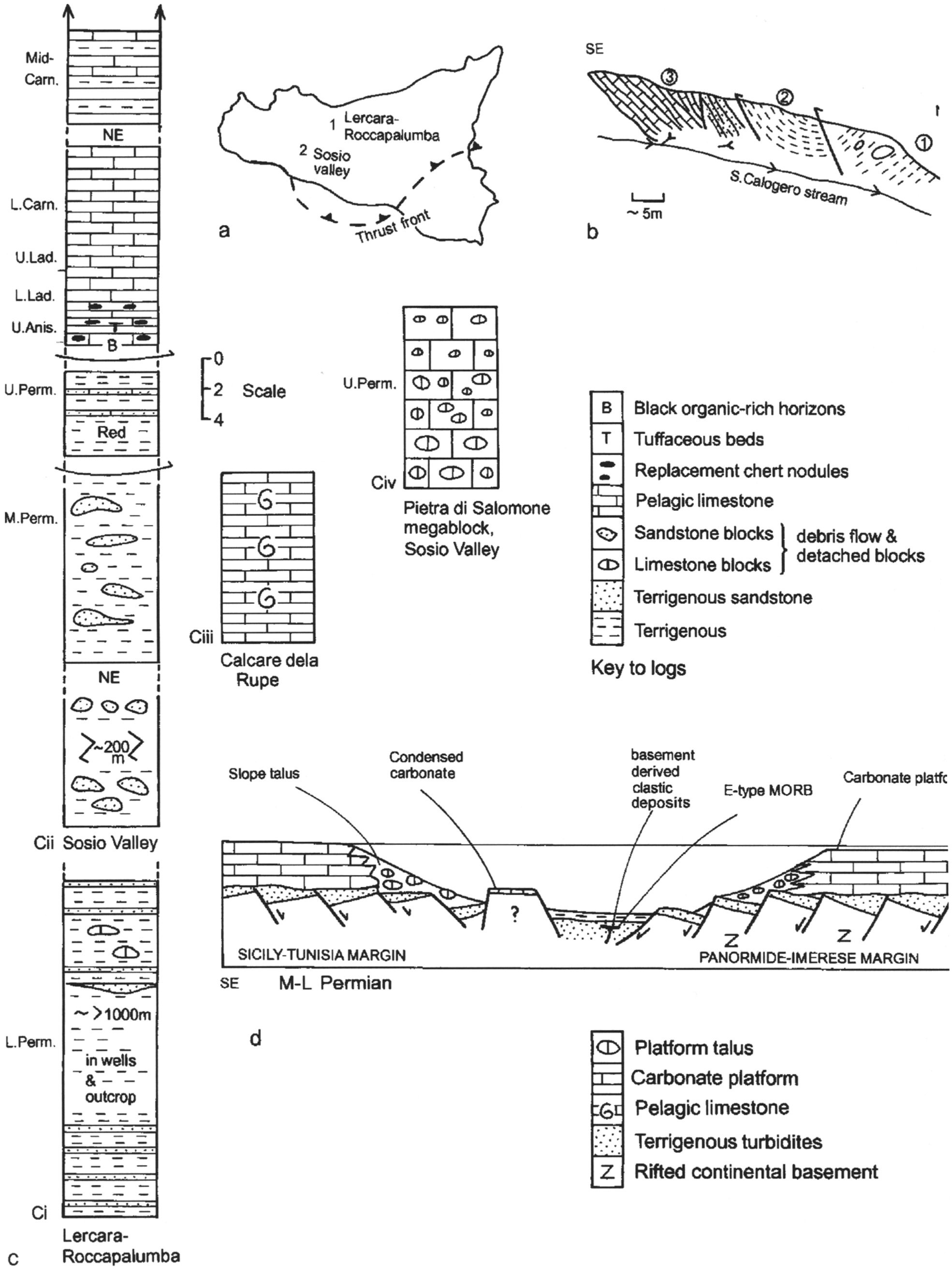


Fig. 3. Upper Palaeozoic–Lower Mesozoic sedimentary and volcanic units exposed in central western Sicily. **(a)** Location; **(b)** sketch section of unit exposed in the Sosio Valley; **(c)** simplified sedimentary logs; **(d)** possible tectonic setting. (See text for explanation and data sources.).

numerous than terrigenous ones. In general, the sandstones contain quartz and muscovite, with subordinate biotite and feldspar and rare zircon (Di Stefano & Gullo 1997).

Thin-section study shows that both the terrigenous and carbonate grains are mainly angular to sub-angular. The terrigenous grains are mainly monocrystalline quartz, polycrystalline quartz (quartzite), mica-schist, plagioclase (mainly altered plagioclase and perthite), muscovite (including large unstrained laths), biotite and rare zircon. Occasional large grains containing plagioclase, quartz and biotite were probably derived from granitic rocks. There are also grains of weakly recrystallized quartzose sandstone, with individual grains set in a matrix of microcrystalline silica. Siltstone rip-up clasts, rich in quartz and muscovite, are also seen, together with rare grains of reworked pelagic micrite with calcite-replaced radiolarians and detrital microcrystalline quartz (?metachert). Some quartz grains are coated with calcareous algae, indicating a shallow-water origin. Carbonate grains are dominated by detrital grains of algal micrite and encrusting calcareous algae together with pisoliths, small oncolites, grapestone intraclasts, echinoids (some coated with calcareous algae), shell fragments, coral (replaced by coarse calcite spar), bryozoans, benthic Foraminifera (e.g. *Miliolina*), ostracodes, gastropods and recrystallized carbonate (marble). The matrix is micritic with scattered diagenetic pyrite.

There are also occasional interbeds of carbonate debris flows (up to 1 m thick), containing clasts (up to 2 cm in size) of mainly neritic carbonate, including coral, encrusting algae, gastropods and fusulinids set in a partially recrystallized calcite spar cement. Rare blocks of deep-water carbonate contain an Early Permian fauna, including ammonoids, trilobites, brachiopods, crinoids and Radiolaria. In general, the siliciclastic turbidites tend to be uniformly fine to medium grained, whereas the associated redeposited carbonates are considerably coarser grained, suggesting a more proximal origin.

In the Leccara area there are several exposures of diabase and basalt, forming sheets up to 30 m thick (e.g. Contrada Rettino body, *c.* 3 km SW of Roccapalumba). These are interpreted as high-level intrusions into wet sediments. They contain phenocrysts of altered plagioclase, olivine, Al–Ti augite and exhibit an enriched mid-ocean ridge basalt (E-MORB) composition (Censi *et al.* 2000). Igneous rocks of similar age and composition are present in the Hyblean area (Bianchini *et al.* 1998).

Middle Permian succession

Key outcrops of deep-water sediments of Mid–Late Permian age occur in the classic Sosio Valley, SW of Palazzo Adriano (Fig. 3a). These units are highly disorganized and were mapped as tectonic mélange, up to 500 m thick, related to Late Neogene thrusting (Di Stefano & Gullo 1996, 1997). The Permian–Triassic deep-sea sediments in this area are sheared and imbricated, and may include the deformed limb of a large inverted isoclinal fold.

The stratigraphically lowest unit is composed of turbidites that are compositionally similar to, but younger (i.e. earliest Mid-Permian) than those of the Lercara–Roccapalumba area, described above (Catalano *et al.* 1991). Local successions, up to several hundred metres thick, in the Sosio Valley are highly disorganized and have been interpreted as olistostromes (Catalano *et al.* 1991). Detached blocks are strewn through a shaly matrix, which locally contains a reworked Early Permian fauna. The individual blocks exhibit layer-parallel extension of thick turbiditic beds to produce classic sandstone 'phacoids'. Similar 'olistostromes' are exposed slightly further upstream to the SE, in tectonic contact with contrasting red mudstones of Late Permian age (Fig. 3b and ci). Several of the large detached sandstone blocks exhibit sheared and slickensided margins, and internal brecciation, showing that they were well lithified before being incorporated into the 'olistostrome'. This questions whether these are really sub-aqueous debris-flows of Mid-Permian age or instead the uppermost levels of a thick Early–Mid-Permian turbiditic sequence that was sheared strongly to create mélange during the Cenozoic thrusting.

Thin-section study of the Middle Permian turbiditic clastic sediments shows that they are more mature, both texturally and chemically, than the turbidites of the Lower Permian interval described above. The terrigenous grains range from sub-angular, to rounded and very well rounded. Grains are mainly unstrained monocrystalline quartz with, in addition, rare strained quartz, plagioclase, fine-grained polycrystalline quartz (quartzite) and rare grains of partially recrystallized siltstone. The carbonate bioclasts are, by contrast, relatively angular, and include all of the neritic grains (e.g. lithoclastic algal micrite) as in the underlying Lower Permian redeposited sediments, with the addition of numerous radiolarians. The matrix is partially recrystallized micrite, with scattered diagenetic pyrite.

Upper Permian succession

The 'olistostrome' is faulted against a younger unit composed of bright red, weakly consolidated, glutinous claystones that contain a rich fauna of Late Permian radiolarians, conodonts and ostracodes (Catalano *et al.* 1991; Kozur 1993; Fig. 3cii). These red mudstones are interbedded with several thin, graded interbeds of redeposited neritic carbonate (<15 cm thick), containing Radiolaria, Foraminifera and conodonts of Late Permian age. The packstone–grainstones also contain minor amounts of quartzose silt and sand (Catalano *et al.* 1991). Thin sections studied reveal mainly redeposited grains of neritic carbonate, as in the underlying Permian coarse clastic facies, especially algal micrite, together with shell fragments, benthic Foraminifera and reworked grains of radiolarian micrite.

Middle–Upper Triassic succession

The Middle-Permian succession is faulted against a contrasting, mainly pelagic carbonate succession of Mid–Late Triassic (Late Anisian–Mid-Carnian) age (Fig. 3b). The Lower Triassic (Scythian) succession is typically absent in Sicily (Catalano *et al.* 1991). The presence of well-graded beds shows that the Triassic succession is inverted. It begins with thin- to medium-bedded, greenish to dark grey radiolarian mudstones, together with tuffaceous and siliceous pelagic limestones (in beds <20 cm thick), of Late Anisian–Early Ladinian age, and then passes into grey to pink nodular and cherty limestones with marly partings (Catalano *et al.* 1991). The lowest interbeds are dark and apparently organic rich. Lenticles and nodules of quartzitic chert of replacement origin are common near the base, but generally decrease in abundance stratigraphically upwards. The highest exposed beds, of Late Ladinian–Early Carnian age, are thin-bedded pinkish pelagic carbonates, with minimal chert.

Thin-section study shows that the carbonates begin with dark grey terrigenous silt, compositionally similar to the material in the underlying Permian terrigenous sediments, dominated by quartz, muscovite and subordinate plagioclase. There are also abundant calcite-replaced radiolarians and recrystallized shell fragments. The directly overlying silty limestone is composed of radiolarians, mainly replaced by microcrystalline and chalcedonic quartz, together with well-aligned straight-shell fragments (*Halobia*), set in a micritic matrix. Pink and grey pelagic carbonates above this are dominated by pelagic micrite (with no terrigenous component), with calcified radiolarians, *Halobia* and *Daonella* pelagic bivalves, pelagic gastropods, ostracodes and rare benthic Foraminifera.

Coeval Permian pelagic and neritic carbonates

Several small limestone blocks are located further NE (downstream) in the Sosio Valley, and may record contrasting slope or fault-block-type facies (Catalano *et al.* 1991; Kozur 1995).

Notably, the ‘Rupe del Passo di Burghio block’ comprises Middle Permian (Wordian), white to grey, ammonoid- and conodont-bearing pelagic limestones (Hallstatt-type facies), with interbeds of redeposited bioclastic calcarenites (Kozur 1995; Di Stefano & Gullo 1997; Fig. 3ciii). In addition, a limestone block (SE of Rupe del Passo di Burgio) includes yellow to green clays with conodonts of Mid-Permian age (Catalano *et al.* 1991). There are also several other blocks of condensed rosso-type facies. Kozur (1995) suggested that these blocks record remnants of Permo-Triassic condensed successions that were derived from widely differing areas of Palaeotethys, characterized by the presence or near absence of a *Pseudofurnishius* (conodont) fauna. Alternatively, the blocks originated within several different palaeoenvironments, but all within a relatively local rift basin setting (Di Stefano *et al.* 1996). The relative abundance of conodonts was instead controlled by facies variation and sedimentation rate variation (i.e. dilution effects) in the latter view.

The deformed Permian–Triassic deep-water succession is structurally overlain by a large block of highly fossiliferous limestone, the well-known Pietra di Salmone (Fig. 3civ). This unit is interpreted as the dismembered limb of a WNW-facing monoclinal fold (Flugel *et al.* 1991). This large block is dominated by a generally fining-upward sequence of redeposited neritic talus (*c.* 70 m thick). The carbonate talus ranges from disorganized, angular blocks and clasts (up to 3.5 m in size), to crudely stratified clast- to matrix-supported debris flows, to pebbly and to gravel-sized calciturbidites with marly tops. The constituents include a rich and diverse fauna of corals, algae, sponges, Foraminifera and conodonts. Some of the clasts, dated as Early–Mid-Permian, were redeposited into a matrix of yellowish, partly dolomitized marls containing conodonts of Mid-Permian age.

Carnian and younger successions

Succession of Late Triassic and younger Mesozoic age are well exposed in a more intact, structurally associated succession (e.g. Pizzo Mondello; Di Stefano & Gullo 1997). These sediments are dominated by marls, claystones and thin-bedded calcilutites and contain Radiolaria, *Halobia* and conodonts (‘Mufara Formation’). Comparable facies, up to *c.* 400 m thick, occur widely in central–western Sicily and include black organic-rich anoxic facies. There are also redeposited neritic carbonates, probably derived from large carbonate platforms (e.g. Hyblean Plateau; Di Stefano *et al.* 1996). Quartzose sediments, including polycrystalline quartz, muscovite and feldspars, are present in some sections and appear to increase in abundance northwards, suggesting derivation from a metamorphic basement exposed in this direction (e.g. Hercynian Kabilo–Peloritani units).

The succession continues upwards into uppermost Triassic pelagic *Halobia*-bearing limestones, commonly siliceous, with continued interbeds of redeposited neritic carbonate. Locally, the Upper Norian interval exhibits huge slump breccias associated with terrigenous clastic input and a low-angle unconformity (e.g. Imerese unit, northern Sicily; Di Stefano *et al.* 1996). A strike-slip controlled setting was suggested by Di Stefana & Gullo 1997), although an associated unconformity in adjacent areas (e.g. Tunisia) might also be explained by a pulse of extension-related flexural uplift. Lower Jurassic cherty pelagic limestones and marls above this are followed by a more or less continuous deep-water succession dominated by radiolarian marls and pelagic limestone, with locally variable intercalations of redeposited neritic carbonate (e.g. Imerse and Sicanian units). Widespread volcanism occurred in the Late Jurassic within some basinal (e.g. Sicanian) and platform (e.g. Hyblean) units. Some Late Triassic Mesozoic platforms were drowned and covered by Ammonitico Rosso (e.g. Hyblean platform), whereas neritic platform deposition persisted elsewhere (e.g. Sicilide units). Passive margin conditions persisted until the end of Mesozoic time, after which convergence began (Catalano *et al.* 2000*b*).

Interpretation: a subsiding deep-water rift

The thick, Lower Permian sandstones record accumulation in a deep-water basin (Fig. 3d), bordered by a relatively local carbonate platform and a metamorphic landmass, presumably the Hercynian orogen to the north (e.g. in the Peloritani Mountains, NE Sicily; Di Stefano & Gullo 1997). The source area included crystalline basement (e.g. mica schist), granitic rocks, low-grade siliciclastic sediments and metacarbonates. The inception of the basin must significantly predate the earliest known Early Permian deep-water sediments. The Middle Permian red claystones accumulated in a fertile relatively deep sea, with deep-water currents connecting to the eastern Tethys (e.g. Crete; Oman; Catalano *et al.* 1991; Kozur 1993). The thin calciturbidites with neritic detritus attest to the proximity of a Late Permian carbonate platform. This platform and related slope possibly included the Pietra di Salomone and several other local limestone blocks in the Sosio Valley. The Pietra di Salomone unit began to accumulate on a carbonate ramp bordering a well-established carbonate platform during Mid-Permian time. The slope later steepened, shedding coarse talus, then stabilized by latest Permian time. The Petra di Salomone 'megablock' specifically represens the fill of a steep-sided channel or canyon cut into the margin of a carbonate platform (Flugel *et al.* 1991). By contrast, the blocks of pink pelagic Hallstatt-type facies accumulated on isolated highs within the basin (e.g. Bernouilli & Jenkyns 1974), but the nature of their basement is not known. It is probably not necessary to invoke tectonic transport of units from quite different oceanic areas, as suggested by Kozur (1993, 1995), but instead different facies may represent deposition in different local palaeo-environments with different sedimentation rates and variable preservation of fossils. The Middle Triassic (Ladinian) interval then documents a quiet pelagic carbonate-depositing setting, in a still-fertile sea, rich in radiolarians.

Although no unbroken succession is preserved, it is likely that, prior to Neogene thrusting, a remarkably long-lived, more or less continuous deep-water succession existed, extending from Early Permian to Early Cenozoic time. No base of the succession is exposed and the Lower Permian siliciclastic turbidites are relatively distal, in contrast to the coarser neritic carbonate material that was supplied from a more local carbonate platform. The overlying 'olistostrome' could indicate a period of tectonic instability during Mid-Permian time according to Catalano *et al.* (1991). By contrast, stable sea-floor conditions are indicated for Mid-Permian time when well-oxidized, fertile hemipelagic sediments accumulated in a deep-water basin (500 m or more) (Catalano *et al.* 1991).

The Lower–Middle Permian turbiditic clastic facies might, in principle, represent different tectonic settings. The first could be a flexural foreland basin related to the Hercynian orogeny (Catalano *et al.* 1991; Ziegler & Stampfli 2001). A foreland basin setting is, however, unlikely as there is no evidence of a foreland basin stratigraphy (i.e. a thickening and coarsening-upward succession), of any overthrust load in the area, or of flexural rebound after a collisional event; rather, deepening continued during the Triassic. Second, the deep-sea sediments might record an accretionary prism (i.e. turbidites) and an associated perched fore-arc basin (i.e. pelagic carbonates) related to subduction. However, there is no evidence within the kilometre-thick siliciclastic turbidites of exotic accretionary material (e.g. pelagic sediments or ophiolites), or thrust repetition of the succession as in an accretionary prism. Third, the succession might record a rift-basin (?transtensional) that post-dated the Early Carboniferous Hercynian orogeny (Di Stefano *et al.*

1996). Of these alternatives, a rift setting best fits the sedimentary record and is consistent with the presence of the intruded 'enriched' basaltic igneous rocks (i.e. E-MORB; Censi *et al.* 2000) and sparse Middle-Triassic silicic tuffs. N-type MORB that could indicate the existence of strongly stretched continental crust or oceanic crust in the vicinity is not recorded.

A rift-related (or transtensional) setting can be related to post-Hercynian extension during Mid?–Late Carboniferous time. By Early Permian time a broad palaeogeographically diversified deep-water basin existed (Fig. 3d), open to Tethys to the east. Condensed pelagic sediments probably accumulated on isolated horsts that were bypassed by gravity flows. By the Early Anisian there was a general switch to hemipelagic carbonate deposition, possibly reflecting increased input of peripelagic carbonate from adjacent carbonate platforms. Continuing high siliceous productivity is suggested by the presence of replacement chert. Variable, deep- or shallow-water carbonate conditions persisted throughout Mesozoic time, punctuated by periodic tectonic instability that triggered mass flows on the margins of carbonate platforms (Di Stefano & Gullo

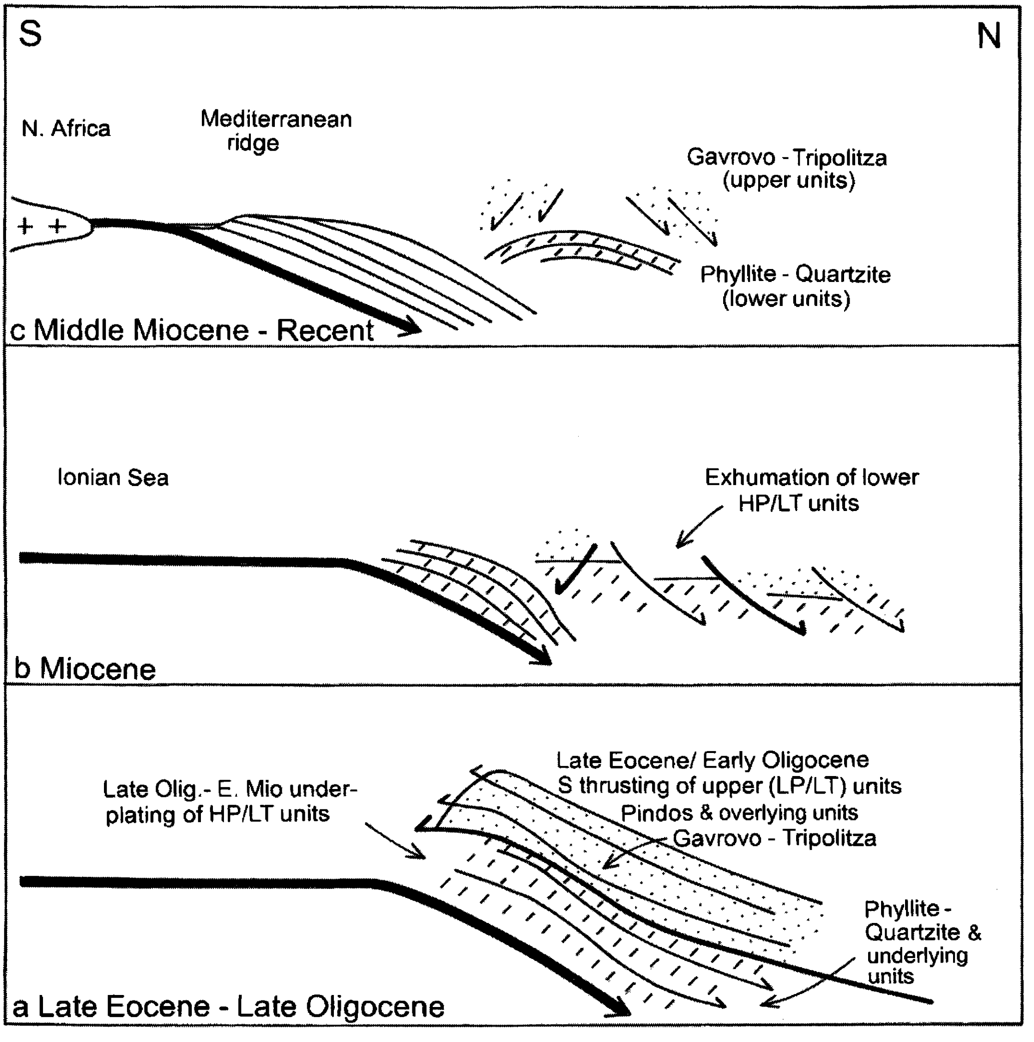


Fig. 4. Interpretative sketch sections to illustrate the Cenozoic Alpine evolution of the South Aegean, which was dominated by collision and exhumation. The effects of the resulting deformation and metamorphism have to be removed to allow interpretation of the Upper Palaeozoic–Lower Mesozoic units discussed in this paper. (See text for discussion and literature.).

1996, 1997). Widespread volcanism and subsidence of carbonate platforms (e.g. Hyblean platform; SE Sicily) during Mid-Jurassic time was broadly coeval with spreading of the central North Atlantic. This rifting could also relate to break-up and the onset of sea-floor spreading to form the Ionian Sea (Catalano *et al.* 2001), with implications for the timing of spreading further east (e.g. Sea of Crete). There is no evidence of the existence of an oceanic basin in the Sicily area until this time.

In summary, the sedimentary facies and igneous rocks exposed in western Sicily are indicative of a rift or failed rift (aulacogen) setting as in Model 1. A deep-water basin open to the Tethys to the east clearly existed but there is no evidence of oceanic crust (ophiolites or related hydrothermal deposits), or of a subsiding passive margin in which an overall thinning- and deepening-upward sedimentary succession would be expected (e.g. similar to the NW African or the eastern USA Mesozoic continental margins). Instead, a slowly subsiding rift persisted until this was reactivated related to opening of the central North Atlantic during Mid–Late Jurassic time. The main conclusion of fundamental importance from Sicily carried over to the south Aegean region, discussed below, is that there is no evidence for the existence of a Late Palaeozoic ocean basin bordering North Africa in this region.

Cenozoic setting of the South Aegean region

Crete and the Peloponnese document palaeotectonic units that were located to the north of a now largely subducted southerly Neotethyan ocean basin that rifted along the North African continental margin. In contrast to Sicily, the Upper Palaeozoic–Lower Mesozoic units there have experienced metamorphism ranging from HP–LT in the structurally lowest units to LP–LT in the higher units.

Any interpretation of the pre-Cenozoic tectonic setting depends critically on correctly restoring the effects of Early Cenozoic–Recent deformation and metamorphism. During the Early Cenozoic, Neotethyan basins progressively closed, affecting, in turn, the Vardar zone to the NE, the Pelagonian zone, then the Pindos zone and finally Apulia–North Africa to the SW (e.g. Dercourt *et al.* 1993; Robertson *et al.* 1999). The higher thrust sheets, the Pindos zone, and the Tripolitza platform and the units above were largely detached from their former Mesozoic–early Cenozoic stratigraphic cover associated with the later stages of closure of Neotethys. They were then tectonically assembled during Late Eocene–Early Oligocene time related to generally northward subduction. The lower, more southerly derived nappes are represented by the Plattenkalk, Tripali and Phyllite–Quartzite units; these units were underplated to the hanging wall of a subduction trench during Late Oligocene–Early Miocene time, resulting in HP–LT metamorphism, internal thrust imbrication, large-scale structural inversion and polyphase folding. Fission-track dating (Thompson *et al.* 1998) indicates that crustal thickening culminated in bi-vergent ductile extensional exhumation during Late Oligocene–Early Miocene time (24–15 Ma) (Fig. 4b). The lower nappes experienced extension-related folding, shearing and large-scale detachment near the contact with the overlying unmetamorphosed Pindos and Tripolitza nappes and also internally (Bonneau 1984; Papanikolaou 1988; Fassoulas *et al.* 1994; Kilias *et al.* 1994, 2002; Jolivet *et al.* 1996; Zulauf *et al.* 2002). The lower nappes in western Crete were rapidly exhumed, whereas exhumation was possibly slower in central Crete, associated with retrograde metamorphism (Fassoulas *et al.* 1994). Subsequently, Mid-Miocene–Recent time was dominated by pulsed extension (Fig. 4c) in an above subduction zone setting, with an important component of strike-slip–transtension in eastern Crete (e.g. ten Veen & Meijer 1999).

In Crete (Fig. 5), the Tripali and Plattenkalk units are discussed first as they restore to a more southerly position than the Phyllite–Quartzite unit when the Cenozoic tectonic effects are removed.

Tectonostratigraphy of Crete

In Crete, the lowest exposed unit in the tectonostratigraphy is the Mesozoic–Lower Cenozoic Plattenkalk unit (Creutzburg *et al.* 1977; Fig. 6). The Plattenkalk is dominated by grey platy pelagic limestones, which contain numerous chert nodules. The lower part of the Plattenkalk unit is represented by the Talea Ori unit, best exposed in central northern Crete (Krahl *et al.* 1988; Fig. 7). The Talea Ori is dominated by Upper Palaeozoic shallow-marine terrigenous and carbonate facies above a probable continental basement. The next unit above the Talea Ori–Plattenkalk unit is the Tripali unit. Based on recent studies in SW Crete (e.g. Lefka Ori) Krahl & Kauffman (2004) suggested that the Tripali unit extends in age from the Early Liassic to Late Cretaceous (or even Early Cenozoic; see below). In the type area, in the Tripali Mountains, the Tripali unit forms a separate tectonostratigraphic entity (Kopp & Ott 1977). However, in western Crete, an important mainly metacarbonate and meta-evaporitic succession of Triassic age was initially referred to the Tripali unit (Gips-Rauhwacke Formation of Wurm 1950) but was later reinterpreted as an

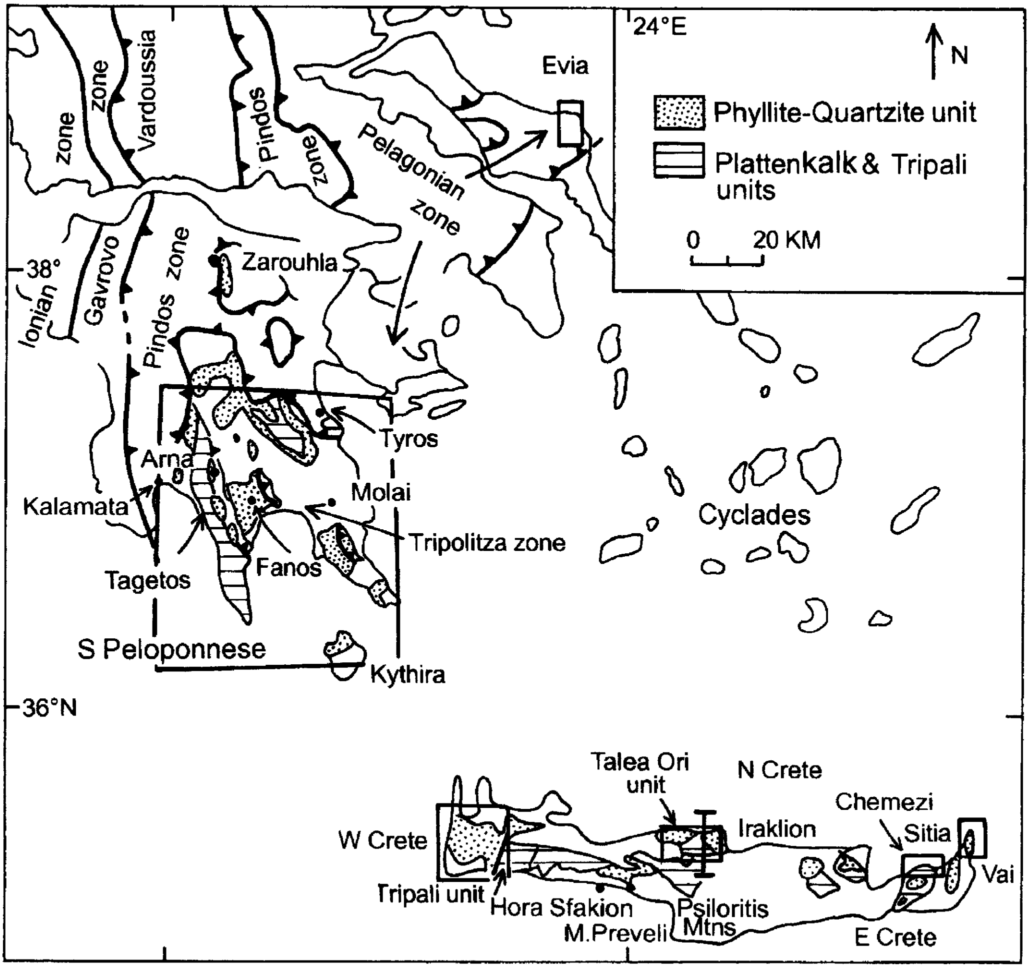


Fig. 5. Outline map of Crete and the Southern Peloponnese showing the main locations of the Upper Palaeozoic–Lower Mesozoic Phyllite–Quartzite unit and related units. A cross-section of north central Crete is shown in Figure 7. More detailed maps of areas within boxes are shown in subsequent figures.

upward continuation of a regionally inverted succession of the Upper Palaeozoic–Lower Mesozoic Phyllite–Quartzite unit (Krahl *et al.* 1983*b*; see below).

The Phyllite–Quartzite unit includes Middle Carboniferous–Lower Cenozoic, mainly siliciclastic successions that are most widely exposed in western Crete (Fig. 5). In addition, contrasting units of 'Hercynian' basement slices, Permian hemipelagic sediments, Triassic volcanic–sedimentary units, and Triassic shallow-marine to non-marine sediments are well exposed in eastern Crete.

The Phyllite–Quartzite unit is, in turn, overlain by the relatively unmetamorphosed Tripolitza and Pindos units (Fig. 6). In Crete, the Tripolitza unit, in places begins with Middle–Upper Triassic dolomites, schists and shallow-marine carbonates ('Ravdoucha beds'; Kopp & Wernado 1983) and then passes into a thick shallow-water platform carbonate unit that culminates in Upper Eocene turbidites (Creutzburg *et al.* 1977; Fleury 1980). The Ravdoucha unit is well exposed in several areas in western Crete (e.g. the south coast at Sellia), where dark mudrocks and carbonates with occasional ammonites pass transitionally upwards into the Tripolitiza carbonate platform (Fassoulas 2001). A comparable but thicker unit in the Peloponnese, known as the Tyros unit, includes extensive Triassic volcanogenic rocks and terrigenous sediments (see below).

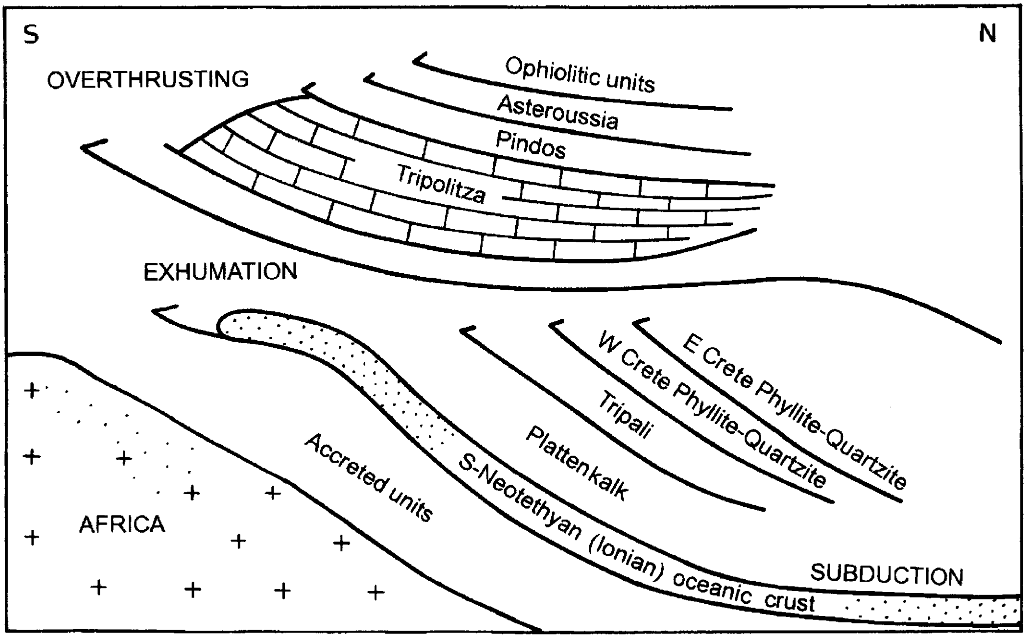


Fig. 6. Stacking order of tectonic units in Crete. (See text for data sources.) The lower thrust sheets experienced Cenozoic HP–LT metamorphism. The overlying Tripolitza platform was detached from its substratum related to Early Cenozoic northward subduction. The detached lower units were deformed and imbricated in the subduction zone, then exhumed.

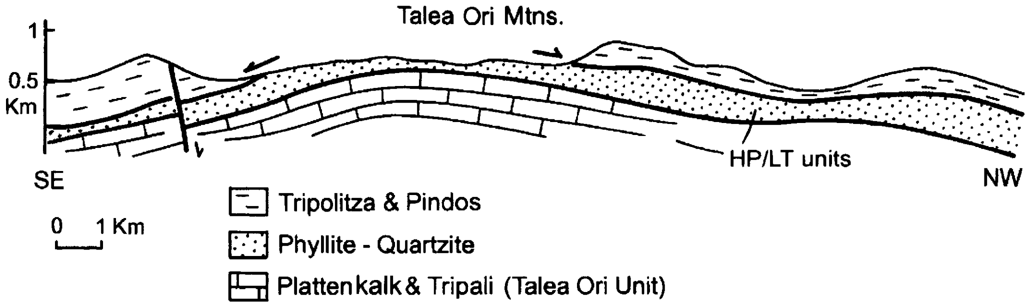


Fig. 7. Simplified cross-section of central northern Crete showing the main thrust sheets, or nappes. (See text for explanation.) Arrows indicate bivergent extensional exhumation. Modified from Fassoulas (2001).

The overlying Pindos unit includes Triassic deep-water sediments (e.g. pelagic limestones, radiolarites) and extrusive rocks (lavas, volcaniclastic rocks and tuffs), Jurassic deep-water sediments and ophiolitic rocks, and culminates in deposition of terrigenous turbidites of Paleocene–Eocene age (Krahl *et al.* 1982; Bonneau 1984). The Pindos and Tripolitza units are locally downfaulted in direct juxtaposition with the Phyllite–Quartzite unit, related to mid-Cenozoic tectonic exhumation, as seen in western Crete.

The Pindos unit is overlain by Upper Cretaceous metamorphic rocks, which are well exposed in central southern Crete (Asteroussia nappe), and finally by Upper Jurassic dismembered ophiolitic rocks (e.g. serpentinitized peridotite and gabbro) at the highest preserved levels of the thrust stack.

The palaeotectonic settings of the lowermost units, the Talea–Ori and Plattenkalk units are next outlined.

Talea Ori and Plattenkalk units

These units include Upper Palaeozoic–Lower Mesozoic continental margin-type sediments and deep-sea sediments that restore to the northern margin of the now largely subducted southern Neotethyan oceanic basin that rifted along the North African margin. In both Models 1 and 2 these two units would be expected to record divergence-related tectonic facies but in a different tectonic context. In Model 1 these units would represent part of the northern passive margin of Gondwana that rifted in Late Palaeozoic time, whereas in Model 2 they represent the southern margin of a rifted 'Cimmerian' fragment, which evolved as a subsiding passive margin during the Mesozoic–Early Cenozoic. The inferred latest Triassic 'Cimmerian' collision to the north could have affected this margin (i.e. by creating a distal foreland basin setting). In Model 3 an entirely Triassic (back-arc) rift would be expected, whereas a Triassic convergent margin might have existed in Model 4. Below, it will be shown that the sedimentary record favours Late Palaeozoic rifting with a further strong pulse of rift-related subsidence in latest Triassic–Early Jurassic time.

The older facies of the Plattenkalk, represented by the much-discussed Talea Ori unit (e.g. Hall & Audley-Charles 1983; Hall *et al.* 1984), is well exposed in the Talea Ori Mountains of central northern Crete, west of Iraklion (Fig. 8) and in the Psiloritis Mountains further south (Fig. 5). The succession exposed in the type area (Fig. 8a) is structurally inverted. The Talea Ori unit is reported to begin with Upper Palaeozoic clastic sediments (Krahl *et al.* 1988) that contain detrital zircons as young as 297–325 ± 5 Ma; i.e. of inferred Hercynian age (Zulauf *et al.* 2002; Brix *et al.* 2002; Romano *et al.* 2002, 2004). The section stratigraphically above comprises phyllites and sandstones with minor cherty horizons (Galinos beds), of inferred Early Permian age. These sediments are then overlain by neritic dolomites and limestones (i.e. Fodele beds), dated as Late Permian. The succession continues with mainly clastic dolomites, limestones and interbedded clastic sediments (Sisses beds), passing into Upper Scythian marbles (Epting *et al.* 1972). The Middle-Triassic interval exhibits a depositional hiatus (Epting *et al.* 1972) marked by subaerial exposure, karst development, diachronous erosion and local bauxite development. This was followed by marine transgression with deposition of Norian stromatolitic dolomites and massive dolomitic carbonates. Carbonate breccias then record a further discontinuity. Champod *et al.* (2004) reported that Upper Triassic facies are transgressive on various different underlying units, down to the level of Late Permian platy pelagic limestones with chert nodules. In most interpretations the Upper Triassic facies are inferred to pass depositionally upwards into the typical deeper water Plattenkalk carbonates. However, Krahl & Kauffman (2004) considered that these pelagic carbonates are not part of the Plattenkalk unit *sensu stricto* but rather representative of a pelagic facies that occurs in several different units. They now consider the Talea Ori unit, of Late Carboniferous–late Early Cretaceous age, to be a separate tectonic unit from the Plattenkalk unit although this is not supported here.

During this work detailed sedimentological observations were made, particularly on the Upper Palaeozoic Fodele and Sisses units and the overlying Triassic carbonate succession that shed light on the depositional and tectonic setting. The Permian Fodele beds (e.g. 1 km east of Sisses) are dominated by regularly dipping, little deformed alternations of darker and lighter coloured, locally pebbly limestones with numerous shelly and bioclastic layers (Fig. 8). Highly fossiliferous dark limestones include well-preserved large brachiopods, coral and gastropods. The bioclastic material is suggestive of a storm-influenced death assemblage.

In the succession studied, north of Fodele village, the base of the succession is in tectonic contact with phyllites of the Phyllite–Quartzite unit. There, the lowest exposed beds of the Sisses unit are very coarsely crystalline marbles intercalated with buff phyllites. These phyllites are lithologically similar to the underlying Phyllite–Quartzite unit so that the exact location of the thrust contact is indeterminate. A similar alternating carbonate–clastic succession is well exposed further east (0.5 km), along the national road, where thick-bedded shales and limestones are interbedded, although the contacts are commonly sheared. A major carbonate conglomerate (*c.* 4.5 m thick) contains well-rounded clasts, up to 3 cm in size, set in a buff-coloured matrix of phyllite. Pebbly phyllitic interbeds contain well-rounded clasts (up to 5 cm in size). These matrix-supported conglomerates form laterally continuous units, up to 6 m thick. Close to the inferred tectonic contact with the Phyllite–Quartzite unit small thrust slices of carbonate rocks and shale include small slices of dark bioclastic and nodular limestone, correlated with the Fodele unit. The matrix-supported conglomerates are suggestive of deposition in a slope setting. Stratigraphically above, the Sisses unit is dominated by neritic carbonates (Fig. 8). The higher levels of the Sisses unit are uniformly fine-grained, thick-bedded to massive carbonates, ranging in colour from white, grey and buff to pink. Massive beds, commonly rich in vugs, appear to include boundstones. Locally, thin beds (*c.* 10 cm thick) exhibit sand-sized clastic interbeds (up to 10 cm thick), showing grading and fine parallel lamination. There are also several thin micro-breccias dominated by small platy

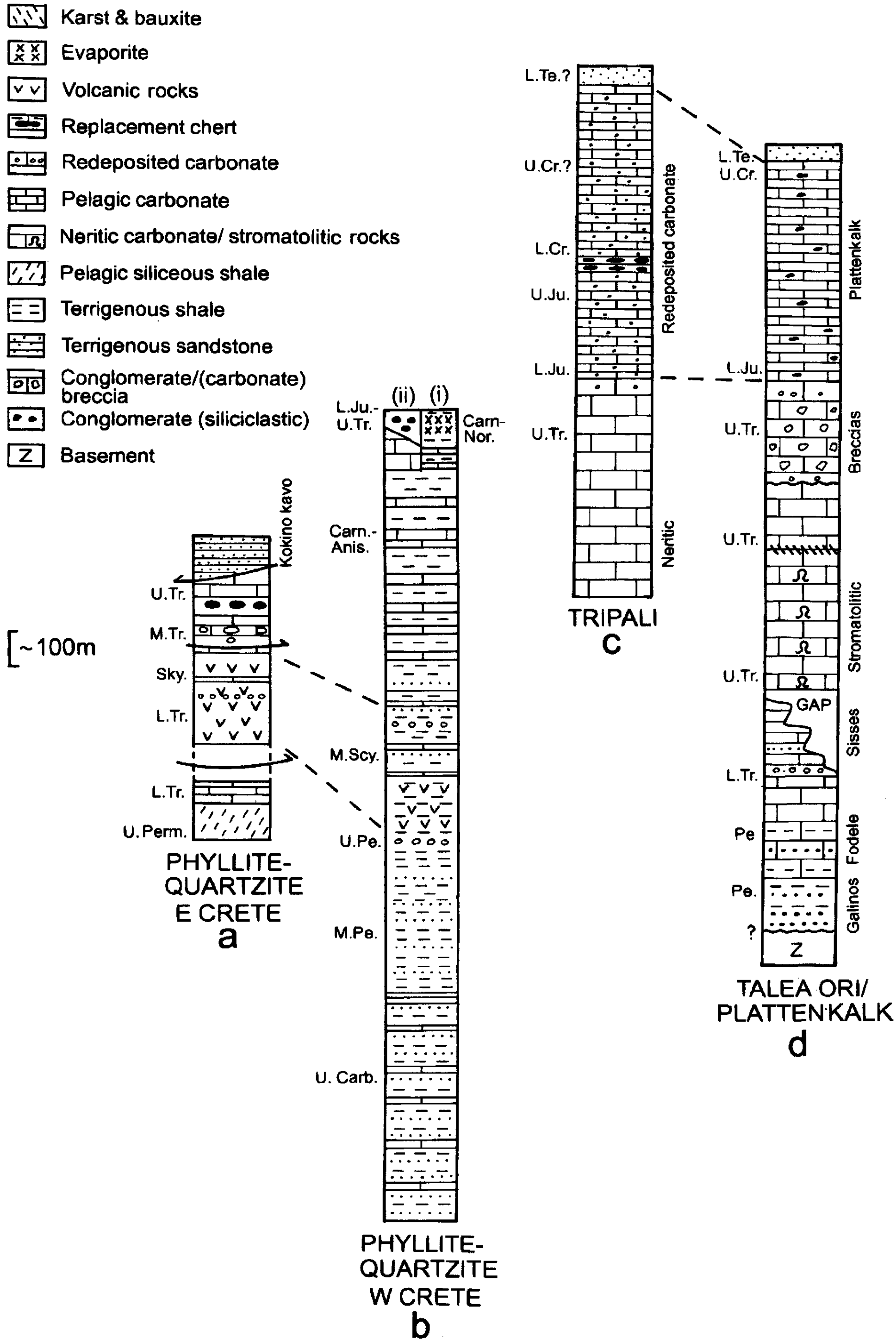


Fig. 8. Summary of the main restored Upper Palaeozoic–Lower Cenozoic successions exposed in the lowest thrust sheets on Crete that experienced HP–LT metamorphism and are discussed in detail in this paper. (See text for discussion and data sources.)

carbonate clasts (<1 cm long) of probable microbial origin (i.e. reworked microbial mats).

The Sisses unit is overlain by very well exposed algal limestones, which are mainly medium to thick bedded, but include thin (several millimetres) lenticular siltstones, suggestive of current reworking. Above the hardground and bauxitic horizons (Epting *et al.* 1972; Fassoulas *et al.* 2004), an upper limestone–dolomite succession is dominated by dark foetid interlayers that increase in abundance upwards and culminate in very dark organic-rich carbonate. Dark stromatolitic carbonates and lighter vuggy carbonates are interstratified as alternating units *c.* 5–8 m thick. Massive vuggy white limestones follow with occasional pebbly horizons containing grey to white finely laminated carbonate clasts, up to 4 cm in size. Above this (near Aloides village), the succession is dominated by a spectacular high-density carbonate turbidites, slumps, carbonate debris flows and coarse carbonate talus. Individual mega-breccia units reach 5 m thickness and include a range of very angular to well-rounded clasts, some > 1 m in size. Several depositional units (up to 2.5 m thick) comprise very well exposed carbonate debris flows grading into high-density type calciturbidites, showing evidence of dewatering, slumping and soft-sediment deformation. Most of the clasts are of light-coloured carbonate facies but a few are dark, similar to the underlying intact carbonate succession. The highest levels of the redeposited facies comprise thick-bedded limestones with limestone conglomerates forming several metre-thick units with well-rounded limestone. Some beds are graded confirming stratigraphic inversion. The typical pelagic Plattenkalk facies then follows without a break (Fig. 8). The lowest of the Plattenkalk sediments are very thick-bedded (up to 0.8 m thick), pale grey crystalline limestones with numerous white chert lenticles and nodules.

Interpretation: a rift setting

The lower part of the succession is consistent with a Permian transgression onto a Late Palaeozoic basement and input of clastic sediment in a tectonically unstable shallow-marine setting. This was followed by a marked regression, probably related to tectonic uplift during Early–Mid-Triassic time, which gave rise to the Mid-Triassic depositional hiatus. The thick organic-rich stromatolitic carbonates accumulated on a subsiding carbonate platform. This was followed, during the Late Triassic, by catastrophic collapse of the carbonate platform leading to the genesis of slumps, mega-breccias, carbonate debris flows and calciturbidites. The deep-water Plattenkalk facies then began to accumulate as calciturbidites and relatively siliceous pelagic carbonates. In Model 1 the above succession records rifting of the North African margin, including a phase of flexural uplift in the Mid-Triassic, followed by a further major rift pulse in the Late Triassic. In Model 2, the succession (Galinos and Fodele units) records extension and subsidence related to opening of a Neotethyan ocean to the south. This was followed by flexural uplift, and crucially collision and post-collisional marine transgression during Mid–Late Triassic time related to and following closure of Palaeotethys (Champod *et al.* 2004).

Here, the sedimentary record of the Talea Ori unit is considered to be consistent with Model 1 (or Model 3). The strong relative subsidence in the Late Triassic is attributed to a pulse of rifting to create a deep-water (Ionian) basin to the south. This pervasive rifting, fault scarp degradation, platform break-up and subsidence from a neritic to a carbonate-depositing setting during the Late Triassic is fully consistent with Model 1. However, these features cannot be explained as a flexural response to a Palaeotethyan 'Cimmerian' collision, as a contrasting flysch-type basin would be expected above the carbonate platform. Also, there is no evidence of the required foreland basin within the Plattenkalk, which restores to a more northerly position, as discussed below.

Tripali unit

The regionally overlying Tripali unit (Fig. 8) is dominated by platform carbonates and is exposed in many mountainous areas (e.g. Levka Ori and Tripali Ori). The Tripali unit was traditionally seen as being restricted to very thick (> 1000 m) successions of Upper Triassic–Lower Jurassic, partly dolomitized platform carbonates (e.g. Jacobshagen 1986). However, Krahl & Kaufmann (2004) reported an upward passage into siliceous metacarbonates (i.e. Plattenkalk-type facies) and metasiliciclastic sediments, extending in age from Early Jurassic (based on ammonites), to at least Albian (based on planktonic Foraminifera). Cenomanian planktonic Foraminifera and rudists occur in debris related to the emplacement of this unit over the Plattenkalk. This suggests that the original succession spanned the Late Triassic to Late Cretaceous–Paleocene.

The higher, post-Triassic, part of the Tripali succession, for example, is well exposed along, and adjacent to, the main north–south road linking Vrisses with Hora Sfakion (Figs 5, 6 and 8). The succession structurally overlies the Plattenkalk in the north. The lowest part of the local succession exposed further south (near the turning to Asfendou) comprises massive to weakly bedded, pale grey crystalline neritic carbonates, of inferred Late Jurassic age (J. Krahl, pers. comm.). Upwards, there is an incoming of medium-grey to dark grey limestone with abundant chert nodules concentrated in finer-grained intervals. The limestones include thick-bedded matrix-supported bioclastic carbonate conglomerates, with clasts up to 5 cm in size, bivalve debris and belemnites. There are also several finely laminated, soft-weathering, shaly intervals. Upwards, the succession becomes increasingly thinly bedded and chert rich, and then passes into *c.* 20 m of brownish shale with chert-rich beds (up to 0.3 m thick), formed by almost complete replacement of thin carbonate beds. Above, there is a return to thicker bedded, purer limestone with abundant chert nodules. The highest levels of the succession, dipping southwards at a moderate angle towards the south coast, comprise medium- to thick-bedded (up to 2.9 m) dark, foetid, marble with abundant cannonball-shaped chert nodules. Individual beds (up to 0.3 m thick) are graded, with scoured bases and shaly tops. Higher parts of the succession, exposed further north (on the main road north of Kares near the col), are dated by a rare occurrence of Upper Cretaceous rudist limestone (J. Krahl, pers. com.). Thick-bedded coarse bioclastic limestones comprise subangular to subrounded carbonate clasts (up to 10 cm in size), including algal carbonate, and thin interbeds of pale grey calcilutite.

Interpretation: a subsiding rift basin

The Upper Triassic–Lower Jurassic limestones of the Tripali unit record a widespread neritic carbonate platform facies. By contrast, the deposition during Jurassic–Late Cretaceous time took place in deep water in a relatively proximal carbonate slope setting. Facies ranged from carbonate debris flows, to high- and low-density turbidity current deposits and siliceous hemipelagic carbonates. The distinctive shaly interval rich in replacement chert may correspond the regional Late Jurassic–Early Cretaceous interval of high siliceous productivity and chert formation (e.g. De Wever 1989). The Tripali unit can be restored as a distal equivalent of the Plattenkalk, hence its similarity in facies. The source of the redeposited carbonate material is inferred to be the Tripolitza carbonate platform, then located to the north.

In Model 1, the Tripali unit records Late Triassic rifting to create a deep-water basin to the south in which deep-water slope and pelagic carbonates accumulated. The Tripali unit can be correlated with the margin of the deep-water Ionian rift basin in the Peloponnese, whereas the Plattenkalk can be correlated with the more distal deep-water pelagic carbonate facies of the Ionian basin which overlies stretched continental crust in the Peloponnese (e.g. British Petroleum Company Ltd 1971). This contrasts with Model 2, in which the Tripali and Plattenkalk units would

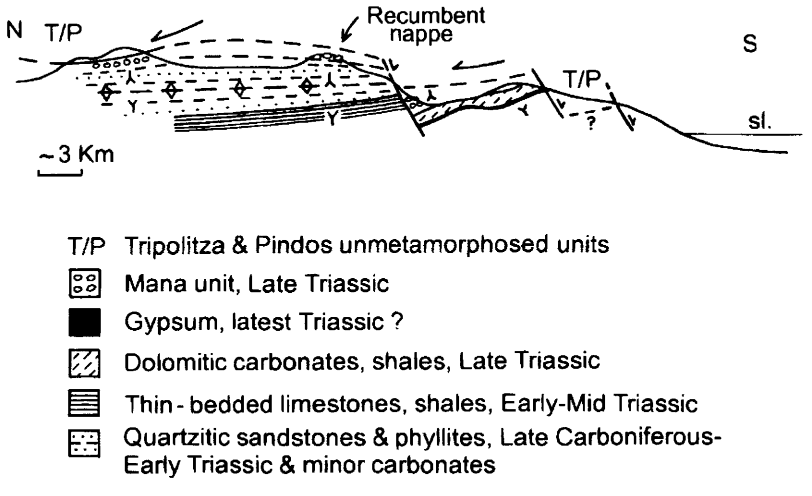


Fig. 9. Simplified cross-section of western Crete showing the division of the Phyllite–Quartzite unit into an upper right-way up succession and a lower inverted succession. This distribution has been explained by the existence of a huge south-facing recumbent nappe (Krahl *et al.* 1983*a*, *b*, *c*). The structure was modified by high-angle faulting. (See text for discussion.)

be expected to overlie south-Neotethyan oceanic crust, for which there is no evidence in either Crete or the Peloponnese.

Phyllite–Quartzite unit of western Crete

Lithofacies of the Phyllite–Quartzite unit as exposed in western and eastern Crete are substantially different and so will be discussed separately below. In general, the Phyllite–Quartzite unit of western Crete comprises a Middle Carboniferous to Lower Triassic deep-water siliciclastic sequence, with alkaline igneous rocks of mainly Early Triassic age. The overall succession shallowed upwards into an Upper Triassic, evaporitic sequence and an Upper Triassic–lowemost Jurassic? conglomeratic sequence in different areas.

In Model 1 (Fig. 2) the siliciclastic sequence represents a divergence-related tectonic setting, namely a Late Palaeozoic deep-water rift. A further pulse of rifting later gave rise to alkaline magmatism in this interpretation. The observed shallowing upwards is interpreted as the effect of flexural uplift related to break-up to form the Pindos ocean to the north. This inferred flexural uplift also affected the Plattenkalk (Talea Ori unit), as noted earlier. In Model 2 the Late Palaeozoic succession would record rifting of a Cimmerian fragment from Gondwana coupled with opening of Neotethys, whereas the Triassic shallowing-upward succession would record a foreland basin or collisional setting. In Model 3, the Triassic volcanic rocks would be expected to show a subduction influence, although the successions might be located too far south from the trench to show such an effect. In Model 4 there would be a greater chance that the volcanic rocks would be subduction related as they should directly overlie a northward-dipping subduction zone.

The key discriminants are again the time and length scales of the inferred tectonic controls on the sedimentary sequences: how proximal or distal are the successions? Is there evidence of contractional deformation coeval with sedimentation, and what is the character and consequent explanation of coeval magmatism? It will be argued that the sequences are consistent with deposition in a rift setting, which then shallowed, rather than with a developing passive margin bordering an already extant oceanic basin or with any form of convergent setting.

One lithological assemblage of the Phyllite–Quartzite unit is widely exposed in western Crete and is now relatively well dated following careful biostratigraphical work, mainly using conodonts, benthic Foraminifera and ostracodes (Krahl *et al.* 1983*a*,*b*,*c*). The dating evidence suggests that both an inverted and a right-way-up succession are present (Fig. 8), which have been interpreted as the lower and upper limbs, respectively, of a regional-scale south-facing recumbent nappe (Fig. 9; Krahl *et al.* 1983*a*,*b*,*c*). The Late Palaeozoic successions are similar in both of these sections, but the facies differed markedly during the Triassic, as outlined below; assuming the large-scale structure is correctly interpreted, this has important implications for sedimentary polarity, when restored.

Carboniferous–Lower Triassic succession

In general, the lower part of the succession (*c.* 600–800 m thick; Fig. 10b) is dominated by siliciclastic sandstones (quartzites), interbedded with shales (phyllites) and subordinate thin-bedded limestones (marbles), ranging in age from Mid–Late Carboniferous to Early Triassic. The thickest-bedded and coarsest-grained siliciclastic sediments are inferred to be of Mid-Permian age (Krahl *et al.* 1983*c*). Several thin conglomeratic intervals, here interpreted as debris flows, occur especially in the Lower and Upper Permian intervals (J. Krahl, pers. com.). The Lower Triassic interval (starting at the Middle–Upper Scythian boundary) exhibits abundant evidence of gravity redeposition of variably consolidated sediment, derived from both deep-water and shallow-water settings, of both Permian and Early Triassic age.

The oldest known part of the succession, of Mid–Late Carboniferous age (Krahl *et al.* 1983*c*), is, for example, well exposed in the NW (e.g. near Sfinari; Fig. 10a), where it dips northwards at a moderate angle. The succession between Sfinari and Kambos, estimated as *c.* 700 m thick, dips generally northwards, with an average dip of 10–15°, and, in places, is strongly sheared, isoclinally folded and faulted. Southwards, a continuous (inverted) succession exposes facies of Early, Mid-? and Late Permian, and then Early Triassic (Scythian) age (Krahl *et al.* 1983*b*; Fig. 10b).

Sedimentary facies

The Middle–Upper Carboniferous part of the succession (near Sfinari, Fig. 10a) comprises medium- to thick-bedded, grey to black, quartzitic meta-sandstones, with dark pelitic intercalations and thin-bedded dark metacarbonates. Individual packets of medium-bedded sandstones, commonly several metres thick, are commonly transitional upwards, over several metres, to grey mudstones, interbedded with subequal thicknesses of relatively fine-grained quartz- and muscovite-rich sandstones (in beds up to 0.25 m thick). These sandstone fade out over several metres and pass into silver–grey mudrocks

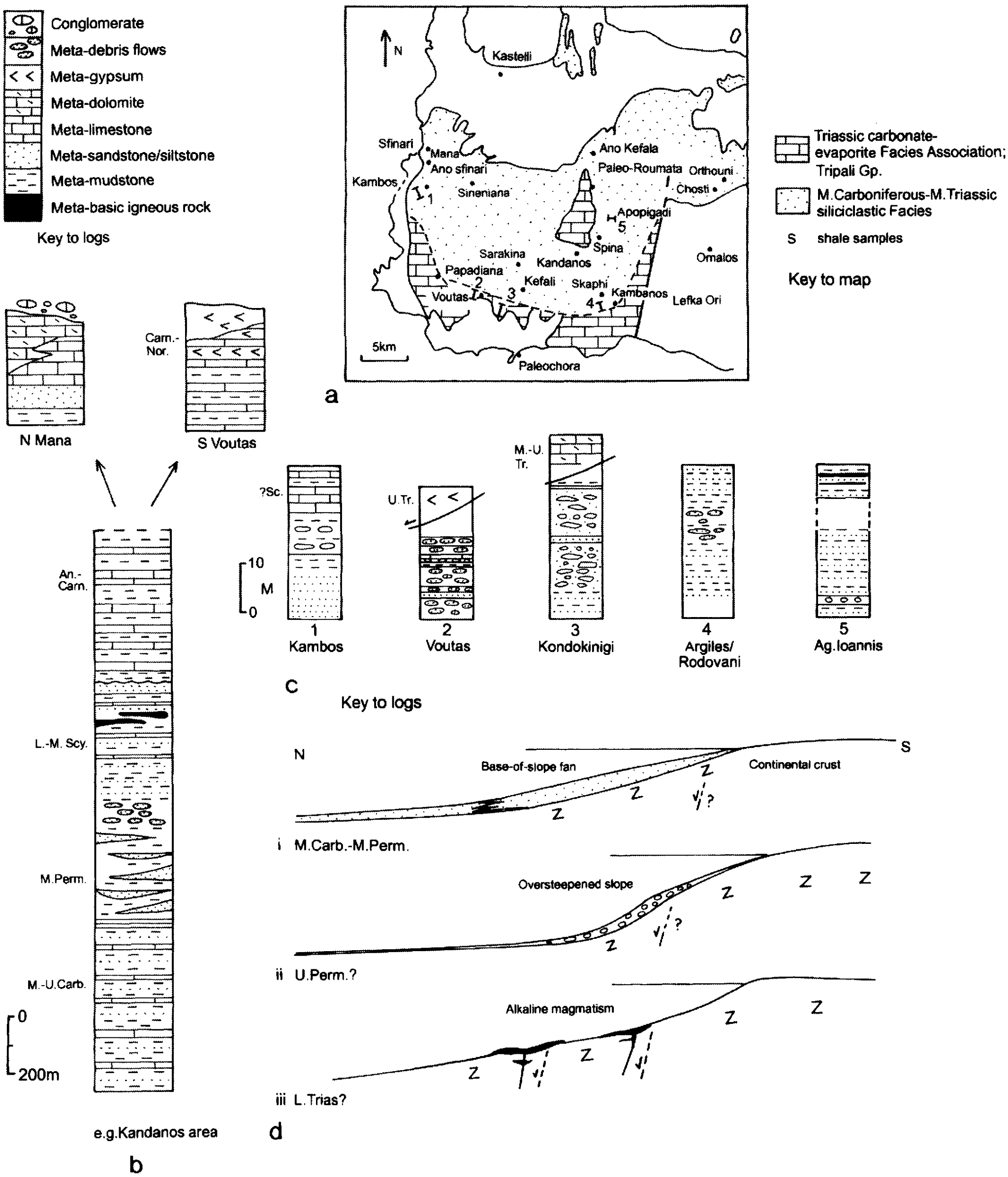


Fig. 10. Late Palaeozoic–Early Mesozoic evolution of western Crete. (**a**) Outline map; (**b**) composite sedimentary log; (**c**) measured logs of matrix-supported conglomerates of late Early–Mid-Triassic age; (**d**) possible interpretation of the units of Late Palaeozoic–Early Mesozoic age. (See text for discussion.)

(phyllites), with only a few thin lenses of fine-grained sandstone (up to 15 cm thick). Occasional carbonate-cemented lenses and concretions are less compacted and retain traces of cross-lamination. A less well-exposed, argillaceous succession further south (disrupted by landslipping) includes several isolated sandstone lenses, individually 3–5 m thick. Further south (near Ano Sfinari) there is a return to thick-bedded sandstones (in beds up to 1 m thick), forming packets up to 20–40 m thick, alternating with finer-grained facies. The beds within individual packets tend to thicken stratigraphically upwards, assuming the succession is inverted.

The Upper Palaeozoic–Lower Triassic succession is cyclical, with alternations of thicker- and

Table 1. *Selected chemical analyses of metashales (schists) from western Crete and of Triassic basaltic igneous rocks from the Phyllite–Quartzite units of western and eastern Crete*

Sample: Rock type:	93/1 shale	93/2 shale	93/3 shale	CRO1 W Crete basalt	CR04 W Crete basalt	CRPS5 W Crete basalt	Vai 04/19 Vai basalt	Vai 04/20 Vai basalt	Vai 04/14 Vai block	Ch/04/24 Vai, W coast	Ch/04/31 Chemezi S
SiO_2	59.22	63.39	57.49	34.62	44.26	44.74	52.25	55.21	53.19	53.75	45.4
Al_2O_3	23.26	19.36	23.32	14.56	14.87	10.65	17.66	18.6	18.83	18.65	19.64
Fe_2O_3	8.65	8.74	9.1	16	16.45	14.54	7.41	7.68	7.87	8.1	10.87
MgO	0.91	1.28	1.14	9.86	12.03	15.73	5.17	5.79	4.93	6.1	10.1
CaO	0.06	0.06	0.05	5.33	4.12	9.35	4.48	1.74	3.43	0.59	1.9
Na_2O	1.89	1.74	1.76	4.03	3.36	1.84	3.91	5.83	6.63	7.57	3.63
K_2O	1.62	1.28	1.99	0.09	0.74	0.54	1.69	0.8	0.11	0.16	0.28
TiO_2	1.14	1.01	1.19	43.52	3.1	1.93	0.68	0.72	0.71	0.72	0.82
MnO	0.05	0.46	0.04	0.25	0.2	0.19	0.68	0.14	0.14	0.17	0.18
P_2O_5	0.06	0.44	0.56	0.75	0.87	0.5	0.12	0.32	0.11	0.11	0.08
LOI							6.07	3.69	3.69	3.36	6.18
Total	96.87	96.95	96.15	100	100	100	99.56	100.32	99.64	99.37	99.11
Zn	100	124	125	206	227	156	85		106	94	100
Cu	15	19	19	133	24	60	7	11	7	7	40
Ni	41	58	61	184	415	515	35	22	27	61	32
Cr	116	97	121	269	508	636	150	115	31	373	105
V	301	287	451	311	226	167	182	150	39	233	232
Ba	40	42	22	31	228	262	288	211	132	142	28
Sc	48	44	46	20	25	20	33	24	7	37	36
Nb	52	45	50	83	97	55	3.5	4	3.5	3	2
Zr	329	340	321	301	336	211	71	72	74	63	44
Y	39	34	25	35	27	20	15	14	14	13	12
Sr	110	101	103	339	138	374	162	83	154	106	43
Rb	91	72	116	1.2	21	14	33	16	1	3	5
Th	18	15	20								
Pb	15	19	10								
La	61	48	62	43	54	36	8	8	8	8	3
Ce	129	103	134	95	114	63	15	16	18.9	15	7
Nd	56	46	67	47	54	30	9	10	9.1	9	6

The analysis was carried out as specified by Fitton *et al.* (1998). Locations of samples: 93/1, 93/2, 93/3, shale from Ayios Zinas, NE of Kandanos; CRR01, CR04, CRPS5, metabasic rocks from Orthouni, Skaphi and Chosti; Vai 04/19 and Vai 04/20, metabasalt from near the base of thrust sheet 1, SW of Vai beach; Vai 04/14, metabasalt block in mélange unit, south of Vai beach; Ch/04/24, metabasalt from well-bedded sequence, north of fish farm, west coast of Vai peninsula; Ch/04/31, metabasalt from near the base of the exposed sequence, south of Paleokambos, 'Chemezi area'. LOI, loss on ignition.

thinner-bedded sediment packets, each around 100 m thick. Despite isoclinal folding, outcrop patterns suggest that the alternating packets are laterally lenticular, on scales of several hundred metres. The thin-bedded packages are dominated by medium-bedded sandstones, up to 25–30 m thick, alternating with thinner-bedded shales, limestones and sandstones. By contrast, the thicker-bedded packets comprise massive sandstones, up to 5 m thick, with subordinate thinner-bedded sandstones, shales and rare limestones. The thickest and most homogeneous Upper Permian? metasandstones (>5 m thick) appear to be the most laterally persistent along strike. Most of the thick-bedded sandstone appears to be massive, possibly reflecting partial recrystallization during Alpine high-pressure metamorphism. However, the bases of some of the thinner-bedded sandstones, interbedded with dark shales, exhibit sharp bases and traces of grading. The thickest-bedded sandstones are locally pebbly, with well-rounded quartzitic pebbles (up to 1 cm in size), and are invariably poorly sorted.

In thin section, medium-coarse grained sandstones (e.g. near Sfinari) exhibit moderately to well-rounded quartz grains set in a matrix of finer grained quartzose sandstones and siltstones. Other sandstones are dominated by subangular grains of quartz and include subordinate lithoclasts of muscovite schist and quartzite. Minor plagioclase is commonly recrystallized. The matrix includes muscovite, probably entirely recrystallized, and scattered heavy minerals (e.g epidote, amphibole and zircon). Higher in the succession (e.g. near Kambos), contrasting calcareous interbeds are composed of sandy limestone with scattered, mainly subrounded grains of quartz within a matrix of recrystallized coarse-grained carbonate. Thinner, dark coloured, sandstone interbeds are well graded with abundant pyrite and organic-rich material within a fine-grained matrix. Overall, the sandstones are mainly quartzarenites and sublitharenites, showing ubiquitous evidence of textural inversion.

Sediment chemistry

To help determine sediment provenance seven samples of dark metashales were collected from the stratigraphically higher part of the succession (of Late Permian?–Early Triassic age) located on a mountainous ridge (Ayias Zinas, NE of Kandanos; Fig. 10a) and were analysed for major- and trace elements by X-ray fluorescence (XRF; see Table 1 for representative analyses). When normalized against the composition of average North American shale (Gromet *et al.* 1984), the samples are compositionally similar to average shale, although relatively enriched in several elements (e.g. La, Ce and Nd), but depleted in Sc, Zn and Ba (Fig. 11). The Ba depletion could relate to mobility of this element during Alpine metamorphism. The shales are compositionally similar to the average shale derived from the north margin of Gondwana in northern Syria (Al-Riyami & Robertson 2002). When plotted on several tectonic discrimination diagrams (Th v. Sc v. Zr/1000; K_2O/Na_2O v. SiO_2; TiO_2 v. Fe_2O_3 + MgO) the samples plot in the oceanic island arc or active margin fields. However, the inferred igneous component is unlikely to relate to contemporaneous volcanogenic input, as sediments are interbedded with siliciclastic sandstones and lack volcanogenic material. In agreement with Romano *et al.* (2006; see also Bojar *et al.* 2002), the petrographic and geochemical evidence suggests that the terrigenous sediments were derived from the North African craton, or possibly from a rifted continental fragment of the same crustal composition.

Magmatic intercalations

Alkaline magmatic rocks, metamorphosed under HP–LT conditions, are locally present, mainly in the higher levels of the siliciclastic–shale succession (Seidel 1978). Small volumes of meta-alkaline volcanic rocks, volcaniclastic sediments and tuffs are interbedded with deep-sea sediments of Early Triassic (i.e. Early Scythian) age (Krahl *et al.* 1983*c*). Using a modern time scale (e.g. Gradstein *et al.* 2004), 'Late Permian' ages, radiometrically determined by Seidel (1978) are reassigned to the Early Triassic, although volcanism may have begun in the Late Permian.

Prominent exposures of basic igneous rocks mainly occur in the higher levels of the siliciclastic-dominated succession. Metabasic igneous rocks were studied from well-exposed ridges to the NE of Kandanos (e.g. Palea Roumata; Spina; Ano Kefala; see Seidel 1978). These form lenticular, competent bodies, ranging from several metres to *c.* 15 m thick, usually traceable laterally up to several hundred metres. The most common lithologies show no obvious extrusive features and are likely to have originated as sills. For example, on a high ridge (near Ayios Ionnis Apopigadi), medium- to thick bedded quartzites and grey phyllites are interbedded with metabasic rocks, in layers up to 3–5 m thick. Further north, at Palea Romata, bluffs of very hard, very resistant metabasites, *c.* 10 m thick, are intercalated with isoclinally folded dark phyllites. Meta-igneous lenses, 10–12 m thick, are traceable laterally for hundreds of metres along the hillside. Similar igneous bodies (5–8 m thick) were also studied further north (at Ano Kefala), within similar NE-east dipping sandstones and phyllites. Resistant bands, up to 10–20 m thick, are also

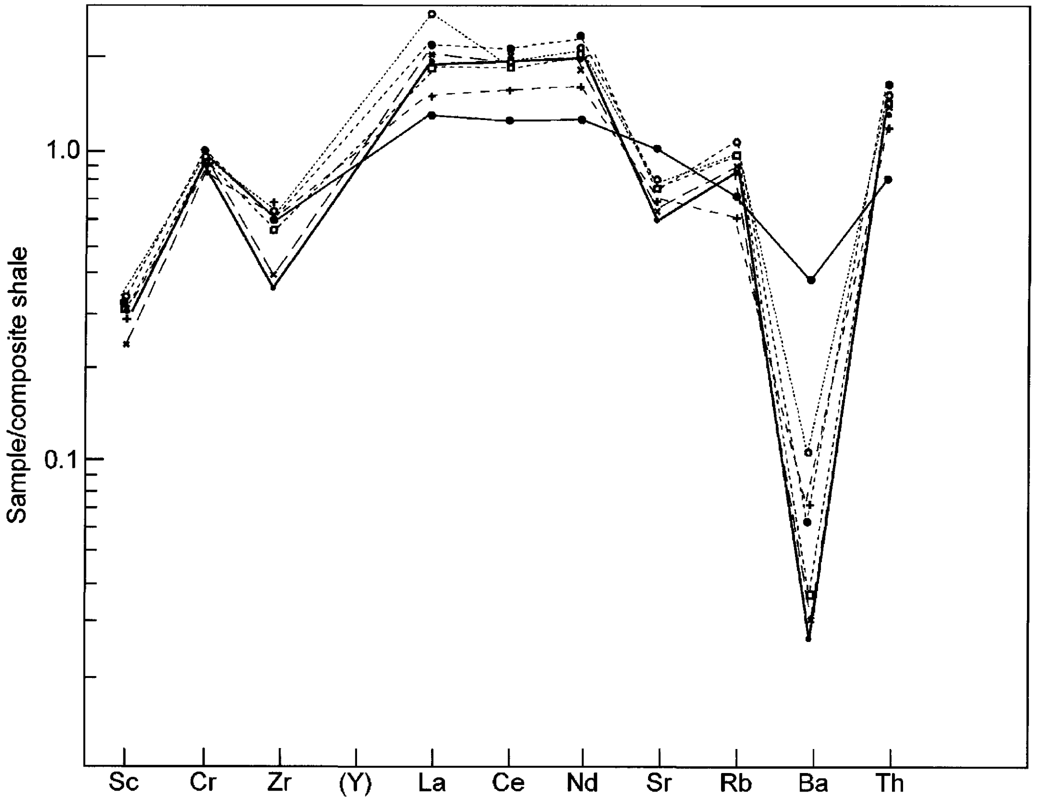


Fig. 11. Sedimentary geochemistry of shales from the Upper Palaeozoic–Lower Triassic interval of the Phyllite–Quartzite unit in western Crete. The results suggest derivation from a continental basement, probably North Africa.

traceable across the hillside, 2 km south of Kandanos (above Anisaraki), within mainly fine-grained metasedimentary rocks.

Meta-basic rocks (17 samples) from several localities in the NW of the area (e.g. Skaphi, Orthouni and Chosti) were analysed for major and trace elements by XRF (see Table 1). When plotted on MORB-normalized 'spider diagrams' (e.g. Pearce 1980), the amphibolites are typical of alkaline rift-related basalts and more fractionated alkaline igneous rocks (Fig. 12), in agreement with previous studies (Seidel 1978). In addition, a further 18 samples of metabasalts were analysed from a small exposure of the Phyllite–Quartzite unit on the hillside above Amoundi Beach on the south coast, SE of Kato Preveli monastery (Fig. 5) and these showed very similar 'enriched' patterns (unpublished data).

Triassic inverted succession

Depositional transitions from the typical siliciclastic facies to the Middle–Upper Triassic more carbonate-rich facies are well exposed in the NW (Kambos area) and in an inlier in the SW (Kandanos area; Fig. 10a). In general, a succession of mainly shales and platy limestones of Early–Mid-Triassic age passes into mainly dolomitic carbonates and shales of Late Triassic age, culminating in shales and dolomitic carbonates with gypsum, of Carnian–Liassic? age. An inferred hiatus during the Late Scythian–Anisian time interval may correlate with the prominent hiatus in the Talea Ori unit (Epting *et al.* 1972; Krahl *et al.* 1983a).

In the NW, near Kambos, alternating sandstones and shales of the Phyllite–Quartzite unit (of Late Permian–Scythian age) pass stratigraphically upwards into dominantly thick-bedded marble, associated with an interval of carbonate conglomerates (*c.* 10 m) comprising clasts of marble (up to 15 cm in size), some of which are well rounded (near the Mid–Late Scythian boundary; Fig. 10c). Individual depositional units, up to several metres thick, are dominated by marble clasts (up to 10 cm in size)

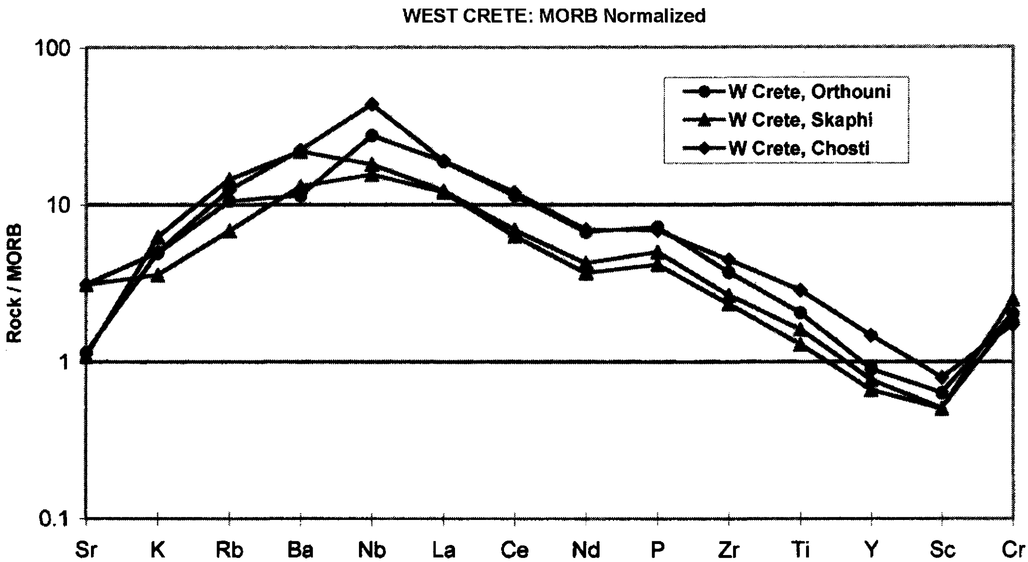


Fig. 12. MORB-normalized 'spider diagrams' of selected metabasic rocks from the Lower Triassic? interval of the Phyllite–Quartzite unit in western Crete. All the samples show 'enriched' trends and plot within a relatively narrow compositional range. (See text for explanation and Fig. 10a for locations.)

that are locally very well rounded. These conglomerates are interbedded with medium- to thick-bedded metalimestones (up to 3 m thick), purple–red shale (phyllite) and several thin interbeds of white sericitic-rich shale (up to 0.12 m thick), possibly tuff. The limestones include reworked ostracodes (Krahl *et al.* 1983*c*). Some beds contain numerous elongate sandstone or siltstone rip-up clasts (up to 0.4 m long), set in a poorly sorted quartz-rich matrix. Intraformational clasts of dark grey phyllite are also present within poorly sorted 'gritty' quartzose metasandstone. Long thin clasts were disrupted while still poorly lithified. Some reworked quartz grains are relatively large and well rounded. Thicker bedded, almost clast-supported conglomerates, occur higher in the succession (in beds up to 2.3 m thick) and contain subrounded limestone clasts (up to 0.6 m long), in which individual clasts range from pure to muddy carbonate. Several of these limestone conglomerates exhibit scoured bases and sharp tops, confirming that this succession is inverted. The succession then passes stratigraphically into well-bedded platy limestones and shales of inferred Mid-Triassic (Anisian–Ladinan) age. A comparable exposure in the Kandanos area extends into Late Triassic dolomitic carbonates and evaporitic facies.

In addition, the inverted Triassic succession is well exposed in an arcuate belt further south, an area strongly affected by neotectonic east–west faulting. However, Krahl (1983*a*,*b*,*c*) has reported a number of locally intact Triassic successions. The Triassic succession is, for example, locally exposed on the footwall (commonly landslipped) of a prominent east–west neotectonic fault zone. Just north of Voutas, typical thick-bedded quartzite (in beds up to 1.5 m thick), of inferred Early Triassic age, is followed northwards (after a short gap in exposure) by thin- to medium-bedded platy granular carbonates with interbedded dark organic-rich phyllites, of inferred Mid-Triassic (Anisian) age. Higher parts of the succession, of inferred Late Triassic age are well exposed in a hilly area to the south (e.g. SW of Voutas). This succession, 100–200 m thick, mainly dips eastwards at moderate angles (*c.* 38°) and is dominated by dark grey dolomitic carbonates, interbedded with thin- to medium- and thick-bedded carbonate-rich shales. The section is locally deformed into south-facing outcrop-scale folds and small duplexes. Individual dolomitic beds, up to 0.6 m thick, are composed of buff sugary carbonate, full of small vugs, apparently created by evaporite dissolution. Thinner-bedded, dark interbeds are finely crystalline, locally with small intraclasts of gypsiferous marl. Fissile intercalations are dark and organic rich. Elsewhere, Triassic evaporites have locally been mobilized to form lenticular masses of ductile-deformed gypsum, tens to several hundred metres thick.

Triassic right-way up succession: Mana unit

A regional right-way up succession, best exposed in the NW (e.g. NE of Sfinari, near Mana and Sineniana; Fig. 10a), exhibits a relatively deep-water setting that persisted during Mid-Triassic time, with Radiolaria and pelagic conodonts, but then shallowed upwards during Late Triassic time (Norian), allowing the deposition of shallow-marine limestones and dolomites (without evaporites). The succession is capped by the Mana unit, which is composed of marbles (Mana marble) of inferred Late Triassic–Early Jurassic? age, followed by undated conglomerates (Mana conglomerate) (Krahl *et al.* 1983*c*).

The Mana unit is critical to the interpretation of the later Triassic palaeoenvironments. For this reason three sections were studied, one in the NW, one further south and one in the far south, to provide a regional overview. In Model 1 this unit would be expected to relate to a rift setting, whereas in Model 2 it should record a foreland basin or post-collisional 'molasse'-type setting.

In the NW, near Sineniana (Fig. 10a), a composite succession was pieced together in several adjacent fault blocks. The exposed succession above the valley floor (*c.* 1 km south of Sineniana; near Tsortsiana chapel; Fig. 13a, locality A) begins with thick-bedded, massive quartzitic sandstones (in beds up to 0.6 m thick) with subordinate mud-rock (dark phyllite). These sediments correlate with the stratigraphically higher parts of the Phyllite–Quartzite succession in the adjacent area. The quartzitic sandstones pass upwards into massive or thick-bedded marble (Mana marble). Similar marbles are downfaulted to the north forming two fault blocks (Fig. 13a). In the more northerly fault block the marbles are locally intercalated with, and then overlain by, coarse quartzitic facies (Mana conglomerate; Fig.13c).

The lowest exposed clastic facies depositionally overlie a white marble that exhibits a fissured upper surface. Fissures are infilled with grey calcareous mudstone, lenticular (0.3 m) sandstone and then conglomerate (*c.* 0.6 m) with very well-rounded, dark grey quartzitic clasts (Fig. 13a, locality B). The clasts (up to 0.35 m in size) are poorly sorted and set in a sandy matrix. There is then further marble (*c.* 80 m thick) before the main succession of the Mana conglomerate comes in depositionally above. This comprises a gently dipping succession of alternating conglomerates, quartzitic sandstones and dark-coloured mud-rocks (phyllites) (Fig. 13c). The succession, although faulted, is locally well exposed along the road directly south of Sineniana (Fig. 13, locality C). There, several individual conglomeratic depositional units (up to 4.5 m thick) exhibit scoured bases, overlain by conglomerate with mainly subrounded to well-rounded clasts (up to 0.5 m in size). The succession grades upwards through medium and fine sandstone to shale. There are also several amalgamated, ungraded intervals of conglomerate and sandstone, again with well-rounded clasts and common intraformational rip-up clasts.

Similar graded, or ungraded, conglomerates are exposed more widely on the hillside to the SW (at Fig. 13, locality D) and can be traced laterally for hundreds of metres with little change in thickness. Interbedded sandstones are commonly graded, fining upwards from coarse sandstone, to fine sandstone, then dark shale. The mudrocks commonly include sandy partings (up to several millimetres thick) with very well-rounded quartz grains. The highest levels of the exposed succession include poorly exposed intercalations of shales, sandstone and conglomerate, and several laterally impersistent intercalations of marble (up to 10 m thick). In this area the Mana unit is structurally overlain by highly recrystallized carbonates and other facies (e.g. shales; radiolarian cherts) of the unmetamorphosed Pindos unit.

At the second, more southerly locality, near Sarakina *c.*14 km further south, the Mana unit is again well exposed as an isolated exposure near the top of a prominent hill (820 m) above Sarakina village (Fig. 14). The succession above the village begins with pale grey, buff or yellow phyllite with occasional thicker quartzitic sandstone beds (up to 0.6 m), of inferred Early Triassic (Scythian) age. South of a prominent col (Fig. 14a) the succession passes upwards without a break into medium-bedded sugary marble with pale phyllite partings. A sheared and folded succession above this is dominated by medium- to thick-bedded grey dolomite and shale of inferred Mid–Late Triassic age.

Above comes thick-bedded marble (Mana marble). The basal contact is marked locally by a *c.* 0.5 m thick zone of sheared and brecciated phyllite and recrystallized marble. The overlying, nearly massive Mana marble is overlain, apparently depositionally, by a veneer of conglomerate (Mana conglomerate) near the hilltop (Fig. 14a). The conglomerate comprises repeated weakly stratified depositional units, each up to several metres thick, made up of densely packed, clast-supported quartzitic conglomerates. The clasts range from well rounded to subrounded (average size 14 cm; maximum 60 cm), with occasional lenticular intercalations of dark sandstone. Despite the strong shearing and localized thrusting it is likely that an originally Lower Triassic–Upper Triassic or Lower Jurassic? right-way up succession is represented at this locality (Fig. 14b and c).

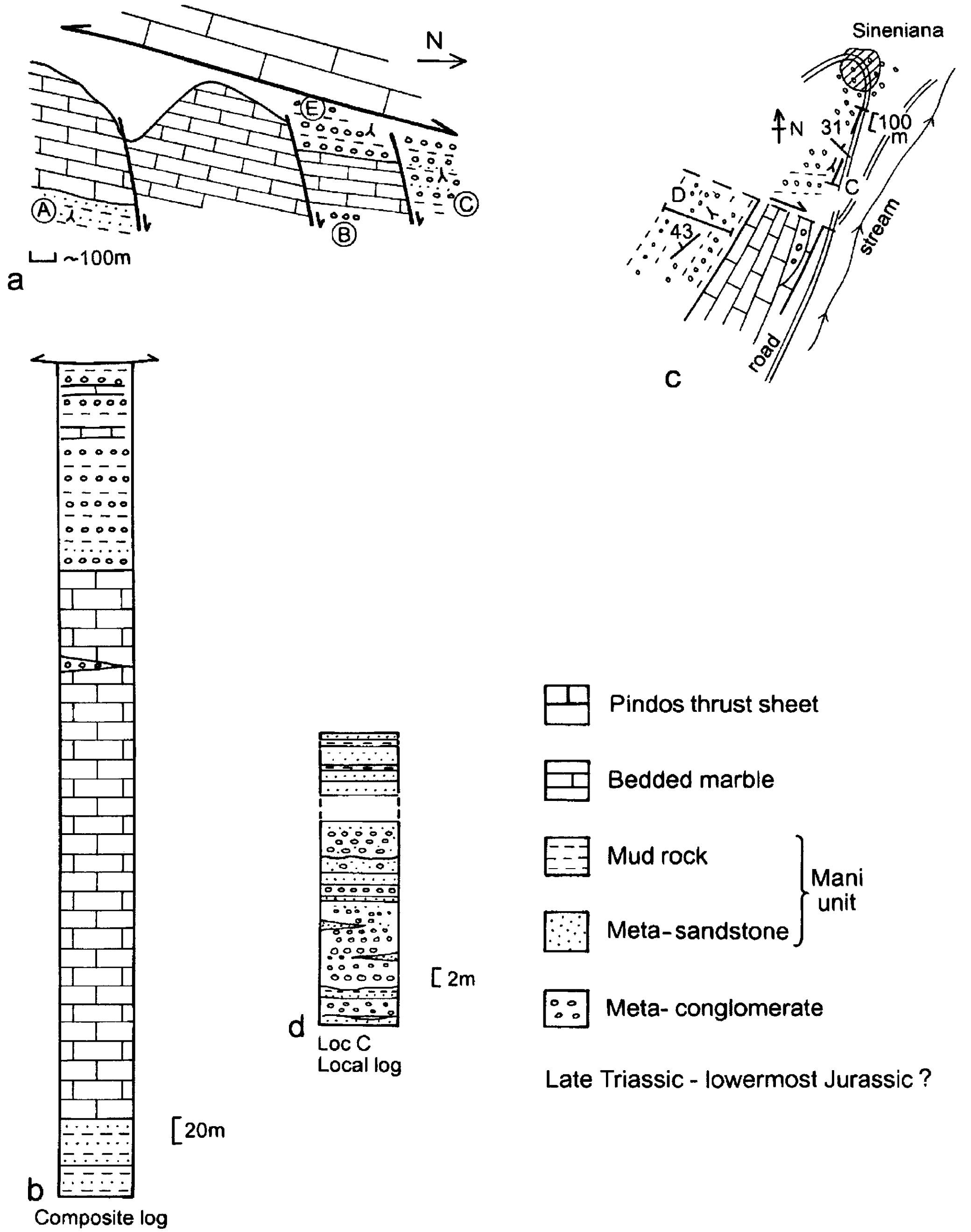


Fig. 13. Tectonic–sedimentary relations exposed in the upper, right-way-up succession in the Phyllite–Quartzite unit in NW Crete (see Fig. 10a). The data presented highlight the importance of the conglomeratic Mana unit (of Late Triassic–Liassic? age). (See text for discussion.)

The third succession studied occurs further south again, in the hanging wall of an east–west-trending zone of major neotectonic downfaulting to the south. Local successions are shown in Figure 10c, logs 2–4. For example, a partially landslipped succession, exposed near

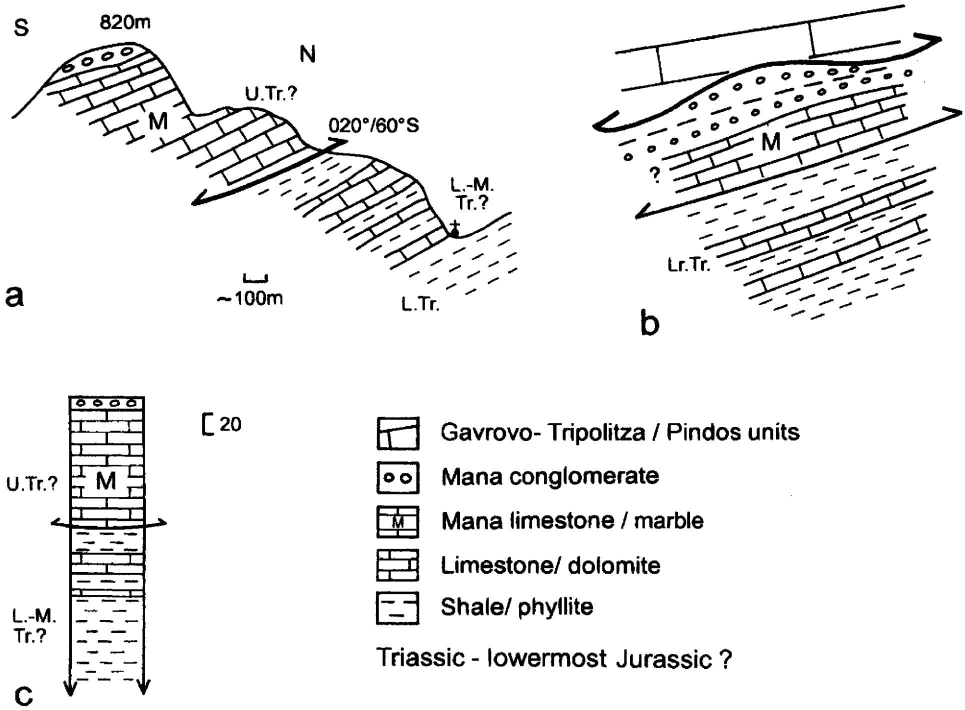


Fig. 14. Additional important tectonic–sedimentary relations exposed in the upper, right-way-up succession in the Phyllite–Quartzite unit in central southern Crete (see Fig. 10a). A Triassic succession is capped by the conglomeratic Mana unit (of Late Triassic–Liassic? age). (See text for discussion.)

Kondokinigi (by the turning to Voutas), passes from green phyllite (of Mid-Triassic? age, into thick-bedded marble (*c.* 16 thick; Mana marble?) and then into a coarsening-upward clastic succession of sandstone (10 m), coarse sandstone (8 m), then conglomerate (>25 m). In the same area (1 km south of Kondokinigi), a gently northward-dipping succession of dark phyllites and thick-bedded quartzitic sandstones passes depositionally upwards into typical Mana conglomerates, *c.* 60 m thick. This conglomerate is dominated by elongate, subrounded, pale yellow to grey, fine-grained sandstone and siltstone, in a matrix of yellowish grey phyllite. Occasional clasts of coarse sandstones are also present. Many of the clasts are elongate (up to 0.6 cm × 15 cm) and set in a sandy and gritty matrix. The conglomerates were in turn overridden by the unmetamorphosed Gavrovo–Tripolitza and Pindos units. Further west, on the westward extension of the fault zone, near Papadiana, local exposures of the Mana unit include conglomerate with subrounded sandstone clasts (up to 0.2 m in size) and quartzitic sandstones affected by soft-sediment deformation.

Interpretation: a developing rift

The overall Middle–Upper Carboniferous to Lower Triassic succession accumulated on a relatively deep-marine slope, to base-of-slope, setting, judging from the redeposited nature of the sandstones and limestones, and the presence of Radiolaria and deep-water conodonts within the interbedded shales. The palaeowater depths are estimated as >500 m (H. Kozur, pers. com.; Fig. 10d). The large-scale (tens to hundreds of metres thick) cycles record the interplay of local tectonics versus eustatic sea-level change, whereas the smaller scale (tens of metres or less) thickening- and coarsening-upward cycles may relate to autocyclic (steady-state) depositional processes (e.g. sand lobe progradation). Sand input peaked during the Mid?-Permian, then waned, whereas carbonate input relatively increased during Early Triassic time. The source of the sandstones was probably Pan-African basement, as exposed in Egypt and Libya, or possibly a rifted continental fragment of the same crustal type. The common well-rounded pebbles within thicker-bedded sandstones were derived from a high-energy shallow-marine or fluvial setting. These sediments were redeposited into deep water by a range of low- to high-density turbidity current and mass-flow processes. Most of the thick-bedded, locally pebbly sandstones are seen as subaqueous sand flows (gravity flows). The prominent interval of matrix to clast-supported conglomerates of Early Triassic age towards the top of the siliciclastic succession records oversteepening of the pre-existing slope, resulting in widespread down-margin gravity transport. The mass movement was possibly triggered by uplift from a deep-sea setting to a shallow-marine carbonate-depositing setting during Mid-Triassic time, resulting in a hiatus in deposition. Material of mostly Permian and Triassic age was derived from settings ranging from shallow shelf (e.g. oolites) to deep water (e.g. micritic clasts with pelagic conodonts).

Little evidence was observed to support the suggestion of sand deposition, within (or close to) an idealized deep-sea submarine canyon (Dornsiepen *et al.* 2001). There are few examples of coarse lenticular, channelized, debris-flow-type conglomerates, typical of such canyon-mouth settings. The sandstones are also dissimilar to typical 'classical' turbidites, as most are poorly sorted and only rarely well graded. These sediments were possibly deposited as sandy mass flows that were transported down a relatively steep, fault-controlled slope, possibly fed from line sources, rather than through regional-scale submarine canyons. The subordinate thinner bedded limestones are interpreted as relatively distal calciturbidites derived from a carbonate platform.

The alternating thinner- and thicker-bedded sand deposition persisted from Mid-Carboniferous to Early Triassic time. This is consistent with gentle subsidence of a rift rather than the long-term subsidence of a passive continental margin bordering an ocean basin, in which an overall thinning and fining-upward and deepening succession would be anticipated. As inferred from western Sicily, above, there is actually no evidence that a Late Palaeozoic ocean actually ever existed in the south Aegean region to the south, adjacent to Gondwana. Also, the Triassic alkaline volcanic rocks (extrusives and sills) that were erupted into deep water mainly during Late Permian?–Early Triassic time would not be expected in a mature passive margin succession. Given their characteristic 'enriched' composition, these volcanic rocks are better interpreted to represent low-degree melts that erupted in an extensional rift setting.

The Phyllite–Quartzite unit of western Crete has also been suggested to represent the distal facies of a south-Palaeotethyan passive margin that accumulated on oceanic crust (Ziegler & Stampfli 2001), but this is opposed by the relatively proximal setting of the sediments. This unit was also suggested to represent Palaeotethyan oceanic sediments preserved as a 'Cimmerian' accretionary prism of pre-Jurassic age (Ziegler & Stampfli 2001). However, this interpretation is opposed by the existence of long intact sedimentary successions and the absence of contemporaneous thrust-imbrication as in accretionary prisms. The observed deformation and metamorphism are instead believed to be entirely of Early Cenozoic age.

The inferred rift basin generally experienced uplift of >500 m during Mid- to Late Triassic time. This uplift took place especially during Late Early Triassic to Mid-Triassic time (Krahl *et al.* 1983*c*). This shallowing was linked to more calcareous, shallower water deposition, although with continuing siliciclastic input. Shallowing continued until deposition was restricted to semi-isolated shallow-marine to lagoonal settings that were possibly fault-controlled. Relative sea-level fall culminated in gypsum deposition in evaporating marginal lagoons. Partial dissolution in response to freshwater leaching gave rise to local solution-collapse breccias (Pomoni-Papaioannou & Karakitsios 2002). By contrast, in the right-way-up succession, open-marine carbonate deposition persisted into the Mid-Triassic, followed by a relatively abrupt upward transition to a shallow-marine carbonate-depositing setting during latest Triassic–Early Jurassic time (Krahl *et al.* 1983*c*); this culminated in the deposition of the Mana marble and Mana conglomerate. The Mana conglomerates are interpreted as shallow-marine to non-marine facies that were deposited in high-energy deltaic settings, possibly including fan deltas. The source of the nearly monomict, remarkably pure quartzitic sandstones is problematic, as derivation from either Pan-African or Hercynian basement would be expected to have produced more polymict material, including schist and gneiss. The sandstones could instead have been derived from an uplifted part of the Phyllite–Quartzite succession, assuming this was already lithified, but again, a more heterogeneous composition, including quartzose, carbonate and basic igneous rocks, would be expected. One other possibility is that the Mana conglomerates were eroded from a succession of texturally mature sandstones, which accumulated on a marginal high (fault block or plateau) bordering a rift basin.

Assuming the regional structure of a south-facing recumbent nappe is correct (Krahl *et al.*

1983*c*), the inverted limb restores to a relatively southerly position, compared with the right-way-up limb. This would imply that the source area of the Mana conglomerate was generally to the north. If correct, the source was a rifted margin or intra-basinal high to the north.

In summary, the western Crete successions are consistent with Model 1 in which rifting took place in pulses, at least during Mid-Carboniferous (post-Hercynian orogeny) and Mid-Triassic time. In Model 2 the turbiditic sandstones of Permian–Early Triassic age would relate to subsidence of the southern margin of a Cimmerian passive margin. However, similar deep-water siliciclastic sedimentation started up to 150 Ma earlier. In Model 2 the Mid-Triassic hiatus and uplift could reflect the passage of a flexural bulge across the basin as Palaeotethys closed to the north. However, such flexural uplift, if significant, should have been followed by flexural downwarping related to southward passage of a thrust load, itself related to the latest Triassic 'Cimmerian' collision with the Eurasian continent to the north. Instead, continued relative uplift is seen through Late Triassic time, culminating in evaporitic deposition. The Mana conglomerate cannot have been shed from the Eurasian margin, in view of its nearly homogeneous quartzitic composition. A more heterogeneous composition, including arc-related or ophiolitic rocks, would be expected if the conglomerates were related to a forearc, foreland basin or collisional setting. Also, if related to a convergent setting an upward-thickening and coarsening succession would be expected, which is not observed.

Phyllite–Quartzite unit of eastern and central Crete

Distinctive lithofacies of the Phyllite–Quartzite unit are exposed in eastern Crete, particularly in the Chemezi and Vai areas (Fig. 5). These lithologies differ from those western Crete in terms of facies and structure. In particular, long intact successions are preserved in eastern Crete whereas successions in western Crete form parts of a tectonic slice complex. In eastern Crete there are two main exposure areas: the Chemezi area and the Vai area in the far east of the island. Here, the main focus will be on the Chemezi area, where most of the relevant units are exposed and the tectonostratigraphy is relatively simple. On the other hand, the local tectonostratigraphy of the Vai area is extremely complicated and to some extent controversial, mainly owing to Cenozoic deformation and metamorphism, such that a fuller treatment is required elsewhere. Most of the units in the Vai area can be generally correlated with those of the Chemezi area, discussed below. However, two critical units in the Vai area are not known in the Chemezi area and these will be discussed below. Some thrust sheets (e.g. high-grade metamorphic rocks) are thin and laterally variable such that no simple tectonostratigraphy is applicable to the area as a whole, and so the various units must be studied and interpreted individually.

In the Iraklion area of central Crete (Fig. 5) an additional outcrop area of Triassic sedimentary and volcanic rocks can be correlated with the units cropping out in eastern Crete. Importantly, these structurally overlie a unit with lithological affinities with the Phyllite–Quartzite unit of western Crete. This allows an assessment to be made of the relative tectonic settings of these two regionally contrasting units.

Taking the Chemezi and Vai areas together, in Model 1 the Upper Permian–Lower Triassic succession represents a rift setting. Upper Permian–Lower Triassic deep-water sediments (i.e. radiolarian shales and hemipelagic limestones) record part of an early rift basin, a counterpart of the pelagic sediments exposed in western Sicily. Middle–Upper Triassic successions then represent a shallow-water to non-marine rift-setting and also deeper marine rift-related settings related to opening of the Pindos ocean to the north. Some Triassic lithologies are associated with volcanic rocks of basic to intermediate composition, which are interpreted as rift related in this model. Fragments of pre-rift continental basement are represented by thrust slices of 'Hercynian' high-grade metamorphic rocks. In Model 2 the successions in the different thrust sheets should record a wide range of tectonic settings, including an ocean basin with volcanic seamounts, a forearc basin represented by radiolarian sediments, a continental crust unit represented by 'Hercynian' basement and a back-arc basin related to the opening of the Pindos ocean. In Model 3, only a Triassic rift would be expected, together with subduction-related magmatic rocks. In Model 4 a supra-subduction zone rift-related setting would be expected, with magmatism possibly affected by either, or both, of a northward-dipping and a southward-dipping subduction zone.

For the central Crete Iraklion area, in Model 1 both structural units exposed would relate to Triassic rift successions. However, in Model 2 this occurrence would record the actual Palaeotethyan suture between a rifted Cimmerian passive margin (i.e. North Africa derived) and an Eurasian active continental margin.

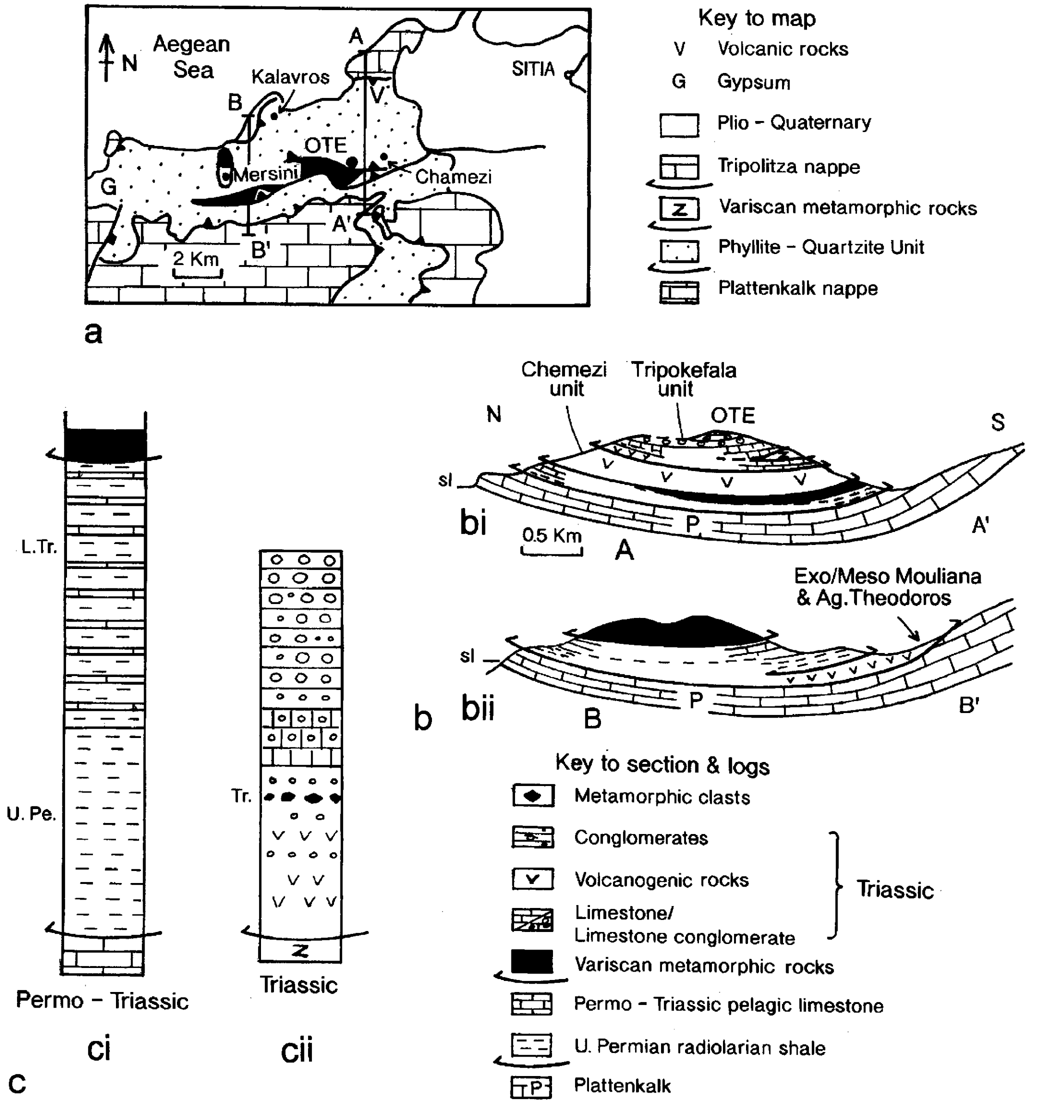


Fig. 15. Upper Palaeozoic–Lower Mesozoic units of eastern Crete from the Phyllite–Quartzite unit in the Chemezi area, west of Sitia. (**a**) Outline map; (**b**) sketch cross-sections; (**c**) measured logs. (See text for explanation and data sources.).

Evidence from the Chemezi area

In this area the Phyllite–Quartzite unit forms a number of thrust sheets that are folded into a broad east–west-trending syncline (Mersini syncline; Krahl *et al.* 1986; Fig. 15a). Two well-exposed traverses were examined during this study (Fig. 15b). Combining these, the following units are recognized from the structural base upwards.

Hemipelagic shale and pelagic limestone

This unit is well exposed in the NW, 0.3 km west of Kalavros (Fig. 15bii). The highest levels of the Plattenkalk unit there, the Lower Cenozoic Kalavros beds, are composed of muddy limestones that are overthrust by a distinctive unit of thin-bedded, relatively undeformed reddish brown platy shale (phyllite), up to 100 m thick, with Mn coatings on fractures. These shales

(Violetschiefer) contain Radiolaria and pelagic conodonts of Early Permian age (Kozur & Krahl 1984). In places, the shales are chemically reduced to a grey or green colour. In thin section, the typical shale is mainly ferruginous mudstone with silty laminae composed of quartz and feldspar. The reddish shale grades upwards into grey– green shale, then into thin-bedded hemipelagic limestone, of up to Early Triassic age (Krahl *et al.* 1986). A succession of relatively undeformed alternating well-bedded pure and muddier limestone with shale partings, *c.* 120 m thick (Agrilos Schichten), is then overthrust by a strongly contrasting mainly volcanogenic unit (Fig. 15ci).

Similar, but faulted and thrust-imbricated, Permian shales and overlying Lower Triassic hemipelagic limestones are well exposed further south, especially along the main road from Ayios Nikolaos to Sitia. Local transitions from red shale to pelagic limestone are again seen. Notably, *c.* 200 m east of Exo Mouliana, a local hemipelagic limestone succession (*c.* 80 m thick) comprises alternations of thin- to medium- and thick-bedded facies, with muddy partings. Further east, near Ayios Theodoros, dark muddy partings are rich in fine plant material. In thin section, the limestones are finely laminated, with numerous small organic-rich grains.

The Lower Permian reddish shales are interpreted as deep-sea hemipelagic sediments, with radiolarians, conodonts and fine-grained terrigenous sediment. The presence of reworked fossils suggests that the deep-water succession may date at least from Late Carboniferous time (Krahl *et al.* 1986), contemporaneous with the deep-water sediments in western Crete. The abundant manganese oxide in the pink shales could reflect continental runoff or even a hydrothermal source of manganese, but there is no known lithological association with igneous rocks. No basement to these deep-sea sediments is exposed. The siliceous sediments gave way to hemipelagic carbonates in the Early Triassic, with continued input of terrigenous silt and the addition of organic matter. The upward colour change possibly reflects a decrease in bottom-water oxygenation, or increase in organic matter, that was possibly climatically controlled.

Volcanic–pelagic carbonate unit

In the SW, near Mesa Mouliana (Fig. 15bii), the Plattenkalk unit dips at 40° to the north and is structurally overlain by volcanic rocks (Seidel 1978), including strongly weathered aphyric, non-vesicular massive basalts and andesites. These volcanic rocks are locally interbedded with grey to pink pelagic limestone, in depositional units up to 8 m thick. The volcanic rocks record eruption in a open-marine relatively deep-water setting, and may correlate with thicker and more intact Lower Triassic successions exposed on the northern flank of the Mersini syncline (see below).

High-grade metamorphic basement

The Permo-Triassic sediments and volcanic rocks are locally structurally overlain by several slices of high-grade basement rocks, mainly gneiss, mica schist, amphibolite and marble (Seidel 1978; Seidel *et al.* 1982). The protoliths include meta-sediments, basic igneous rocks and granitic rocks. For example, near Chemezi, the metamorphic rocks are dominated by dark grey, brown weathering, mica-schist intercalated with dark amphibolite. The metamorphic rocks are cut by numerous quartz veins and show evidence of extensive brittle deformation (e.g. small-scale duplexes), especially near the thrust contact with the underlying deep-sea sediments. Recent radiometric dating shows that the high-grade metamorphic rocks are divisible into several units, one with Cadomian (late Precambrian) and Late Carboniferous ages (Mirsini crystalline complex), and another with Late Permian ages (Kalavros crystalline complex; Finger *et al.* 2002; Romano *et al.* 2002, 2004, 2006). Some of the amphibolites exhibit MOR-type protoliths suggesting that they could have been accreted from a 'Hercynian' ocean prior to metamorphism. South-verging small-scale structures (e.g. C–S fabrics) have been reported by Romano *et al.* (2006), although these need to be treated with caution as they occur within relatively thin thrust sheets that have undergone Early to Mid-Cenozoic subduction and exhumation.

Volcanogenic–limestone–shale–conglomerate unit

In the NW, on the northern flank of the Mersini syncline (Fig. 15bi), the Permian–Early Triassic deep-water radiolarian shale–pelagic carbonate succession or, locally, high-grade metamorphic rocks are structurally overlain by a contrasting succession that begins with metavolcanic rocks (mainly basaltic andesite), *c.* 100 m thick (Fig. 15cii). Occasional massive lavas near the exposed base of the succession are overlain by lava breccias, in units up to 8 m thick. These are mainly composed of pebbly volcaniclastic conglomerate or breccia (with stretched pebbles up to 10 cm long), volcaniclastic sandstones, volcanogenic shales. Repeated thin (<1 m) lava flows are interbedded with limestone conglomerates. The lavas

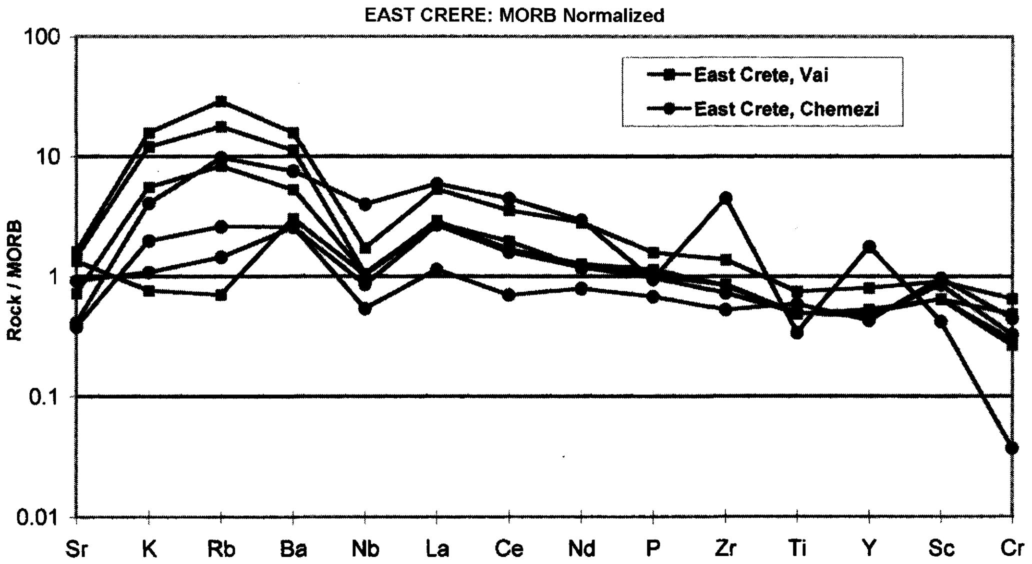


Fig. 16. Geochemical plots of Triassic basalts from the Phyllite–Quartzite unit in the Vai and Chemezi areas of eastern Crete. (See text for explanation and Table 1 for locations.)

and associated relatively deep-water sediments are inferred to be of mainly Late Scythian (Late Olenekian) age (Krahl *et al.* 1986). Several samples of basalt of sub-alkaline composition were analysed by XRF and when plotted on MORB-normalized spider diagrams (Fig. 16) reveal a distinctive subduction influence, as shown by a relative depletion of immobile incompatible elements (e.g. Pearce 1980), and notably a pronounced negative niobium anomaly.

The mainly volcanogenic lithologies are, in turn, overlain by thin- to medium-bedded limestones (*c.* 60 m thick). The thicker-bedded limestones are deformed into 'mega-boudins' (up to 10 m thick × 80 m long) by extensional shear, whereas the thinner beds are stretched to form detached ' phacoids' within shale.

Above this follows a strongly sheared interval, 10 m thick, probably also of Early Triassic age, made up of crudely stratified conglomerates and breccias, in beds up to *c.* 3 m thick, with clasts (up to 0.6 m in size) including schist and gneiss ('Chemezi beds'). Intercalated green volcanogenic shale (1 m thick) includes blocks of mica-schist (up to 0.4 m long). Overlying conglomerates are dominated by stretched pebbles composed of limestone, intercalated with sandstone lenses (up to 1.5 m thick).

Higher levels of the succession, *c.* 150 m thick, comprise intercalations of siliciclastic sandstone, andesitic lava breccia (with clasts up to 0.35 m in size), volcanogenic debris flows, pebbly sandstones (with angular limestone clasts), volcaniclastic sandstone (in <1 m thick interbeds) and volcanogenic shale. Occasional competent beds of limestone debris flows (in units <1 m thick) are strongly stretched to form phacoids in an incompetent volcanogenic matrix. Grading and scour structures indicate that the succession is mainly the right way up. The highest levels are isoclinally folded, suggesting the presence of a thrust, or at least strongly sheared contact, above which there are slightly metamorphosed brilliant reddish purple meta-mudrocks ('violet schists'), with thin (<10 cm), fine-grained sandstone interbeds (*c.* 80 m thick altogether). Fresh cuttings reveal the presence of dark organic-rich mudrocks prior to surface oxidation.

The mixed volcanogenic–limestone–shale–conglomerate unit records andesitic volcanism in a highly unstable setting marked by redepositional processes, with volcanogenic and carbonate debris flows and turbiditic volcaniclastic sandstones. Volcanism later gave way to the accumulation of volcanogenic sediments and limestone conglomerates with clasts of locally derived metamorphic basement lithologies. This unit is inferred to be late Early to Early Mid-Triassic in age (Chemezi beds; Krahl *et al.* 1986). The overlying dark organic-rich muds and minor volcaniclastic turbidites possibly represent an original upward continuation of this setting in a quiescent relatively deep-water, reducing environment.

Coarse limestone conglomerate unit

The above unit is structurally overlain by a thick (<250 m) succession of very coarse limestone conglomerates and breccias (Tripokefala beds), well exposed in the vicinity of the OTE tower; Fig. 15a). To the north, the underlying volcanogenic unit is bounded by a major high-angle fault. According to Kozur (pers. com.), the limestone conglomerates are no younger than late Early to early Mid-Triassic in age, whereas Champod *et al.* (2004) suggested a Ladinian to Carnian age. Near the basal thrust contact coarse limestone conglomerates predominate, with stretched, but originally well-rounded clasts (Fig. 15cii). These clasts are locally fused, giving the impression of massive carbonate rock. The limestones locally include small slices (duplexes) of metavolcanic rocks (up to 4 m thick × 10 m long), as seen on a steep slope 0.4 km WNW of the OTE tower. Above this there is a generally thinning- and fining-upward succession of limestone rudites, sandstones and shales (Fig. 15cii). The limestone are mainly lenticular conglomerates and breccias, mostly composed of stretched limestone clasts. The interbedded sandstones are packed with elongate limestone clasts (up to 35 cm long) and include occasional large blocks of schist, gneiss (up to 0.6 m in size) and occasionally andesite. Several individual units grade from microconglomerate to sandstone with shale partings and occasional pebbly bands, suggesting that the succession is the right way up. The highest exposed levels (near the OTE tower) are dominated by brown-weathering shales and sandstones, with only occasional conglomerates (up to several metres thick). Most clasts are <10 cm in size but a few intervals with larger clasts are also is present.

Thin sections show that the typical sandstones are dominated by polycrystalline and monocrystalline quartz, with subordinate mica-schist, muscovite and plagioclase, set in a variably altered, chloritic fine-grained, or siliceous matrix. Some sandstones contain abundant relatively fresh basalt, mainly as angular elongate grains (shards) in a dark mesostasis of altered volcanic glass. Other sandstones contain coarse plagioclase (diabase or gabbro) and large clear tabular quartz grains (possibly reworked phenocrysts).

The coarse limestone conglomerates (of Early–Mid-Triassic age) are interpreted as a proximal deltaic unit that accumulated in a shallow-marine to possibly non-marine setting. The clasts were derived from metamorphic basement, shallow-water carbonates and volcanogenic units. The structurally underlying units (volcangenic units and high-grade metamorphic slices) could have supplied most of this material. However, a suggestion that the conglomerates represent a 'Verrucano-type facies' unconformably overlying metamorphic basement (Champod *et al.* 2004) could not be confirmed, as all the observed contacts are tectonic.

Exotic gypsum

A notable feature of this area is the presence of large lenses of gypsum, up to several hundred metres thick, in which Triassic fossils are recorded (Krahl *et al.* 1986). A good example is exposed in, and around, a large working quarry in the west of the area (SSW of Mochlos; marked G in Fig. 15a), just above the Plattenkalk unit. This gypsum is entirely recrystallized, shows locally steep banding and is deformed by ductile folds. The sugary recrystallized gypsum includes numerous angular fragments of dark dolomitic carbonate (up to 15 cm long). The gypsum is exotic to the intact stratigraphic successions in the Chemezi area and is interpreted as having been extruded along the regional tectonic contact between the underlying Plattenkalk and the overlying Permo-Triassic units from a source of Triassic evaporite elsewhere in the basin.

The possible regional tectonic significance of the Chemezi area will be considered after a summary of the comparable Vai area, below.

Evidence from the Vai area

The outcrops in the Vai Peninsula (Fig. 5) are located on the western and eastern parts of the peninsula with intervening Neogene–Recent sediments (Fig. 17). An initial non-trivial problem is how to correlate the tectonostratigraphy of these two outcrop areas.

Recent workers agree that a stack of thrust sheets is present in the Vai Peninsula as a whole, as in the Chemezi area discussed above. The entire peninsula was mapped and described by Haude (1989). He interpreted the peninsula as a pile of thrust sheets, which were folded into a huge NNE–SSW-trending recumbent isoclinal fold, with the western limb mainly the right way up and the eastern limb inverted. This interpretation was largely followed by Krahl & Kauffman (2004). During this work it was found that by no means all of the successions within individual thrust sheets in the eastern Vai exposures are stratigraphically inverted, questioning the existence of a simple west-facing nappe structure.

In the western peninsula area (Fig. 17d) most of the contacts between lithological units are sheared such that it is commonly difficult to

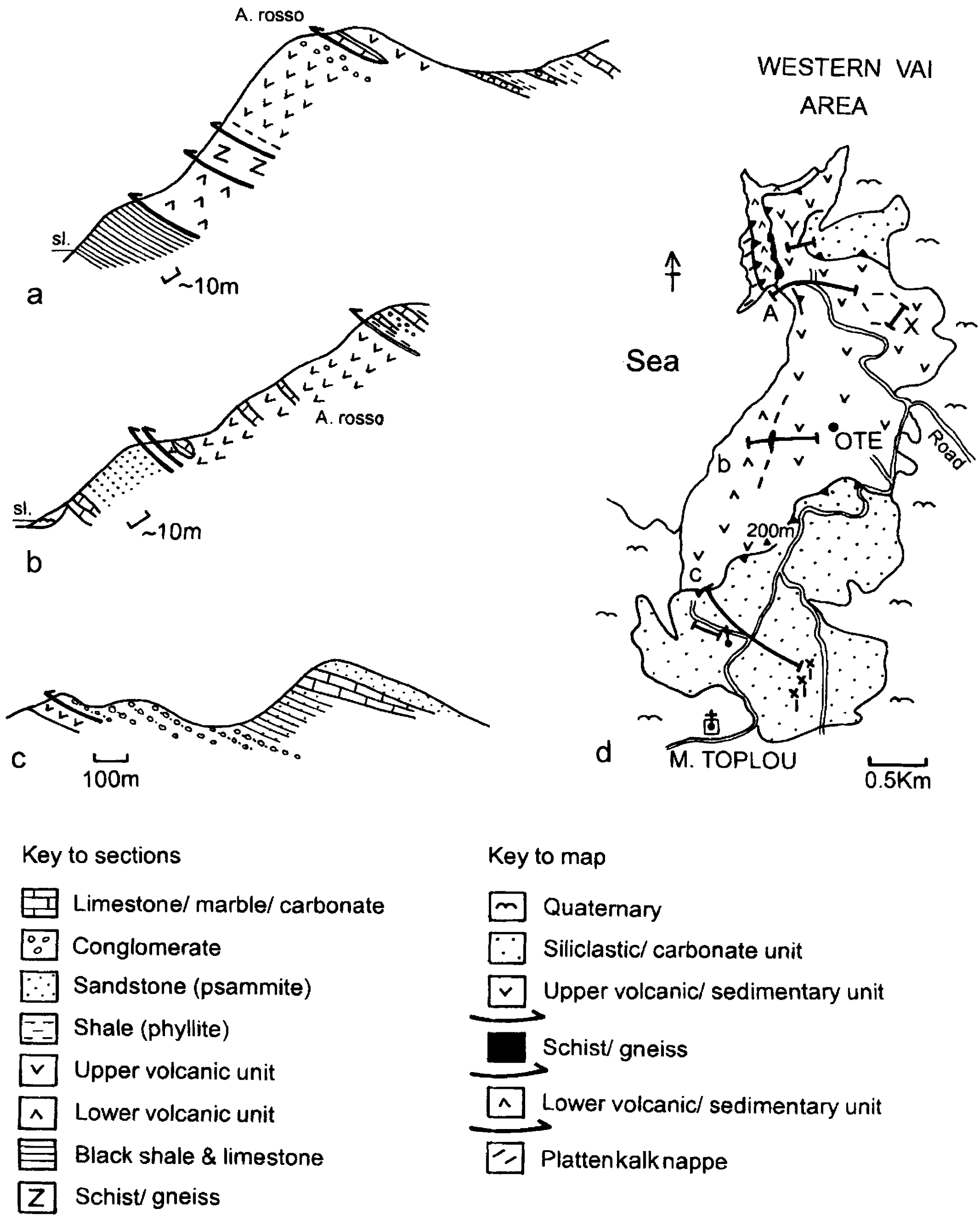


Fig. 17. Tectono-stratigraphy of the Phyllite–Quartzite unit exposed on the western side of the Vai peninsula, eastern Crete. (**a**–**c**) Representative cross-sections; (**d**) sketch map showing the locations. (See text for discussion and data sources.).

distinguish between primary thrust contacts and sheared lithological contacts. However, several intact successions can be inferred in the western peninsula area (Fig. 17a–c). Details of these successions and measured logs will be given elsewhere. In general, moving structurally upwards there is a lower unit dominated by relatively deep-marine, mainly andesitic volcanogenic rocks, including pelagic limestones (e.g. Ammonitico Rosso) and rare red chert, of inferred Early Triassic age, mainly based on dating of conodonts within the interbedded Ammonitico

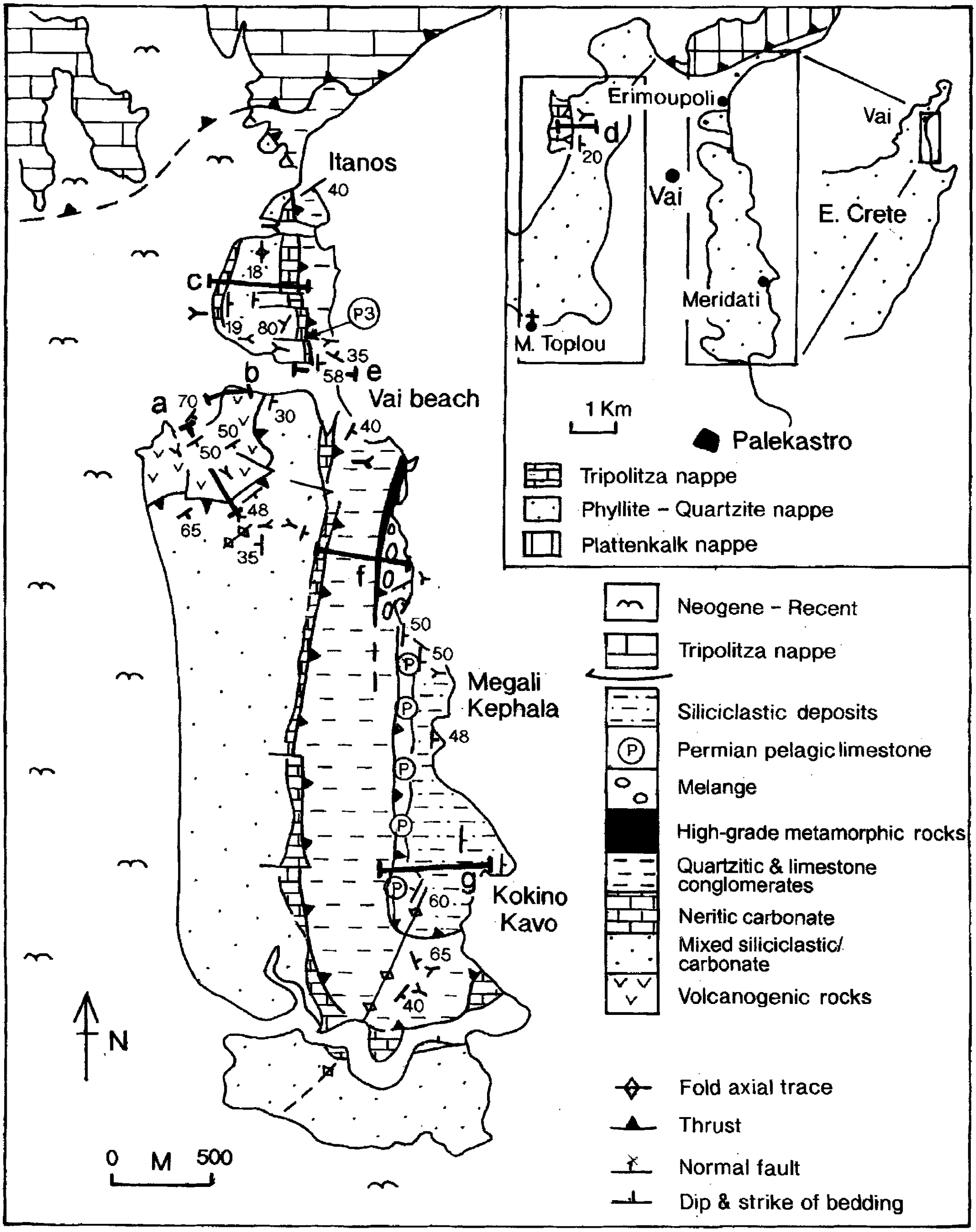


Fig. 18. Tectonostratigraphy of the Phyllite–Quartzite unit exposed on the eastern side of the Vai peninsula, eastern Crete. Inset: larger area including Toplou monastery and the west coast of the Vai peninsula (Fig. 17d). (See text for explanation and data sources.)

Rosso (J. Krahl, pers. comm). This volcanogenic unit is locally intersliced with schist and gneiss, and may include older (?Late Permian–Early Triassic) dark organic-rich shales and carbonates near the base. The volcanogenic unit is overthrust by a mainly shallow-water succession of carbonates interbedded with siliciclastic sandstones and conglomerates of inferred deltaic origin. Clasts

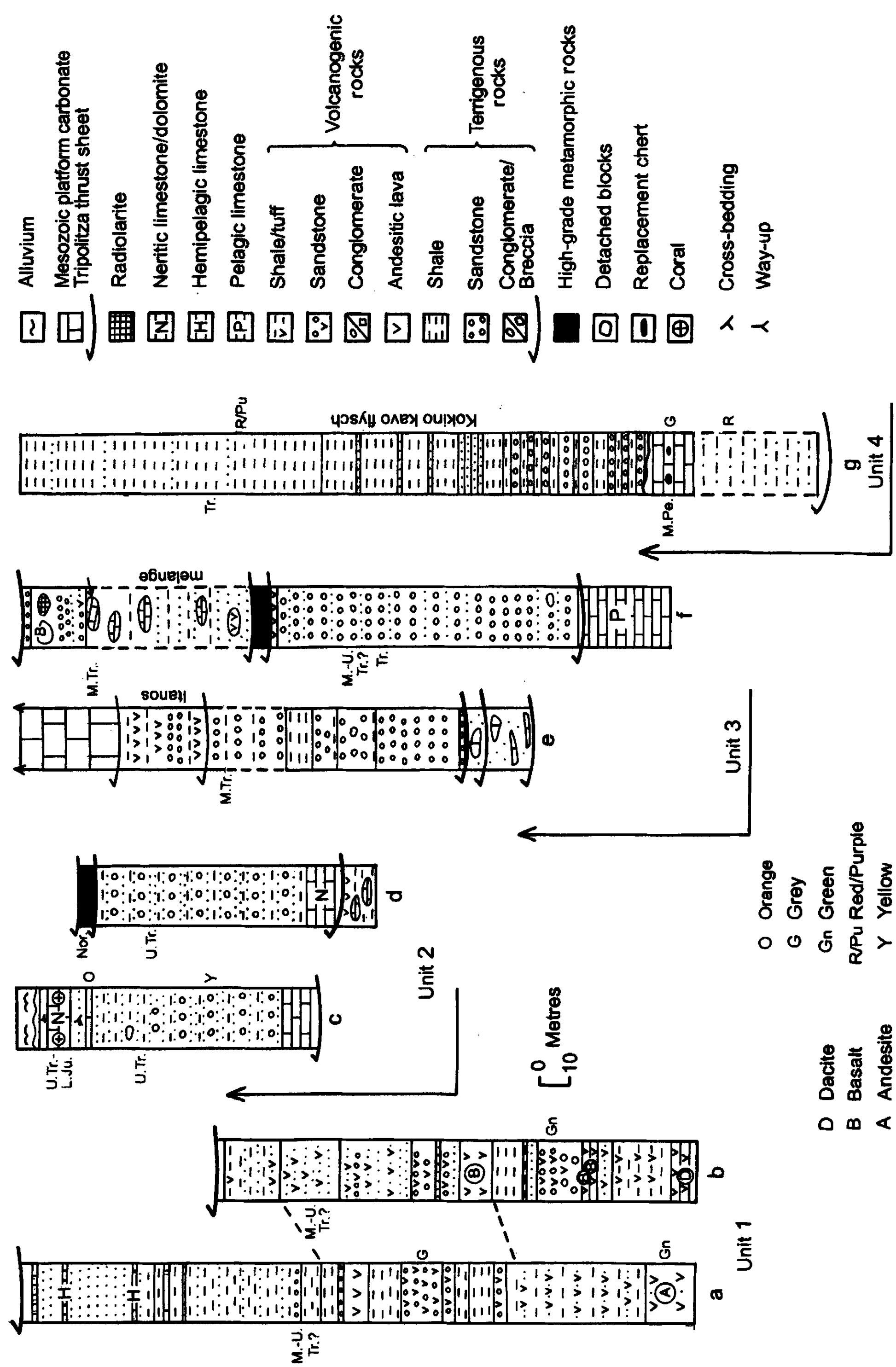


Fig. 19. Measured logs of sedimentary–volcanic successions exposed on the eastern side of the Vai peninsula, eastern Crete. The logs are arranged in overall structural order from the base to the top. (See Fig. 18 for locations and the text for explanation.)

range from mainly carbonate to metamorphic basement-derived. The age of this succession is possibly Mid–Late? Triassic based mainly on lithological correlation with similar units exposed in the eastern side of the peninsula.

The eastern peninsula area (Fig. 18) is more complex, lithologically and structurally. In general, four main eastward-dipping units were recognized during this work in ascending (present-day) structural order. Sedimentary logs were measured in each of these units (Fig. 19) and will be described in detail elsewhere.

First, a mixed volcanogenic–siliciclastic unit ('arc unit' of Stampfli *et al.* 2003) is exposed in a small area (several square kilometres.) in the eastern part of the peninsula inland from Vai beach (Figs 18a, b & 19a, b). This outcrop correlates with a much larger volcanogenic succession of Early Triassic (Olenekian) age of the western area of the Vai Peninsula (Fig. 17a, b). Several samples of relatively fresh basaltic lava of sub-alkaline composition were chemically analysed from this unit (see Table 1). Samples from near the base of the succession (Fig. 16) show a subduction influence, as indicated by a pronounced negative Nb anomaly, similar to the lavas from the Chemezi area.

Second, a mixed siliciclastic–carbonate unit is extensively exposed (Figs 18c, d & 19c, d), especially SE of Vai beach on a broad ridge running southwards *c.* 3 km, as far as near Meridati (Fig. 18d inset), and is also well-exposed NW of Vai beach. The exposures, north and south of Vai beach, are likely to be offset by a down-to-the-north normal fault (Haude 1989). The succession is equivalent to the 'Vai flysch, Upper forearc' unit of Stampfli *et al.* (2003). However, the succession reported here differs, as it necessary to take account of folding on north–south axes as indicated by changes in the younging direction (Figs 18 and 19).

Third, a coarse quartzitic and carbonate conglomerate unit (Figs 18e, f & 19e, f) is exposed north of Vai beach, near the coast. This comprises siliciclastic conglomerates and breccias equivalent to the 'Vai flysch upper forearc unit' of Stampfli *et al.* 2003. Units 1–3 can be generally correlated with the Triassic coarse clastic units exposed in the Chemezi area.

Fourth, there is a slice of mélange, *c.* 300 m thick, equivalent to the 'olistostrome' of Stampfli *et al.* 2003 (Figs 18f and 19f). This is associated with a slice of high-grade metamorphic rocks (Fig. 18f). There is no known equivalent of this mélange unit in the Chemezi area. However, it is critical to the interpretation of alternative settings, as it has been interpreted as a Palaeotethyan accretionary complex in Model 2 (Stampfli *et al.* 2003) and so will be discussed in some detail below.

The sedimentary matrix of the mélange, which is well exposed along the coast, comprises intercalations of coarse clastic sediments and mud rocks (phyllites), ranging from reddish to pink and green, owing to local oxidation–reduction effects. Individual matrix-supported conglomerates, typically *c.* 1 m thick, and with scoured bases, grade from pebbly conglomerate into coarse quartzitic sandstone, then shale. Other interbeds include abundant limestone talus. Other coarse clastic sediments are interpreted as high-density turbidites and classical turbidites. Sedimentary way-up evidence is indicative of outcrop-scale isoclinal folding on north–south axes. However, the eastward-dipping mélange fabric appears to be mainly inverted based on graded bedding and basal scour structures in most debris flows and sandstone turbidites.

Individual mélange blocks are best exposed several hundred metres inland. The most common mélange block is limestone, which shows local transitions to sedimentary limestone talus. There are also rare small blocks of pelagic sediment. For example, a small block (0.6 m × 0.3 m) of red radiolarite was observed within quartzitic and carbonate-rich debrites. Small lenses and blocks of red shale are also present, together with several blocks of pink pelagic limestone (Ammonitico Rosso), up to *c.* 15 m in size, as seen in the east. Several blocks of andesitic lava and basaltic lava breccia are also present. The Ammonitico Rosso is inferred to be of Anisian age, whereas the red radiolarite was dated as Late Anisian (Mid-Triassic) (Stampfli *et al.* 2003). Blocks of both chemically andesitic and within-plate-type basalt were reported (Stampfli *et al.* 2003). One lava block analysed during this study shows a subduction-influenced signature (Fig. 16).

In addition, in the highest structural position along the coast there is a well-exposed, remarkably intact, relatively undeformed succession of pelagic limestone, shale and sandstone (Figs 18g & 19g). No direct equivalent of this is known in the Chemezi area, although the pelagic limestones at the base of the succession in the eastern part of the Vai Peninsula could be equivalent to the pelagic limestones in the upper part of the Upper Permian–Lower Triassic succession at Chemezi.

This eastward-dipping composite unit is well exposed between Megala Kephali and Kokino Kavo (Figs 18g and 19g), the latter name reflecting a brilliant red–brown–violet colour of metashales and metasandstones exposed on headlands. The basal contact with the mélange is associated with strong shearing, isoclinal folding, duplex formation, tension gashes and carbonate veining. Kinematic indicators indicate normal fault displacement (in present orientation).

Above this contact, the unit begins with strongly deformed thin- to medium-bedded black limestones and black shales (*c.* 8 m). Grading and sharp bases in less deformed medium-bedded limestones near the contact are indicative of (local) stratigraphic inversion. Associated folded phyllite varies from green to purple, probably related to diagenetic alteration. Southwards, the succession passes into alternating thin-, medium- and locally thick-bedded grey limestone–phyllite alternations (*c.* 80 m thick). The limestones contain deep-water conodonts of Mid-Permian age (Stampfli *et al.* 2003; Krahl & Kauffman 2004). The individual pelagic limestone beds are laterally continuous and exhibit internal fine parallel lamination, suggestive of an origin as fine-grained calciturbidites, together with scattered nodules and lenticles of black chert of diagenetic origin. There is then a sharp, but apparently depositional contact between the fine-grained limestone and very coarse breccia–conglomerate above, composed mostly of quartzitic clasts. Traced laterally, this contact appears to cut stratigraphically downwards into the hemipelagic limestones and is therefore interpreted as a low-angle erosional unconformity.

Above this contact, individual well-bedded (eastward-dipping) lenticular conglomerates and sandstones ('Vai flysch, lower forearc unit' of Stampfli *et al.* 2003) can be traced laterally (up to several hundred metres) before wedging out. Individual depositional units, up to 3 m thick, begin with clast-supported, quartzose conglomerates (clasts up to 4 cm in size), grading upwards into sandstones with muddy tops. Excellently developed grading in many beds confirms that this succession is the right way up. In all, nine graded packages of conglomerate–sandstone, each up to 2.5 m thick, were observed. The average thickness and grain size of each depositional package decreases upwards, until the exposed succession (in coastal cliffs) culminates (at Kokino Kavo) in up to 100 m of reddish meta-mud-rocks (phyllites), interbedded with thin- to medium-bedded, graded metasandstones (Fig. 19g). Although this succession generally thins and fines upwards, several thick, relatively coarse beds persist to the highest exposed levels along the coast. In thin section, the sandstones are composed of relatively uniform, well-sorted grains of mainly quartz and quartzite, with minor muscovite and granular iron oxide. This spectacular succession remains undated but is probably of Triassic age based on regional comparisons with other facies.

Interpretation: rift-related settings

In the eastern Vai exposure area, the volcanogenic–siltstone–carbonate unit (Unit 1) of Early–Mid-Triassic (Late Scythian–Anisian?) age records the extrusion of mainly subduction-influenced andesites, as massive flows, volcanic breccias and rare pillow lavas, interbedded with volcaniclastic sediments, tuffs, pelagic carbonates, with deep-water conodonts (of Early Triassic age) and rare metacherts (jaspers). This volcanogenic succession (or a lateral equivalent) is a possible source for Mid-Triassic island-arc tholeiite (IAT)-type basalt blocks and also the Ammonitico Rosso and chert in the mélange, assuming volcanism continued into the Mid-Triassic. A possible exception is one recorded instance of a block of within-plate basalt in the mélange, which Stampfli *et al.* (2003) interpreted as remnant of an oceanic seamount. However, geochemically similar basalts also occur in rift settings, e.g. as in western Crete, discussed above.

The mixed siliciclastic–carbonate unit (Unit 2), widely exposed in both the eastern and western outcrop, is interpreted as a shallow-marine to locally non-marine, channelized deltaic sequence of Early Triassic (Early Olekenian)–Late Triassic (Carnian–Norian?) to Early Jurassic (Rhaetian?) age (Stampfli *et al.* 2003). The conglomerates accumulated in a proximal deltaic setting where metamorphic basement was exposed. The interbedded limestone conglomerates probably record the erosion of a carbonate platform, in response to relative sea-level changes or local tectonics. A more intact carbonate platform, rich in coral, became established during the Norian–Rhaetian. The conodont assemblage within these limestones is reported to be similar to that elsewhere at the base of Tripolitza carbonate platform succession, which may suggest that the Phyllite–Quartzite succession originally continued upwards into the Tripolitza platform (Stampfli *et al.* 2003). However, the actual contact is now a major structural and metamorphic discontinuity, probably a major Alpine thrust that was reactivated by exhumation.

The structurally overlying coarse quartzitic and carbonate conglomerate unit (Unit 3), of inferred late Early–Mid-Triassic age, accumulated on relatively steep subaqueous slopes dominated by gravity-flow processes, probably as proximal fan deltas on a linear margin. These

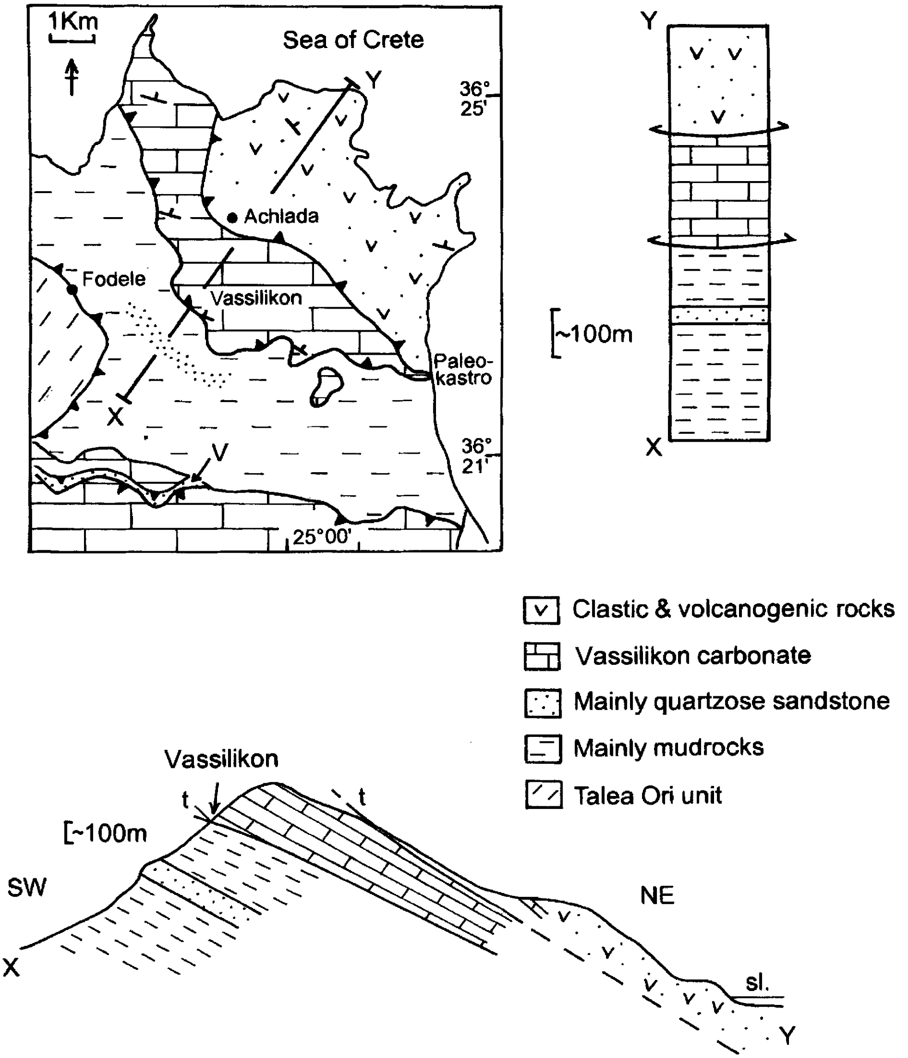


Fig. 20. Occurrence and tectonostratigraphy of the Phyllite–Quartzite unit, west of Iraklion in central northern Crete. (See text for explanation and data sources.)

sediments are unlikely to represent a channelized mid-fan of an idealized deep-sea fan, as suggested by Stampfli *et al.* (2003), especially as the conglomerates are tabular to broadly lenticular rather than markedly channelized; finer grained inter-channel fan muds and turbidites, as expected in mid-fan settings, appear to be absent. The abundance of quartzitic clasts may reflect preferential preservation of erosionally resistant lithologies derived from a nearby continental basement. The abundance of gneiss clasts confirms the existence of a metamorphic basement. This unit may be broadly coeval with, but more distal than the deltaic, to shallow-marine mixed siliciclastic–carbonate unit (unit 2), and a moderate depth (e.g. outer shelf-type) rather than abyssal setting is preferred here.

The mélange unit includes pelagic carbonate, radiolarite and both within-plate basalt (WPB)-type and subduction-influenced basalts within a sheared, siliciclastic and volcaniclastic matrix. The age of the mélange is likely to be Mid-Triassic, of similar age to, or slightly younger than, the exotic blocks present.

The structurally highest unit, exposed only on the east coast (Unit 4), is interpreted as turbidites and mass flows derived from a continental source area, controlled by relative sea-level changes. A relatively proximal setting is particularly suggested by the presence of occasional relatively thick-bedded debris flows and high-density turbidites towards the top of the exposed succession. This does not support the suggestion that these sediments accumulated on the distal outer part of a deep-sea fan (see Stampfli *et al.* 2003), where thinner bedded and more hemipelagic sediments would be expected. The small scale and thickness of the sedimentary units is inconsistent with regional-scale settings such as deep-sea fan in a fore-arc basin, as implied in Model 2. The bright red–purple colour ('violet schists') probably reflects secondary oxidation of iron, possibly controlled by a high amount of organic matter originally deposited in these sediments (as noted in the Chemezi area).

Evidence from the Iraklion area, central Crete

The Phyllite–Quartzite unit is also extensively exposed in central Crete, west of Iraklion (Fig. 20), where two contrasting units are separated by an undated interval of coarsely crystalline marble (Vassilikon unit; Epting *et al.* 1972; Krahl *et al.* 1988). It is important to note that the lower of these units shows similarities to the Phyllite–Quartzite unit of western Crete, whereas the upper is more similar to the successions exposed in eastern Crete. This is one area in Crete where the two contrasting units of the Phyllite–Quartzite unit are known to occur together.

The lower part of the exposed succession, structurally above the Talea Ori–Plattenkalk unit, is dominated by soft-weathering, pale micaceous phyllite with alternating packages of medium- to thick-bedded sandstones and dark shale. Laterally continuous medium beds of soft-weathering, fine-grained micaceous sandstones are interbedded with dark phyllite. Several cliff-forming intervals of medium- to thick-bedded, laterally continuous quartzitic sandstones (in beds up to 3 m thick) include subordinate interbeds of black phyllite. The highest exposed levels beneath the Vassilikon marble comprise strongly cleaved grey phyllites with psammitic alternations. The gently north-dipping Vassilikon marble is virtually massive, with evidence of pervasive anastomosing shear zones in the lower part. A sharp contact with the underlying phyllites is indicative of a tectonic contact. The upper contact of the Vassilikon marble, exposed on the dip slope to the north, is also interpreted as tectonic. Above this, a contrasting Phyllite–Quartzite succession includes thin-bedded limestones, volcaniclastic sandstones and tuffaceous sediments, followed by massive, locally porphyritic andesitic lava and then further marble intercalations. Higher levels of the succession, exposed towards the coast, include massive andesitic lavas, lava breccias, volcaniclastic sandstones and tuffaceous sediments, >250 m thick.

Interpretation: additional rift-related settings

The lower Phyllite–Quartzite unit exposed in the south is lithologically comparable with the Phyllite–Quartzite unit in western Crete, as discussed above, although inferred sandstone turbidites are thinner bedded and the succession is more shaly overall, suggestive of a relatively distal setting. The undated Vassilikon marble might represent a slice of Hercynian basement, Triassic neritic limestones, or a strongly recrystallized fragment of later Mesozoic carbonate platform rocks, the last-mentioned being plausible as the marble is homogeneous, comparable with the Mesozoic Tripolitza carbonate platform. The overlying mixed terrigenous–volcanogenic succession includes andesitic lavas that have geochemical affinities with the Triassic volcanic rocks exposed in eastern Crete (Seidel 1978) and is interpreted as part of a similar rift-related setting.

Discussion of tectonic settings in eastern and central Crete

Most of the evidence from the Chemezi, Vai and Iraklion areas of the Phyllite–Quartzite unit is consistent with Model 1, in which all of the units, including the volcanic rocks, developed in a proximal to more distal rift setting, dating from the Permian, or earlier. In this interpretation the volcanic rocks relate to a pulse of Early–Mid?-Triassic extension. The pre-existing sediments were uplifted related to flexural uplift of the rifted margin during Mid-Triassic time, ushering in shallow-water to non-marine deposition adjacent to Hercynian continental basement during Late Triassic–earliest Jurassic? time. This could be same regional flexural effect that also affected the Phyllite–Quartzite unit in western Crete and the Talea Ori (Plattenkalk) unit. The possible cause of this inferred flexural uplift is considered in the discussion section near the end of the paper.

In Model 1, the contrasting exposures in the Iraklion area could be explained as different depocentres within a palaeogeographically varied rift setting. The Vassilikon marble might record the remnant of a former intra-rift high on which neritic carbonates accumulated, possibly during Late Triassic–later Mesozoic time. On the other hand, there is little evidence that the Iraklion outcrops record an actual Palaeotethyan suture, as in Model 2, as there is no evidence there of accreted oceanic material (deep-sea sediments or oceanic crust), or of any more intense deformation than seen elsewhere. Relatively intact successions are preserved in both the upper (eastern Crete-like) and lower (western Crete-like) thrust sheets there.

In Model 2, the Vai, Chemezi and Iraklion areas restore as a series of volcanic arc, basement (backstop), fore-arc and accretionary tectonic settings. The andesitic volcanic rocks and related sediments record part of the arc. The high-grade metamorphic basement slices represent part of the backstop of a subduction zone. The deep-water clastic and carbonate sediments ('Vai flysch') accumulated in a proximal to distal fore-arc basin and finally the mélange unit (eastern Vai area) represents a preserved fragment of an accretionary wedge created by northward subduction of Palaeotethys. The younger, coarser conglomerates and shallow-water carbonates accumulated in a post-collisional, transgressive setting in this model.

Some problems with Model 2 include the following.

(1) The inferred Permian–Lower Triassic 'fore-arc' sediments exposed in eastern Crete (i.e. Vai and Chemezi areas) contain terrigenous silt, but no arc-derived volcaniclastic sediment, as expected for a fore-arc basin. These sediments appear to have accumulated in a quiet, deep-water environment, unlike fore-arc basins that are typically unstable, and generally include turbidites, debris flows and slump deposits, rich in volcaniclastic debris.

(2) There is no definite record of any related accretionary wedge (e.g. trench-type sediments, or slices of oceanic crust). The blocks of within plate-type basalt within the Middle Triassic mélange unit in the eastern Vai outcrop could be related to Triassic rifting, rather than fragments of Palaeotethyan seamounts. 'Subduction–erosion' (e.g. von Huene & Scholle 1991) might account for the lack of an accretionary wedge. However, many comparable modern settings, including the eastern Mediterranean Sea south of Crete (Chaumillon *et al.* 1996) and the Gulf of Makran (Glennie *et al.* 1990) are associated with the development of accretionary prisms, which have a high potential for preservation in the stratigraphical record.

(3) No major Triassic magmatic arc, e.g. involving large-scale central-type volcanism, is known anywhere in the region. Triassic volcanic rocks that exhibit a subduction-related chemistry are interbedded with continentally derived subaqueous slope material, including coarse terrigenous debris flows and turbidites; there is no evidence of volcanic build-up, typical of continental margin arcs (e.g. Cascades, Andes). Also, air-fall tuffs typical of Andean-type margins (e.g. Andes) are sparse. The absence of large central-type volcanoes is surprising, as modern back-arc rifts typically develop by the splitting of developed volcanic arcs (e.g. the Marianas and Tonga arcs; and the SW Pacific; see Robertson 1994, for literature). The fact that both the Pindos and Vardar Triassic basins are inferred to be back-arc basins in Model 2 would lead one to expect the existence of a substantially developed magmatic arc, which in reality does not exist. On the other hand, the volcanic rocks are mainly andesitic and do show a subduction-related geochemical influence. This is consistent with Models 2, 3 and 4, but apparently not with Model 1. Possible reasons for this discrepancy are considered in the discussion and conclusion section.

(4) The expected >60 km width of crust between the inferred trench, arc, continental backstop and back-arc rift appears to be absent. For example, in the Vai area, all four of these units, as inferred by Stampfli *et al.* (2003), appear locally as thin thrust sheets one above the other. Even taking into account Cenozoic deformation, too many plate-scale processes are inferred in too small an area. In the convergent Model 2, tiny slivers from a range of tectonic settings (oceanic, fore-arc continental) originally at least tens of kilometres apart, were preferentially incorporated in a 'Cimmerian' thrust belt during pre-Jurassic time

(5) Collision of a 'Cimmerian' continent with the Eurasian margin during latest Triassic time would be expected to result in flexural collapse of the continental margin to form a regional, turbidite-filled foreland basin, yet neritic deposition prevailed, after a break in deposition during Mid-Triassic to Early Jurassic time.

(6) Perhaps aware of some of the above difficulties, Ziegler & Stampfli (2001) suggested that only a 'soft collision' and 'docking' took place of the Cimmerian continent with the Eurasian (Pelagonian) margin, leaving no perceptible stratigraphic or structural record. This was, however, contradicted by Stampfli *et al.* (2003), who inferred more pervasive collision and metamorphism.

However, there is, as yet, no firm evidence of a significant regional 'Eo-Cimmerian' compressional or metamorphic event dating from Late Triassic (Carnian–Norian) time in Crete or the Peloponnese (see below). For example, detailed structural studies in eastern Crete show that the main regional D_2 event affects both the Phyllite–Quartzite unit and underlying Plattenkalk (Lower Cenozoic Kalavros beds) and thus must be Alpine. D_1 is represented by rare east–west-trending, isolated folds and a bedding-subparallel cleavage (Zulauf *et al.* 2002) of uncertain origin. There is no obvious petrographic evidence of an Eo-Cimmerian metamorphic event (e.g. relict textures) of upper greenschist facies, or higher grade, despite reported conodont colour indices suggesting temperatures of *c.* 500 °C, i.e. *c.* 100° in excess of the temperature estimated for the reported regional HP–LT alpine metamorphism (Stampfli *et al.* 2003). However, the metamorphic grade of the lower thrust sheets (Plattenkalk, Tripali and Phyllite–Quartzite) varies considerably throughout Crete (e.g. Zulauf *et al.* 2002) and so this may not be a problem.

Evidence from the Peloponnese

The tectonostratigraphy of the Peloponnese is similar to that of Crete, with a similar pile of thrust sheets exposed in the same order. Most relevant outcrops are located in the southern and central Peloponnese, but there are also small isolated exposures in the NW Peloponnese (e.g. Zarouhla–Feneos area; De Wever 1975). Counterparts of the Plattenkalk and the Phyllite–Quartzite units, and possibly the Tripali unit, are present, and above this there are counterparts of two different facies-associations representing the Phyllite–Quartzite unit. In general, these units are less well dated than in Crete. As in Crete, the HP–LT metamorphic units are structurally overlain by a low-grade metamorphosed to unmetamorphosed shelf to carbonate platform succession.

There are, however, several differences between Crete and the Peloponnese, which make some discussion useful here in the attempt to discriminate amongst regional tectonic settings.

First, in Crete the Mesozoic Gavrovo–Tripolitza carbonate platform that overrides the entire underlying thrust stack is depositionally underlain only by a thin intact sedimentary succession (Ravdoucha beds). However, in the Peloponnese the equivalent Gavrovo–Tripolitza carbonate platform is stratigraphically underlain, albeit with a sheared contact, by a much thicker Triassic unit, known as the Tyros beds, which include both volcanic and terrigenous sedimentary units. These are critical to an understanding of the Triassic rift history of the Pindos basin to the north. Second, fragments of unmetamorphosed Palaeozoic sediments, known locally in the Peloponnese, could record part of a pre-existing continental basement. Third, there have been reports of a possible Palaeotethyan accretionary prism in the Peloponnese, which if correct would constitute important evidence for Model 2.

In the discussion below evidence from the equivalents of the Plattenkalk unit and equivalents of the western Crete Phyllite–Quartzite unit will be summarized, highlighting features that are relevant to understanding the tectonic setting, although a wealth of new information available warrants a fuller discussion elsewhere. It will be concluded in this section that most of the evidence is again consistent with the rift-related Model 1, although as in eastern and central Crete some of the Triassic volcanic rocks appear to be anomalous, as they record a subduction-related geochemistry.

Metasediments of the structurally lower units in the Peloponnese

Tectonostratigraphy

The Plattenkalk in the Peloponnese, as in Crete, is dominated by platy pelagic metalimestones with replacement chert, of inferred Jurassic–Cretaceous age, passing into Eocene flysch (Lekkas & Papanikolaou 1980; Papanikolaou & Skarpelis 1986). The typical Plattenkalk facies is underlain by poorly dated 'Permo-Triassic' phyllites, quartzite and conglomerates (Psonias 1981). According to some workers (Dittmar *et al.* 1989; Dittmar & Kowalczyk 1991) these facies form the stratigraphic base of the Plattenkalk unit. However, in the SW Mani Peninsula, marbles (Mani marbles) and associated metaclastic facies, correlated with the Plattenkalk, are reported to be locally tectonically inverted and thrust over the Phyllite–Quartzite unit (Alexopoulos & Lekkas 1999). This raises the possibility that metaquartzose sediments underlying the Plattenkalk generally in the SW Peloponnese could represent thrust slices of the Phyllite–Quartzite unit rather than a true stratigraphic basement, as some workers have previously assumed.

The Phyllite–Quartzite unit in the Peloponnese is traditionally divided into three 'nappes', although these are only partially exposed in individual areas and are separated by Cenozoic

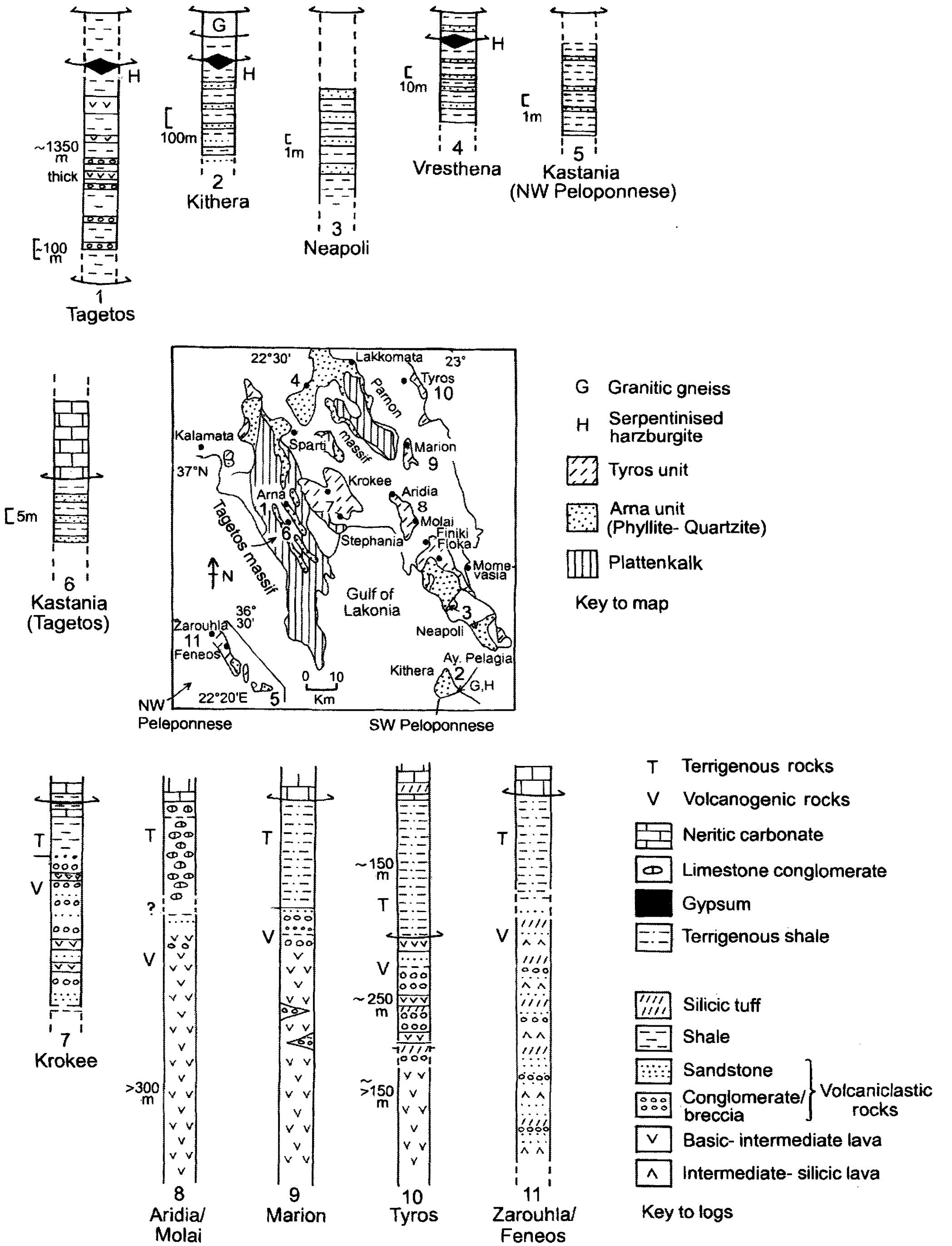


Fig. 21. Measured logs of the Plattenkalk unit, including the Arna unit, and the Phyllite–Quartzite unit (Tyros unit) in the Peloponnese with locations (inset). (See text for explanation and data sources.)

sedimentary basins (Ktenas 1924, 1926; Thiébault 1982; Triboulet & Bassias 1986). The 'lower nappe', dated as partially Triassic (Brauer *et al.* 1980; Doert *et al.* 1985), together with the Lower Cenozoic? meta-flysch (Papanikolaou & Skarpelis 1986), has a total structural thickness of *c.* 2000 m in the Parnon massif and *c.* 1300 m in the Taygetos massif (e.g. Doutsos *et al.* 2000; Xypolias & Doutsos 2000). This unit has experienced intense deformation and HP–LT metamorphism (*c.* 400°C and 10 kbar), although mineral assemblages indicative of *P–T* conditions are commonly lacking; also *P–T* conditions may vary regionally as in Crete. The Phyllite–Quartzite unit is dominated by meta-mudrocks and metaquartzitic sandstones. However, exposures in the Tagetos Mountains, termed the Arna unit, include metashales (phyllites), metaquartzose conglomerates, MORB-type tholeiitic metabasalt and rare harzburgite (Skarpelis 1982). Occasional occurrences of harzburgite and metabasic igneous rock occur elsewhere in the Peloponnese (Thiébault 1991; see below). In NE Kythira (Fig. 1), the Phyllite–Quartzite unit is associated with granitic gneiss of Hercynian age (Xypolias *et al.* 2006). Also on Kythira, according to Stampfli *et al.* (2003), Danamos (1992) has reported the presence of metabasic rocks and metacherts (lydites) that are said to exhibit primary contacts; if correct, this could be indicative of an origin as accreted Palaeotethyan oceanic crust, consistent with Model 2 (Stampfli *et al.* 2003). The ultramafic rocks and basic lavas within the Phyllite–Quartzite unit generally (including the Arna unit) have been interpreted to include exotic accretionary material (Skarpelis 1982), and could thus represent part of a Palaeotethyan accretionary complex (Stampfli *et al.* 2003). An alternative view is that these thrust-intercalated mafic–ultramafic rocks were derived from the Pindos ocean to the east, in the Cycladic region, during Early Cenozoic Alpine deformation and metamorphism (Papanikolaou 1996–1997; Pe-Piper & Piper 2002).

In the literature a 'middle nappe' of the Phyllite–Quartzite unit is reported to comprise meta-clastic rocks, metacarbonates, metashales and sparse metavolcanic rocks, of possibly Carboniferous, Permian and Triassic age (Thiébault 1982; Triboulet & Bassias 1986; Bassias & Triboulet 1994). Lithologies of no higher metamorphic grade than greenschist facies attributed to this unit are reported to be exposed in areas including the NW Peloponnese (Zarouhla and Feneos areas; De Wever 1975; Pe-Piper 1983), the SE Peloponnese (near Molai) and near Kalamata (Verga). However, these successions (Fig. 21) remain poorly dated and are reported to be overlain by carbonates similar to the Gavrovo–Tripolitza unit (i.e. without any overlying 'upper nappe'), thus questioning the reality of a 'middle nappe' as a regionally significant tectonic unit.

The mainly volcanogenic 'upper nappe' (Tyros beds), of greenschist-facies metamorphic grade, is widely exposed and less deformed. Dating is primarily based on an inferred depositional passage upwards into well-dated shallow-marine carbonates of the Gavrovo–Tripolitza zone of Late Triassic (locally Carnian) age (Thiébault & Kozur 1979; Lekkas & Papanikolaou 1980; Skarpelis 1982). Successions in the SW Peloponnese (Stephania–Krokee area) are commonly assumed to be Early Triassic in age (Doert *et al.* 1985), but remain poorly dated. In the SE Peloponnese volcanic rocks are reported to be interbedded with carbonates and clastic sediments that are well dated as Carnian in age (Thiébault & Kozur 1979; Brauer *et al.* 1980). According to Gerolynatos (1994) the Tyros unit as a whole comprises a Permian to Upper Triassic, mainly sedimentary succession with volcanic intercalations mainly of Scythian and Late Triassic (Carnian–Norian) age. However, it is questionable whether an intact stratigraphic succession is anywhere present.

The volcanogenic rocks range from basalts and basaltic andesites to dacites in different areas, with commonly pyroclastic and tuffaceous sediments, previously believed to be mainly non-marine, together with minor intrusive rocks (Pe-Piper 1983). Extensive geochemical studies have showed that the Triassic volcanic rocks are largely calc-alkaline, but locally tholeiitic, to alkaline in composition (Pe-Piper 1983; Pe-Piper & Piper 2002; see Degnan & Robertson 2006). The tectonic settings of eruption were seen as either intra-continental extension-related (Dornsiepen & Manutsoglu 1996), or subduction-related (Pe-Piper & Piper 2002).

Plattenkalk (Mani unit)

The lowest unit, the Plattenkalk, locally termed the Mani unit, is structurally overlain, first by the HP–LT Phyllite–Quartzite unit (including the Arna unit) and then by the LP–LT Tyros unit, followed in turn by a deformed sedimentary transition to the Gavrovo–Tripolitza unit. These two units are separated by a regional low-angle tectonic contact, interpreted in different areas as a thrust fault (related to subduction), or a low angle-extensional detachment (related to exhumation).

The base of the Plattenkalk unit, as exposed in the eastern Tagetos (e.g. near Kastania, south of Arna; Fig. 21, log 6), is a low-angle tectonic contact, underlain by thick-bedded quartzose sandstones and phyllites that are lithologically very similar to the Phyllite–Quartzite unit of western Crete or the SE Peloponnese (see below). These rocks are thus unlikely to represent a true stratigraphic basement (Doert & Kowalczyk 1985), or to be equivalent to the shallow-water carbonates of the Talea Ori unit in Crete (i.e. Sisses and Fodele units).

Phyllite–Quartzite unit

The Phyllite–Quartzite unit shows considerable lithological variation throughout the Peloponnese, as follows.

Metasediments. In most areas the succession is dominated by quartzose sandstones, shales (locally carpholite-bearing) and subordinate quartzose conglomerates. The sandstones are most thickly bedded and coarse grained in the far SE (i.e. on Kythira) and in the Napoli area (Fig. 21, logs 2 and 3). However, sandstones are generally thinner, finer grained and more thinly bedded further north (e.g. Vresthena; Fig. 12, log 4) and in the NW Peloponnese (Fig. 21, log 5), although even in these areas occasional intercalations of relatively thick-bedded (>0.5 m) and coarse-grained (conglomerate grade) sandstones are present. In all cases, the sandstones are mainly quartzitic and comprise well-rounded grains, as seen in western Crete. The Arna unit in the type area of the Tagetos Mountains (Fig. 21, log 1) is unusual, as the background grey or black (chloritoid-bearing) phyllites include numerous conglomerate intercalations composed almost entirely of quartzite clasts, ranging from clast-supported to locally matrix-supported conglomerates. Clasts (up to 80 cm in size) vary from angular to rounded, and locally to well rounded. Conglomeratic horizons, up to *c.* 20 m thick, can be traced laterally for hundreds of metres along strike, but no intact succession can be recognized.

Metabasic igneous rocks within the Arna unit. In the type Arna unit in the Tagetos, the metasediments (mainly phyllites and quartzite conglomerates) are locally intercalated with metabasic rocks of MORB type (Skarpelis 1982). At one locality (Malevos, near the Neohori–Georgitza road; see Papanikolaou & Skarpelis 1986) metabasalt, *c.* 100 m thick occurs within dark pelitic metasediments but the upper and lower contacts are poorly exposed. Further south (e.g. near Gorani, 14 km south of Arna village) metabasic rocks include definite volcanic breccias and appear to be interbedded with metasediments, including quartzitic conglomerates; some meta-sills may also be present. Individual, metabasic units are mapped as lenses up to 4 km long (see Papanikolaou & Skarpelis 1986). In addition, a small exposure of glaucophane-bearing metabasic rocks (several metres thick) with WPB chemical affinities occurs in the NW Parnon massif (near Lakkomata), associated with small bodies of serpentinized ultramafic rock (Tribolet & Bassias 1986).

Metaserpentinite. The Phyllite–Quartzite unit includes small bodies of serpentinized ultramafic rocks at five well-documented localities, the first three of which mentioned below were studied in the field. All of these occur near the overlying tectonic contact with the Gavrovo–Tripolitza unit and appear to have undergone similar HP–LT metamorphism and deformation as the enclosing metasedimentary rocks. First, in the Tagetos massif, the Arna unit includes a lenticular, north–south-trending exposure of antigoritic harzburgite, several hundred metres long by several tens of metres wide (Skarpelis 1982; Fig. 21, log 1). Second, on Kythira, harzburgite is located (south of Ayia Pelagia; Fig. 21, log 2) within terrigenous metasediments, near the contact with the overriding Gavrovo–Tripolitza carbonate platform. Third, in the NW Parnon massif (at Vresthena; Fig. 21, log 4) serpentinized harzburgite forms a lens (*c.* 10 m thick) within terrigenous sediments. This is located several tens of metres beneath the overriding Gavrovo–Tripolitza platform carbonates. Fourth, a smaller body (a few metres) elsewhere in the Parnon Massif (Agios Petra) is mapped as occurring directly along the tectonic contact of the Arna unit with the Gavrovo–Tripolitza carbonates (Skarpelis 1982). Finally, a small body of sheared serpentinized ultramafic rocks (possibly including dunite) is associated with phyllites and minor metabasic igneous rocks in the NE Parnon massif, at Lakkomata (Triboulet & Bassias 1986).

Granitic gneiss. On the island of Kythira (Fig. 5) the Arna unit is associated with a small body of granitic gneiss (near Ayia Pelagia) that was recently shown to be of Hercynian age (Xypolias *et al.* 2006), similar to eastern Crete (see Romano *et al.* 2006). Metagranitic rocks, affected by neotectonic extensional faulting, occur near the coast, just north of Agios Pelagia; these are rocks locally overlain by carbonates of the Gavrovo–Tripolitza unit and are in faulted contact with unmetamorphosed, Triassic? sandstone, shale and limestone of the Pindos zone.

Interpretation: a rifted continental basement

The quartzitic sandstones were derived from a continental setting, as in western Crete. Also, the Upper Palaeozoic gneissic rocks exposed on Kythira are suggestive of the existence of a Hercynian basement, as in eastern Crete.

The serpentinized ultramafic rocks (mainly harzburgites) are likely to represent remnants of oceanic mantle (either oceanic or sub-continental). The MORB-type rocks in the Tagetos appear to be at least partially interbedded with metasedimentary rocks, rather than entirely exotic units. These associated sediments include quartzose conglomerates that are lithologically very similar to the Mani unit of western Crete, of inferred latest Triassic age. These volcanic rocks might conceivably relate to opening of the adjacent Pindos ocean during the Late Triassic. However, an origin as allochthonous thrust slices related to Alpine deformation cannot be ruled out in view of the strong deformation and high pressure metamorphism (Skarpelis 1982). In addition, the WPB-type metavolcanic rocks in NW Parnon (Lakkomata) might be of oceanic origin (seamounts), as they are associated with serpentinite, although Thiébault (1991) favoured an intracontinental origin. Indeed, the ultramafic rocks in all cases occur enclosed within terrigenous clastic sediments and lack evidence of significant amounts of other possible ophiolite-related rocks (e.g. cherts, gabbro, etc.). Also, the existence of metacherts and metabasic volcanic rocks on Kythira was not confirmed.

In all cases the meta-ultramafic rocks occur in lenses or pods just beneath the major thrust fault or extensional detachment at the base of the Gavrovo–Tripolitza carbonate platform. They are thus entrained within or close to a profound tectonic and metamorphic discontinuity. It is, therefore, probable that these lithologies represent exotic units that were exhumed from a deep subduction environment related to exhumation of the HP–LT Mana (Phyllite–Quartzite) unit. It is possible that the serpentinites ultimately originated as fragments of subducted Pindos ocean (in the Cycladic region) that were later exhumed as diapiric pods in response to deep-seated out-of-sequence thrusting. There is, thus, no clear evidence that the ultramafic rocks record parts of a Palaeotethyan accretionary complex, which is otherwise not supported by field evidence within the Phyllite–Quartzite unit in the Peloponnese.

Summarizing, the structurally lower HP–LT unit in the Peloponnese are consistent with a rift-related setting, as in Models 1 and 3. The claimed evidence of a Palaeotethyan accretionary complex, apparent evidence supporting Models 2 or 3 (e.g. on Kythira island), can now be discounted.

Evidence from the structurally higher units in the Peloponnese

Triassic volcanic–sedimentary Tyros unit

During this work it was found that an overall succession in this unit divides into a lower part, which is mainly volcanogenic, and an upper part, which is mainly terrigenous. The ‘intermediate nappe’ is here not considered to be a regionally significant unit (see below), but can instead be correlated with the traditional ‘upper nappe’. The lower part of the volcanogenic succession is dominated by massive lavas, with, in addition, subordinate volcanic rudites (breccias and conglomerates; e.g. Aridia; Fig. 21, log 8). The upper part of the volcanogenic succession is commonly more varied and includes numerous thick (up to tens of metres) intercalations of poorly sorted volcanogenic rudites, volcaniclastic sandstones, occasional shales (typically pink coloured) and silicic tuffs (e.g. Tyros, Fig. 21, log 10; Marion, log 9). The succession in the SW Peloponnese (Krokee area) is mainly volcaniclastic (Fig. 21, log 7), although there exposure is limited by low topography; potentially deeper levels of the succession are not exposed. In the NW Peloponnese (e.g. at Zarouhla, Feneos and Kastania; Fig. 21) tuffaceous sediments and volcaniclastic sediments are relatively more abundant, together with common intermediate to silicic composition lava flows.

The upper part of the Tyros unit, where exposed, comprises strongly contrasting terrigenous sediments, mainly lithologically homogeneous quartzitic shales and mica-schists. The contact with the underlying volcanogenic unit is typically sheared, but is interpreted here as an deformed normal contact rather than a major tectonic break. The upper terrigenous unit is relatively thick (>100 m) in most areas (e.g. Tyros, Marion and NW Peloponnese; Fig. 21, logs 9–11). Elsewhere, the upper terrigenous unit is much thinner (tens of metres at Krokee; Fig. 21, log 7), down to only several metres (e.g. near Aridia and Floka). However, parts of the original succession may have been tectonically removed. Locally, evaporite (gypsum) has been reported from the upper terrigenous unit near the contact with the overlying Gavrovo–Tripolitza platform carbonates (e.g. Krokee and near Verga, Kalamata area; N. Skarpelis, pers. comm).

The upper terrigenous unit, or locally the volcanogenic unit (e.g. near Aridia and Floka), is overlain by neritic carbonates, commonly stromatolitic, forming the base of the Mesozoic Gavrovo–Tripolitza carbonate platform succession. In all areas the contact is moderately to strongly sheared, with much evidence of layer-parallel extension and other features indicating at least partial detachment from the underlying Tyros unit. However, in some areas (e.g. south of Aridhia; Tyros) facies are transitional indicating that a normal contact was originally present (Lekkas & Papanikolaou 1980).

The 'intermediate' nappe is problematic. In the NW Peloponnese a two-part volcanogenic–terrigenous succession, as elsewhere, is structurally overlain by the Gavrovo–Tripolitza carbonate platform, providing no basis for the recognition of a separate, higher tectonostratigraphical unit. A large exposure in the SE Peloponnese (south of Aridia) includes folded phyllites, mica-schists and limestone conglomerates with subordinate volcanic intercalations. Local contacts are not well exposed. However, directly east of Molai (Fig. 21, log 8) comparable limestone conglomerates appear to pass depositionally into the Gavrovo–Tripolitza platform carbonates. On the other hand, well-dated meta-clastic 'Permo-Carboniferous' sediments with Late Carboniferous sporomorphs structurally underlie the Gavrovo–Tripolitza carbonate platform in the SE Peloponnese (south of Monemvasia; Paraskevopoulou 1951; Fytrolakis 1971; N. Skarpelis, pers. comm), suggesting that the Triassic volcanic rocks in this area could have a continental basement. It, therefore, seems likely that the 'intermediate nappe' is a composite unit including lithologies underlying, laterally equivalent to, and overlying the overall volcanogenic–terrigenous succession.

Interpretation: rift settings

The Triassic Tyros unit volcanogenic succession formed in a regionally extensive, subaqueous rift setting. Fragments of a sedimentary basement may be represented by the occurrences of Palaeozoic terrigenous and calcareous sediments (south of Monemvasia; Fytrolakis 1971). The rift basin was partially filled by flood basalt. The presence of abundant volcaniclastic sediments, largely subaqueous debris flows, with little terrigenous material, is suggestive of mass wasting on a subaqueous fault scarps. More fractionated (intermediate–silicic) volcanism predominated in the NW Peloponnese (Feneos–Zarouhla). After volcanism largely ended the inferred rift was covered by terrigenous muds and shallowed, culminating in the accumulation of varied carbonates, organic-rich muds and local evaporites. The presence of volcanogenic horizons interbedded with transitional neritic carbonates (e.g. at Tyros) is suggestive of volcanism during the Carnian. This volcanism can be directly related to the opening of the Pindos ocean basin to the NE (in present coordinates). Where locally intact, the succession passes transitionally upwards into the Gavrovo–Tripolitza platform of Late Triassic–Early Jurassic age. Localized limestone conglomerates beneath the carbonate platform (i.e. west of Molai) are indicative of mass wasting of a nascent carbonate platform, as the pre-existing rift-related accommodation space was filled prior to regional covering by a thick Bahama-type carbonate platform.

The Phyllite–Quartzite unit in the Peloponnese is generally similar to counterparts in western and central Crete. The intercalations of quartzitic conglomerates in the Arna unit (Tagetos massif) are lithologically similar to the Mana conglomerate in western Crete, of inferred latest Triassic–earliest Jurassic? age there. The clastic sediments and carbonates exposed in the transition between the Tyros unit and the Tripolitza platform are similar to the Ravdoucha Beds of western Crete, although Late Triassic volcanic rocks are not exposed in the latter unit.

Despite differences in metamorphic grade (relatively high grade in eastern Crete, but lower grade in the Peloponnese) the Tyros unit shows some similarities to the Phyllite–Quartzite unit of eastern Crete (Vai–Chemezi areas) and central Crete (upper structural unit). In both the Peloponnese and eastern Crete, intact succession include a thick basaltic–andesitic volcanogenic sequence (lavas and volcaniclastic sediments) that passes upwards into shallow-water carbonates of Late Triassic–earliest Triassic age, correlated with the Tripolitza carbonate platform. In addition, slices of Hercynian granitic gneiss occur locally in both areas. However, conglomerates (with abundant basement-derived material) are much more extensive in eastern Crete than in the southern Peloponnese. Also, the Late Triassic volcanism in the Peloponnese is unknown in Crete (although dating remains limited). Similar volcanic rocks of the Tyros beds locally underlie the most proximal of the Pindos–Olonos nappes (Degnan & Robertson 1998, 2006), confirming that the Late Triassic Tyros volcanic rocks relate to opening of the Pindos ocean.

Most of the evidence, outlined above, is consistent with Model 1, as for Crete. However, the presence of subduction-influenced Triassic volcanic rocks could also be consistent with Models 3 and 4, which invoke a southward-dipping subduction zone, although as noted below there is little or no evidence independent of geochemistry that such a south-dipping subduction zone existed in the Triassic.

Evidence from the Pindos zone

Additional relevant evidence comes from the regionally overlying Pindos zone, which is fragmentary in Crete but better exposed in the Peloponnese (Fig. 5) and in Greece further north. The Gavrovo–Tripolitza platform was overthrust by the relatively unmetamorphosed Pindos unit during Early Cenozoic time (e.g. Bonneau 1984; Jacobshagen 1986; Papanikolaou 1996–1997). Much evidence already exists in the literature, which can be used to test the alternative tectonic models.

In Model 1 (divergence-related), the Pindos zone originated as a continental rift in the Triassic (Dercourt *et al.* 1986) but then developed into a subsiding passive margin as the Pindos ocean opened (Smith *et al.* 1975; Robertson & Dixon 1984; Robertson *et al.* 1991). The Pindos thrust sheets restore as an east-facing deep-water slope to abyssal plain (Degnan & Robertson 1998) that probably accumulated on 'transitional' crust within a continental–ocean transition zone (Degnan & Robertson 2006).

In Model 2 (convergence-related) the Pindos ocean originated as a Late Triassic back-arc basin related to the later stages of northward subduction of Palaeotethys (Stampfli *et al.* 2003). This subduction culminated in the collision of a rifted 'Cimmerian' fragment with a Eurasia-related unit represented by the Pelagonian zone during Late Triassic (Carnian–Norian) time. In principle, any such Cimmerian suturing related to northward subduction need not have affected a related marginal basin to the north, which could have remained isolated. However, Stampfli *et al.* (2003) specifically argued that a collision-related compressional event is, indeed observed within the Pelagonian zone further north; this implies that stress was transmitted across the Pindos deep-sea basin from a suture zone to the south to a Pelagonian continent to the north.

Is such a compression-related event actually recorded in the Late Triassic sedimentary fill of the Pindos basin? A regional 'Cimmerian' suturing to the south could have resulted in uplift and increased supply of clastic sediment to the basin during latest Triassic–earliest Jurassic time. Also, if the basin was internally deformed, sediment redeposition, slumping, or an intra-basin unconformity might be present: however, none of these features are apparent within the Triassic–Early Cenozoic Pindos succession (Degnan & Robertson 1998). The field evidence instead supports continuing passive margin subsidence of the Pindos basin from Late Triassic to Early Cenozoic, punctuated by clastic influxes that can be mainly related to the effects of eustatic sea-level change.

The Pindos thrust sheets are locally underlain and intercalated with a mélange including blocks of volcanic rocks, some of which are dated as Triassic from associated sediments. Discrete thrust sheets including Triassic volcanic rocks are also locally present (see Pe-Piper & Piper 2002; Degnan & Robertson 2006). Extensive geochemical studies indicate that the Triassic igneous rocks commonly show a geochemical subduction influence that could be consistent with Models 2, 3, or 4. In addition, some 'enriched' basalts are present that could represent fragments of emplaced seamounts (Degnan & Robertson 2006). An origin related to a northward-dipping subduction zone (Models 2 and 4) is, however, unlikely as no ocean to the south has been identified, as discussed earlier.

Pe-Piper & Piper (1998, 2002) have invoked an additional Triassic subduction zone (an intra-oceanic one that dipped southwards) to explain, in particular, localized occurrences of high-magnesian andesites (boninites) and plagiogranites. In this interpretation (Model 4) subduction would have culminated in collision of a trench with a Pelagonian passive margin to the north. This would have been expected to emplace Triassic (and younger) ocean crust (ophiolite) over a Triassic or younger accretionary prism. However, the overriding ophiolites are Mid-Jurassic in age (Liati *et al.* 2004) and underlying accretionary units document a Triassic rifted margin (e.g. in Evia, Othris and Pindos; Robertson *et al.* 1991). There is thus no sedimentary or structural evidence for southward Triassic intra-oceanic subduction, as in Model 4. On the other hand, the presence of Mg-andesites locally that imply remelting of previously depleted mantle, clearly needs an explanation.

Evidence from the Pelagonian zone

The Pelagonian zone, in turn, structurally overlies the Pindos zone; it is restricted to high-level fragments in Crete and the Peloponnese but is much more intact and widely present in central and northern Greece. Key areas include those NE of Athens, such as in Evia, where the Pelagonian zone has experienced only low-grade metamorphism, in contrast to northern Greece where the grade is higher (Mountrakis 1986). The Pelagonian zone comprises a pre-Triassic continental basement, which includes Upper Palaeozoic granites. Transgressive platform carbonate deposition was punctuated by ophiolite emplacement and deformation during Late Jurassic–Early Cretaceous time (see Rassios & Moores 2006). This was followed by renewed platform deposition until emplacement as part of the Hellenide nappe pile during Early Cenozoic time (Mountrakis 1986).

In Model 1 (divergence-related) the Pelagonian zone is interpreted as a microcontinent rifted from Gondwana in the Triassic (Dercourt *et al.* 1986) related to opening of a Pindos ocean (Robertson *et al.* 1991). In Model 2 the Pelagonian zone is interpreted as a microcontinent that was rifted from Eurasia related to opening of a Vardar back-arc oceanic basin to the north and a Pindos back-arc basin to the south (De Bono *et al.* 1998; Vavassis *et al.* 2000; Stampfli *et al.* 2001). In Models 3 and 4 the Pelagonian zone is seen as a microcontinent that was detached from Gondwana associated with opening of a back-arc marginal basin, over either a south- or a north-dipping slab.

Evidence to test the above alternatives mainly comes from the western (Pindos) and eastern (Vardar) margins of the Pelagonian zone. The evidence from the western Pelagonian margin is clearly consistent with Model 1, as there is evidence of Triassic rift-related sedimentation and alkaline volcanism (Mountrakis 1986), as is well exposed in the Othris area (Smith *et al.* 1975). Available evidence from the eastern margin of the Pelagonian zone (Vardar margin) also points to the existence of a Triassic rifted margin (see Sharp & Robertson 2006). Triassic basalts in the Vardar zone (i.e. within the Eastern Almopias zone) lack geochemical evidence of a subduction influence, opposing Models 2 and 3, in which the adjacent Vardar basin is seen as an above-subduction zone back-arc rift or oceanic basin.

In presenting evidence that, if valid, would support Model 2, Stampfli *et al.* (2003) argued that the Pelagonian zone experienced a pulse of regional 'Cimmerian' compression related to suturing of Palaeotethys to the south in latest Triassic time. Stress from this collision was transmitted across the Pindos, inferred back-arc marginal basin, triggering a stratigraphic inversion event within the Pelagonian zone during latest Triassic time. Stampfli *et al.* specifically argued that a Permian–Triassic rift succession exposed on the island of Evia (Fig. 5), termed the Liri unit, experienced compression-related uplift, associated with mass-wasting of 'olistostromes', and that this was then unconformably overlain by a Jurassic carbonate platform (Stropones Limestone). During the present work, this interpretation was tested in the field and it was found that the evidence is instead consistent with the extension-related Model 1.

The Liri unit is divisible into southerly and northerly exposures, separated by an inaccessible mountainous area (Fig. 22a). The Liri Unit has experienced greenschist-facies metamorphism and extreme layer-parallel extension, with the development of ubiquitous 'phacoidal' fabrics. Sedimentary structures (e.g. grading) are relatively well preserved, especially in the higher stratigraphic levels, and show that the sequence is mainly the right way up. The Jurassic shallow-water Stropones Limestone ('cover unit') is, in fact, located structurally beneath rather than above the Liri Unit; consequently, the Liri unit lacks any preserved overlying depositional cover in this area (Fig. 22b and c). The Liri unit is instead structurally overlain, above a major low-angle thrust contact, by a regionally extensive Pelagonian thrust sheet. The local Pelagonian sequence, of Late Permian? to Mid-Triassic age, includes metasiliciclastic sandstones, shale, ribbon chert, redeposited carbonates (including debris flows), andesitic–rhyolitic metavolcanic rocks and tuffaceous–volcaniclastic sediments, consistent with a rift-related origin. The succession passes upwards into a several-kilometre-thick unit of platform carbonates of Late Triassic–Jurassic age, typical of the Pelagonian zone generally (Fig. 22d). This overall succession is stratigraphically underlain by schists and granitic rocks ('Hercynian basement') and coarse 'basal' clastic sediments derived from these lithologies (Fig. 22b).

Petrographic study (19 samples) shows that the meta-sandstones of the lithic unit are mainly arkoses and lithic arkoses, mainly derived from granitic and metasedimentary lithologies, as widely exposed within the 'Hercynian' basement beneath the Jurassic platform carbonates throughout the Pelagonian zone (Mountrakis 1986).

The Liri unit was mainly deposited by turbidity currents and mass-flow processes that were active during an inferred Permo-Triassic rift setting. However, there are few indications of water depths, which could have been relatively shallow (tens to several hundred metres). Radiolarian cherts or other evidence of pelagic deposition are absent. Some localized 'cherts' represent secondary alteration, of possibly hydrothermal origin.

The uppermost part of the Liri unit, mostly <10 m below the overriding Pelagonian thrust sheet, includes scattered small outcrops of highly fossiliferous shallow-water carbonate (Fig. 22b). This limestone is well dated as Late Carboniferous–Late Permian based on shallow-water calcareous fossils (e.g. benthic foraminifera) (Stampfli *et al.* 2003). These limestones apparently represent fragments of a long-lived

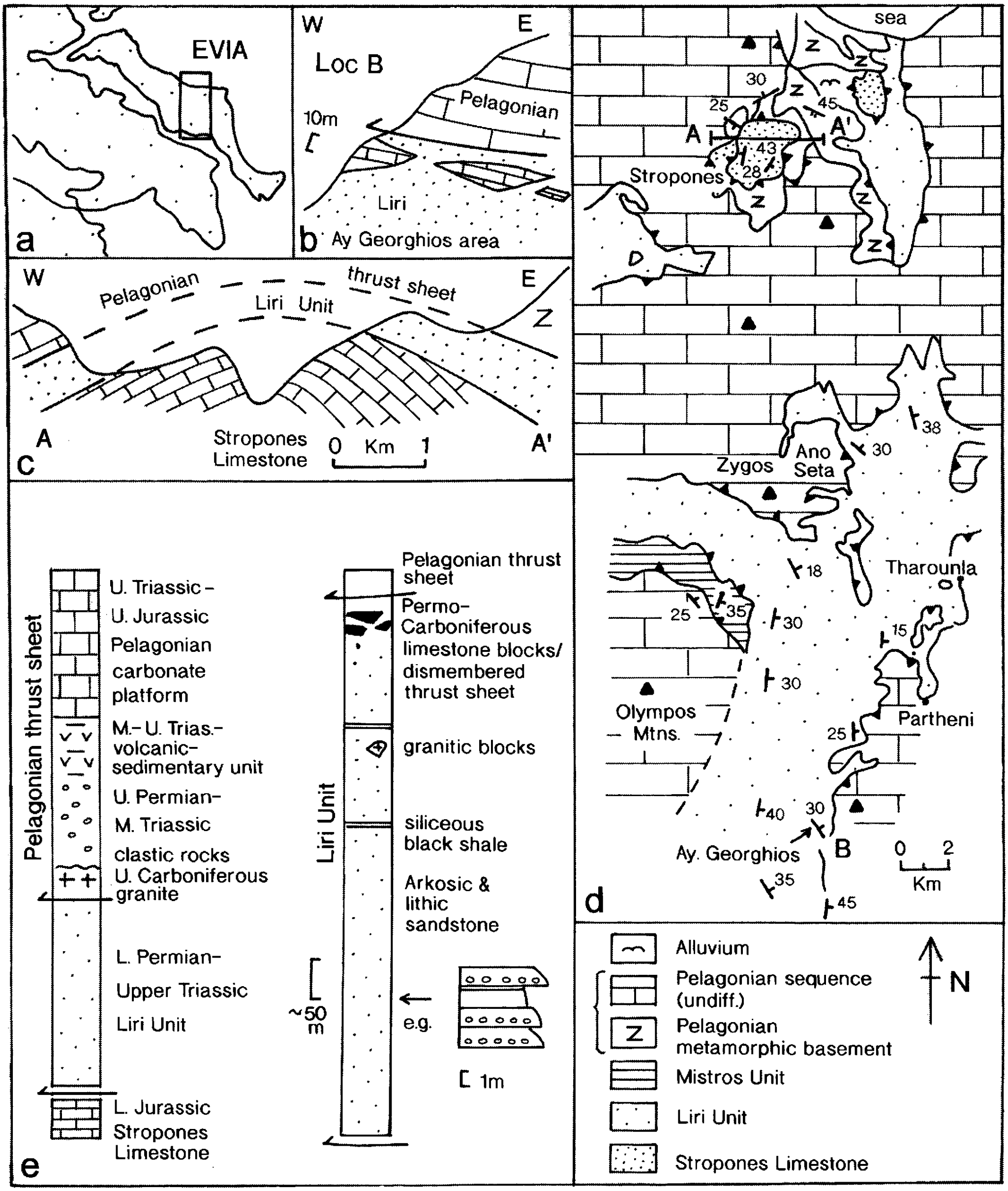


Fig. 22. Upper Palaeozoic–Lower Mesozoic sedimentary and volcanic units exposed on the Island of Evia. (**a**) Regional location; (**b**) Cross-section at location B (see d); (**c**) Cross-section at A–A; (**d**) Outline geological map; (**e**) Tectonic-stratigraphy in this area (left) and specific sedimentary log of the Liri Unit (right). (See text for explanation and data sources.)

carbonate platform that was possibly constructed on Hercynian basement within the Pelagonian zone. Individual blocks are typically less than a metre to several tens of metres in size. Small outcrops of distinctive dark grey shallow-water limestone can be traced for tens to hundreds of metres along strike, suggesting the existence of dismembered sheets, in addition to detached blocks. Where smaller blocks (several metres across) are seen lower down the sides of valleys these are commonly landslipped.

Measurements of bedding dip within the limestone blocks and sheets show that the bedding is everywhere moderately inclined

(Fig. 22d), subparallel to the tectonic contact with the overlying Pelagonian thrust sheet. Where the contact with the shale–sandstone matrix is rarely exposed this is seen to be a sharp tectonic contact. The margins of the limestone blocks are commonly brecciated and calcite veined. The adjacent matrix sediments do not contain sedimentary fragments of the same limestone (although some entrained phacoidal fragments are locally present). Petrographic study of the sandstone matrix enclosing the limestone blocks did not reveal sedimentary limestone clasts, but rather the composition remains unchanged from the underlying sandstones.

Interpretation: rift basin deformed in Early Cenozoic time

In Model 1 the exotic blocks and sheets were shed into a rift basin related to rifting (i.e. rift shoulder uplift) that immediately preceded break-up to form the Pindos ocean in Late Triassic time (Fig. 23b). In Model 2 (convergent margin–collision) the upper part of the Liri unit is viewed as an 'olistostrome' containing blocks shed into a back-arc basin, an event that was triggered by stratigraphic inversion related to 'Eo-Cimmerian' orogeny in latest Triassic time (Fig. 23a). However, as noted above there is no evidence of a critical Jurassic sealing unconformity, which is essential to validate this interpretation. Without this, there is no evidence of pre-Jurassic deformation and instead the deformation is likely to be Cenozoic, associated with the SW-directed regional emplacement of Pelagonian thrust sheets.

It is also possible that neither of the above alternatives is correct and that the 'olistostrome' instead relates to Cenozoic overthrusting. In this interpretation (Fig. 23c) the limestones could represent the dismembered remnants of a thrust sheet of neritic limestone of Pelagonian affinity that was entrained, together with the overriding Pelagonian thrust sheet, during regional Early Cenozoic deformation. Several observations are consistent with this interpretation: first, most of the blocks occur in highly sheared shales just beneath the overlying Pelagonian thrust sheet; second, many blocks join up as larger dismembered sheets at the same structural level; third, the dips are parallel to the overlying Pelagonian thrust sheet and are not variable as expected for a debris flow (olistostrome) origin; finally, there is an absence of associated limestone-derived debris flows and other gravity-flow deposits within the Liri unit, in contrast to typical large-scale debris-flow deposits (olistostromes).

In summary, it is concluded that there is no evidence for latest Triassic 'Cimmerian' compression within the Pelagonian zone, as implied by Model 2.

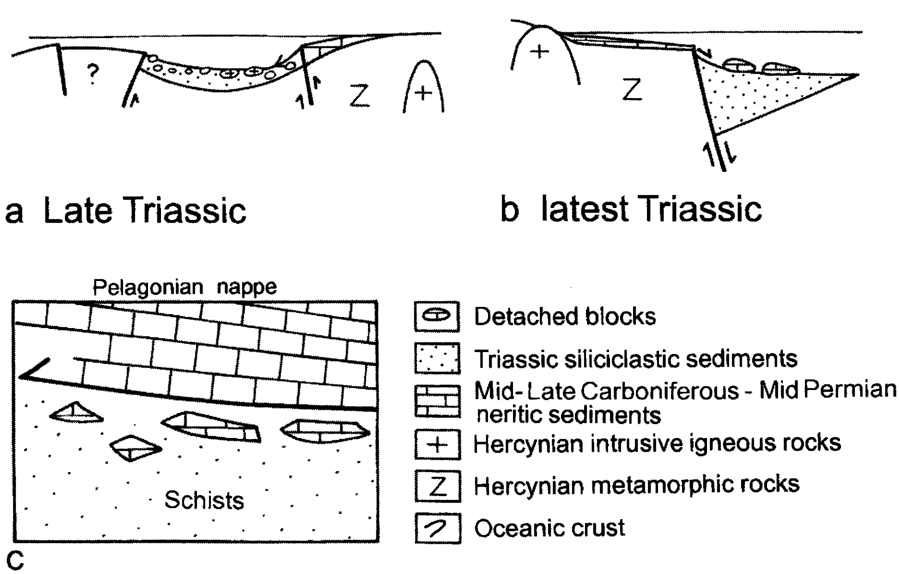


Fig. 23. Alternative tectonic setting for the Triassic of central Evia (Pelagonian zone). (**a**) Compression (stratigraphic inversion interpretation) (Stampfli *et al.* 2003); (**b**) rift-related interpretation; (**c**) formation by layer-parallel extension (boudinage) of a thrust sheet beneath the Pelagonian nappe during the Cenozoic. A thrust-related interpretation is favoured, as discussed in the text.

Possible objections to rift-related settings

Although the rift-related Model 1 explains most features of the geological evolution, there are several potential arguments against this interpretation, which are discussed below. However, each of these can be countered, as will be seen.

(1) The Mid-Carboniferous–Early Triassic basins, as documented in Sicily and western Crete, were sufficiently deep (c. 500 m or more) (Kozur 1993, 1995) to require the existence of a contemporaneous ocean basin even if no oceanic crust is preserved

There is, however, no requirement for an oceanic basin to have existed adjacent to Gondwana during Late Palaeozoic time, as in Models 2 and 4. Similar broad, deep basins existed widely around the margins of the Atlantic prior to spreading. These include the Jean d'Arc basin off Newfoundland (Reid & Keen 1990), the Hatton Bank and adjacent basins off Ireland (Fowler *et al.* 1989), the Lusitanian basin off Portugal (Wilson 1988), and comparable rift basins bordering the Central and South Atlantic and Indian oceans, and the Red Sea (Purser & Bosence 1998). There are also numerous examples of deep-water rifts marginal to now-sutured oceans that are exposed on land, notably around the Western Alps (Lemoine *et al.* 1986).

(2) No viable mechanism for the Triassic rifting of continental fragments exists other than back-arc extension

If true, this would favour Models 2, 3 and 4 over Model 1. However, similar rift settings are known from non-emplaced passive margins, notably the Exmouth Plateau off the NW Australia margin (Von Rad *et al.* 1992). Also, similarly rifted fragments appear be embedded in accretionary margins of Indonesia (Pigram & Pannabean 1984). A plausible mechanism might involve calving of weak marginal rift units, up to several hundred kilometres in size from a parent continent. The driving force could be slab-pull related to regional subduction, in this case northward subduction under the Eurasian margin during Late Palaeozoic–Early Mesozoic time. Although slab-pull by itself might be insufficient to initiate continental break-up (Smith 1999), it is possible that break-up could result from multiple rift events, especially once a rift was weakened by rift magmatism (Buck 1993). In the south Aegean there is indeed a history of pulsed rifting starting in Mid–Late Carboniferous time, with an extensional pulse associated with magmatism in the Early Triassic, and final break-up to form the Pindos ocean in the Late Triassic.

(3) The change from deep-water to shallow-water deposition during the late Early–Mid Triassic, as documented within the Phyllite–Quartzite unit of western Crete and eastern Crete implies uplift of >500 m and so favours a convergence-related, foreland basin or collisional setting

Undeformed rifts worldwide, including the Red Sea (Purser & Bosence 1998), the Gulf of Aden (Robertson & Bamakhalif 2001), the Avalon margin (e.g. Avalon platform; Tuckolke *et al.* 2004), the Indian ocean (e.g. off East Africa–Madagascar; Hankel 1994), and many other examples are known to have undergone hundreds of metres (to several kilometres) of marginal uplift related to extension, prior to the onset of sea-floor spreading.

There are several possible mechanisms for such uplift. First, a model of inhomogeneous crustal stretching with depth predicts flank uplift of 1–2 km (e.g. Braun & Beaumont 1989; Steckler & Omar 1994), although this would be hard to test using field geological evidence. Second, a thermal pulse could cause regional uplift. A plume influence related to Triassic rifting has been suggested for the Balkan region (Dixon & Robertson 1999). The presence of ocean island basalt (OIB)-type basalts in many areas (e.g. western Crete) could reflect a plume influence but could alternatively be explained by low-degree melting of potentially inhomogeneous subcrustal mantle. As yet, there is no definite evidence of a plume-related setting in the south Aegean region. Third, a pre-existing rift basin could be flexurally uplifted related to a pulse of extension that was focused elsewhere in the rift zone. Such a change in the locus of rifting could cause a change in the dip of the related extensional faults such that the pre-existing rift footwall was transferred to the hanging wall of the subsequent rift. Such an effect alone would be capable of explaining the relatively rapid change from relatively deep-sea (>500 m) to neritic depositional conditions, as observed in the south Aegean region.

(4) The geochemical evidence of Triassic igneous rocks requires coeval subduction in the south Aegean region

The Triassic volcanic rocks of eastern Crete, the Peloponnese, many other parts of Greece and also NW Turkey exhibit a chemical signature that is widely believed to require a subduction setting, at least locally (Pe-Piper & Piper 2002). Key points are the presence of Triassic arc-type granites (e.g. Cyclades, northern Menderes, eastern Crete; see Romano *et al.* 2006), the local occurrences of shoshonitic and high-K andesites (Lakmon Mtns.) and high-K andesitic intrusions (i.e. Kokkino, SW Peloponnese), the rare occurrence of boninitic-type lavas (Othris and Edipsos), and the presence of pyroclastic rocks (implying a high volatile content). Following an extended discussion, Pe-Piper & Piper (2002) concluded that the chemistry of some of the Triassic igneous rocks requires the involvement of subduction-derived fluid in the melt process and 'that subduction may be either Hercynian or of Triassic age' (p. 103). An inherited subduction influence, presumably related to Hercynian subduction in the south Aegean region, was previously proposed by various workers (Robertson & Dixon 1984; Dixon & Robertson 1993, 1999; Capedri *et al.* 1997; Pe-Piper & Piper 1998). Implicitly, Pe-Piper & Piper (2002) have acknowledged that these two alternatives, a coeval Triassic versus a Hercynian inherited subduction signature, cannot be resolved by geochemical evidence alone. Thus, the decisive factors can only be the geological evidence for subduction zones of the requisite age and location. No independent evidence for such Triassic subduction zones was found in the south Aegean region during this study and, therefore, the model of subduction zone inheritance is preferred. In keeping with this, volcanic rocks within units that restore further south (e.g. Phyllite–Quartzite unit, western Crete) are enriched in incompatible elements (with no subduction influence), similar to modern rift basalts (e.g. Fitton *et al.* 1998). By contrast, volcanic rocks extruded through Hercynian basement units, generally located further north, are relatively depleted in incompatible elements (e.g. Nb), possibly reflecting the extraction of a lithosphere-hosted subduction component of probable Hercynian age.

It was similarly suggested that the presence of radiometrically dated Late Triassic calc-alkaline granitic rocks (orthogneiss) in the Vai area, eastern Crete, implies a convergent margin (subduction) setting during the Triassic, possibly related to southward subduction (Romano *et al.* 2006; Model 3). At least some of the granitic rocks in this area crystallized, then were exhumed and eroded throughout Mid–Late Triassic time, as similar granitic rocks are found as clasts within associated coarse clastic sediments of this age. The small Triassic granitic bodies might relate to melting in an extensional setting, followed by rapid exhumation, as inferred, for example, for the Oligocene granites of northern Greece (Kolokotroni & Dixon 1991).

(5) There is no evidence of a Triassic subduction zone in northern Greece and thus the convergence of Africa and Eurasia must be accommodated in the south Aegean

However, several studies of units associated with the southern margin of Eurasia, including the Pontides (e.g. Ustaömer & Robertson 1997; Okay 2000), the Caucasus (e.g. Adamia *et al.* 1995) and the south margin of Eurasia generally (Nikishin *et al.* 2001) have concluded that a subduction zone dipped northwards beneath Eurasia and was active especially during Carboniferous to Mid-Jurassic time (Nikishin *et al.* 2001; Ustaömer *et al.* 2005; Kazmin & Tikhonova 2006). Palaeotethyan units have also been identified in former Yugoslavia (see Karamata 2006). Some workers have suggested that a Palaeotethyan suture is located in the Vardar zone of northern Greece (Robertson & Dixon 1984; Mountrakis 1986). However, the assumption that the Serbo-Macedonian and Rhodope zones formed part of the southern margin of Eurasia by Early Mesozoic time is now questioned by radiometric dating and structural studies that suggest that independent terranes existed until docking with Eurasia during Alpine (Jurassic) deformation (Himmerkus *et al.* 2006). A Palaeotethyan suture may thus be buried within northern Greece, removing the need for a more southerly located Palaeotethyan subduction zone. Such a subduction zone would have extended eastwards into the Pontides and westwards into former Yugoslavia.

(6) Regional plate reconstructions favour the existence of a Palaeotethyan ocean in the south Aegean region

Garkunkel (2004) favoured a reconstruction akin to the convergence-related Model 2. He argued that between Late Permian and Late Triassic time, the regional palaeogeography evolved from a Pangaea A to a Pangaea A-2 type setting (see Smith *et al.* 1981; Smith 2006; Fig. 24). If correct, this would imply convergence along the southern margin of Eurasia of several hundred kilometres during this time. This reconstruction assumes *c.* 500 km of right-lateral motion between Africa and Eurasia, and that this was translated into clockwise tightening of a Palaeotethyan gulf in the east (comparable with the setting of the modern Gulf of Makran). Ziegler & Stampfli (2001) argued that up to 400 km of right-lateral displacement did indeed take place, but placed

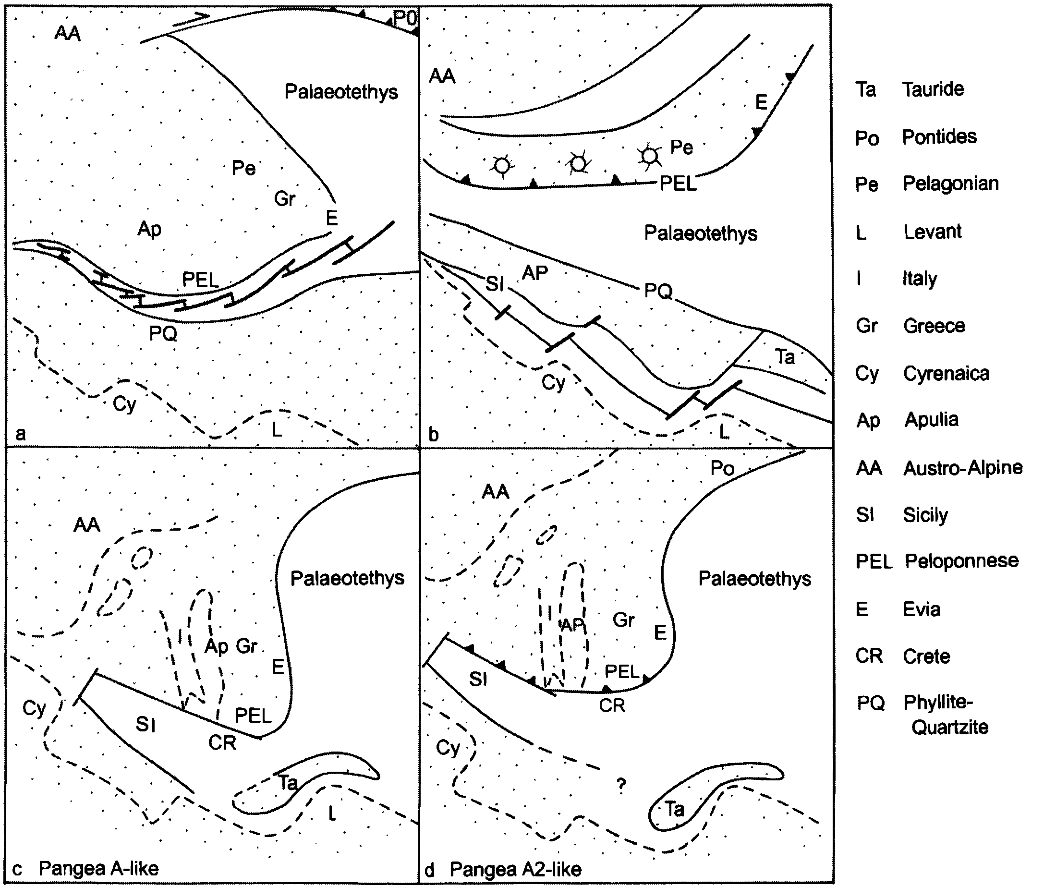


Fig. 24. Alternative models for the Late Palaeozoic–Early Mesozoic tectonic evolution of the south Aegean region. (**a**) A single spreading axis propagated from the wider Tethys to the east (Ricou 1996); (**b**) northward subduction zone and continental rifting (Stampfli *et al.* 2001); (**c**) extensional setting, assuming a Pangaea A-like reconstruction (Garfunkel 2004); (**d**) convergent setting assuming a Pangaea A2-like reconstruction (Garfunkel 2004). AA, Austro-Alpine; Ap, Apulia; CR, Crete; Cy, Cyreanaica; E, Evia; Gr, Greece; I, Italy; L, Levant; Pe, Pelagonian; PEL, Peloponnese; Po, Pindos; PQ, Phyllite-Quartzite; SI, Sicily; Ta, Tauride.

this within the latest Carboniferous–Early Permian time interval, associated with transtensional collapse of the Hercynian orogen and the development of related pull-apart basins (see, e.g. Scotese & Langford 1995). Regional magmatism (granitic intrusion and calc-alkaline extrusion) climaxed in Early Permian time. As a result, the preferred Pangaea A-2 type setting could have been established prior to Triassic time. A several hundred kilometre wrench offset might have been dissipated along several lineaments within the Hercynian orogen rather then being directly translated into regional clockwise rotation of the Eurasian margin in the Central and Western Mediterranean. Garfunkel (2004) considered an alternative, extension-related setting, which could be valid for the Triassic assuming Pangaea reorganization largely occurred in pre-Triassic time, as favoured here (Fig. 24).

Regional tectonic development

One of the remaining problems is the relationship between Hercynian deformation and metamorphism, as documented by the fragmentary high-grade units scattered around the south Aegean region, and the Triassic rift setting (see Romano *et al.* 2006). How did the North African passive margin to the present south escape this deformation and metamorphism? By contrast, Hercynian compressional deformation affected the North African margin west of Tunisia (Guiraud *et al.* 2001). Also, what was the nature of the contact between these metamorphosed (e.g. Crete) and unmetamorphosed (e.g. North Africa) domains?

In Model 1 (divergent setting) the Talea Ori–Plattenkalk unit in Crete and the Southern Peloponnese formed the distal edge of the North African margin during pre-Triassic time. Assuming that the Hercynian-age detritus within the Talea Ori–Plattenkalk unit (Brix *et al.* 2002) was locally derived, these sedimentary units are likely have been deposited on Hercynian basement, which was detached during Early Cenozoic subduction and is thus mainly not now exposed. In this model it is possible that the deformed and metamorphosed northern edge of the Hercynian orogeny, located along the North Gondwana margin, was later rifted to open the Neotethyan ocean basin to the north (Triassic or younger), stranding it entirely to the north.

Models 2 and 4 (convergent settings) are problematic as no evidence of a northward-dipping Palaeotethyan subduction zone has been identified in the south Aegean region, ruling out any juxtaposition of metamorphosed and unmetamorphosed units as a result of subduction–collision (pre-Jurassic). In Model 2 the Talea Ori–Plattenkalk unit rifted from North Africa as the Cimmerian continent, well south of areas affected by Hercynian orogenesis, yet contains Hercynian-age detritus. To explain this, Champod *et al.* (2004) suggested that Hercynian-aged detrital zircons were transported hundreds of kilometres southwestwards through a continental rift system from the well-established Hercynian orogen in the central–west Mediterranean region. However, a local provenance from exposures of high-grade 'Hercynian' basement is more consistent with the sedimentological evidence outlined earlier in the paper.

In Model 3 (southward subduction), the Hercynian-age granitic rocks (e.g. Pelagonian and South Aegean) relate to southward subduction of Palaeotethys (e.g. Şengör 1984; Romano *et al.* 2006; Xypolias *et al.* 2006). Units affected by Hercynian metamorphism (e.g. Chemezi and Kythira) were close to the trench in the north relative to the North African continent further south. The deformation and metamorphism could then simply have tailed off southwards, with the original transition now being hidden beneath the Sea of Crete. In this interpretation the Carboniferous radiometric ages of the Mersini basement complex, eastern Crete, and associated structural evidence (Romano *et al.* 2006) are consistent with southward 'Hercynian' subduction. However, in this interpretation the Permian and Triassic ages from other crystalline units in the area (Romano *et al.* 2006) are surprising, as it is generally believed that Hercynian orogeny had given way to extension-controlled exhumation by the Late Carboniferous (Ziegler 1988; Ziegler & Stampfl 2001). Also, the Triassic rift-related basaltic rocks of western Crete do not exhibit a subduction influence or contain arc-derived detritus, as would be expected if they represented back-arc marginal basins above a coeval southward-dipping subduction zone.

There is evidence of Carboniferous subduction-related magmatism further east, in Turkey, but only in the north (e.g. in the NW Pontides; Ustaömer *et al.* 2005), which implies the existence of northward subduction beneath Eurasia. If southward subduction in the south Aegean region is also accepted, this would require the existence of two subduction zones, one dipping northwards beneath Eurasia and the other dipping southwards beneath Gondwana, both active during Late Palaeozoic time. However, in Turkey there is as yet no convincing evidence of Late Palaeozoic southward subduction; for eample, along the northern margin of the Tauride–Anatolide platform, where passive margin conditions persisted (Robertson *et al.* 2004). This suggests that any south-directed subduction would have mainly affected areas in the west, in the central and western Mediterranean regions and elsewhere in southern Europe but not Turkey further east. Devonian–Carboniferous southward subduction as well as northward subduction have indeed been inferred for the Hercynian basement in central and western Europe (e.g. Eastern Alps), giving rise to a doubly vergent orogen (Neubauer & Handler 1999).

The apparent absence of Hercynian deformation and metamorphism within both North Africa and Gondwana-derived units (e.g. Taurides and Anatolides) raises the possibility that the Hercynian units of the south Aegean region might represent exotic terranes that were emplaced from the central Mediterranean region by right-lateral strike-slip (Dornsiepen *et al.* 2001). In this scenario, the fragmentary 'high-grade' metamorphic units of the south Aegean region (e.g. eastern Crete) formed in response to collisional suturing of Palaeotethys, some way westwards of their present position during Carboniferous time (Fig. 25). Open-ocean conditions (i.e. Palaeotethys) persisted further east, from western Turkey eastwards. Palaeotethyan exotic terranes were displaced eastwards in response to tectonic escape from the Hercynian suture zone to an open ocean to the east, during or soon after diachronous closure of Palaeotethys further west (SW and central Europe; Alps) during Late Devonian–Early Carboniferous time. This process would be comparable with the westward

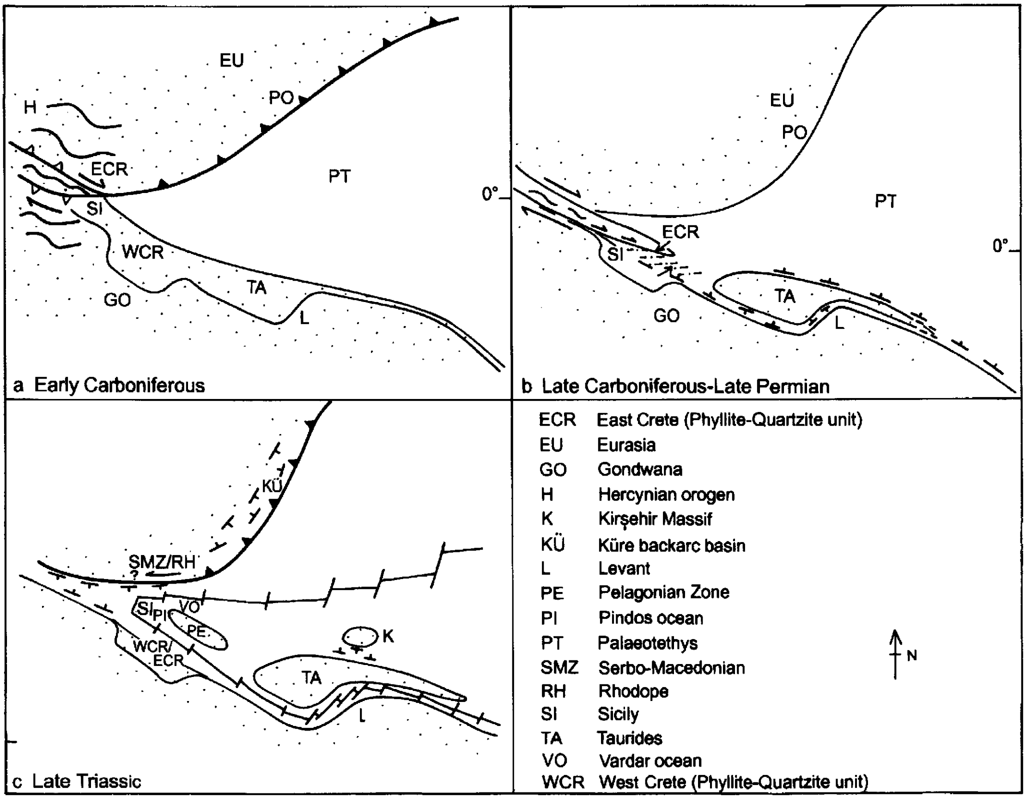
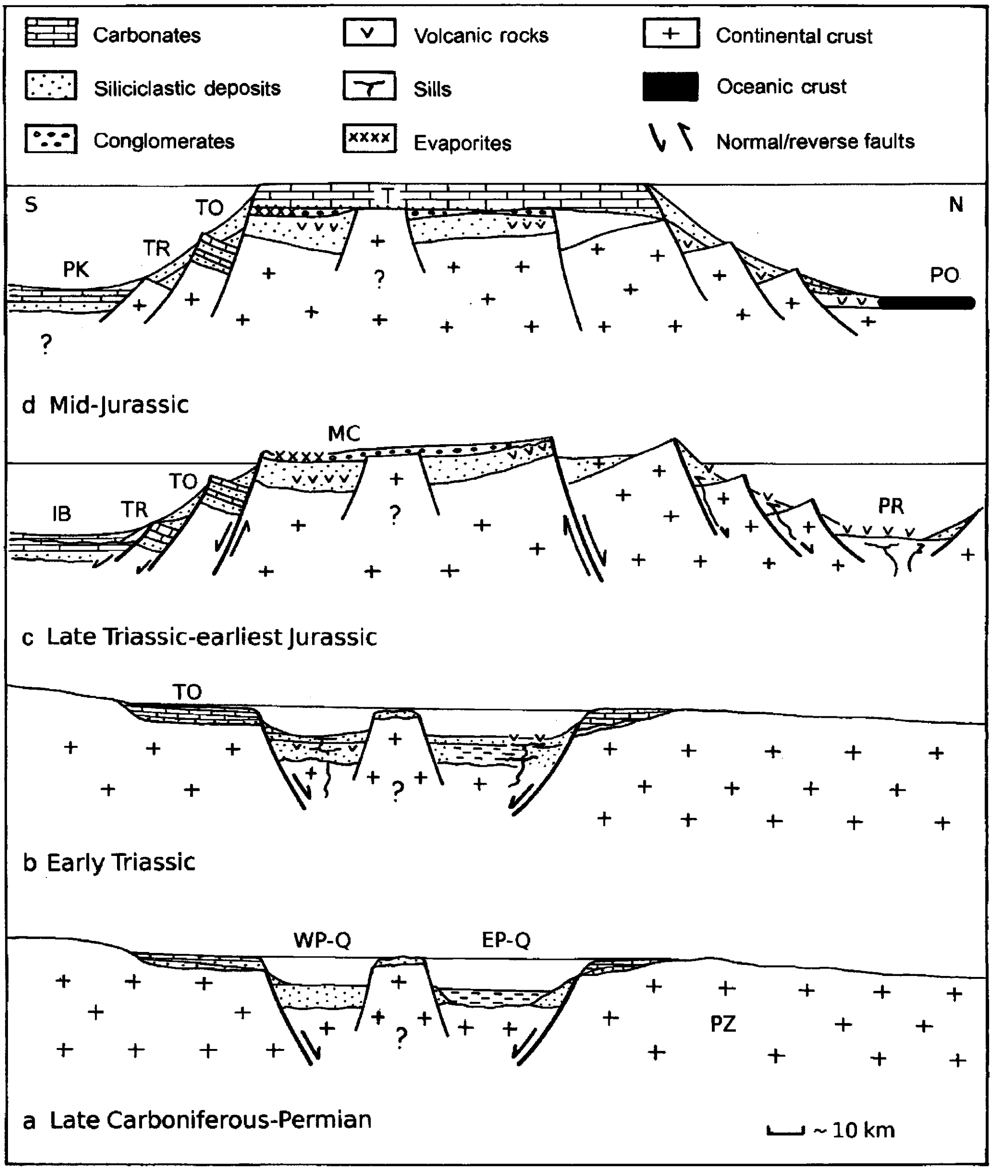


Fig. 25. Proposed tectonic evolution of the Upper Palaeozoic–Lower Mesozoic units of the south Aegean region. (**a**) Diachronous closure of Palaeotethys. (**b**) Syn-post-collisional right-lateral wrench faulting displaces exotic Hercynian terranes into the south Aegean region. Deep-water sediments accumulate in transtensional basins open to Palaeotethys to the east. (**c**) The south Aegean margin undergoes continental break-up to form the Pindos ocean and counterparts in the easternmost Mediterranean region.

tectonic escape of Anatolia after the Miocene. Such dextral displacement might have occurred at any time during Late Carboniferous–Early Triassic time, associated with reorganization to a Pangaea A-2 type assembly. Depending on the timing of any such displacement, the Late Palaeozoic deep-water basins of the south Aegean could have been strike-slip controlled. More evidence is needed to discriminate between the above alternatives, but the tectonic escape interpretation is promising.

Following the Hercynian orogeny, the Permian–Triassic deep-sea sediments of western Sicily and eastern Crete accumulated in a broad, relatively deep, possibly transtensional rift basin. In Sicily, terrigenous turbidites sourced in the exhumed Hercynian orogen were deposited in deep water, followed by open-marine radiolarian sediments and pelagic carbonates. Condensed pelagic carbonates accumulated on intra-basin highs and carbonate platforms developed around the periphery of the basin. Further east, in Crete, the Late Palaeozoic deep-sea siliciclastic sediments of the Phyllite–Quartzite unit and the shallow-marine siliciclastic sediments of the Talea Ori unit were deposited within, and along, the margins of a broad deep-water rift basin, fed from the North Africa craton and possibly from the detached northern margin of this basin. The coeval deep-sea sediments of eastern Crete formed in a relatively distal part of the rift, isolated from coarse terrigenous input. During this time the south Aegean region remained open to Palaeotethys further east.

The Triassic Pindos rift basin in Greece extended northwestwards through the Budva zone of former Yugoslavia to connect with the Lagonegro zone in southern Italy. The rifted pelagic basin in the Lagonegro zone dates from the Mid-Triassic, but rifting apparently



EP-Q	E Crete Phyllite-Quartzite	PO	Pindos ocean
I	Ionian Basin	PR	Pindos rift
MC	Mana Conglomerate	T	Tripolitza platform
NA	North Africa	TO	Talea Ori
PZ	Pelagonian Zone	TR	Tripali
PK	Plattenkalk	WP-Q	West Crete Phyllite-Quartzite

Fig. 26. Inferred tectonic evolution of the Upper Palaeozoic–Lower Mesozoic units of Crete and the southern Peloponnese. Rifting along the north margin of Gondwana gave rise to a relatively wide and deep rift basin. During the Mid-Triassic, extensional faulting of this margin resulted in flexural uplift, whereas the basins to the north and south rapidly subsided culminating in opening of the Pindos ocean to the north. The former rift zone was capped by the Gavrovo–Tripolitza carbonate platform during Mesozoic passive subsidence.

commenced in the Late Permian, based partly on the evidence of reworked neritic fossils (see Ziegler & Stampfli 2001 for review). However, there is little evidence for the existence of a pre-existing, Late Palaeozoic deep-water basin in the Lagonegro zone similar to the Sicanian basin. This, in turn, implies that the Pindos ocean did not simply widen the pre-existing Late Palaeozoic rift, but instead created a new basin, which reactivated an older rift in the east (e.g. Crete) but left it abandoned in the west (Sicanian basin).

The Late Palaeozoic deep-water rift basin in the south Aegean region was reactivated in the Triassic as a precursor to opening of the Pindos ocean. A pulse of rifting, probably focused along the Pindos rift to the NE, resulted in flexural uplift of part of the pre-existing rift basin (i.e. rift-shoulder uplift) in the south Aegean region (Fig. 26). By contrast, the Permo-Triassic rift basin further west, in Sicily (Sicanian basin), was abandoned and gently subsided until Mid-Jurassic time when it was reactivated related to opening of the central North Atlantic. Extension, however, reached as far west as this area and resulted in episodic destabilization of bordering carbonate platforms and localized Triassic volcanism.

After spreading of the Pindos ocean began in Late Triassic time, passive margin subsidence was accommodated by the growth of large carbonate platforms bordering the Pindos ocean and the abandoned Permian rift basin in the Sicily area. The platforms were constructed right across the former rift basins represented by the Phyllite–Quartzite unit after their Mid-Triassic flexural emergence and some erosion (e.g. to form the Mana conglomerate; Fig. 26).

During Late Triassic time, Crete, the Peloponnese and south Aegean as a whole experienced passive margin subsidence, building up the kilometres thick Gavrovo–Tripolitza carbonate platform and its passive margin units, including the Talea Ori and Tripali units. The platform was detached when the basement was subducted during the Early Cenozoic, followed by exhumation, whereas the platform cover and its local substratum (Tyros and Ravdoucha units) were accreted to the overriding plate. In Crete, the Triassic carbonate platform represented by the Talea Ori and Tripali units rifted and foundered, followed by deposition of the pelagic Plattenkalk, a counterpart of the deep-water Ionian zone in western Greece and Albania.

In western Sicily, the Sicanian basin was reactivated during Mid-Jurassic time, related to opening of the Central Atlantic. A spreading centre possibly migrated eastwards to open the oceanic Ionian basin during Late Jurassic time (see Catalano *et al.* 2001), possibly even extending eastwards to open or widen the southernmost Neotethyan oceanic basin between Crete and North Africa.

During Cenozoic subduction in the south Aegean region the Tripolitza platform was detached from its pre-Jurassic rift-related substratum that was subducted, accreted to the overriding plate, and then was exhumed as the HP–LT units of the lower thrust sheets (Phyllite–Quartzite, Talea Ori–Plattenkalk and Tripali units).

Conclusion

Of the alternative models for the Late Palaeozoic–Early Mesozoic setting of the south Aegean region, a pulsed rift model best fits the evidence, based on new field-based observations in western Sicily, Crete, the Peloponnnese and Evia combined with a review of the literature (Figs 2 and 26). A deep-water rift opened along the northern margin of Gondwana during the Mid–Late Carboniferous, followed by a further pulse of rifting in the Early Triassic, preparatory to opening of the Pindos ocean to the NE (present coordinates) during the Late Triassic. Mid-Triassic uplift and erosion in Crete is explained by upward flexure of the preceding Late Palaeozoic rift zone, related to renewed rifting to form the Pindos ocean in the south Aegean region. In the absence of evidence for contemporaneous Triassic subduction, it is inferred that the observed subduction signature in many the Triassic rift-related basalts (e.g. eastern Crete, Peloponnese) relates to melting of heterogeneous subcrustal mantle. The subduction fluids were probably introduced during Hercynian orogenesis.

Our present understanding of the tectonic development of the south Aegean region owes much to the detailed biostratigraphical studies of J. Krahl in Crete. I would like to thank him for information on the literature and relevant outcrops in Crete. Thanks are also due to H. Kozur for written and verbal discussions during this work. I am grateful to R. Catalano for a helpful field introduction to the geology of western Sicily, and to N. Skarpelis for a similar introduction to the SW Peloponnese. J. Dixon is thanked for continuing helpful discussion. G. Karner provided useful insights into modern rifted margins. The manuscript benefited from comments by J. Dixon, P. Degnan and D. Mountrakis.

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