Synthesis of the tectonic–sedimentary evolution of the Mesozoic–Early Cenozoic Pindos ocean: evidence from the NW Peloponnese, Greece

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Abstract: The tectonic development of the western part of the Pindos ocean in southern Greece is exemplified by the mountainous Pindos thrust belt in the NW Peloponnese. A Late Triassic–Early Cenozoic succession exposed within imbricate thrust sheets records a range of deep-water siliciclastic, redeposited carbonate and siliceous sediments, which in general become more distal oceanwards towards the east. Igneous rocks, locally dated as Triassic, occur within a mélange that is entrained beneath and within the Pindos thrust stack; these igneous rocks and related sediments are interpreted as remnants of a continent–ocean transition zone. 'Immobile' element geochemistry is explicable by rifting of a compositionally heterogeneous subcontinental mantle, possibly related to pre-existing Hercynian subduction, although coeval Triassic subduction cannot be excluded based on evidence from this area alone. Localized, 'enriched' basalts are interpreted as fragments of oceanic seamounts formed in a relatively distal setting. Late Paleocene–Early Eocene (locally Mid-Eocene) siliciclastic turbidites, derived from the north, record the latest deposition prior to incorporation of the sedimentary succession into a westward-migrating accretionary wedge during post-Early Eocene time in the NW Peloponnese. Structural restoration of the well-ordered thrust stack indicates a minimum of 201 km (55%) of shortening at an average rate of 5.8 mm a^{-1}. As the Pindos allochthon approached the Apulian continent, the Gavrovo–Tripolitza foreland underwent flexural upwarp during the Mid-Eocene, followed by collapse to create a foreland basin by the Late Eocene. This basin was infilled with generally upward-thickening and -coarsening deep-water turbiditic sediments of Late Eocene–Early Oligocene age. The foreland was, in turn, overthrust by the Pindos accretionary prism during post-Early Paleocene time, and was then imbricated and thrust over the Ionian foreland basin to the west by Pliocene time.

Forming the backbone of mainland Greece and extending into Albania, the Pindos Mountains have attracted considerable interest since the classic regional mapping and stratigraphical studies of Dercourt (1964), Aubouin *et al.* (1970), British Petroleum (1971), Jacobshagen *et al.* (1978), Fleury (1980) and Dercourt *et al.* (1986, 1993) (Fig. 1). Western Greece is traditionally subdivided into a series of tectono-stratigraphic units, originally termed isopic zones (Aubouin *et al.* 1970). The more westerly of these include the Gavrovo–Tripolitza zone and the Pindos (or Pindos–Olonos) zone (Pindos suture in Fig. 1). The Gavrovo–Tripolitza zone forms the foreland to a thrust belt represented by the Pindos zone. The Pindos zone has been variously interpreted as a Triassic rift located close to Gondwana (Aubouin *et al.* 1970; Dercourt *et al.* 1986; Yılmaz *et al.* 1996), a mid-ocean ridge-type basin (Smith *et al.* 1975; Robertson & Dixon 1984; Robertson *et al.* 1991, 1996; Smith 1993), or a marginal basin variously related to southward subduction (Şengör *et al.* 1984), northward subduction (Stampfli & Borel 2002; Stampfli *et al.* 1998, 2001), or partially related to westward intra-oceanic subduction (Pe-Piper & Piper 2002).

The main objective here is to synthesize the constructive and destructive evolution of the Pindos ocean based mainly on evidence from the NW Peloponnese (Fig. 2). First, we summarize the rift and passive margin evolution of the western margin of the Pindos ocean. We then discuss the importance of a mélange unit beneath and within the thrust stack for understanding the nature of the continent–ocean transition zone, utilizing previously unpublished igneous geochemical data. We then present detailed structural information for a well-exposed structural traverse in the NW Peloponnese and interpret this as a westward-migrating accretionary prism. We also outline evidence for the emplacement

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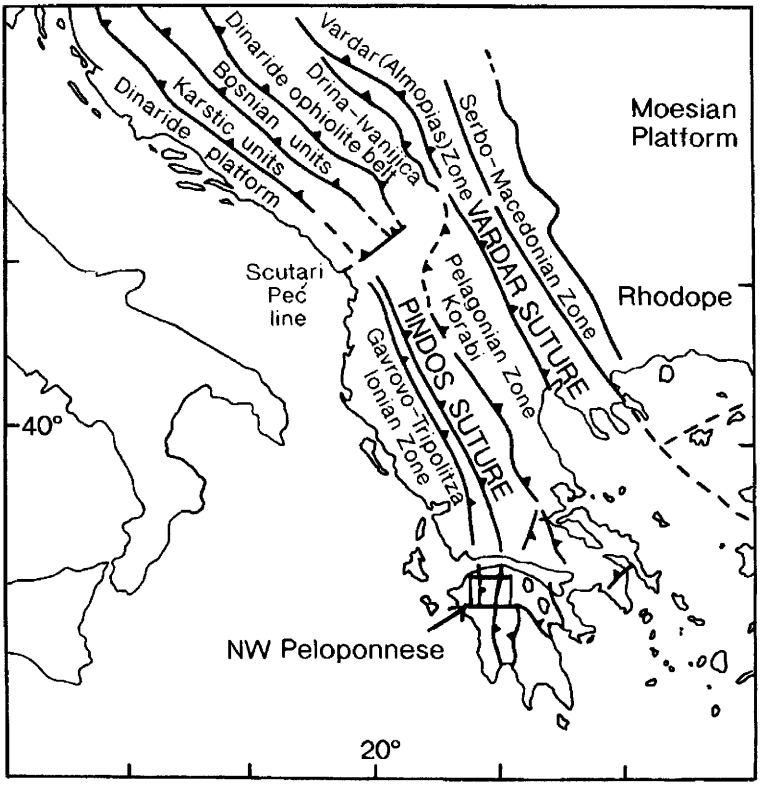

Fig. 1. Setting of the Pindos suture and the location of the study area in the NW Peloponnese.

of the Pindos allochthon over the Gavrovo–Tripolitza platform to the west following the development of a related foreland basin. Finally, we summarize the history of opening and closure of the Pindos ocean in southern Greece in the light of the wider regional tectonic setting.

Rift–passive margin development

The rifted passive margin of the Pindos ocean is recorded by the Triassic–Early Cenozoic deep-water sediments of the Pindos Group that are exposed as a stack of imbricate thrust sheets (Robertson 1994; Degnan & Robertson 1998; Fig. 2). The differential thicknesses and facies of successions exposed in more proximal (westerly) to more distal (easterly) areas have influenced the style of subsequent deformation, as discussed later in the paper.

The oldest sediments, mainly exposed in the westerly, structurally lower thrust sheets, known as the Priolithos Formation (Degnan & Robertson 1998) are mainly sandstones (litharenites), siltstones, shales, calcilutites and nodular limestones. The sandstones contain abundant detritus from metamorphic (e.g. quartzite, mica schist) and plutonic igneous (e.g. granitic units) sources and include sparse to locally abundant volcanic material of mainly intermediate to silicic composition. The Triassic sandstones plot in the litharenite field of the Q:F:L diagram (McBride 1963) and in the recycled orogen field on the Qm:F:L diagram (Dickinson & Suczek 1979). The Triassic limestones contain *Holobia* sp. and conodonts, indicative of a Late Triassic (Carnian–Norian) age (Flament 1973). The Triassic mixed carbonate–siliciclastic sediments pass depositionally upwards into debris flows, carbonate turbidites and hemipelagic carbonates of the Carnian–Liassic Drimos Formation. In the Early Jurassic there was a regional change to contrasting argillaceous sediment deposition, known as the Kastelli Mudstone Member (Leesteena Formation). These sediments pass upwards into reddish, mainly siliceous facies, including well-developed ribbon radiolarites

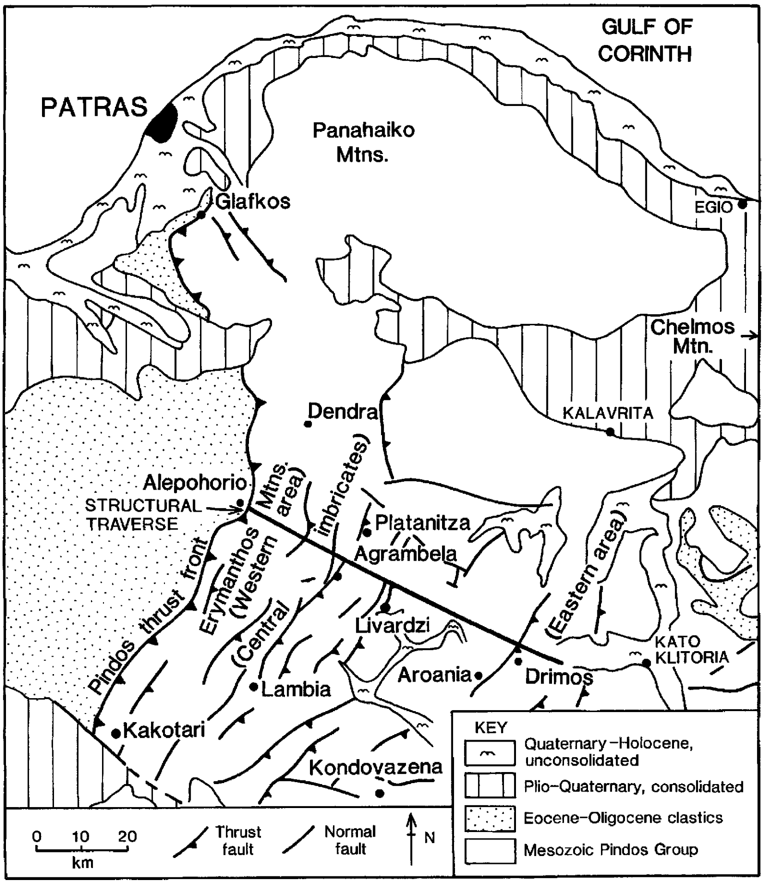

Fig. 2. Outline tectonic map of the NW Peloponnese, showing approximate trajectory of the structural traverse studied and location of places discussed in this paper.

of the Middle–Upper Jurassic Aroania Chert Member, associated with manganese of probable hydrothermal origin (Pe-Piper & Piper 1989; Robertson & Degnan 1998). Local developments of fine-grained calpionnellid limestones of Late Jurassic (Tithonian) age mainly occur in the south Peloponnese. Manganese-rich cherts of Late Jurassic age are locally interbedded with volcaniclastic sediments and overlie strongly altered volcanic rocks (e.g. at Aroania, Kombigadi and Drimos; Robertson & Degnan 1998). By the Tithonian, there was a swich to the accumulation of fine-grained carbonates with varying amounts of pelitic material in relatively distal settings represented by the Paos Limestone Member (Lambia Formation; Neumann *et al.* 1996; Neumann & Zacher 2004). There are occasional dark, locally organic-rich intervals (Neumann *et al.* 1996; Wagreich *et al.* 1996; Neumann 2003). During Cenomanian–Turonian time terrigenous turbidites, known as the Klitoria Sandstone Member (Premier Flysch de Pinde), are exposed within thrust imbricates located in the central parts of the thrust traverse. These sediments lack sand-sized ophiolitic debris as seen in equivalent sediments north of the Gulf of Corinth, but clay mineral studies of interbedded fine-grained sediments suggest an ophiolitic source within the Peloponnese (Thiébault & Fleury 1980). A thick sequence of Late Cretaceous pinkish–greyish pelagic carbonates, represented by the Erymanthos Limestone Member (Lambia Formation) follows, interspersed with minor greyish organic-rich layers (Neumann *et al.* 1996; Neumann 2003).

Around the Cretaceous–Cenozoic boundary there was a return to periodic siliceous and organic-rich sedimentation, known as the Katataktis Passage Member of the Lambia Formation (Robertson & Degnan 1998), or the 'Couches de

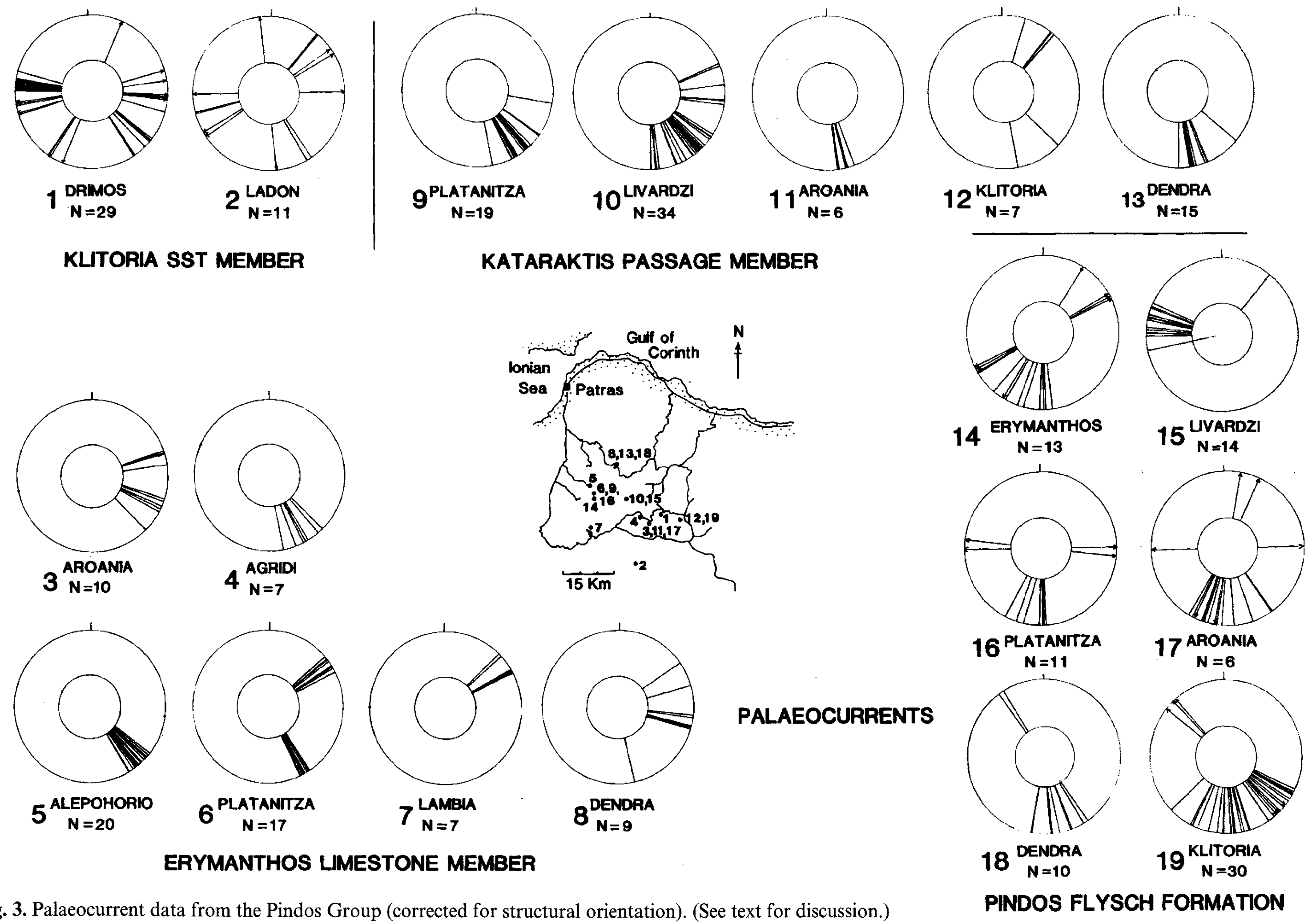

Fig. 3. Palaeocurrent data from the Pindos Group (corrected for structural orientation). (See text for discussion.)

Passage' (Flament 1973; Fleury 1980). The age of the base of this member ranges from Mid- to Late Maastrichtian or even Campanian in different areas (see Piper 2006).

Siliceous sediments of the Kataraktis Passage Member are interbedded with, and overlain by, relatively fine-grained carbonate conglomerates derived from the Gavrovo–Tripolitza platform to the west. More proximal successions of Cretaceous age, located in structurally lower thrust sheets, contain abundent channelized limestone conglomerates, interpreted as a base-of-slope apron (Degnan & Robertson 1998). Piper (2006) confirmed that successions become more distal from west to east across the Peloponnese as a whole. He also showed that palaeocurrents are generally towards the SE, although data are sparse. Results obtained during this study support this interpretation (Fig. 3).

Sedimentation culminated in the deposition of siliciclastic turbidites known as the Pindos Flysch Formation ('Flysch de Pinde'), of mainly Late Paleocene–Early Eocene age, but extending to Mid-Eocene age in the extreme SE of the Peloponnese (see Piper 2006). Degnan & Robertson (1998) reported mainly southward palaeocurrents in the NW Peloponnese (Fig. 3). These results are based on measurements of micro-cross-lamination, primary current lamination, groove casts and sole marks. Differences in current orientation were occasionally noted between sole marks and bedding–internal structure; sole-mark data were used preferentially in such cases. Piper (2006) noted generally SE-directed palaeocurrents in the NW Peloponnese. South-directed palaeocurrents were reported from the west central and eastern Peloponnese (Richter 1993). By contrast, sparse palaeocurrents in the SW Peloponnese are to the NE (see Piper 1984, 2006 for literature review).

Degnan & Robertson (1998) interpreted the Pindos flysch in the NW Peloponnese as turbiditic sedimentation in a subduction trench setting with sand supply mainly from the north. Transitional facies represented by carbonate debris flows in the west record a continuation of input from the adjacent Gavrovo–Tripolitza carbonate platform (Degnan 1992). Piper (2006) envisaged sea-level fall rather than tectonics as the dominant control of Pindos flysch deposition in the Peloponnese. Sediment was potentially derived from both the Apulian foreland and the overthrusting Pelagonian active margin in northern Greece, followed by southward transport into the Peloponnese area. Both tectonic setting and sea level are likely to have contributed to the deposition.

Nature of the continent–ocean transition zone

Clues to the nature of the basement beneath the deep-sea sediments of the Pindos Group are provided by the presence of igneous rocks that, where dated, are mainly of Triassic age (see Pe-Piper & Piper 2002, for literature review). Some of these volcanic rocks occur as local thrust sheets intercalated with shallow-water to deep-water sediments in different areas. They are also more extensively represented by a volcanic–sedimentary mélange that is entrained beneath and within the Pindos thrust stack.

In several areas, the Pindos thrust sheets include coherent units of igneous rocks. For example, north of the Gulf of Corinth (near Mesophyton) Triassic *Halobia* limestones include a Jurassic? dolerite sill (Green 1982). Further west in the Lakmon Mountains shallow-marine sedimentary rocks are interbedded with lavas, agglomerate and tuff; extrusive rocks include andesites and shoshonites (Pe-Piper 1983; Pe-Piper & Piper 2002).

Scattered throughout the Peloponnese the Pindos thrust sheets are underlain by, or intercalated with, a thin (tens of metres) unit of mélange, termed the Formation à Blocs by De Wever (1976*a*). This typically occurs between the Gavrovo–Tripolitza foreland unit and the overlying Pindos thrust sheets at a number of localities (Fig. 4) throughout the Peloponnese and also to the north of the Gulf of Corinth (De Wever 1975, 1976*a*,*b*; Richter & Lensch 1977, 1989; Dercourt *et al.* 1978; Pe-Piper & Piper, 1998). In the central Peloponnese the mélange includes basalt blocks (near Methedrio and Lagadi), and in the southern Peloponnese, a local mélange exposure (at Sellas) includes basic extrusive blocks resembling pillow basalt. In the Peloponnese, most of the igneous rocks occur as blocks in a mélange, although primary sedimentary contacts with deep-sea sediments are preserved locally.

In the far SW Peloponnese, at Kokino, blocks of volcanic and volcaniclastic rocks (up to 20 m thick) are associated with hyaloclastites and sills. Above this, a sequence of volcanic rocks, volcanogenic sediments and minor pink *Halobia*-bearing limestones passes depositionally upwards into well-bedded *Halobia*-bearing hemipelagic limestones of Late Triassic age. This unit can be correlated with the locally exposed base of the Pindos Group (Degnan & Robertson 1994), and is important as it establishes that the oldest sediments exposed in the Pindos thrust sheet (Priolithos Formation) accumulated on a volcanic basement.

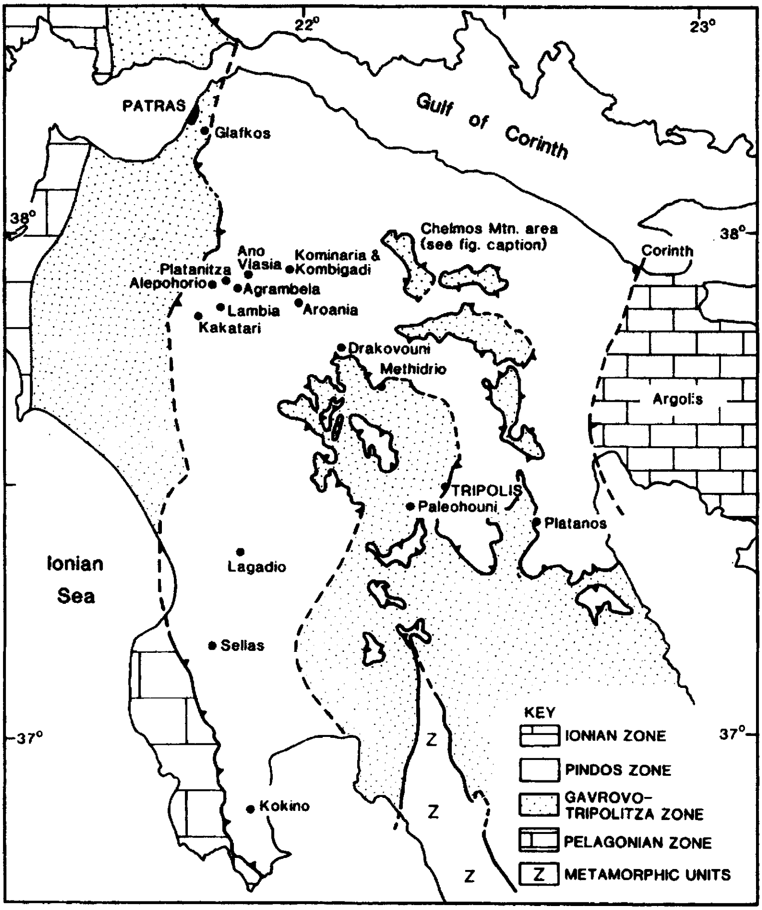

Fig. 4. Outline map of part of the Peloponnese showing the main locations of the mélange that is entrained at the base of and within the Pindos thrust sheets. From Degnan & Robertson (1994).

Where locally present beneath the Pindos thrust sheets in the NW Peloponnese, the area of main concern here, the mélange typically includes large volumes of strongly sheared sedimentary rocks, especially near the structural base. The extrusive blocks (mostly < 1 m^3 in size) are mainly basic–intermediate–siliceous volcanic rocks, volcaniclastic rocks and rarely intrusive igneous rocks (e.g. dolerite). Plagiogranite has also been reported (Pe-Piper & Piper 1991, 2002). Most of the mélange blocks are undated, but at one locality (Mati) limestone crusts attached to an igneous block contain Triassic fossils (De Wever 1975).

Sheared sedimentary rocks occur in the matrix of the mélange and as strongly dismembered thrust sheets in some places. These sediments can be correlated with lithologies of both the Gavrovo–Tripolitza zone beneath (e.g. sandstones, redeposited limestones with benthic Foraminifera, brown shales) and with the Pindos zone above (e.g. pelagic and redeposited limestones, quartzose sandstones, radiolarian cherts, red shales). The matrix of the mélange was created by tectonic fragmentation of mainly sedimentary material from both the Gavrovo and Pindos units. The igneous lithologies (e.g. basalt) are preserved as relatively large competent blocks (up to 20 m in size), whereas the less competent sedimentary rocks (e.g. volcaniclastic sandstone) have experienced pervasive layer-parallel extension on all scales, giving rise to phacoid-shaped blocks.

A prominent mélange exposure in the eastern part of the NW Peloponnese, at Drakovouni (Fig. 4) includes numerous metre-scale blocks of basalt that were created by the fragmentation of pre-existing basalt flows. Pillow interstices are infilled by pelagic sediment, either carbonate or manganiferous chert. Generally smaller blocks of pyroclastic tuff, lava breccia and lava conglomerate are also present. Within several of the larger igneous blocks, short intact successions can be recognized, e.g. basalt passing into dolerite and tuff or volcaniclastic sediment. This locality is notable as successions within the thrust stack are mainly restricted to Cretaceous and Palaeogene ages; also, the volcanic rocks (e.g. at Drakovouni) are commonly of enriched, within-plate-type (see below). Elsewhere in this eastern area the mélange occurs extensively around Chelmos (Fig. 4), where it contains blocks derived from the Gavrovo footwall, the overlying Pindos thrust sheets and from a basic–intermediate volcanic sequence associated with volcaniclastic sediments (Degnan & Robertson 1994).

Geochemical fingerprinting of basaltic rocks

The geochemistry of the extrusive rocks within the mélange may help to indicate the nature and tectonic setting of the crust beneath the deep-sea Pindos sediments. As the rocks are all altered only 'immobile' major elements and trace elements were used for tectonic discrimination (Pearce & Cann 1973; Pearce 1980). The lithologies are mainly basaltic andesites but range from basalt, to andesite, to rhyodacite. A previous study of the stable trace elements (*c.* 20 samples) suggested that most are subalkaline; a few are enriched, akin to within-plate ocean-island basalts (OIB) (Pe-Piper & Piper 1991). The extrusive rocks in the mélange fall within the range of compositions of Triassic rocks from numerous other areas of Greece, as summarized by Pe-Piper & Piper (2002).

During this work a large number of lava samples (*c.* 100) were collected to determine if any significant geochemical differences exist along or across the strike of the mélange outcrops throughout the Peloponnese as a whole. Sixty-two samples remained after removal of markedly fractionated and highly altered rocks (i.e. SiO_2 outside the range 35–80%; MgO outside 3–9% and CaO outside 5–15%). In a Ti v. Zr diagram most samples plot in the overlapping mid-ocean ridge basalt (MORB), volcanic arc basalt (VAB) and within-plate basalt (WPB) fields (Fig 5a). However, samples from Drakovouni plot in the WPB field. In the Zr/Y v. Zr diagram (Pearce 1980) many sample from Drakovouni also plot in the WPB field (Fig. 5b). In the Cr v. Y diagram unfractionated basalts plot in the overlapping VAB–MORB and WPB–MORB fields, or the VAB field alone (Fig. 5c). Samples from Drakovouni again plot mainly in the WPB field. In the Ti/Cr v. Ni diagram (Beccaluva *et al.* 1979) samples are spread over the MORB and IAT fields. The least fractionated basalts were also plotted on MORB-normalized 'spidergrams' (Pearce 1980) and were found to exhibit near-MORB compositions of 'immobile' trace elements, commonly with a negative Nb anomaly (Degnan 1992). In several samples (sample 605, central Peloponnese and sample 421, eastern Peloponnese (Table 1)) no relative Nb depletion is apparent.

Many, but not all of the basalts from the mélange exhibit an apparent subduction influence, as shown by a depletion of high field strength elements (HFSE), relative to an enrichment in large ion lithophile elements (LILE). The basalts show a wide range of largely overlapping compositions. Overall, two types of basalts can be recognized: one is similar to MORB, but typically with a Nb depletion suggestive of a subduction influence; the other, as at Drakovouni, is more WPB-like.

The limited available evidence suggests that many of the basalts are Triassic in age, although some could be younger. For example, sparse extrusive igneous and pyroclastic rocks directly underlie Late Jurassic radiolarian cherts and interbedded hydrothermal manganese deposits, notably at Aroania (Robertson & Degnan 1998); these volcanic rocks could be post-Triassic. Noting the common subduction influence and the rare presence of plagiogranites, Pe-Piper & Piper (1991) suggested eruption in an intra-oceanic subduction setting within the Pindos ocean. Pe-Piper & Piper (1991) suggested that the Triassic volcanic rocks formed in a back-arc marginal basin related to a generally eastward-dipping subduction zone located in the south Aegean region (present coordinates). However, they suggested more recently (Pe-Piper & Piper 2002) that some of the Triassic volcanic rocks (e.g. high-Mg extrusive rocks and shoshonites) could relate to an additional westward-dipping intra-oceanic subduction zone, located within the eastern part of the Triassic Pindos ocean, near the Pelagonian zone (i.e. two opposing subduction zones would have existed).

As a simpler alternative we suggest that the geochemical variation in the Triassic basalts relates to variable-degree melting of heterogeneous subcontinental mantle that was modified by the addition of subduction fluids, possibly related to Hercynian subduction (Robertson & Dixon 1984; Dixon & Robertson 1993). This, however, would not apply to the WPB-like basalts (e.g. at Drakovouni). These are associated with oceanic (rather than rift-related)

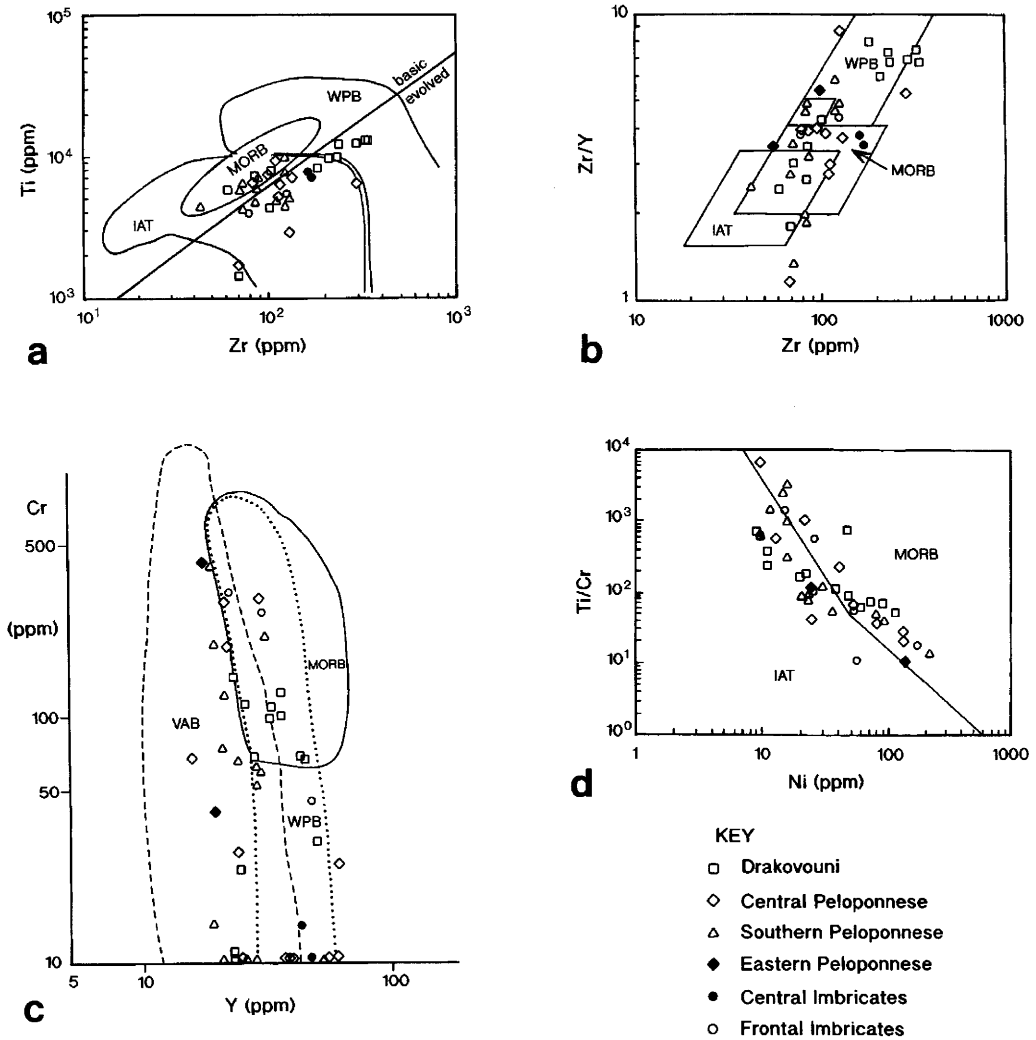

Fig. 5. 'Immobile' element geochemical plots of basic lava blocks from the mélange beneath the Pindos thrust in different areas; (**a**) Ti v. Zr; (**b**) Zr/Y v. Zr; (c) Cr v. Y; (**d**) Ti/Cr v. Ni. (See text for discussion.)

sediments and are found in a relatively easterly, distal location, and so these basalts might record fragments of accreted oceanic seamounts.

Assuming the above interpretation of the Triassic subduction-influenced volcanic rocks is correct, a restoration of the Pindos basin represented by the imbricate thrust slices containing mélange at the base (see below), suggests that these basalts are likely to have formed within 60 km oceanwards of the rifted continental margin edge. These basalts could, therefore, have erupted within a continent–ocean transition zone. Modern continent–ocean transition zones remain poorly understood, except for several examples such as the Atlantic margin of Iberia (e.g. Whitmarsh *et al.* 2001) and the Alps (Manatschal *et al.* 2003). This has been mainly explored by the drilling of topographic highs and it is not known whether extrusive igneous rocks are present in the intervening basins.

Genesis and emplacement of the Pindos thrust sheets

The Pindos ocean was consumed in an east-dipping subduction zone during Late Cretaceous–Palaeogene time (Robertson *et al.* 1991). The subduction zone reached the westernmost

Table 1. Selected chemical analyses of Triassic basic extrusive igneous rocks from the mélange beneath and within the Pindos thrust sheets at several localities in the Peloponnese. Analysis was by X-ray fluorescence according to the method of Fitton & Dunlop (1985). The data are given in full in Degnan (1994). LOI, loss on ignition.

Locality:	Drakovouni	Central Pelop.			E. Pelop.		S. Pelop.		
Sample no.:	276	453	454	588	605	421	423	356	399
SiO_2	51.36	49.09	50.13	46.64	48.7	45.56	52.28	50.65	50.19
Al_2O_3	18.3	17.12	16.02	14.81	18.1	14.8	15.4	17.48	16.55
Fe_2O_3	8.09	10.09	9.42	8.74	6.19	7.85	4.46	5.42	7.33
MgO	8.5	5.74	5.62	4.21	9.05	6.36	2.54	8.38	3.04
CaO	2.17	8.45	7.07	11.46	7.32	11.16	9.97	7.96	8.41
Na_2O_3	6.39	4.77	5.5	4.81	3.06	4.34	4.08	2.97	6.92
K_2O	0.12	0.15	0.19	0.95	0.63	0.16	2.69	0.26	0.06
TiO_2	1.37	2.01	2.07	1.06	1.18	0.79	0.75	1.21	0.69
MnO	0.15	0.2	0.18	0.11	0.15	0.15	0.1	0.12	0.23
P_2O_5	0.35	0.56	0.64	0.18	0.18	0.16	0.3	0.23	0.23
LOI	4.18	3.83	3.24	7.45	4.9	9.3	7.64	5.57	6.25
Total	101.6	99.99	100.08	100.42	99.46	100.62	99.95	100.24	99.9
Ni	114	38	20	136	85	141	25	83	10
Cr	157	107	73	328	201	475	41	149	7
V	152	248	210	310	244	267	194	233	172
Sc	24	38	28	37	40	42	16	46	37
Cu	34	36	59	20	60	4	44	85	18
Zn	64	79	69	53	73	61	54	67	107
Sr	324	166	388	403	397	145	307	357	156
Rb	0	1	2	13	7	3	22	0	1
Zr	183	237	297	82	89	58	101	122	73
Nb	9	10	14	75	4	3	7	6	4
Ba	53	83	28	3	93	42	53	99	15
Pb	1	0	3	3	4	2	1	0	5
Th	0	0	3	1	3	1	2	2	1
La	13	22	9	3	5	8	16	8	11
Ce	32	50	51	28	12	15	40	26	30
Nd	15	28	36	19	16	9	24	15	15
Y	23	35	43	21	23	17	19	21	55

part of the Pindos ocean in the NW Peloponnese soon after the accumulation of the Pindos flysch during Late Paleocene to Early or Mid-Eocene. The process of subduction–accretion, modified by post-accretionary tightening, gave rise to the present Pindos thrust stack.

Previously, the structure of particularly the Gavrovo–Tripolitza footwall in the NW Peloponnnese was considered in detail by Xypolias & Doutsos (2000). Also, within the Pindos thrust stack of the southern Peloponnese, Skourlis & Doutsos (2003) recognized a dense array of moderate-angle imbricate thrusts in the west, with broader, upright to moderately inclined thrust sheets in the centre, and inclined to recumbent folds in the east of the thrust stack. In the orogenic model of Doutsos *et al.* (2006), the Pindos thrust stack was subdivided into a 'pro-wedge' in the west, an 'uplifted plug' in the central Peloponnese (i.e. Chelmos massif) and a 'retro-wedge' in the east.

During this work (Degnan 1992) a detailed structural transect was carried out across the NW Peloponnese, combined with reconnaissance of adjacent areas (Fig. 2). This showed that the thrust belt in this area can be usefully subdivided into three subareas: the Frontal Imbricates, the Central Imbricates and the Eastern Imbricates, based on distinctive structural features. A detailed cross-section given in Figure 6.

Frontal Imbricates

The Frontal Imbricates are bounded to the west by the Pindos basal thrust, which at Alepohorio (Figs 2 and 7) dips at 20–30° eastwards, although

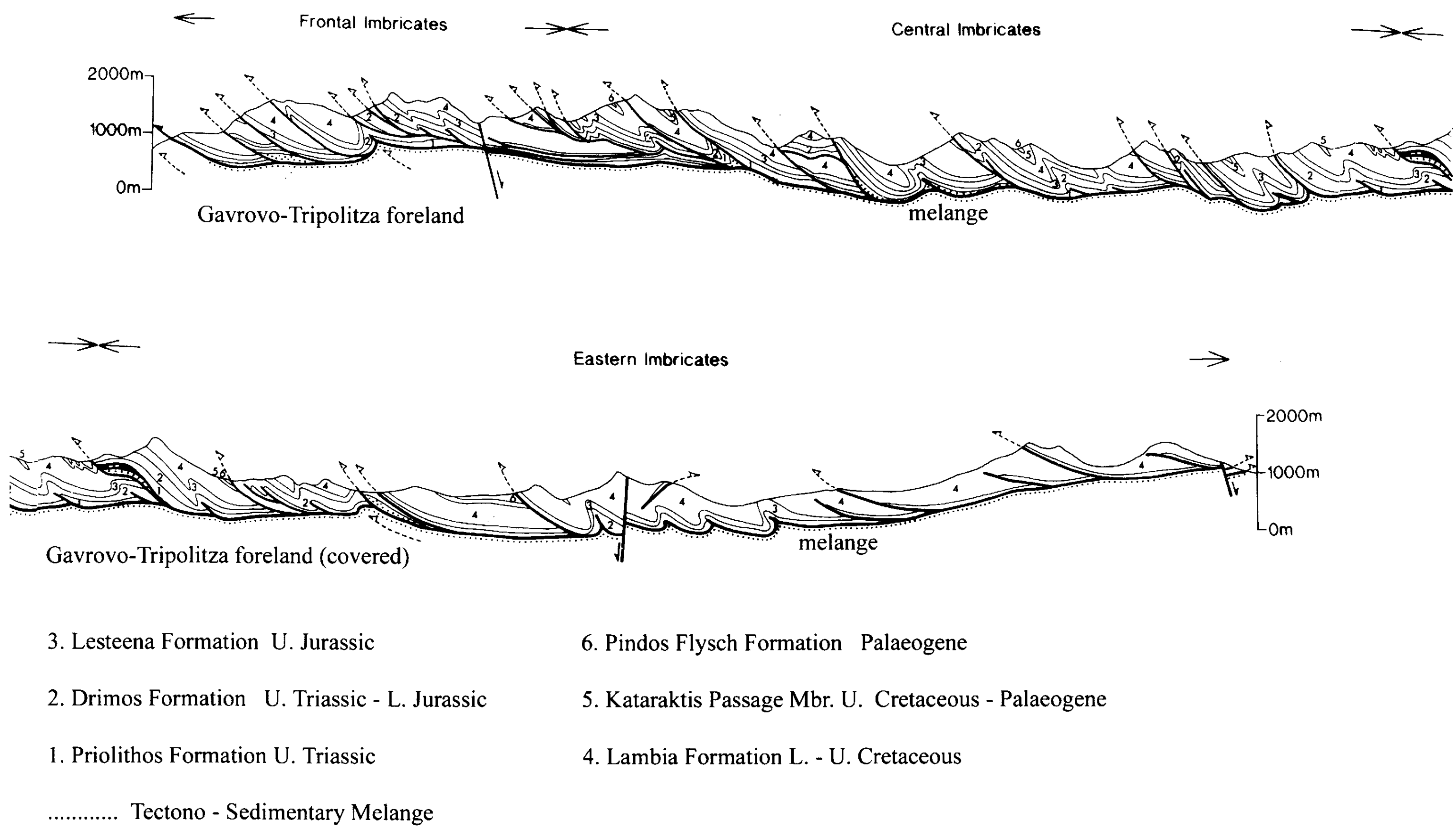

Fig. 6. Detailed structural cross-section of the traverse shown in Figure 3.

a discrete thrust plane is rarely exposed. The structurally lowest strata of the Pindos Group comprise a 'broken formation', but form more coherent thrust sheets upwards. There are marked variations in age of the stratigraphic units exposed in the lowermost thrust units. For example, Triassic sediments (Priolithos Formation) are exposed at Alepohorio in the north, whereas Upper Triassic–Liassic (Drimos Formation) and Upper Jurassic (Lesteena Formation) sediments are seen further south. A possible reason is that the basal thrust cuts variably up and down section within the Pindos hanging-wall section along strike.

Exposures in the frontal region are dominated by Cretaceous limestones (Lambia Formation). Throughout the frontal imbricates, bedding is generally moderately to steeply eastward dipping. The typical fold style varies from open concentric folds to medium-scale chevron folds, with gently plunging fold hinges (both NE- and SW-trending). The axial planes are moderately to steeply dipping. The trend of the fold-hinge lineations implies that the mean emplacement direction in the frontal imbricates was towards the NNW on the transect studied (Fig. 8).

The strata are disharmonically folded and numerous highly contorted folds and small thrusts are seen in the thicker-bedded units. Numerous fold trains of low-amplitude concentric folds are exposed. Fold inter-limb angles are generally wide, with wavelenghs of up to *c.* 5 m; axial surfaces are steeply inclined eastwards, to vertical. The thicknesses of largely intact sequences in the Cretaceous interval (Erymanthos Limestone Member) are estimated at 250–300 m. The apparent thicknesses of Cretaceous limestone reaches 500–600 m and reflect thickening as a result of folding and the development of intraformational duplexes.

The structural style of the frontal imbricates is exemplified at Alepohorio, where a large-scale synform (amplitude >300 m) is present in the footwall to a major thrust at the head of the Alepohorio valley (Fig. 7). The axial surface of this structure is horizontal to weakly eastward dipping and strikes towards 020°. Both limbs of the major synform are preserved on the southern side of the valley. The oldest sediments are exposed beneath the lower limb, whereas equivalent lithologies of the eastern limb may be dissected by a later thrust. Medium-scale folds are preserved in the lower limb. By contrast, an east-vergent, isoclinal fold with chevron folds is developed on the upper limb. On the north side of the valley, the upper limb of the macroscopic fold has been largely thrust-out, but the enveloping surfaces of a fold train (a series of chevron folds) define part of the synform. Mapping of the thrust sheets in the eastern part of the Frontal Imbricates demonstrates that a splay of minor thrusts is connected by branch lines to a major thrust fault in the footwall that converges to the south and north. The smaller thrust sheet corresponds to a hanging-wall cut-off at the present level of erosion.

Central Imbricates

Within the Central Imbricates (from just east of Platanitza to the Aroania thrust; Figs 2 and 6) the sediments are generally thinner bedded than in the Frontal Imbricates, reflecting more distal deposition (Degnan & Robertson 1998). The Central Imbricate folds range in amplitude and wavelength from <1 cm to >200 m. Disharmonic, medium-scale folds dominate the Jurassic interval (Lesteena Formation) and the overlying Cretaceous interval (Erymanthos Limestone Member). Chevron folds exhibit *c.* 60° interlimb angles, although some folds are in places strongly flattened. Larger-scale folds are well exposed in thickly bedded, competent, intraformational limestone beds within the Jurassic siliceous interval (Aroania Chert Member; e.g. at Livardzi and Drimos, Fig. 2). Folding there is highly irregular, with wavelengths of *c.* 100–300 m. Small-scale chevron folds and local box folds apparently formed prior to larger-scale folding, as the enveloping surfaces of fold trains are themselves folded. The longer-wavelength folds correspond to tightening of a 'single layer' to form irregular buckle folds.

Major thrust planes that duplicate large parts of the stratigraphy are generally not well exposed in the Central Imbricates. Mapping indicates that most of the major thrust faults are hanging-wall and footwall flats. Local hanging-wall or footwall cut-offs probably represent thrust deformation of pre-existing folds rather than thrusts cutting up through an undeformed 'layer-cake' sequence. Thrusting of previously folded strata also accounts for local younger-on-older stratigraphic juxtaposition. Many of the thrust faults die out along strike as doubly plunging structures. The maximum stratigraphic separation is in the centre of the mapped thrust traces, with displacement disappearing at both ends. Asymmetrical antiformal folds are seen at the surface (e.g. west of Livardzi). Shortening directions derived from mean fold-hinge lineations indicate movement towards 297–298° in the west and 284–293° in the east (Fig. 8).

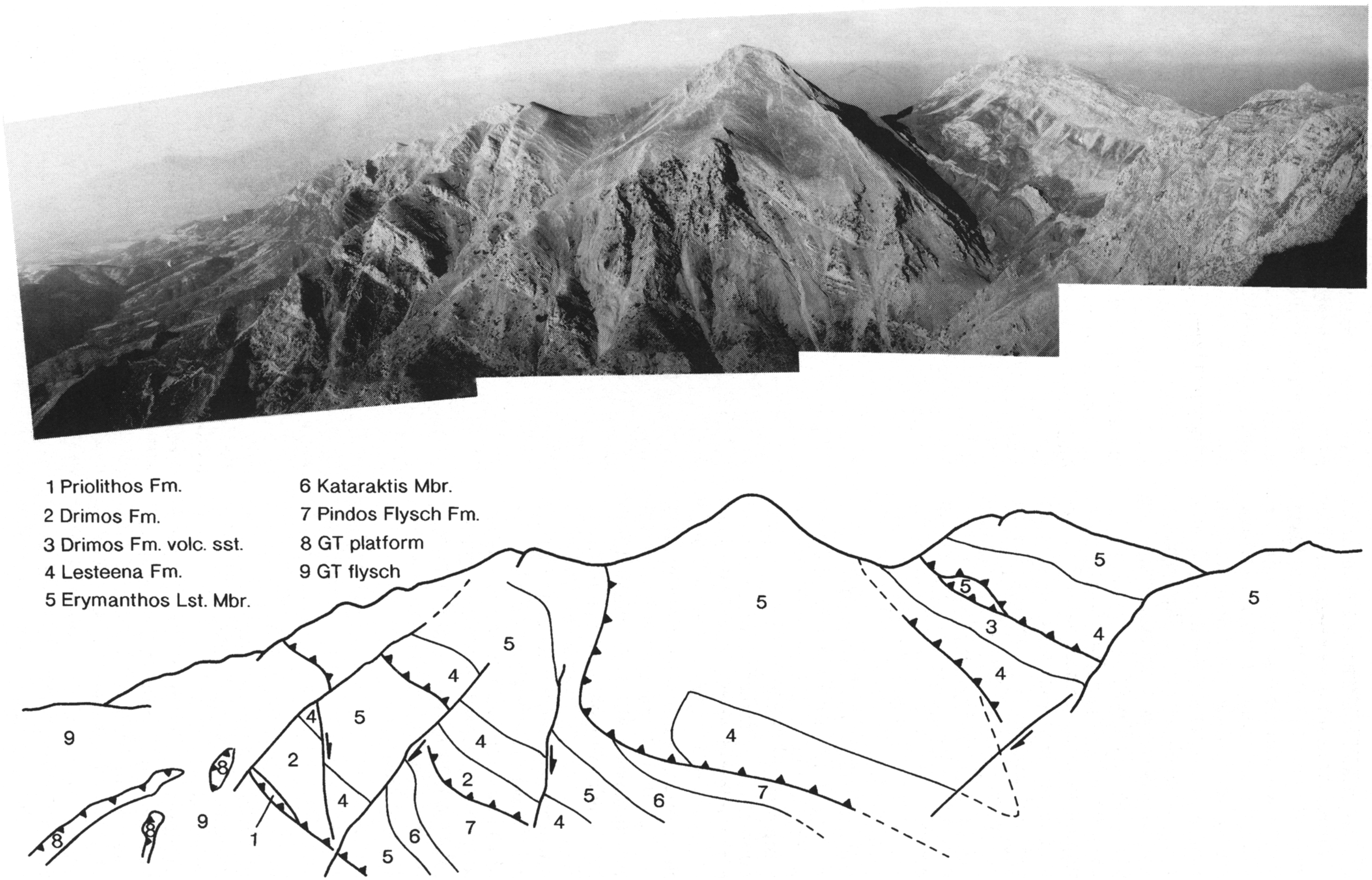

Fig. 7. Panorama of the north side of the Alepohorio valley section. Field of view *c.* 4 km across. The low lying sediments of the foreland basin, to the left of the collage, were deposited on neritic carbonates of the Gavrovo–Tripolitza zone and later overthrust by the Pindos zone. The well-defined hanging-wall cut-off and antiform in the centre of the photograph should be noted.

Eastern Imbricates

The Eastern Imbricates (from east of Aroania to west of the Chelmos antiform; Figs 2 and 6) differ from those further west as the spacing between major thrust faults that duplicate the stratigraphy is wider there. Bedding flattens on a map scale into more gently folded antiform–synform couplets, and higher stratigraphic levels (Cretaceous) are more widely exposed. Thrust planes are steep in the west, but become shallower eastwards (e.g. 20–30°; Fig. 6).

Medium-scale folds again form fold trains, with enveloping surfaces connecting fold inflection points and surfaces that were seen to be folded at several localities. Folded enveloping surfaces define the limbs of outcrop-scale folds, as in the Central Imbricates, although some folds may have resulted from later south-to-north directed compression; this would account for the observed wide spread of bedding and hinge lineation orientations.

In the eastern part of the Eastern Imbricates (Fig. 7) bedding is inclined both ESE and also westwards. Shortening directions derived from fold-hinge orientations imply a mean emplacement direction of 270°. Further west there is a 23° change in the inferred emplacement direction (Fig. 9). Minor thrusts show a wide spread of fault plane orientations, but with a concentration dipping ESE. Normal fault trends are also widely dispersed, although WSW–ENE-striking, high-angle fault planes are statistically significant.

One other notable feature is that Jurassic siliceous sediments (Aroania Chert Member) are generally the oldest unit exposed in the Eastern Imbricates (east of Kato Klitoria), compared with the Western and Central Imbricates in which the complete Triassic–Early Cenozoic stratigraphy is eposed.

Accretionary processes

The style of deformation records the processes of detachment of the sedimentary cover from the transitional–oceanic crust that was being subducted eastwards. We interpret the sequence of events as indicated below.

Initial compression within the Pindos ocean resulted in the generation of medium-scale folds. Continued shortening produced serial sinusoidal folds, which tightened into chevron folds within competent units. By contrast, incompetent units experienced largely bedding-parallel shear. Fold amplification and tightening continued until strain accommodation was achieved by flattening of sedimentary multilayers, and meso-scale folds were incorporated into larger-scale folds. During decoupling of the sedimentary cover from oceanic crust, thrust faults propagated upwards from a basal detachment along axial-planar lines of weakness, exploiting the previous large-scale folding. In places, it can be seen that transverse ramps and high-angle faulting have controlled the local structural style. These faults reflect differential stresses within the thrust sheets and are clearly coeval with the main phase of deformation. Some additional deformation may relate to fault-propagation folding. A good example of this sequence of events is suggested by the thrust/fold relations near Drimos (Fig. 2) in the Eastern Imbricates (Fig. 9). It should be noted that more recent NW–SE normal faulting is also present, reflecting neotectonic extension focused on the Gulf of Corinth.

A minimum estimate of the shortening across the Pindos thrust–belt can be made. To achieve an effective restoration, the base of the Cretaceous limestones (Lambia Formation) is extrapolated to link hanging-wall and footwall cut-offs (with minimum values). The restored line-length also takes into account the shortening caused by mesoscale folding and the volume loss through dissolution (estimated from stylolite occurrences). Medium-scale folding over a 100 m section was measured and shortening was estimated at three representative localities in each of the three imbricate regions (Alepohorio, Livardzi and Aroania). Taking account of the observed variation in pressure solution across the thrust belt the volume loss is estimated as 3.5% (with a possible range from 1 to 18%). The overall results suggest that there has been at least 60% shortening in the 43 km-wide traverse of the imbricate thrust sheets (Degnan 1992).

The Pindos allochthon extends eastwards from the Eastern Imbricates (east of Chelmos) for a further 99 km. Of this, the 56 km wide 'Table d'Arcadie' (Dercourt 1964) consists almost entirely of the flat-lying, strongly deformed, Late Cretaceous pelagic carbonates (Erymanthos Limestone Formation). It is difficult to evaluate the shortening in this area; 50% was suggested by Fleury (1980) but it may be much more. The total shortening across the exposed Pindos thrust belt is estimated as at least 55% and more probably *c.* 67%, suggesting that this part of the pre-existing Pindos basin was at least 300 km wide. Green (1982) estimated orogenic contraction at 63% for the Pindos thrust stack north of the Gulf of Corinth. A similar result was obtained by Xypolis & Doutsos (2000). In addition, an unknown amount of oceanic crust was subducted, and so the Pindos ocean basin

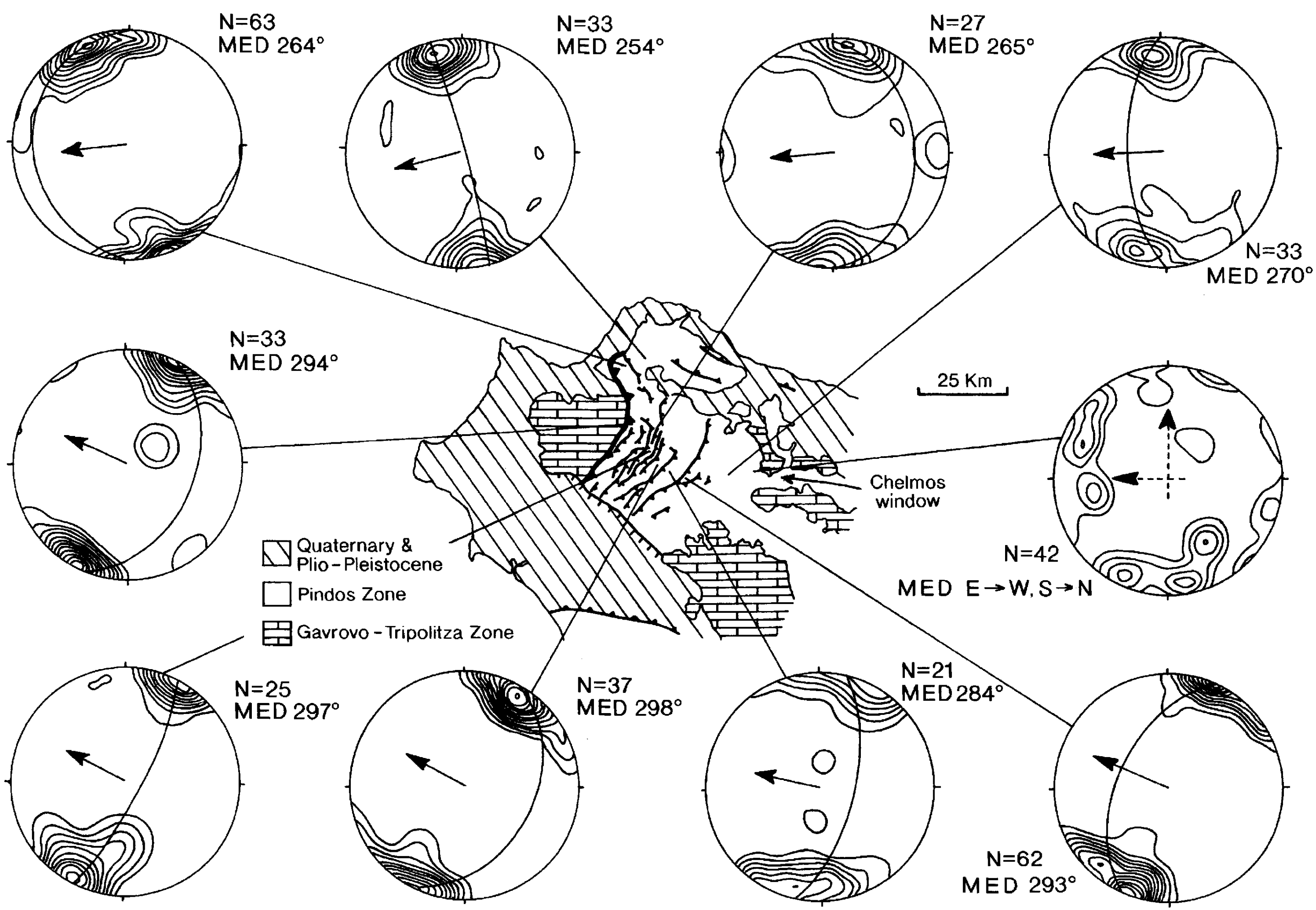

Fig. 8. Summary of fold data indicating direction of emplacement of the Pindos thrust sheets.

was probably much wider (Robertson *et al.* 1991).

The period of deformation when the preserved part of the Pindos ocean represented by the thrust belt in the NW Peloponnese was telescoped from >300 km to 99 km is equivalent to an average shortening rate of *c.* 5.8 mm a^{-1}. This rate is similar to that for some other mountain systems; e.g. 5 mm a^{-1} in the Canadian Rockies (Elliott 1976); or 6 mm a^{-1} in the Pyrenees (Williams & Fischer 1984).

The development of the imbricate fan in the west in contrast to the 'Table d'Arcadie' region in the east probably relates to the presence of relatively competent thick and homogeneous Cretaceous pelagic limestones in the east. This compares with the more heterogeneous, generally older sequence further west; in these, incompetent layers (e.g. shale) provided the rheological contrasts necessary for the development of a linked thrust stack system.

In the frontal region of the thrust belt (Western Imbricates) there is a curvature of the thrust traces and a radial arrangement of both transverse faults and emplacement vectors (Fig. 8). A possible explanation is that the orogenic front was buttressed by a promontory of the Apulian margin, now located in the Chelmos region. This collapsed when it was overridden by the Pindos thrust sheets; insufficient topographical relief perhaps remained to influence thrust geometry. Emplacement structures were influenced by similar continental margin promontories elsewhere, including the Appalachians, the Rocky Mountains and Oman.

Evolution of the foreland

The Pindos thrust complex was emplaced over the eastern margin of the Apulian continent in post-Early Oligocene time, as dated by the youngest sediments exposed in a thick underlying clastic foreland basin succession of the Gavrovo–Tripolitza zone (Kamberis *et al.* 2006; Fig. 1). The underlying carbonate platform documents a Triassic rift history, coeval with the opening of the Pindos ocean, and a passive margin history contemporaneous with the deep-sea sedimentation in the Pindos ocean. The Gavrovo subzone forms the footwall of the Pindos thrust sheets in the NW Peloponnese, whereas the Tripolitza subzone is exposed in tectonic windows through the Pindos thrust stack further east (e.g. Chelmos). During the Mesozoic the Gavrovo–Tripolitza zone was bordered to the west by the Ionian zone, an intracontinental rift basin (Blumenthal 1933; Renz 1955; Dercourt 1964; De Wever 1975; Fleury 1980; Thiébault 1982; Clews 1989).

Rift–passive margin evolution

In the NW Peloponnese the lowest exposed stratigraphic unit is composed of Triassic low-grade siliciclastic metamorphic rocks, rich in mica (Zarouhla Group; De Wever 1975). This sequence includes metavolcanic rocks (tens to several hundred of metres thick), mainly composed of trachyandesite, metabasalt and metasiliceous tuffs (Aghios Ilias Eruptive Formation; De Wever 1975). Chemically, the extrusive igneous rocks typically show a trace-element composition suggestive of a subduction influence, similar to those of the mélange, discussed above, although again interpretations vary (e.g. Pe-Piper 1982, 1998; Dornsiepen & Manutsoglu 1996). The low-grade metamorphic sequence in the NW Peloponnese can be generally correlated with the mainly Triassic Tyros Beds exposed in the southern and central Peloponnese (Ktenas 1926; Lekkas & Papanikolaou 1980; Dittmar & Kowalezyk 1991).

The Mesozoic carbonate sequence of the overlying Gavrovo carbonate platform in the NW Peloponnese begins with poorly dated shallow-water limestones, gypsum and brecciated dolomite ('gargneoule'), commonly folded and thrust imbricated. This is then overlain by a well-documented shallow-water carbonate platform sequence, several kilometres thick (Dercourt 1964; De Wever 1975). A Late Triassic–Early Eocene sequence (Tripolitza subzone) is well exposed east of Chelmos Mountain. This accumulated on a carbonate platform, with numerous breaks in deposition, erosion surfaces and bauxite development, especially during the Late Jurassic (De Wever 1975) and Cretaceous (Tsalia-Monopolis 1977). Hardgrounds and bauxites that developed in Mid-Eocene time in some areas indicate emergence at the highest levels of the platform sequence (Fleury 1980).

Foreland basin development

A thick Upper Eocene–Lower Oligocene clastic sedimentary sequence overlying the Gavrovo–Tripolitza zone documents the development of a foreland basin related to flexural warping and then collapse of the carbonate platform as the Pindos allochthon was overthrust westwards.

Within the more westerly areas of the carbonate platform (Gavrovo subzone), neritic carbonates of Late Mesozoic–Early Cenozoic age pass through an interval of redeposited carbonate breccias and then into a thick

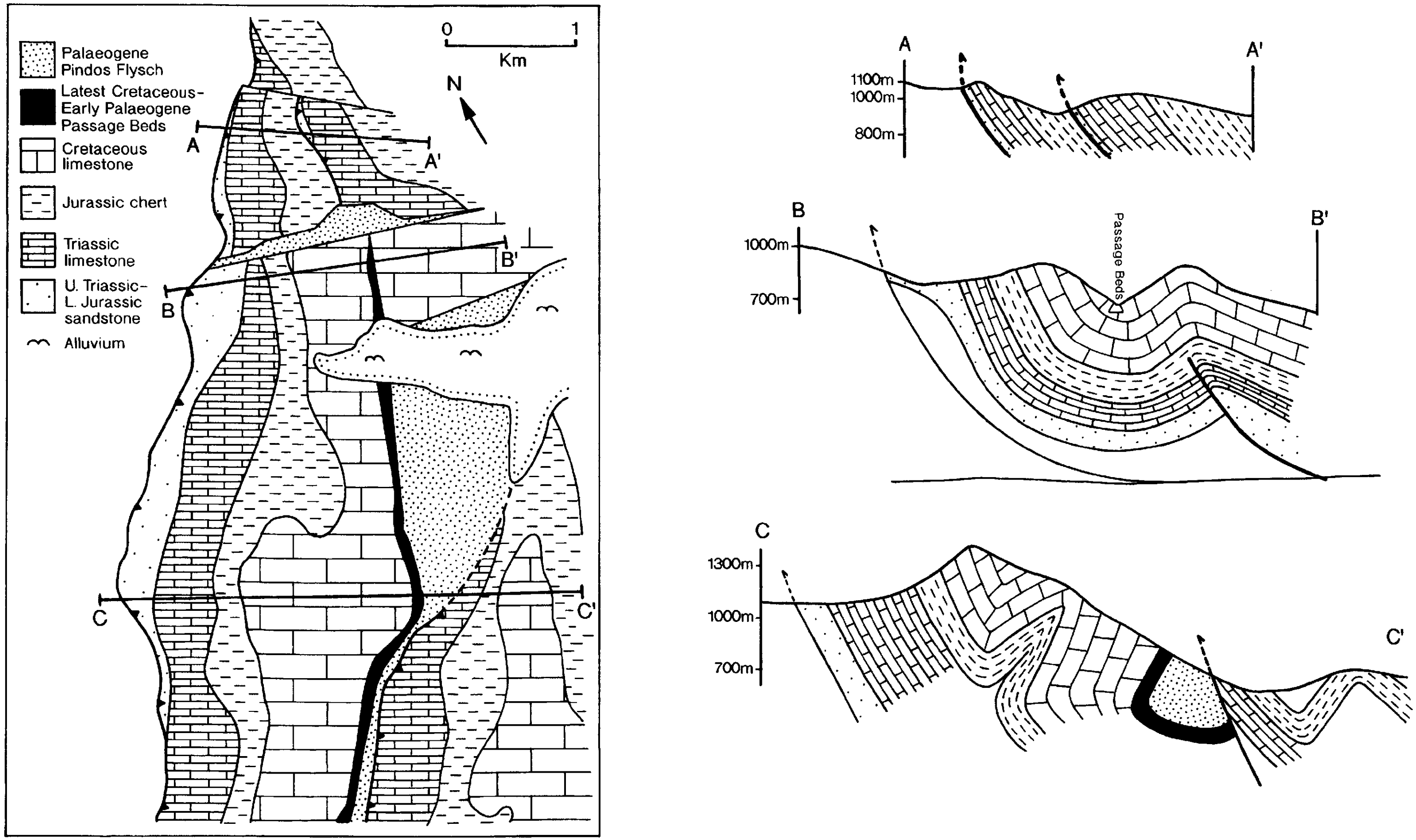

Fig. 9. Geological map and cross-sections of the thrust stack near Drimos in the eastern area (Fig. 2). The lines of section show the structural variation along strike over several kilometres. These sections also illustrate the style of folding associated with faulting and detachment that is thought to characterize the Pindos thust stack.

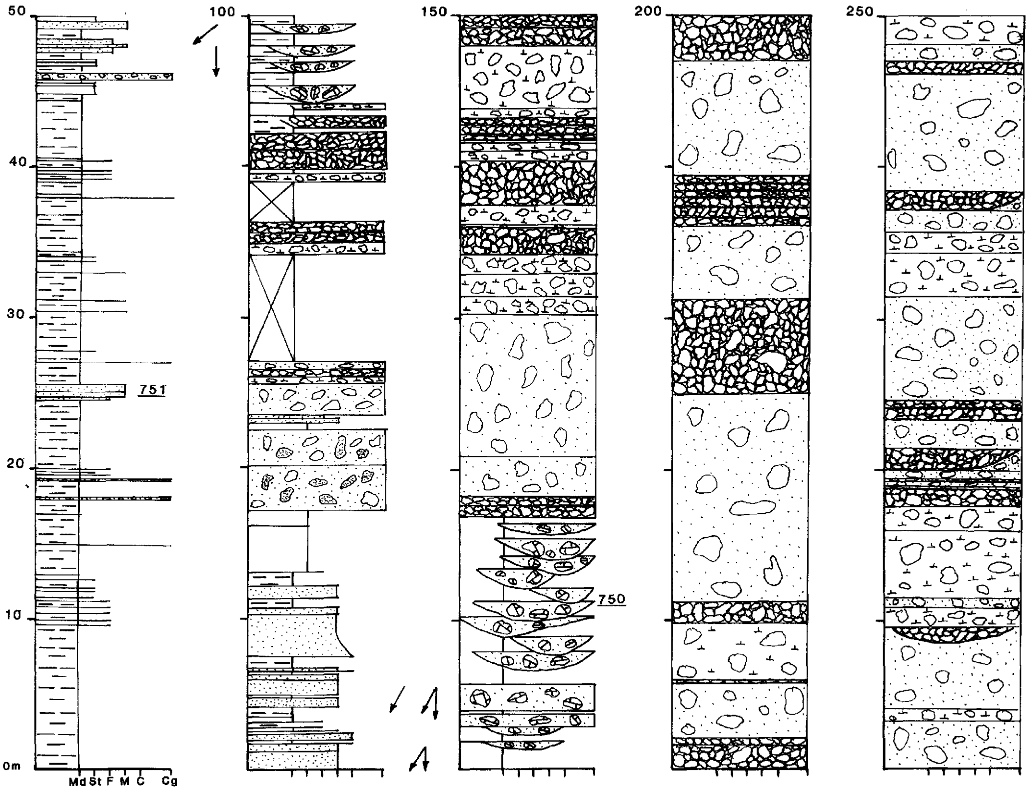

Fig. 10. Sedimentary log of the Early Tertiary succession of the highest levels of the Gavrovo–Tripolitza zone beneath the Pindos thrust in the Alepohorio area. Key: mudstones (broken horizontal lines) pass upwards into sandstones (stipple) then into conglomerates. Clasts range from clast-supported (densely packed) to matrix sorted (with a sandy matrix). Depositional units are tabular or channelled (with scoured bases). The conglomerate matrix towards the top is calcareous in places.

siliciclastic sequence, known as the Gavrovo–Tripolitza Flysch, of Eocene–Oligocene age (Lekkas 1980; Bassias & Lekkas 1988).

Near the Pindos thrust front (e.g. at Alepohorio; Fig. 2), a thick siliciclastic sequence above platform carbonates (Fig. 10) begins with mudstones, in beds up to 1.2 m thick, interbedded with sandstone, minor siltstone and rare pebblestone with well-rounded limestone clasts. Sandstone turbidites exhibit well-developed grading, flute casts, groove casts and micro-cross-lamination. The main lithology is litharenite, with abundant grains of quartz, carbonate and subordinate feldspar, chert and volcanic rocks. The siliciclastic sequence culminates in a 15 m thick interval of mainly medium-grained, thick-bedded sandstone, in beds up to 2.7 m thick.

The turbiditic sequence passes upwards into channelized conglomerates and breccias (*c.* 230 m thick). Individual beds are mainly massive (up to 14 m thick) and include large clasts. Three end-member conglomerate types are present. First, there are clast-supported conglomerates, in beds up to 6.2 m thick with well-rounded clasts (up to 90 cm × 50 cm in size), composed of micritic limestone and carbonate grainstone. Second, there are matrix-supported conglomerates, in beds up to 4 m thick, that contain large clasts (up to 50 cm in size), set in a fine-grained carbonate matrix. Third, there are matrix-supported conglomerates that are lithologically variable with clasts of sandstone, carbonate, shale and chert, set within a green pelitic matrix. A Late Eocene, or younger, age for these conglomerates is indicated by the presence of *Actinocyclina* sp. and *Nummulites* sp. within a clast of neritic limestone (C. Betzler, pers. comm.). This sequence (Fig. 10) terminates with >200 m of strongly deformed siliciclastic turbidites beneath the Pindos thrust.

Similar turbidites elsewhere in the NW Peloponnese were dated as Oligocene (Izart 1976). In the central Peloponnese, other conglomerates are dated as Late Oligocene (Mansy 1971). The massive conglomerates in the NW Peloponnese contain few palaeocurrent indicators. However, facies equivalents elsewhere in western Greece typically indicate north–south paleocurrents (Izart 1976), as seen north of the Gulf of Corinth (Lash 1988; Leigh 1991).

The redeposited carbonate breccias can be explained by mass wasting related to break-up and subsidence of the platform that was probably fault controlled. Breccia clasts locally contain a Triassic or Jurassic fauna (Fleury 1980), suggesting that relatively deep levels of the platform were exposed and eroded.

Elsewhere, within the more distal Tripolitza subzone (e.g. at Chelmos), a transition from neritic carbonate to sandstone turbidites is marked by alternations of carbonates and siltstones–fine sandstones ('couches de passage') of Eocene (upper Lutetian–lower Priabonian) age (Dercourt 1964). The uppermost neritic carbonates contain only benthic Foraminifera (e.g. *Discocyclina* sp.), whereas the overlying siltstones contain a mixture of benthic and planktonic Foraminifera, indicating deeper-water deposition. In some areas neritic carbonates of Mid-Eocene age are directly overlain by medium- to coarse-grained turbiditic sandstones and mudstone, up to several hundred metres thick, e.g. around Drakovouni.

Recently, Kamberis *et al.* (2006) have carried out a detailed micropalaeontological and sedimentological study of the turbiditic sediments of the both the Gavrovo–Tripolitza and Ionian basins in the NW Peloponnese using a combination of field, well and seismic data. The thickness of the Gavrovo foreland basin sediments is estimated as up to 3.7 km from subsurface information. Deposition of the Gavrovo turbiditic sediments was found to range from Late Eocene (locally), to Early Oligocene. As noted above, deposition began with relatively fine-grained deep-water sediments, followed by a much coarser succession of channelized conglomerates and sandstones, disorganized conglomerates and massive sandstones; an upper finer-grained interval is also reported across the NW Peloponnese.

Following a prolonged period of passive subsidence, the Gavrovo–Tripolitza carbonate platform emerged during Mid-Eocene–Late Eocene time, probably reflecting westward migration of a flexural forebulge ahead of the advancing Pindos allochthon. Erosion locally reached deep levels of the platform that were locally exposed, possibly accompanied by faulting. Carbonate breccias were shed into deep water, especially near the former shelf-slope break. The appearance of deep-water siliciclastic turbidites in Late Eocene-Early Oligocene time indicates that flexural collapse of the platform had by then taken place to create a foredeep (Leigh 1991; Kamberis *et al.* 2006). The main sediment source was the Pindos allochthon in view of the abundance of chert and other lithologies typical of the deep-water Pindos sequences. The coarsening- and thickening-upward trend of the siliciclastic turbidites as a whole reflects the advance of the overthrust load. The coarse massive conglomerates rich in neritic carbonate clasts in the upper part of the sequence probably record advanced break-up and erosion of the carbonate platform prior to overthrusting. A similar flexural response to overthrusting is seen elsewhere, including many comparable Tethyan settings (e.g. Oman; Glennie *et al.* 1973; Robertson 1987).

Deformation of the Gavrovo–Tripolitza foreland

The high-level structure of the Tripolitza subzone in the east is well exposed in a large tectonic window on Chelmos Mountain where the Pindos basal thrust is broadly folded (Fig. 2). This large-scale fold developed adjacent to an extensional fault in the platform, possibly related to flexural collapse; this was later inverted prior to, or during, emplacement of the Pindos thrust sheets. The Pindos units were, therefore, finally transported over an already folded and internally thrust-imbricated foreland, which could have helped trigger the development of frontal ramps within the overriding Pindos allochthon. Following emplacement, large-scale folding of the Pindos and Tripolitza units probably reflects tightening of the Pindos suture zone. The Pindos basal thrust almost directly overlies the platform carbonates, as at Chelmos, or is first underlain by several tens of metres of intervening turbiditic sediments, as at Drakovouni, ~20 km to the south. The basal décollement forms a thrust-flat in the south, and cuts up-section further north.

Further west, deep-level thrust imbrication of the Gavrovo carbonate platform and its overlying foreland basin has been revealed by well and seismic reflection data (Sotiropoulous *et al.* 2003; Kamberis *et al.* 2006). Large listric thrusts are inferred to root towards the basal décollement of the underlying Ionian zone sediments. Thrusting of the Pindos thrust stack over the foundered Gavrovo foreland basin took place following

Early Oligocene turbidite deposition within this basin. The Gavrovo–Tripolitza zone was, in turn, thrust over the Ionian zone by the Pliocene following the termination of clastic sedimentation in the Ionian basin (Underhill 1989; Clews 1989).

Discussion: regional evidence for the Pindos ocean

The Pindos ocean rifted in the Early–Mid-Triassic and began to spread in Late Triassic time bordered to the west by the Gavrovo–Tripolitza carbonate platform. The continental–ocean transition zone comprised mainly subduction-influenced basic–intermediate–acidic lavas and volcaniclastic sediments in our interpretation. Deep-water sedimentation was in progress from the Early–Mid-Triassic onwards, as documented by well-dated sediments from, for example, Vardoussia north of the Gulf of Corinth (Celet 1962; Bernoulli & Laubscher 1972; Ardaens 1978). In Albania (Krasta–Cukali zone), Permian clastic sediments (Verucanno facies) and Lower Triassic pelagic carbonates–volcanic rocks occur at the base of a Mesozoic–Lower Cenozoic succession, similar to the Pindos Group in the Peloponnese (see Robertson & Shallo 2000). In former Yugslavia, correlative units, notably the East Bosnian–Durmitor unit in northen Montenegro, Serbia–Kosovo, eastern Bosnia and Croatia, again show evidence of Triassic rifting, with a peak of volcanism in Ladinian time (Karamata & Vujnović 2000; Knezević & Cvetović 2000; Karamata 2006). In Croatia, Late Permian rifting, with block faulting and localized evaporite deposition was followed by mainly shallow-water carbonate deposition. The eruption of rift volcanic rocks (basalts, andesites and rhyolites) climaxed in Ladinian time, and had mainly ended prior to Norian time (Pamić 1984). Platform rocks are locally cut by rift-related syenite, diorite and gabbro (Pamić *et al.* 2002). A subduction influence on volcanism is widely inferred (see Karamata 2006, and references therein), as for the Triassic rocks of Greece (Pe-Piper & Piper 2002).

In Greece, large blocks of ammonite-rich pelagic limestone occur beneath the Pindos thrust front, at Glafkos in the NW Peloponnese and Meghdovas to the north of the Gulf of Cornith (Fleury 1980). This Ammonitico Rosso has been interpreted as accumulating on marginal highs (fault blocks) located along the western margin of the Pindos ocean, within the continent–ocean transition zone (Degnan & Robertson 1998; Pe-Piper & Piper 2002). In former Yugoslavia, Upper Triassic shallow-water carbonates are overlain by Lower Jurassic condensed Ammonitico Rosso in a similar setting.

Sedimentation in the western Pindos ocean became generally more distal eastwards, which, combined with palaeocurrent data, indicates sediment derivation from Apulia. Widespread Mn deposition within non-calcareous cherts of Late Jurassic–Early Cretaceous age (Leestena Formation) possibly relate to hydrothermal activity at a spreading centre (Pe-Piper & Piper 1989). Supra-subduction spreading is inferred by some workers to have taken place above a westward-dipping subduction zone in the Mid-Jurassic, followed by northeastwards emplacement of ophiolites (e.g. Pindos, Vourinos) onto the Pelagonian continent by Late Jurassic time (Robertson *et al.* 1991; see also Rassios & Moores 2006; Smith 2006, and references therein). However, other workers continue to believe that the Jurassic ophiolites were emplaced from the Vardar zone further east (e.g. Dercourt *et al.* 2000; Stampfli *et al.* 2001).

The width of the Pindos ocean remaining after the ophiolite emplacement depends on the relative width of the Vardar ocean in the NE versus the Pindos ocean in the SW, both being located between the Eurasian and North African continents. The Pindos ocean remaining after the Mid–Late Jurassic ophiolite emplacement was relatively narrow in the north but widened southwards into southern Greece (Robertson *et al.* 1991). Support for this includes the apparent absence of a Pindos ocean west of the Scutari–Peć transverse lineament, the presence of ophiolitic detritus (e.g. chrome spinel) in Upper Cretaceous (Cenomanian–Turonian) turbidites associated with the Apulian margin (e.g. Klitoria Sandstone Member or Premier Flysch de Pinde), north of the Gulf of Corinth, and the apparent absence of Late Cretaceous–Early Cenozoic arc magmatism related to subduction of the Pindos ocean. Assuming the Pindos ocean was relatively narrow by the Early Cretaceous, most of the remaining separation between Eurasia and North Africa was accommodated by the Vardar ocean (see Sharp & Robertson 2006).

The Pindos ocean was closing in Albania and northern Greece in the Late Cretaceoous–Paleocene, giving rise to the Pindos Flysch in these northerly areas. This sediment contains material derived from an exhumed accretionary wedge (i.e. blue amphibole) and the overriding Pelagonian continent (e.g. ophiolitic debris) (Faupl *et al.* 1996, 1998, 2002; Richter & Müller 2002). The timing of onset of eastward subduction in southern Greece (south of the Gulf of Corinth) remains poorly constrained but it is likely that there too subduction was active at least from Mid-Cretaceous time. We infer that the Pindos ocean was subducted during an eastward, scissors-like closure of an originally southward-widening ocean, and thus the time of final collision and suturing decreased southwards (Robertson *et al.* 1991).

In the Peloponnese there is little record of the eastern Pindos ocean, possibly because the subduction décollement was located at a high level resulting in complete subduction, or an early accretionary wedge was overridden by the Pelagonian continent, a form of subduction erosion. As subduction continued the décollement level lowered to near, or above, the igneous basement–sediment interface, allowing the Cretaceous pelagic carbonates of the Eastern Imbricates (Lambia Formation) to be accreted while oceanic basement was subducted. A thickening wedge of passive margin sediments, located *c.* 60 km from the rifted margin, then entered the trench and the décollement level moved to a relatively lower stratigraphical level again, allowing the entire sedimentary succession and fragments of 'transitional' igneous crust and possible seamounts (e.g at Drakovouni) to be accreted within the mélange.

There is a surprising absence of thick clastic sediments (turbidites and debris flows) derived from the overriding accretionary prism in a subduction–trench setting, in contrast to the Pindos flysch in northern Greece. Such sediments were perhaps originally deposited in the subduction trench but were later detached and bulldozed ahead of the advancing accretionary prism and later reworked as Pindos-derived sediment into the very thick Gavrovo foredeep.

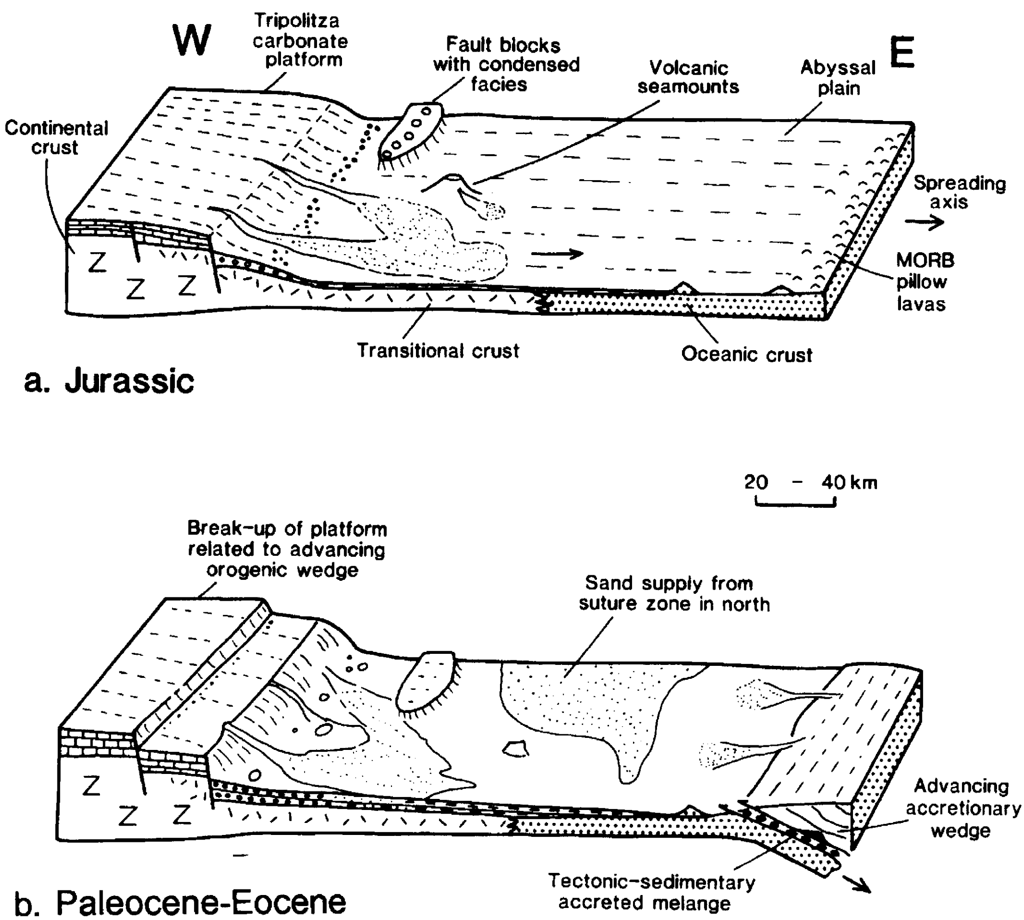

Fig. 11. Schematic reconstruction of the western margin of the Pindos ocean in the NW Peloponnese. (**a**) Mesozoic rifted margin of the Pindos ocean; (**b**) Early Cenozoic eastward subduction. (See text for explanation.)

Conclusions

The Pindos zone in the NW Peloponnese is interpreted as an oceanic basin that rifted during the Triassic, reaching its maximum width in Jurassic–Cretaceous time (Fig. 11). This was followed by eastward subduction beneath the Pelagonian continent during Mid-Cretaceous?–Palaeogene time. Triassic to Palaeogene sedimentary successions of the rifted passive margin become more distal towards the east away from the Apulian continent. The latest sediments to accumulate, turbidites of Paleocene–Late Early Eocene age (Pindos flysch), were mainly derived from north of the Gulf of Corinth. The geochemistry of Triassic basaltic rocks within the mélange, like the Triassic volcanic rocks in many adjacent areas, including the foreland, is indicative of a subduction influence, related to either contemporaneous or pre-existing (Hercynian?) subduction. We suggest that this volcanism was related to variable melting of a heterogeneous subcontinental mantle located within a *c.* 60 km wide continent–ocean transition zone. Igneous crust and deep-sea sediments were detached from the continent–ocean transition zone, along with probable seamount fragments, and incorporated as blocks into a mélange entrained at the base of, and within, individual thrust sheets. The deep-sea sedimentary cover of the Pindos ocean was accreted to form a imbricate stack of thrust sheets in post-Early to Mid-Eocene time. As the Pindos allochthon approached the Apulian continental margin, represented by the Mesozoic–Early Cenozoic Gavrovo–Tripolitza carbonate platform, the footwall flexed upwards in the Mid-Eocene, then collapsed beneath the advancing thrust load. Siliciclastic turbidites accumulated in a foredeep constructed on the downflexed Gavrovo carbonate platform during the Late Eocene–Early Oligocene, prior to overthrusting by the Pindos allochthon. The Gavrovo–Tripolitza zone was itself imbricated and thrust westwards over the former Ionian rift basin by the Pliocene. Post-accretion compression and strike-slip related to post-accretion tightening of the Pindos thrust stack was followed finally by neotectonic extensional faulting.

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