BEITRÄGE ZUR REGIONALEN GEOLOGIE DER ERDE

Georgia Pe-Piper
David J. W. Piper

The igneous rocks of Greece

The anatomy of an orogen

GEBRÜDER BORNTRAEGER

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Preface

This book has developed from our many years of field and laboratory work on a wide range of Greek igneous rocks: from Carboniferous granites to Triassic ophiolites to Miocene shoshonites to andesitic lavas erupted in historic times. Within the borders of Greece, there is a complete Wilson tectonic cycle of oceanic rifting, spreading, subduction and collision. Greece is a microcosm of the tectonic processes and igneous products that occur in many orogenic belts, particularly those of the former Tethyan oceans. Many geologists have been attracted to Greece not only by its rocks, but by the delights of the country and its people, but as a result, the scientific literature on Greek rocks is widely scattered and in many languages. We hope that this account of the igneous rocks of Greece will be of interest to all those who seek to understand the distribution and origins of igneous rocks in orogenic belts. At the same time, we hope to perform a valuable service in synthesising a literature that is relatively inaccessible to many workers. At the time of writing, the understanding of the tectonics of Greece is advancing rapidly. On the other hand, we hope that our descriptive synthesis of the actual rocks of igneous origin will stand the test of time. In general, we have not included information appearing in the literature after the year 2000.

We owe thanks to many friends who have worked with us in the field and laboratory over the years. We thank G. Christofides, M. Higgins and V. Tsikouras for their critical review of individual chapters and V. Jacobshagen for reviewing the entire book. Sharon Weaver, Don Horton, Gilles Dessureau and Steve Ingram prepared or edited many of the figures and Steve Ingram, Chrissoula Baltas and Lila Dolansky assisted with the final checking of the book. DJWP thanks the Geological Survey of Canada for their broad-minded tolerance of his scientific work in Greece, which has been largely carried out during vacations. GPP thanks Saint Mary's University for provision of research facilities and the Natural Sciences and Engineering Research Council of Canada for research funding. We dedicate this book to our children, Tina and Liza, who during their childhood saw almost as many rocks as we did.

GEORGIA PE-PIPER
DAVID J.W. PIPER
Halifax, 2001

Contents

Abbreviations and acronyms

AFC	Assimilation with fractional crystallisation
ASI	Aluminium saturation index
HFSE	High field strength elements
HREE	Heavy rare-earth element(s)
IAT	Island-arc tholeiite
IGME	Institute of Geology and Mineral Exploration (Greek Geological Survey)
IUGS	International Union of Geological Sciences
Ma	Million years (age)
Myr	Million years (duration)
MORB	Mid-ocean ridge basalt
NHRL	Northern hemisphere reference line (for Pb isotopes)
LILE	Large-ion lithophile element(s)
LREE	Light rare-earth element(s)
Mg#	Magnesium number
OIB	Ocean-island basalt
P-T-t	Pressure-temperature-time
QAPF	Quartz - alkali feldspar - plagioclase - feldspathoid
REE	Rare-earth element(s)
SSZ	Supra-subduction zone (ophiolite)
WPB	Within-plate basalt
$element_N$	Normalised abundance of an element
ε_{Nd}	Standardised Nd isotope ratio
β	Stretching factor
T_{DM}	Model age based on the depleted mantle model for Nd isotopes

Mineral abbreviations generally after Kretz (1983), including

Ab	albite
Ac	acmite
Amp	amphibole
An	anorthite
Bi	biotite
Cpx	clinopyroxene
Di	diopside
En	enstatite
Fs	ferrosilite
Hbl	hornblende
Hd	hedenbergite
Ilm	ilmenite
Kfs	potassium feldspar
Mgt	magnetite
Ms	muscovite
Ol	olivine
Opx	orthopyroxene
Or	orthoclase
Phl	phlogopite
Pl	plagioclase
Prg	pargasite
Qtz	quartz
Sal	salite
Spl	spinel
Ts	Ca-Tschermak's molecule
Wo	wollastonite

1 Introduction

1.1 Igneous rocks within the Hellenide orogen

Greece lies at the southern end of the Balkan peninsula (Fig. 1) and the Hellenide orogen forms part of the Alpine-Himalaya belt that formed by the destruction of Tethys. The present Hellenic subduction zone and South Aegean island arc (Fig. 2) mark the subduction of the African beneath the Eurasian plate. Behind this subduction zone, the Aegean area is now one of the most rapidly extending areas of continental crust in the world.

Within its borders, Greece has a wide range of igneous rocks that record a varied tectonic history through a complete Wilson cycle of oceanic rifting, spreading, subduc-

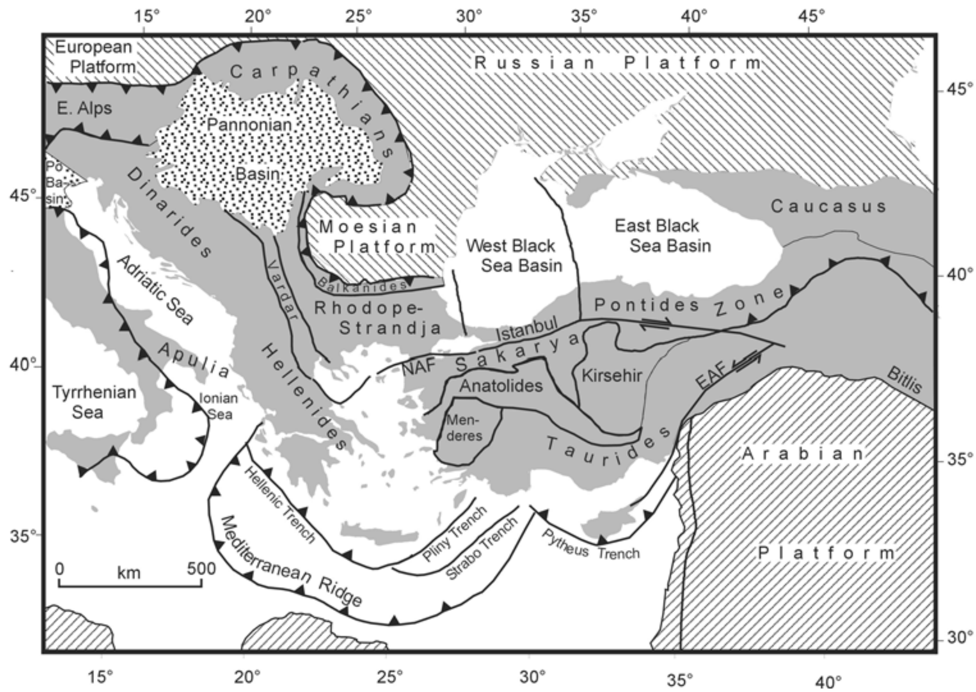

Fig. 1. The Hellenide orogen in its regional context. EAF = East Anatolian Fault, NAF = North Anatolian Fault.

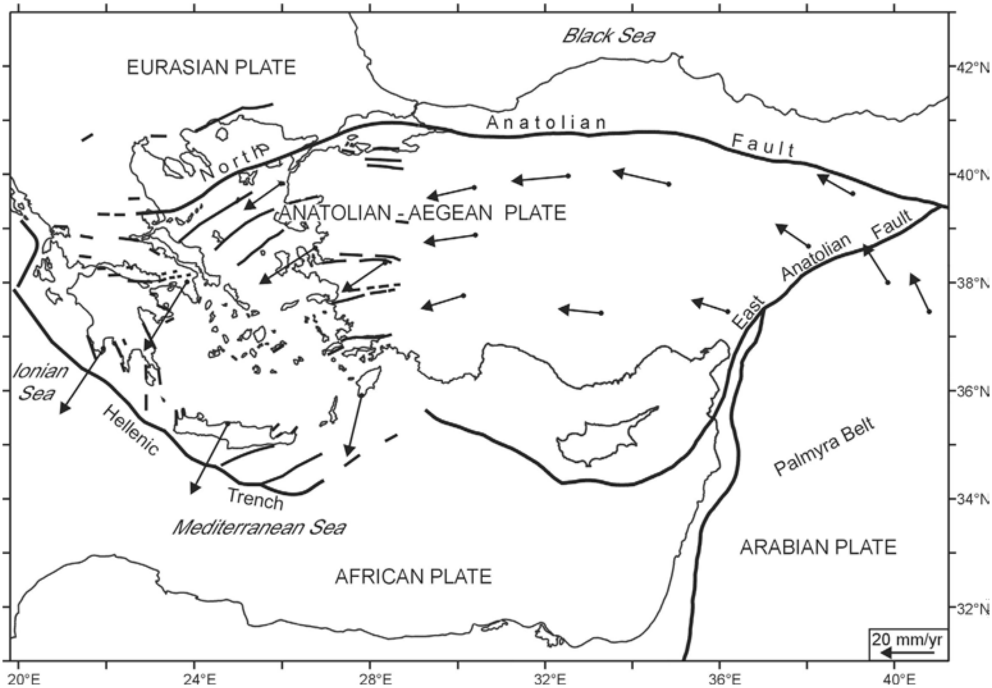

Fig. 2. The present plate-tectonic setting of Greece showing major faults. Seismically-active faults in Greece and western Turkey from Hatzfeld (1999). Instrumentally-determined motion vectors relative to Eurasia from Reilinger et al. (1997).

tion and collision. Basement rocks record an episode of Hercynian subduction and plutonism on the collisional/transtensional northern margin of Gondwana. The eastern Mediterranean Neotethys ocean opened as a result of rifting of this margin of Gondwana. Ophiolitic remnants of Neotethys are widespread. Early Tertiary collision produced a Hellenide mountain chain similar to the Alps and Himalayas. Rapid Neogene extension of the Hellenides behind the South Aegean arc has formed the Aegean Sea. In many ways, Greece is a microcosm of the tectonic processes and igneous products that occur in many orogenic belts, particularly those of the Tethyan region.

Young orogens, such as those in Greece, provide actualistic or near-actualistic analogues for interpretation of more ancient rocks. Many igneous rocks in Greece have suffered little alteration and are well constrained by biostratigraphy. The tectonic and geophysical environment of Neogene subduction-and extension-related rocks can be directly determined. This book, then, is more than an account of the rocks of one country: rather, it describes and interprets an important component of the anatomy of an orogen.

1.2 Organization of the book

1.2.1 Introduction

We have adopted a broadly chronologic approach to the organisation of this book. Individual chapters are organised around common geologic themes. Information on general geology and tectonics is presented in outline in this first chapter, and introduced in more detail at appropriate places in later chapters. This has, in places, required some repetition or cross-referencing. Some general thematic issues are briefly explored in the final chapter of the book.

For the igneous rocks of Greece, we have tried to provide a reasonably complete bibliography, even of older studies. However, in referencing work on regional geology, we have preferred to cite only synthesis papers rather than the regional studies on which they were based. The whole of Greece has been mapped by IGME[1] at a scale of 1:50 000 and we have extensively used the 1:500 000 map of the whole of Greece (Bornovas & Rondogianni-Tsiambaou 1983). However, we have cited these IGME maps only where they are the sole reference to a particular locality. Conventional topographic maps have been generally unavailable, particularly to foreign geologists, and place names commonly vary between maps. As a result, many localities are known under several different names, depending on the age of the map used[2].

In two appendices, we summarize the available radiometric geochronology of the igneous rocks of Greece (Appendix 1) and provide representative geochemical analyses of the principal igneous rocks described in this book (Appendix 2). The latter are selected principally from our personal data base of over 1300 rock geochemical analyses but include selected analyses from our data base of approximately 5700 analyses by others in the literature. In preparing plots to illustrate geochemical variability of the igneous rocks of Greece, we have used the entire data base of 7000 analyses. Many of these analyses, however, report only major elements and a few trace elements. Information on which analyses are illustrated in individual figures is given in Appendix 3.

1.2.2 Rock and mineral nomenclature

We use the IUGS classification based on the total alkali-silica diagram (Le Bas et al. 1986) for volcanic rocks that are not altered (Fig. 3). Because of the wide use of the term "rhyodacite" in previous work on the South Aegean arc, we have in places used this term

[1] Institute of Geology and Mineral Exploration, the Greek geological survey. See table of abbreviations and acronyms, p. XVI.

[2] Where a single locality is known by several names, we have adopted the name that seems to be most widely used. We have attempted to use place names in the readily available ELPA Tourist Maps (26^{th} Greek edition, 1999), which includes a gazetteer. We have generally adopted the spelling in this gazetteer transliterated to the latin alphabet. The principal exceptions are the use of standard English translations of Greek provinces (e.g., Thrace, Thessaly, Cyclades) and the city of Athens. Because of continuing political instability, we use the term "Former Yugoslavia" for the area north of Greece between Albania and Bulgaria.

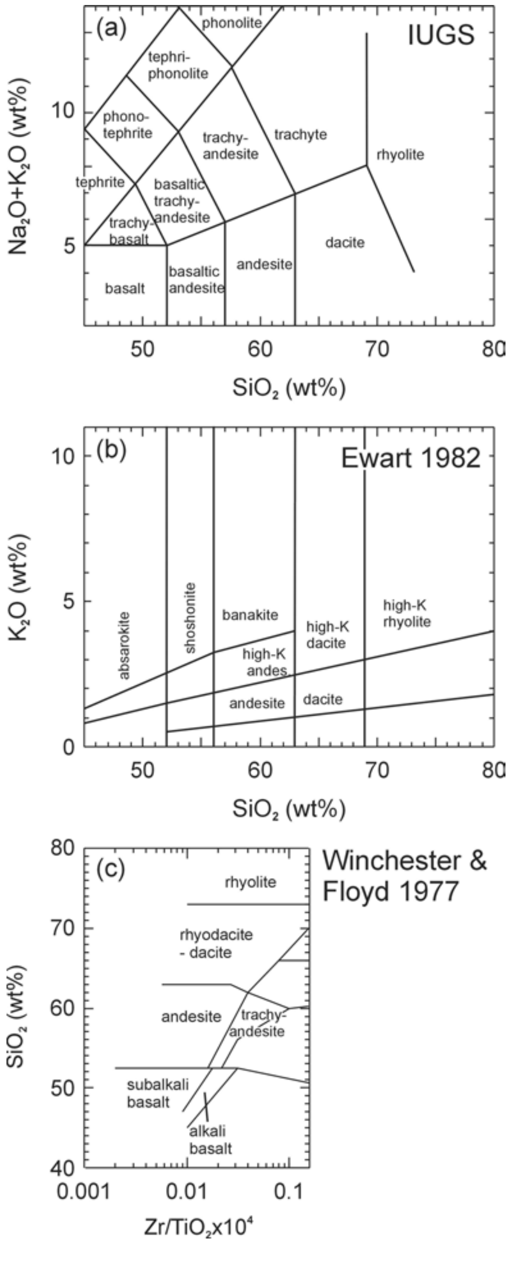

Fig. 3. Geochemical nomenclature of volcanic rocks used in this book. (a) The IUGS total alkalies vs. silica classification (Le Bas et al. 1986). (b) Classification of Ewart (1982) based on potash vs. silica. (c) Classification for older altered rocks based on immobile elements (Winchester and Floyd 1977).

for IUGS dacites with > 68% SiO_2. In general geological descriptions, the term andesite may include the IUGS basaltic andesite and andesite. In some cases, where the potassium content of rocks is important, we use the nomenclature of Ewart (1982) (Fig. 3). For plutonic rocks, we use the method of de la Roche et al. (1980), modified by Batchelor & Bowden (1985) (Fig. 4). Where modal data exist, we use the QAPF classification (Le Bas & Streckeisen 1989). The IUGS clast size classification of pyroclastic rocks has been used, with size boundaries at 64, 2, and 0.063 mm (Le Bas & Streckeisen 1989). For

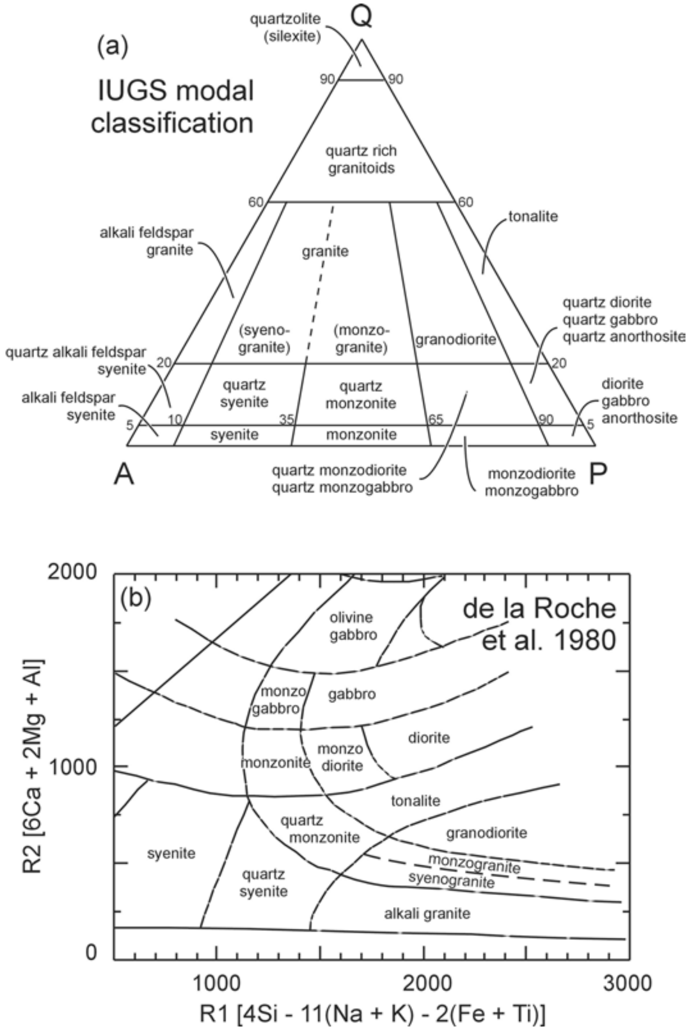

Fig. 4. Nomenclature of plutonic rocks used in this book. (a) The IUGS modal classification of plutonic rocks (Le Bas & Streckeisen 1989). Q = modal quartz, P = modal plagioclase, A = modal alkali feldspar. (b) Geochemical nomenclature of plutonic rocks, based on de la Roche et al. (1980) and Batchelor & Bowden (1985).

older altered rocks, where alkali mobility renders the IUGS scheme questionable, we use the SiO_2 vs. Zr/TiO_2 nomenclature of Winchester & Floyd (1977) (Fig. 3). In general, for mineral nomenclature, we have followed the usage of Deer, Howie & Zussman (1992). Amphibole nomenclature is after Leake (1978).

Various commonly used diagrams have been used to illustrate chemical variation in igneous rocks. Trace element variation, particularly in mafic rocks, is illustrated by element abundances normalised to primitive mantle and plotted as spidergrams (Sun & McDonough 1989). For rare-earth element patterns, the chondrite normalising factors of Sun & McDonough for C1 chondrite are used. In reporting radiometric dates, all old K-Ar dates have been recalculated to the decay constants of Steiger & Jäger (1977). Standard errors are generally not cited in the text, but are provided in tables and in some cases have been rounded up. We report pressures in kilobars (kbar), where 1 kbar = 0.1 GPa.

1.3 Outline of the geology of Greece

1.3.1 Introduction

The Hellenide orogen in Greece links the orogenic belt of the Balkan peninsula to Turkey (Fig. 1). It is bordered to the southwest by the Tethyan oceanic remnant of the Ionian Sea. What is now Greece underwent extension in the Permian and Triassic, following the Hercynian orogeny, resulting in the creation of the oceanic seaways of Neotethys. The main deformation of the Hellenide orogen was "alpine", with destruction of seaways of the Tethyan ocean in the Mesozoic and major collisional uplift in the early Tertiary. The later Tertiary was a time principally of extension to the north of the Hellenic subduction zone, where the African plate is subducted beneath Eurasia.

The land is mountainous, with the higher mountains above 2000 m, culminating in Mount Olympos at 2911 m. Outcrop is good in mountainous areas, although frequently rendered difficult of access by goat-resistant vegetation. Molasse-type sediments are accumulating in the extensional basins on land, obscuring older rocks. Only the highest land was glaciated in the late Pleistocene. Only small areas of the Aegean Sea, in the Cretan Basin and the North Aegean Trough, are deeper than 1000 m (Fig. 5). In the Ionian Sea, water depths exceeding 4000 m are found in the Hellenic Trench.

1.3.2 A brief history of geological studies in Greece

Although classical authors made important geological observations, systematic geological studies began following Greek independence in 1821 with the Expédition scientifique de Morée (Boblaye & Virlet 1833). The work of Alfred Phillipson (1892, 1896, 1901) laid the foundation for the stratigraphic and tectonic framework of Greece, much of which remains valid today.

In the first half of the twentieth century, Carl Renz (1940, 1955) developed the modern understanding of the biostratigraphy and tectonic zones of Greece. Since the found-

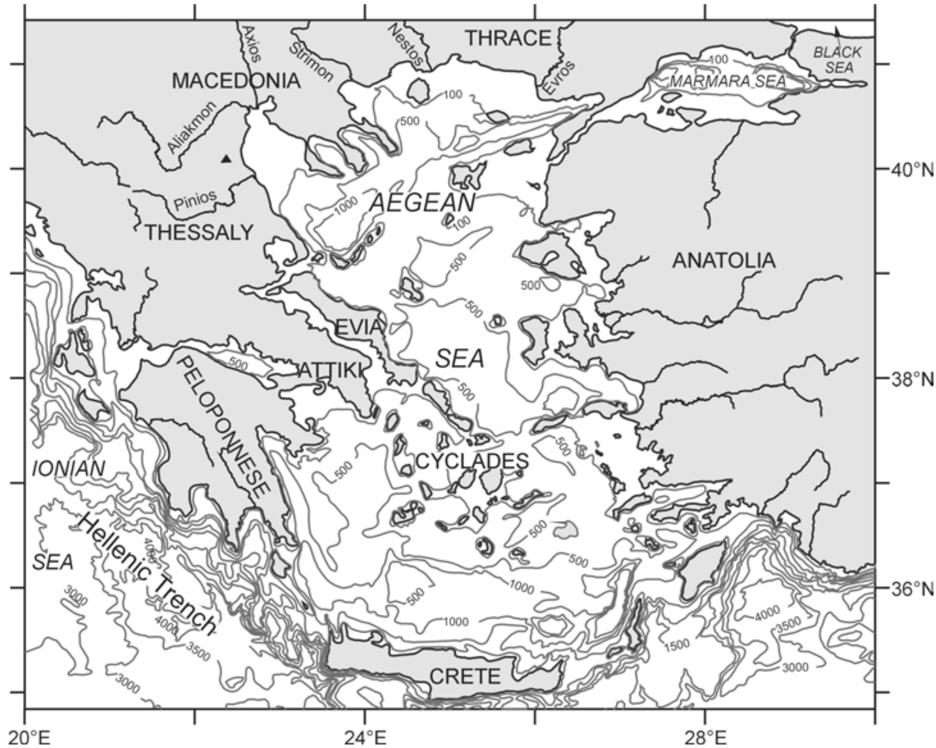

Fig. 5. Major rivers of Greece and bathymetry of the Greek seas. Depths in metres.

ing of the Greek geological survey (IGME) in 1952, the whole of Greece has been mapped at a scale of 1:50 000 and systematic regional studies have been carried out, principally by Greek, French, German, British and Dutch groups. Important syntheses have been prepared by Aubouin (1959), Jacobshagen et al. (1978) and Jacobshagen (1986).

Detailed study of the igneous rocks of Greece began with the monograph on Santorini by Fouqué (1879). In the first half of the twentieth century, K.A. Ktenas made a remarkably wide range of contributions concerning the igneous and tectonic history of Greece and the petrography of the igneous rocks. The island of Santorini has remained a centre of geological attention in the second half of the twentieth century, and has been related to the destruction of the Minoan civilisation, Plato's Atlantis and the biblical plagues of Egypt (e.g., Luce 1969). In the study of older rocks, the work of Moores (1969) on the Vourinos ophiolite ushered in the explosion of work on ancient oceanic crust in the last 25 years.

1.3.3 Terranes of the Hellenides

In Greece, the nappe pile of the Hellenide orogen has traditionally been divided into tectono-stratigraphic "isopic[1] zones" (Renz 1940), which are laterally persistent Mesozoic facies belts generally bounded by Cenozoic thrust faults (Fig. 6). Isopic zones correspond approximately to the modern concept of terranes, although in some cases several isopic zones make up a single terrane. From west to east, the **Pre-Apulian**, **Ionian**, and **Gavrovo-Tripolitsa** zones are remnants of the continental margin of Apulia: a promon-

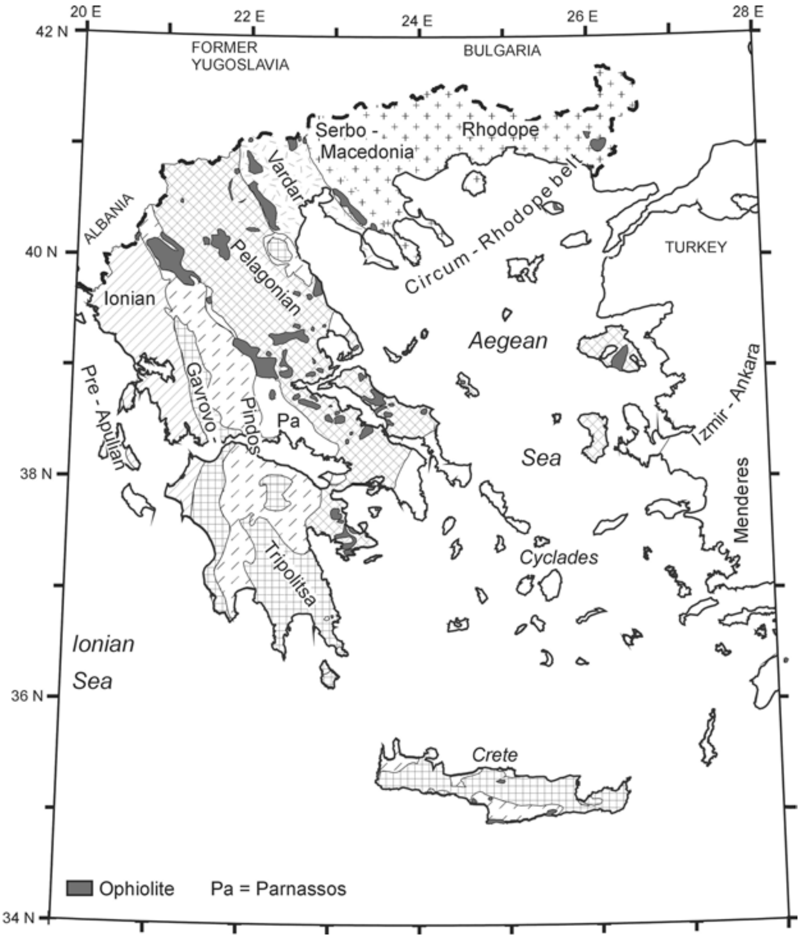

Fig. 6. Summary geological map of Greece showing the main isopic zones (modified from Jacobshagen 1986).

[1] "similar-looking", term first used by Mojsisovics (1879). Tripolitsa was originally spelled Tripolitza by Renz (1940).

tory of Gondwana (Laubscher & Bernouilli 1972, Ricou et al. 1986). They form the base of the nappe pile in western Greece.

East of the Apulian margin, the **Pindos** zone formed as an extensional ocean basin in the Early Mesozoic (Smith 1979, Pe-Piper & Piper 1992). The **sub-Pelagonian** zone is the complexly deformed eastern margin of the Pindos zone against the Pelagonian microcontinent. It is represented in the Ultrapindic (Boeotian of Jacobshagen 1986) and Maliac nappes and the overlying ophiolite sheet (Eohellenic nappe of Jacobshagen 1986). The **Parnassos** zone of central Greece is a large carbonate platform unit that lies between the Pindos and sub-Pelagonian zones.

The **Pelagonian** zone comprises a stack of nappes of Hercynian basement rocks overlain by Mesozoic cover. Late Jurassic – early Cretaceous deformation was associated with ophiolite emplacement. Late Cretaceous extension resulted in accumulation of mesoautochthonous deep-water sediments.

The **Vardar** zone represents the formerly oceanic suture zone between the Pelagonian microcontinent and the Rhodope massif. It has a complex assemblage of basement rock, deep-water sedimentary rock, volcanic rock, ophiolite and melange. It was divided into the Almopias, Païkon and Peonias subzones by Mercier (1966a). The Peonias subzone and its eastward continuation in Thrace are termed the **Circum-Rhodope belt**.

Northeast of the Vardar zone, the **Serbo-Macedonian** and **Rhodope** zones (or massifs of many authors) are nappe piles including pre-Alpine continental basement. They have a complex Mesozoic and early Tertiary deformational history, apparently resulting from collision of a Rhodope microcontinent with the Moesian platform of Eurasia (Fig. 1) and northward subduction of the intervening ocean. The Serbo-Macedonian zone of Greece has classically been correlated with rocks to the north in Serbia, but recent work suggests that the Serbo-Macedonian zone in Greece may be a continuation of the Rhodope nappe pile (Godfriaux et al. 1996).

The principal isopic zones can be traced from mainland Greece southeastward to the islands of the Aegean Sea. In the Cyclades, rocks equivalent to the Pindos and Pelagonian zones, together with basement rocks further outboard than the Pindos zone, experienced early Tertiary high-grade metamorphism. This area is referred to as the **Attico-Cycladic complex**.

The tectono-stratigraphic zonation of western Turkey differs in several respects from that of Greece, with important Triassic and Cretaceous oceanic sutures. Some components of the distinctive geological history of western Turkey are recognisable in the eastern Greek islands. The Vardar zone continues through Turkey in the İzmir-Ankara zone. North of this zone, the Sakarya microcontinental block formed by closure of a Karakaya ocean in the Triassic, and is represented in the northeast Aegean islands of Lesbos and Chios. South of the İzmir-Ankara zone, the Menderes massif is a core complex with similarities to the Pelagonian zone, although the lower nappes have African affinities. Farther south, the Lycian nappes and the Taurides (Fig. 1) show similarities to the Pindos and Ionian zones.

1.3.4 Pre-Mesozoic continental basement

Several pre-alpine (Paleozoic or older) continental basement blocks are recognised within the Hellenides (Fig. 7). Most of these are generally interpreted to be fragments of Gondwana that rifted apart in the Permian or Triassic to create the Neotethyan oceanic seaways. Their late Paleozoic rocks are discussed in detail in Chapter 2. From south to north, these blocks are:

Menderes lower nappes. These experienced the late Precambrian Pan-African orogeny and appear to be a fragment of Gondwana. Similar rocks may be present in Ios and perhaps other islands of the Cyclades.

Apulia. Slivers of amphibolite, micaschist and gneisses in the lowest nappe of Crete are interpreted as Hercynian basement (Seidel & Wachendorf 1986). Pre-Alpine gneissic basement occurs in the islands of Ios, Sikinos, Paros and Naxos in the Cyclades (Henjes-Kunst & Kreuzer 1982).

Pelagonia. In the Pelagonian zone of Thessaly and western Macedonia, a sequence of paragneiss and amphibolite is intruded by late Carboniferous granodioritic plutons and unconformably overlain by Permo-Triassic sedimentary rocks (Mountrakis et al. 1983, Kilias & Mountrakis 1987).

Sakarya. The Sakarya basement block, sutured during the Triassic by closure of the Karakaya ocean[1], is represented in the metamorphic basement rocks of Lesbos, but is much more clearly developed in adjacent areas of NW Anatolia.

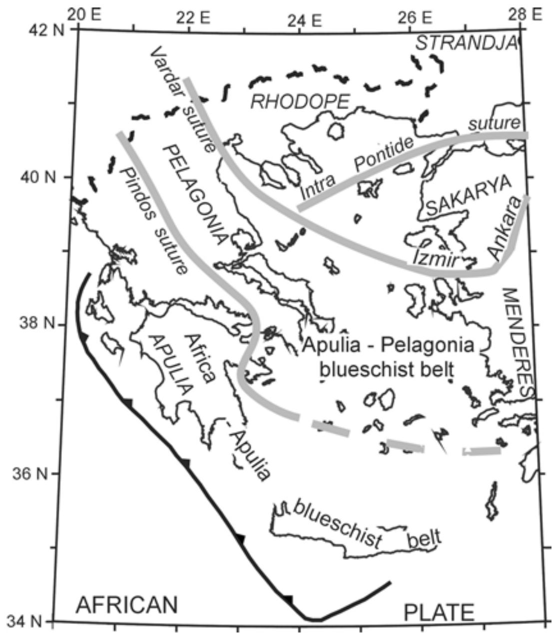

Fig. 7. Interpreted oceanic suture zones and microcontinental blocks of Greece.

[1] but for a contrary interpretation, see Okay & Tüysüz (1999).

Rhodope. Sparse fossil control and radiometric ages in northern Greece and Bulgaria suggest that the lowest tectonic unit of the Rhodope ("Lower Terrane" of Burg et al. 1996) includes pre-Mesozoic basement, notably in the Kerdilion sequence which is unconformably overlain by Permo-Triassic rocks, and in Thassos. Whether there was a discrete "Rhodope microcontinent" or whether the Rhodope nappe pile includes a number of continental blocks accreted during subduction (Barr et al. 1999) is unclear.

1.3.5 Tectono-stratigraphy of the Hellenides

The classic interpretation of the Hellenide orogen is of a complex nappe pile between an Apulian microcontinent foreland and a Rhodope hinterland (Aubouin et al. 1970, Jacobshagen 1986). In western mainland Greece, Jacobshagen (1986) distinguished the West Hellenic nappes (Ionian and Gavrovo-Tripolitsa isopic zones), the Central Hellenic nappes (Pindos and Pelagonian isopic zones) and inner Hellenic nappes including the Eohellenic ophiolite nappe and nappes of the Vardar zone (Table 1). Jacobshagen (1986) recognised several major phases of deformation, notably an Eohellenic (Late Jurassic – Early Cretaceous) and a Mesohellenic (Eocene) orogenic phase.

The sense of earliest nappe movements is one of the historical controversies of the Hellenides, in particular with respect to the source oceans for ophiolites. Aubouin (1959) and Jacobshagen (1986) interpreted the Hellenide orogen as showing progressively younger deformation westward, with the Eohellenic ophiolite nappe originating in the Vardar ocean. In recent years, there has been increasing realisation that the Hellenides formed by sea-floor spreading and subsequent complex closure of several Mesozoic oceanic seaways, with the classic Eohellenic nappe originally emplaced in a northeasterly direction onto the southwest margin of the Pelagonian microcontinent with a Pindos ocean. Only later were the ophiolites thrust westward with final closure of the Vardar and Pindos oceans in the earliest Tertiary.

Local nappe sequences are summarized in Table 1. Crustal stacking patterns have been complicated by probable mid-crustal delamination during Paleogene collision and by reactivation of older thrusts as major extensional detachment faults in the Neogene. The deeper parts of the nappe pile of the internal Hellenides experienced early Tertiary blueschist metamorphism followed by mid Tertiary greenschist metamorphism as a result of the Paleocene collision of Pelagonia and Apulia to create a crustal duplex following closure of the Pindos ocean. The history of closure of the Vardar ocean is less well understood, largely because so little of the Vardar zone is exposed. The Rhodope and Greek Serbo-Macedonian zones resulted from Mesozoic collision of Moesia and the Rhodope microcontinent.

1.3.6 Mesozoic rifting and sedimentation

Permian and Triassic rifting (Fig. 8) was widespread throughout Greece: related igneous activity is described in Chapter 3. This rifting continued with the creation of Late Triassic to Jurassic ocean basins, remnants of which are preserved as ophiolites. Early Permian

Table 1. Tectono-stratigraphic sequences in the Hellenides.

General tectono-stratigraphy of the Hellenides of western mainland Greece (after Jacobshagen 1986)
Vardar and circum-Rhodope nappes
Eohellenic ophiolite nappe
Maliac nappes
Pelagonian nappes
Parnassos nappe
Pindos nappe
Upper West Hellenic nappes (= Ionian and Gavrovo Tripolitsa zones, Olympos window)
Lower West Hellenic nappes (= Phyllite-Quartzite series)
Apulian foreland (= Pre-Apulian zone)

Tectono-stratigraphy of Crete (after Jacobshagen 1986)
4th stockwerk: composite nappe or accretionary melange, including ophiolites
3rd stockwerk: deep-water Mesozoic sediments = Pindos zone
2nd stockwerk: upper Triassic-mid Eocene platform limestone, then flysch = Tripolitsa zone (Apulia)
1st stockwerk: slices of Hercynian basement, Phyllite-Quartzite series
Autochthon (Permian – Oligocene) = Ionian zone (Apulia)

Tectono-stratigraphy of the Cyclades (after Dürr 1986)
Uppermost unit: non-metamorphic upper Paleozoic – Jurassic limestones, ophiolites, Lower Cretaceous neritic limestones = Pelagonian zone
Upper unit: low grade metamorphic rocks of the Athens area
Intermediate unit: deep-water Mesozoic metasedimentary and metaigneous rocks = Pindos zone
Basal unit: Hercynian basement, Mesozoic carbonate platform, Paleocene flysch = Apulia

Tectono-stratigraphy of the nappe pile of Thessaly (after Schermer 1993, and Jacobshagen 1986)
(see also Table 2)
Lechonia and Skyros klippe
Eohellenic ophiolitic nappe: probably derived from Vardar ocean
Flambouron unit: Pelagonian Permian-Mesozoic sediments overlying Hercynian basement, divided into
Infrapierien subunit
Pierien subunit
Ambelakia unit: metasedimentary and metavolcanic rocks (?Pindos zone)
(divided by Doutsos et al. 1993 into Ambelakia and Kokkinopilos units)
Olympos unit: Mesozoic neritic carbonate, Eocene flysch (Apulian continental fragment)

Tectono-stratigraphy of the Rhodope and Greek Serbo-Macedonian zones (after Burg et al. 1996)
Upper Terrane = Vertiskos unit in part
Intermediate nappes:
a) Gneiss-marble sequence
b) Gneiss sequence
c) Eclogite-metabasite-gneiss sequence
d) Migmatite-orthogneiss sequence = Vertiskos unit in part
Lower Terrane = Pangeon unit, Kerdilion unit

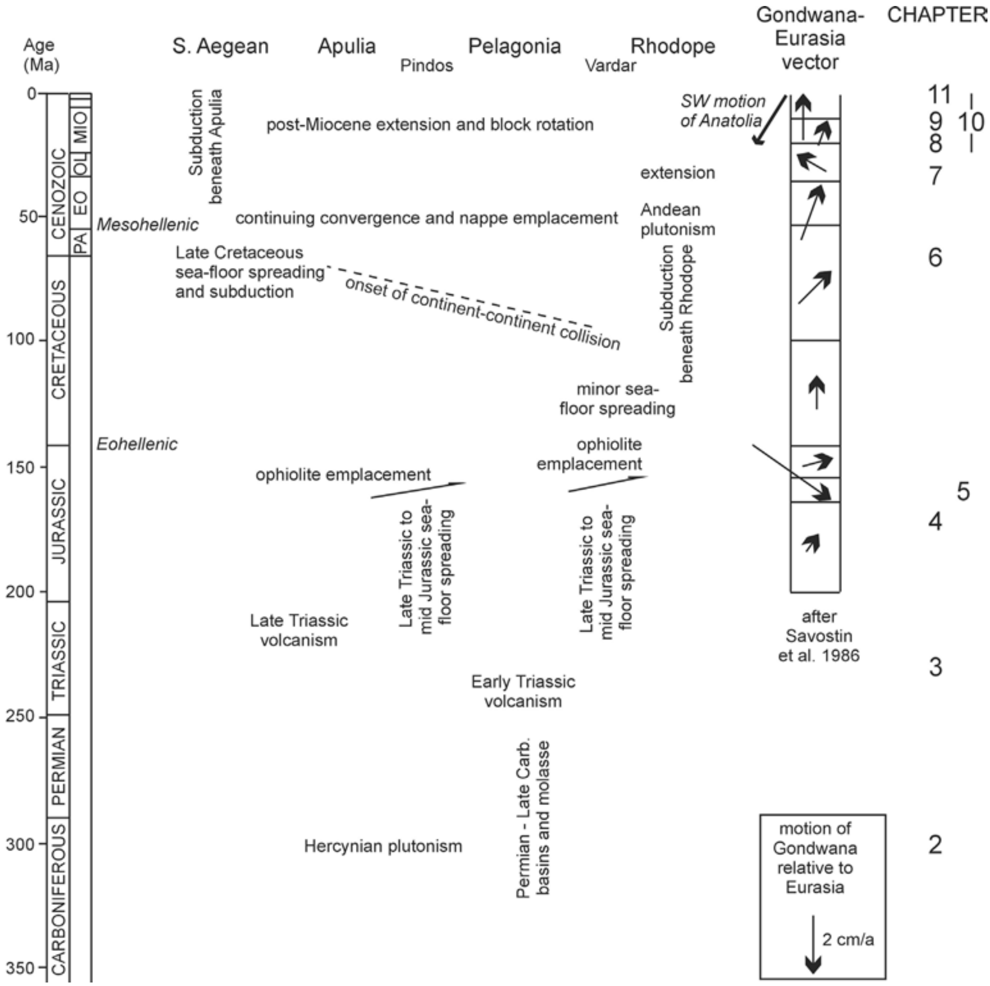

Fig. 8. Summary chronology of the geological history of Greece.

rifting is represented by deep water marine sediment facies in Sicily and Crete (Robertson et al. 1991). In the Pelagonian and Tripolitsa zones, the Permian consists of molasse-type terrestrial clastic sedimentary rocks with shallow marine transgressive limestones and clastic sedimentary rocks. This Permian rift is interpreted by Jacobshagen (1994) as a precursor of Jurassic and Cretaceous subsidence in the Ionian zone.

The main rift event took place in the early to mid Triassic (Robertson et al. 1991). Deep water sediment facies developed in the Ionian and Pindos zones and locally within the Pelagonian zone of Argolis. Early Triassic subaerial and submarine lavas are overlain by transgressive mid-Triassic limestones in the Tripolitsa, sub-Pelagonian and Pelagonian isopic zones. Volcanic activity in places continued to the late Triassic. Final continental break-up and initial ocean spreading in the Pindos ocean took place in the Carnian – Norian (late Triassic). Deep water sediments continued to accumulate in the Ionian and Pindos zones, while thick platform carbonates accumulated in the Gavrovo-Tripolitsa and Pelagonian zones.

In the Vardar zone, the Almopias subzone includes Triassic and Jurassic deep-water radiolarites tectonically intercalated with mafic lavas and ophiolites, indicating that it was a rifted oceanic seaway. The Peonias subzone also records Permo-Triassic rifting and the formation of late Triassic oceanic crust with overlying radiolarites.

1.3.7 The ophiolites

The ophiolites of Greece (and adjacent areas to the north in the Balkan peninsula) outcrop as two N-S trending belts. The western belt outcrops in the sub-Pelagonian zone of Greece and in the central Dinarides, and is dominated by lherzolites, ferrogabbros and olivine tholeiites. Mid-Jurassic radiometric dates (Spray et al. 1984) from metamorphic soles to ophiolites record early high-temperature displacement of the ophiolites. Final emplacement onto the eastern edge of the Pelagonian microcontinent is stratigraphically constrained to the Kimmeridgian (Robertson et al. 1991). Although many authors have interpreted these ophiolites as originating from the Vardar zone (Aubouin 1959, Jacobshagen 1986), fabrics within some of the ophiolite complexes (Rassios 1991) and a variety of structural evidence (Robertson et al. 1991) indicate eastward emplacement from an ocean in the Pindos zone. Later nappe movement transported some ophiolites westwards above Lower Eocene flysch of the Pindos zone (Jacobshagen 1986). These ophiolites are considered in Chapter 4.

The eastern belt, bordering on the Vardar zone and continuing northward to the inner Dinarides, comprises harzburgites and dunites, olivine gabbros, tonalites and quartz tholeiites. These ophiolites appear to be of various types and are derived from ocean crust and upper mantle in the Vardar zone. Ophiolites of the Almopias subzone were emplaced on the eastern margin of the Pelagonian zone in the early Cretaceous (Berriasian) and are unconformably overlain by a deepening succession of upper Cretaceous limestones and flysch. At least some ophiolites in the Peonias subzone appear to be autochthonous and are of middle Jurassic age. These ophiolites are considered in Chapter 5. A late Cretaceous extensional event recognised in the Almopias subzone and part of the eastern Pelagonian zone is represented further south by ophiolites in Evia and Crete: these ophiolites are described in Chapter 6.

1.3.8 Alpine orogenesis

Several major orogenic phases can be recognised in the geological evolution of Greece. The term "Alpine" is generally applied to deformation of Mesozoic and Tertiary age. Of the pre-Alpine evolution, only a Hercynian (mid-Carboniferous) orogenic phase is well documented. The Permian to Jurassic was largely characterised by extension, although the history of the Rhodope massif at this time is not well constrained.

The Eohellenic (late Jurassic to early Cretaceous) orogenic phase of Jacobshagen (1986) is represented in the Vardar zone by the closing through eastward subduction of the Almopias ocean beneath the Païkon island arc in the upper Jurassic and the deformation of the Circum-Rhodope belt. At the same time, the eastern part of the Pindos ocean was consumed by westward subduction beneath a spreading ridge, which was then obducted onto the Pelagonian continental margin as the ophiolitic Eohellenic nappe.

Not all the Pindos ocean was obducted in the Eohellenic event, and in southern Greece and Turkey, Cretaceous sea floor spreading widened the Pindos ocean. Cretaceous oceanic crust was also created by pull-apart oceanic basins in the Vardar zone and more widespread extension in the İzmir-Ankara zone. During the Cretaceous, however, both the Pindos and Vardar oceanic zones were consumed by subduction which culminated in the late Cretaceous – early Tertiary collision of the continental fragments of Apulia, Pelagonia and the Rhodope massif (the Mesohellenic orogenic phase of Jacobshagen 1986).

The blueschist metamorphic complexes from Thessaly to the Cyclades formed in this collision, probably diachronously from mid Cretaceous in Thessaly to Paleogene in the Cyclades. During subsequent early Miocene extension, these rocks in the Cyclades were affected by a Barrovian high-temperature metamorphism. Compression continued through the late Oligocene to early Miocene, with major shortening in the external Hellenides. Continuing continental subduction and subsequent delamination resulted in early Miocene blueschist metamorphism in Crete and the Peloponnese and subsequent uplift. Other nappes of the external Hellenides are essentially unmetamorphosed. Deformation has gradually progressed westward, reaching the Ionian zone in the Pliocene (Underhill 1989). Crustal thickening in the orogen was probably geographically limited, since large volumes of erosional detritus are lacking. Although Avigad & Garfunkel (1991) argue that crust was thickened to 80 km, P-T paths of metamorphic rocks in Thessaly, the Peloponnese and the Cyclades generally require crust only 45–50 km thick (Schermer 1990, Bassias & Triboulet 1994).

1.3.9 Neogene extension and block rotation

Neogene crustal extension in the Aegean area has been generally interpreted as a result of escape tectonics, with the Anatolia – Aegean microplate "squeezed" westwards between the Arabian indentor and Eurasia (Jackson 1994) (Fig. 2). Collision of Anatolia with Eurasia in the Miocene resulted in crustal thickening of eastern Anatolia and subsequent westward gravitational collapse leading to the escape of Anatolia and the development of the North Anatolian Fault, which has progressively propagated eastward in the past 10 Myr, reaching the north Aegean at about 5 Ma (Armijo et al. 1999). Decreased rates of convergence between Africa and Eurasia in the Oligocene (Savostin et al. 1986) resulted in roll-back of the subducting plate boundary (Royden 1993) and extension of the Aegean. Extensional collapse probably began in the Rhodope orogen (Burg et al. 1990) and continued later as a result of the elevation of the Hellenide orogen. The P-T paths of metamorphic rocks in the Hellenide orogen are characterised by near isothermal decompression (Avigad 1993). Sonder & England (1989) and Taymaz et al. (1991) have found good agreement between earthquake motions and models involving simple extension of the lithosphere in the Aegean area, although the details remain unclear.

Today, the Aegean Sea and the surrounding land areas are one of the most rapidly extending areas of continental crust in the world, with the southern Aegean moving at about 35 mm/yr relative to Eurasia (Jackson 1994). The stretching parameter, β, for the post-mid Miocene Aegean is about 1.7; lesser values are estimated for the Oligocene of northern Greece (e.g., Sonder & England 1989). Structural evidence shows that plutons of the Cyclades and North Greece were emplaced during active extension and the formation of core complexes (Boronkay & Doutsos 1994, Dinter et al. 1995).

Subduction of Mediterranean ocean crust has been synchronous with extension since at least the early Miocene, based on the length of the tomographically imaged subducted slab (Meulenkamp et al. 1988, Bijwaard et al. 1998) (Fig. 9) and the deformation of the Mediterranean ridge (Kastens 1991).

Block rotations in the Cenozoic inferred from paleomagnetic data (data reviewed by Walcott & White 1998) range from 25° to 50° in much of western Greece, to near zero in Crete, Rhodes and Thrace, to 15° to 30° counterclockwise in the eastern Aegean and western Anatolia (Fig. 10). This rotation is inferred to be related to subduction roll-back at the Hellenic subduction zone and large scale extension in the Aegean. Throughout the Miocene, a discrete West Aegean crustal block rotated 30° clockwise, with the Mid-Cycladic Lineament marking the junction with the eastern Aegean-Anatolian block (Walcott & White 1998). Relative motion between these two blocks ended in the Early Pliocene, when the North Anatolian Fault propagated into the Aegean Sea.

1.3.10 Cenozoic igneous activity

Volcanism and plutonism have been widespread in both time and space throughout the extensional area of the Aegean in the Cenozoic. Although formerly ascribed to southward migrating subduction (e.g., Fytikas et al. 1984), most of these igneous rocks are now recognised as resulting from extension (Pe-Piper et al. 1994).

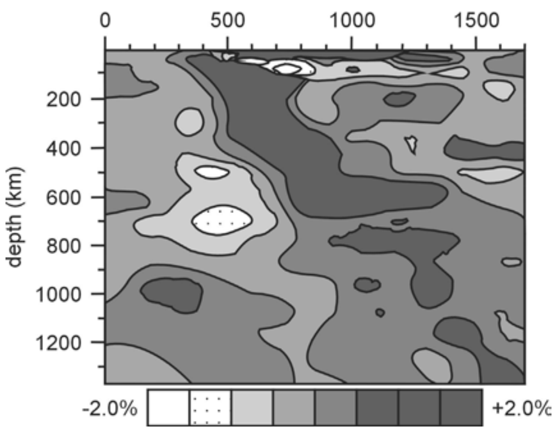

Fig. 9. Cross-section of the Aegean region showing seismic travel-time tomography of the mantle beneath Greece (modified from Figure B7 of Spakman et al. 1993). Based on EUR89B model, profile located on Fig. 13.

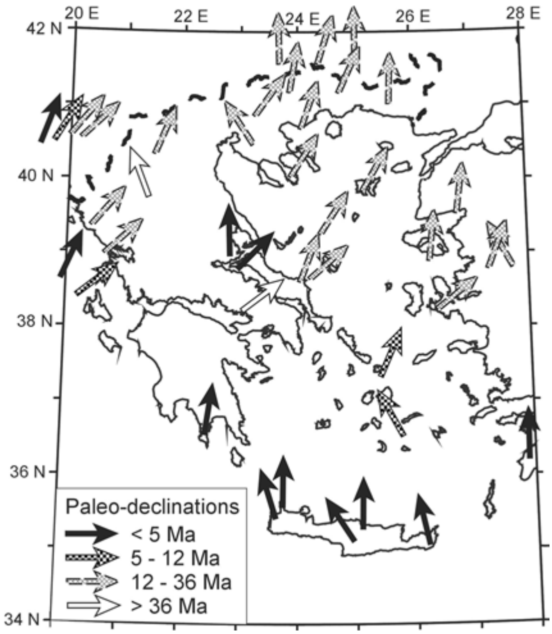

Fig. 10. Summary of Cenozoic paleomagnetic declination data (modified from Walcott & White 1998) for the Cenozoic of Greece.

Several types of igneous rock can be distinguished in the Cenozoic of the Aegean area (Fytikas et al. 1984, Pe-Piper et al. 1994), forming belts that change from north to south (Fig. 11) in age and rock type.

1. In North Greece, Oligocene plutonic rocks range from granite to monzonite with subordinate gabbro. They are synchronous with extensional basins containing terrigenous and volcanoclastic sedimentary rocks and calc-alkaline to shoshonitic lavas. The oldest lavas are middle Eocene and volcanism culminated in the late Oligocene, with minor early Miocene plutonism.
2. Early Miocene calc-alkaline to principally shoshonitic volcanism is voluminous and widespread in the islands of the northeast Aegean and adjacent areas of Anatolia.
3. Minor mid Miocene high-Mg andesite lavas (with chemical composition of adakite) are found in islands of the central Aegean Sea.
4. Minor "back-arc" volcanic centres of late Miocene to Quaternary age are found throughout the western and southern Aegean. They include a range of rock types, generally with older rocks being shoshonites, followed by potassic trachyte and then sodic nepheline-normative basalt.
5. Mid- to late Miocene plutons of the Cyclades are mainly granodiorite, with lesser gabbro and granite. Mafic rocks are K-rich and in the eastern Cyclades appear cogenetic with "back-arc" volcanic rocks.
6. Calc-alkaline volcanism in the South Aegean arc began in the early Pliocene and continues today, with several volcanoes active in historic times (Fig. 12).

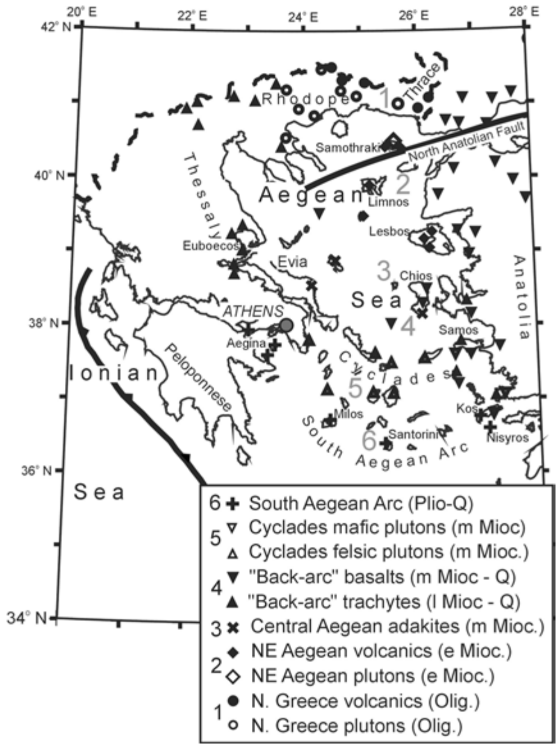

Fig. 11. Cenozoic belts of igneous rocks in Greece.

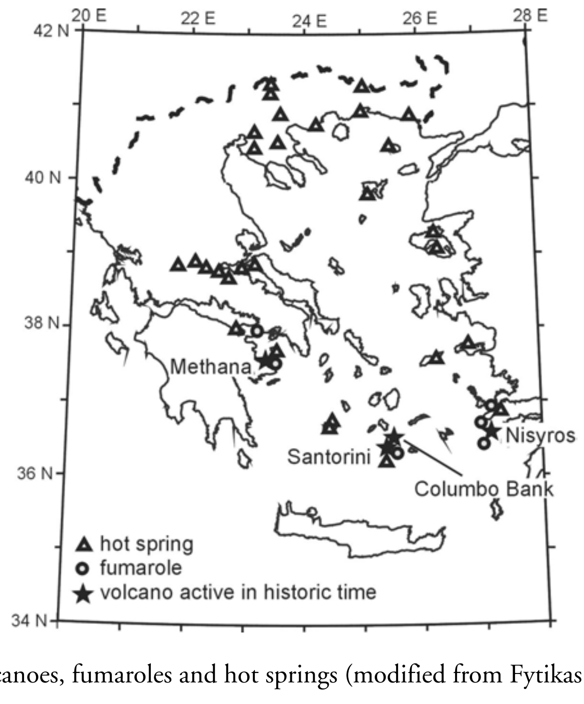

Fig. 12. Active volcanoes, fumaroles and hot springs (modified from Fytikas & Kolios 1979).

1.4 The present geophysical fields

Bouguer gravity data has been summarized by Makris & Röwer (1986) (Fig. 13). Regional magnetic data are presented by Dercourt (1980, p. 353). Depth to the Moho (Fig. 14) has been estimated from seismic refraction and gravity data by Makris (1977) and Makris & Röwer (1986). Crust is thin beneath the Aegean Sea (ca. 30 km, reaching less than 20 km in the Cretan Sea) and thick beneath the mountains of northern Greece (45–50 km) and the Peloponnese (30–45 km). Seismic tomography of the mantle (Spakman et al. 1988, Ligdas et al. 1992, Bijwaard et al. 1998) (Fig. 9) shows a high-velocity anomaly dipping northwards from the Hellenic trench, with a marked steepening below 150 km. The anomaly can be detected to a depth of 1200 km. Regional heat flow data have been compiled by Fytikas & Kolios (1979) (Fig. 15).

The distribution of earthquake epicentres (Fig. 16) shows that shallow seismicity is complex and widespread, but concentrated along the active plate boundary in the Hellenic Trench, the area between Kefallinia and Pelion, and the Corinth graben (Hatzfeld 1999). First motion studies show strike-slip motion on the North Anatolian and Kefallinia faults, reverse faulting along the Hellenic Trench and normal faulting elsewhere. The western termination of the North Anatolian Fault is connected to the Kefallinia fault by a diffuse system of grabens with normal faulting. Deep earthquakes define a Benioff zone dipping about 30° (Comninakis & Papazachos 1980). Motion on the Anatolia-Aegean microplate has been demonstrated by satellite laser ranging and Global Positioning System measure-

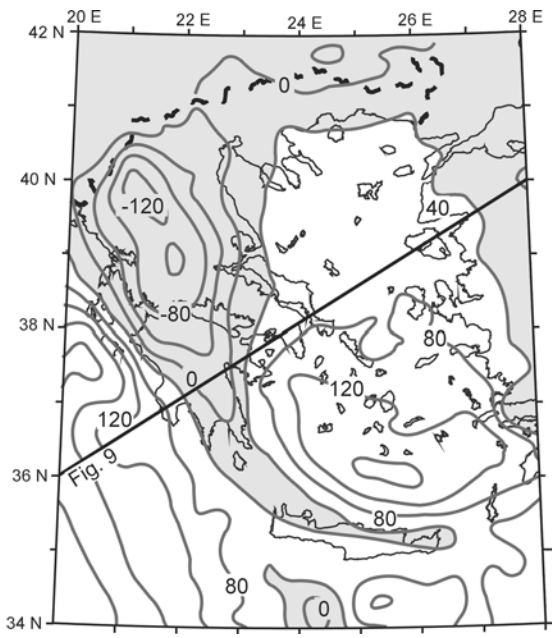

Fig. 13. Bouguer gravity anomaly map of Greece (in mgal) (modified from Makris & Röwer 1986).

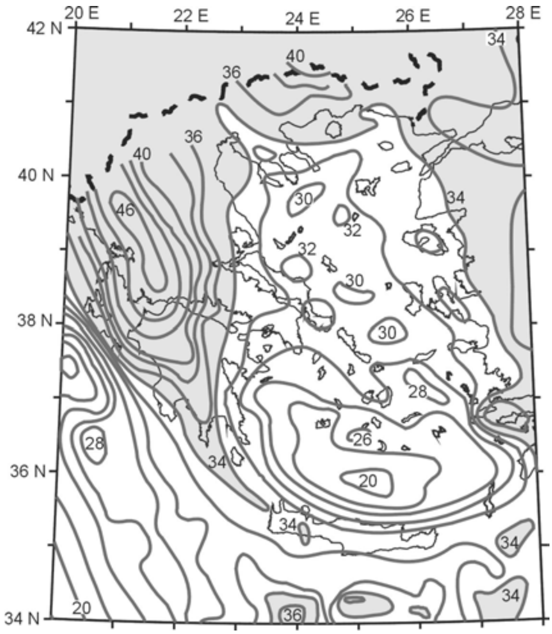

Fig. 14. Map showing depth to the Moho (in km) (modified from Makris & Röwer 1986).

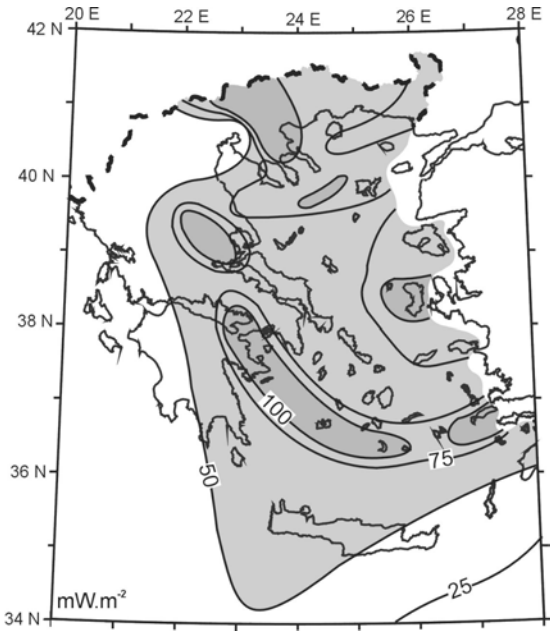

Fig. 15. Preliminary heat flow map of Greece (modified from Fytikas & Kolios 1979).

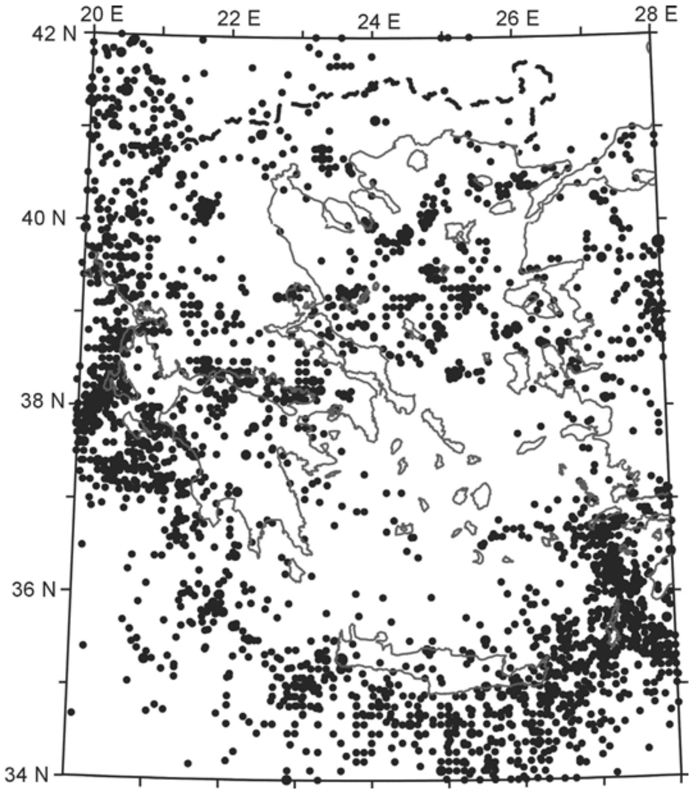

Fig. 16. Map showing modern seismicity (cf. Makropoulos & Burton 1984). All earthquakes ≥ M = 4 are shown.

ments (Reilinger et al. 1997) (Fig. 2). Such data suggest that the entire Anatolian – Aegean block is currently rotating as a coherent unit and that there is not a separate Aegean plate.

1.5 Plate tectonic evolution of Greece

The history of the Greek region since the late Paleozoic has been dominated by the relative motion of the large Gondwana (African) and Eurasian plates, ultimately resulting in the destruction of the intervening ocean, comprising the Paleozoic Paleotethys and the Mesozoic Neotethys. Numerous small continental fragments were rifted off the northern margin of Gondwana, but paleomagnetic data suggest that they remained within 1000 km of the Gondwana margin (Turnell 1988). The relative motions of Gondwana and Eurasia since the Triassic (Fig. 8) are independently known from the magnetic anomaly record in the Atlantic and Arctic oceans (Savostin et al. 1986) and have had a major influence on the regional geology and igneous activity of Greece.

This following brief synthesis of the plate-tectonic evolution of Greece is based on large-scale reconstructions of the Tethyan belt by Dercourt et al. (1986), Şengör (1989), Scotese (1991) and Stampfli (2000) (Fig. 17) and detailed syntheses of the Aegean area by Robertson & Dixon (1984), Şengör et al. (1984) and Robertson et al. (1991, 1996) (Fig. 18). Many uncertainties of detail remain in understanding the plate tectonic evolution (see, for example, the four different Permian reconstructions shown in Fig. 37). More paleomagnetic information is required from the small continental fragments of the southern Balkans and Anatolia. A definitive synthesis is not possible at the time of writing.

In the late Paleozoic, the supercontinents of Laurasia and Gondwana converged and collided along the Hercynian or Variscide orogeny of Europe (Rey et al. 1997). Greece lay at the western end of the Paleotethyan ocean between the two supercontinents and experienced Hercynian deformation. This Paleotethyan ocean was destroyed by Mesozoic subduction along the southern margin of Eurasia, although the detailed history is not clear (see discussion in Ch 2.7). The Pontides of northern Turkey (Fig. 1) are fragments of Laurasia, whereas the microcontinental blocks of Greece (with the possible exception of Rhodope) appear to be fragments of Gondwana. Permian to Triassic rifting of the northern margin of Gondwana resulted in the opening of the eastern Mediterranean

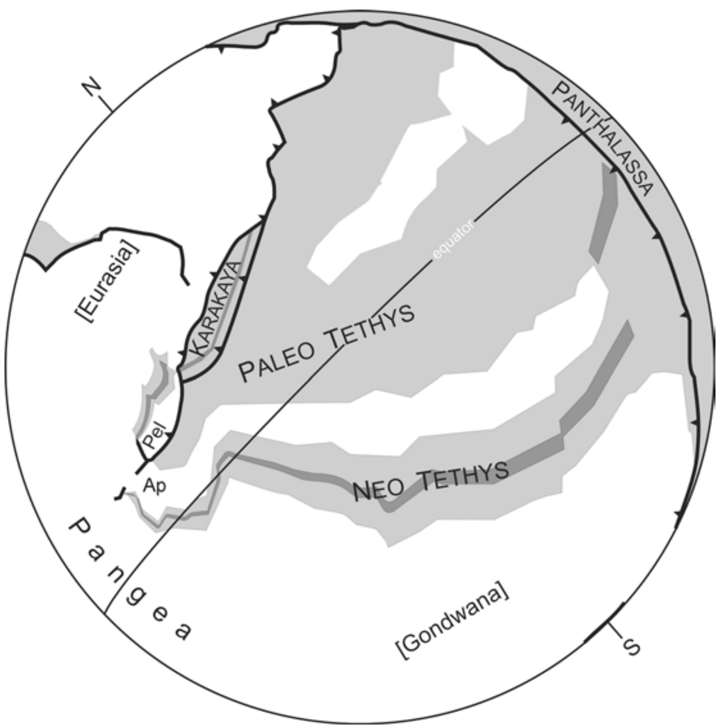

Fig. 17. Cartoon maps modified from Stampfli (2000) showing one interpretation of the Tethyan region between Gondwana and Eurasia at the Permian-Triassic boundary. Ap = Apulia, Pel = Pelagonia.

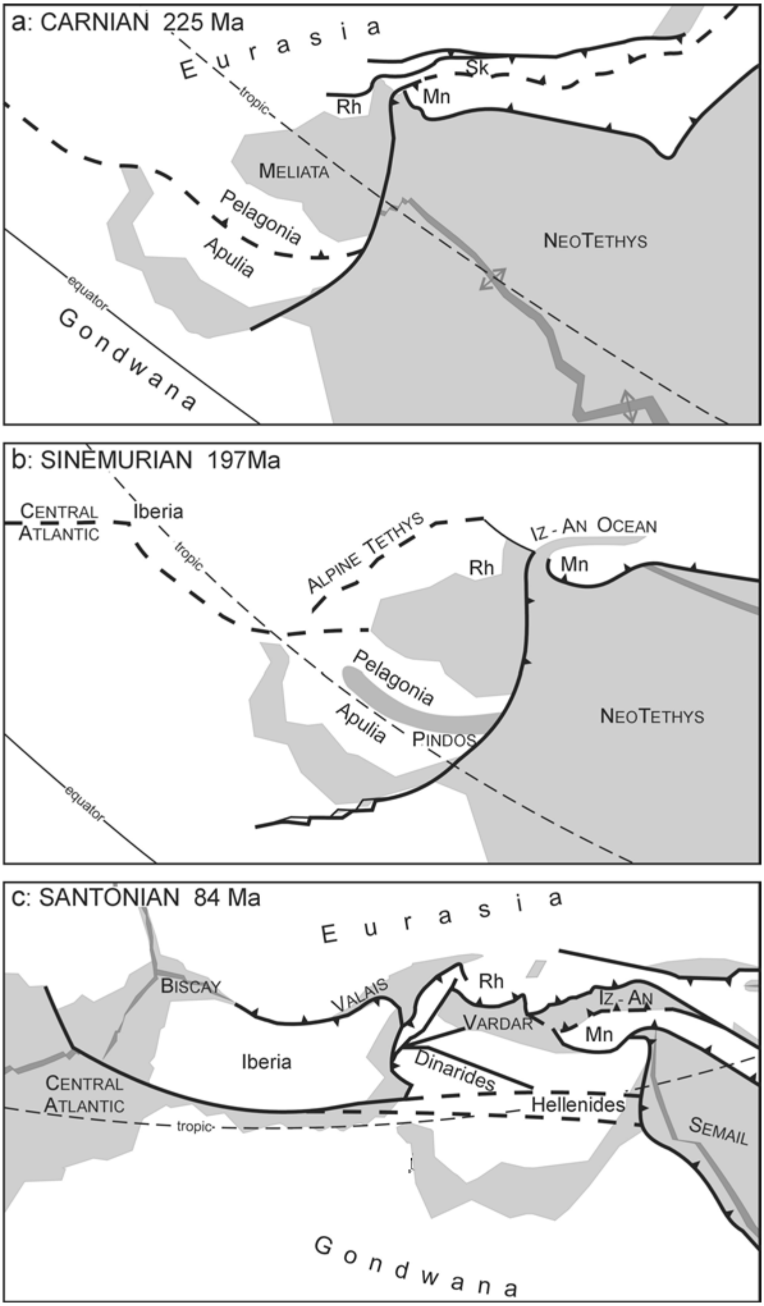

Fig. 18. Plate tectonic reconstruction of the Greek region proposed by Stampfli (2000) in mid Triassic, early Jurassic and late Cretaceous. Iz-An = İzmir-Ankara, Mn = Menderes, Rh = Rhodope, Sk = Sakarya.

Neotethys ocean. At least three oceanic basins are represented in Greece: what is now the oceanic crust of the eastern Mediterranean north of the African passive margin, the Pindos ocean between Apulia and Pelagonia, and the Vardar ocean, between Pelagonia and Rhodope (Fig. 18). Partial closure of the Pindos and Vardar oceans took place in late Jurassic to early Cretaceous time, with the emplacement of ophiolites. Late Cretaceous extension resulted in formation of oceanic crust in central and southeastern Greece, Cyprus and Turkey, but most of these ocean basins closed in the late Cretaceous.

Collisional orogenic uplift began in the early Cretaceous, but much of this deformation is obscured by the main late Eocene collisional event that produced a major mountain chain (Fig. 19). Extensional collapse of this orogen began in the Oligocene of northern Greece and the early Miocene in the south, while subduction of the eastern Mediterranean remnant of Neotethys continued from the south.

The kinematics of modern deformation in Greece is controlled by the westward tectonic escape of Turkey from the Arabian-Eurasian collision zone in eastern Turkey (McKenzie 1970) (Fig. 2). This results in some convergence between NW Greece and Apulia (Taymaz et al. 1991) and rollback of the South Aegean subduction zone (Royden 1993). The change in deformation style from strike-slip in the Aegean Sea to mainly normal faulting in mainland Greece is related to the collision of northern Greece and Albania with Apulia, so that western Greece cannot rotate rapidly enough to take up the western motion of Turkey. This results in E-W shortening, which is compensated by N-S extension as the southern Aegean margin can move easily over the South Aegean subduction zone.

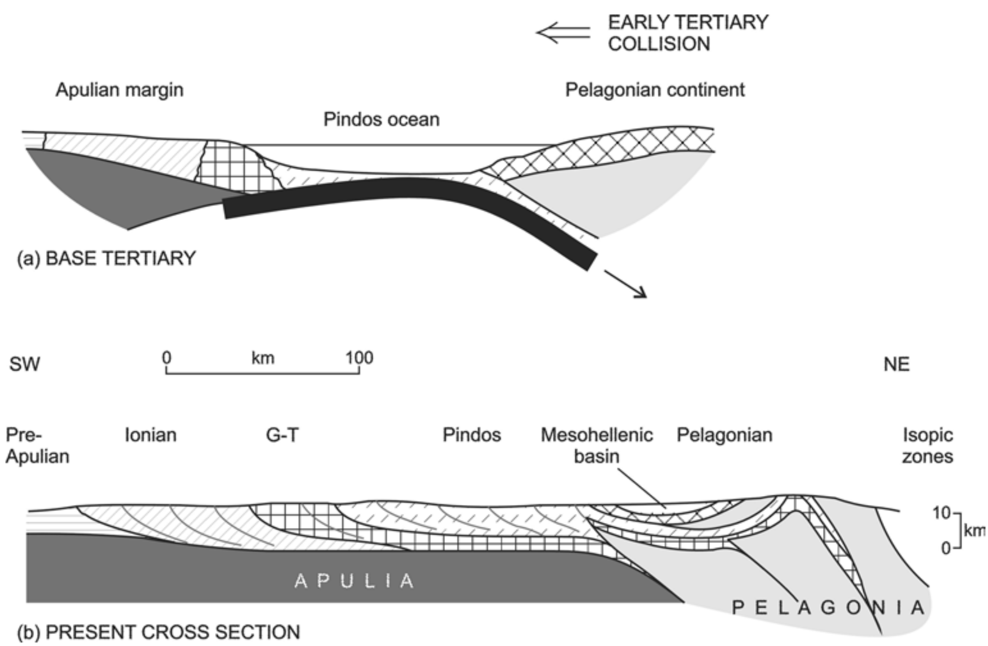

Fig. 19. Cartoon cross-sections illustrating probable plate-tectonic evolution of western Greece (modified from Jacobshagen 1986). G-T = Gavrovo-Tripolitsa zone.

2 Late Paleozoic plutonism and volcanism

2.1 Introduction: the Hercynian basement of Greece

Pre-alpine basement rocks of Greece occur sparsely within the continental fragments whose Mesozoic dispersal led to the creation of Neotethyan oceans. In many areas, these basement rocks comprise paragneisses metamorphosed in the late Paleozoic Hercynian or Variscide orogen and intruded by syn- or post-orogenic plutons. The paragneisses include metavolcanic rocks of both subduction and ophiolitic origin, reflecting the closure of Paleozoic oceans. In places both in northern Greece (Rhodope) and the southern Aegean (Ios) there is evidence of older basement rocks mobilised in the late Precambrian or early Paleozoic. Unmetamorphosed Silurian and Devonian sedimentary rocks outcrop in the accretionary melange of western Chios. The pre-Hercynian history of Greece cannot be unravelled without much more U-Pb zircon geochronology than is available at the time of writing (2000). On the other hand, the Hercynian rocks of Greece are relatively well known. They provide important constraints on the paleogeographic setting of the various continental blocks of Greece in the late Paleozoic. We describe first the well-known Hercynian plutons of the Pelagonian microcontinent and then compare them with Hercynian rocks of Rhodope, fragments in the Vardar zone, and basement in various islands of southern Greece.

The Hercynian or Variscide orogeny of Europe resulted, in a general manner, from the early Paleozoic convergence and Devonian to early Carboniferous collision of Gondwana and Laurasia. The orogen trends east-west through central Europe, extending north of the Alpine front from Bohemia to the Iberian peninsula. In many areas, Hercynian metamorphic and plutonic rocks are incorporated into the Tertiary Alpine orogen (Franke 1989, Frisch & Neubauer 1989, Finger & Steyrer 1990). Most proposed models for the tectonic evolution of the Hercynian fold belt favour subduction-collision processes (e.g., Behr et al. 1984, Liew & Hofmann 1988, Rey et al. 1997). In the eastern part of the orogen, Finger & Steyrer (1990) proposed that convergence and collision of Gondwana and Laurasia in the early Carboniferous resulted in an Alpine type collision, replaced by a Cordilleran-type margin in the late Carboniferous as a result of opening of the Carboniferous Paleotethyan gulf and its subduction beneath the Hercynian fold belt. This produced an outboard belt of I-type granites and an inboard belt of K-rich S-type granites. In the western part of the orogen, authors such as Ziegler (1986) and Matte (1991) considered the Hercynian fold belt to have formed by collision in the Devonian of several microcontinental terranes between Laurasia and Gondwana, followed by a prolonged late Carboniferous to Permian period of post-collisional extension, at which time there was voluminous granitic magmatism.

2.2 Plutons in the Pelagonian zone of Thessaly and western Macedonia: type example

2.2.1 Regional structural setting

The "Pelagonian" nappe pile of northern mainland Greece (sensu Jacobshagen 1986) structurally overlies the Olympos window and underlies the Eohellenic ophiolites (Table 2). The Pelagonian nappes consist principally of slices of continental basement of the Pelagonian zone overlain by late Paleozoic-Triassic terrigenous metasedimentary rocks and Mesozoic neritic limestones. These rocks were variably metamorphosed, in places to blueschist facies, during mid-Cretaceous to early Tertiary continental collision (Schermer 1993). The Olympos window consists of Mesozoic platform carbonates, interpreted as a fragment of Apulia, overlain by Eocene flysch. It is structurally overlain by metasedimentary and metavolcanic rocks of the Ambelakia unit, which may be derived from subduction of the Pindos ocean.

Table 2. Nomenclature of the nappe pile of Thessaly; based on Schermer (1993) and Doutsos et al. (1993), showing correspondence to nomenclature of various other authors.

Nappe	brief description	usage of other authors
1. Ophiolitic nappe	Ophiolites derived from Vardar ocean	Eohellenic nappe of Jacobshagen (1986)
2. Flambouron unit, where possible divided into:		
2a. Infrapierien unit	Hercynian gneiss and schist, meta-sedimentary and metavolcanic rocks, Mesozoic neritic carbonate rocks.	Pelagonian basement nappes and parts of Blueschist nappes of Doutsos et al. (1993)
2b. Pierien unit	Hercynian granitic gneiss, thin meta-sedimentary rocks, Mesozoic neritic carbonate rocks.	
3. Ambelakia unit		
3a. Ambelakia unit (s.s.)	Marbles, quartzites, intermediate volcanic rocks	Ambelakia unit of Jacobshagen (1986) and Schmitt (1983) (in the Ossa area only)
3b. Kokkinopilos unit	Limestone and flysch overlying volcanic melange with blueschist metamorphism	Ossa unit of Jacobshagen (1986) and Schmitt (1983)
4. Olympos unit	Triassic, lower Cretaceous-Eocene neritic carbonate rocks, Eocene flysch	

The Pelagonian continental basement includes paragneisses, minor amphibolites and orthogneisses, deformed during alpine orogenesis. Locally, the orthogneisses are little deformed and cross-cut the paragneisses and amphibolites, demonstrating a pre-Alpine, probably Hercynian, deformation. Two tectonostratigraphic units of basement are distinguished. In the upper unit, the Kastoria unit, plutonic rocks and orthogneisses have been described from Varnountas and its southward continuation at Kastoria in northwestern Macedonia. In the lower unit, the Flambouron unit of Papanikolaou (1984), plutonic rocks include the Pieria granodiorite in the High Pieria Mountains and the Livadi granite to the south; the Olympiada area west of Kato Olympos; the Deskati series of the Kamvounia mountains; and the Verdikoussa area (Fig. 20). U-Pb dates on zircons from the granitoids cluster around 302 Ma (Appendix 1) indicating Hercynian emplacement. $^{39}Ar/^{40}Ar$ and K-Ar ages (Appendix 1) suggest unroofing and closure to argon diffusion at 295–290 Ma. Earliest thrust deformation of the unit (D1 of Schermer

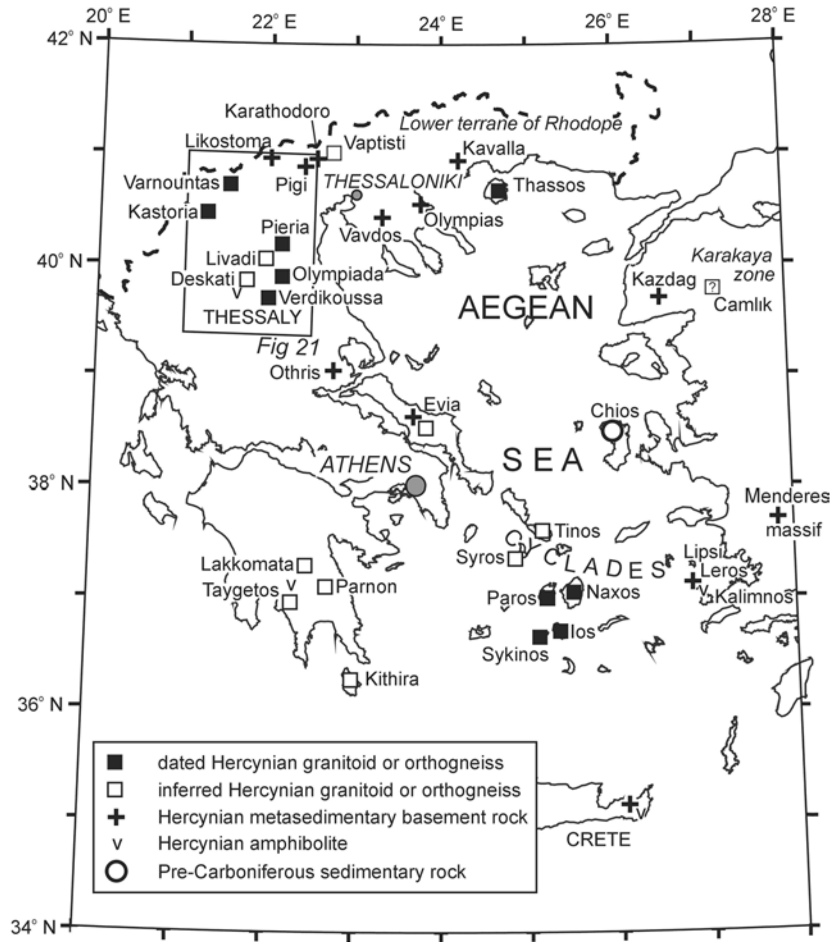

Fig. 20. Map showing general distribution of Hercynian rocks in Greece.

1993) has been dated as Cretaceous by K-Ar on muscovite at 98–107 Ma (Appendix 1). This thrusting is associated with metamorphism close to the greenschist-epidote-amphibolite facies boundary at high structural levels (Yarwood & Dixon 1979) and greenschist to close to the blueschist-greenschist transition at lower structural levels (Schermer 1990). The latter is visible on some $^{39}Ar/^{40}Ar$ spectra as ~100 Ma dates (Schermer et al. 1989, Lips et al. 1998).

Schermer et al. (1990) distinguished two early Tertiary deformational events, dated by $^{39}Ar/^{40}Ar$ to 53–61 Ma and 36–42 Ma, that occurred under blueschist facies conditions (T = 200°–350°, P = 4.5–8 kbar: Schermer 1990; see also Lips et al. 1998). Mylonitic shear zones with a top to SW sense of shear were interpreted as thrusts by Schermer (1993) and Godfriaux & Ricou (1991) and as extensional shear zones by Kilias (1991) and Sfeikos et al. (1991). In contrast, Barton (1976) and Doutsos et al. (1993) interpreted structures to indicate NE-directed compression; Schermer (1993) interpreted the same structures as a later shallower-level deformation (D4). These early Tertiary deformational events were associated with the subduction of the residual Cretaceous Pindos ocean basin and thrusting of Pelagonian rocks over the margin of the Olympos micro-continent.

2.2.2 Geology of the Verdikoussa pluton

The granitoid Verdikoussa pluton of northern Thessaly (Fig. 21), described by Pe-Piper et al. (1993a), is taken as a type example of Hercynian plutons of the Pelagonian basement. It consists principally of granodiorite, with some areas of granite, and marginal diorite. It occurs as an antiform with two structural units separated by a major mylonite zone (Fig. 22): the upper unit comprises mostly orthogneisses whereas the lower unit is mostly undeformed granodiorite. The pluton is in tectonic contact with structurally higher orthogneiss of the Pelagonian basement (Kilias & Mountrakis 1987), probably with a Hercynian granitoid protolith. The only intrusive relationships seen are where granodiorite intrudes older diorite.

In the less deformed areas, the granodiorite shows irregular patches and layers of more felsic and more mafic lithologies, with both granodiorite and tonalite present. Some rocks contain K-feldspar megacrysts. Earlier pegmatite and later aplite veins with a parallel orientation cut the granitoid rocks. Some later diorite sheets within the granodiorite are also cut by aplite. Rocks identified as granites from chemical analyses appear to be small bodies with tectonised contacts against granodiorite. Stocks at the margin of the pluton include early diorite (intruded by granodiorite with a veined igneous contact) and hornblende diorite pegmatite.

2.2.3 Geology of other plutons of the Pelagonian zone

Other plutons within the Flambouron unit include the **Pieria** (= Kataphygion, Kastania, Flambouro) granitoid, described by Yarwood & Aftalion (1976) and Kotopouli et al. (2000), which outcrops over about 100 km^2 of the High Pieria Mountain northwest of Mount Olympos. Parts of the granitoid are almost undeformed, other parts are ortho-

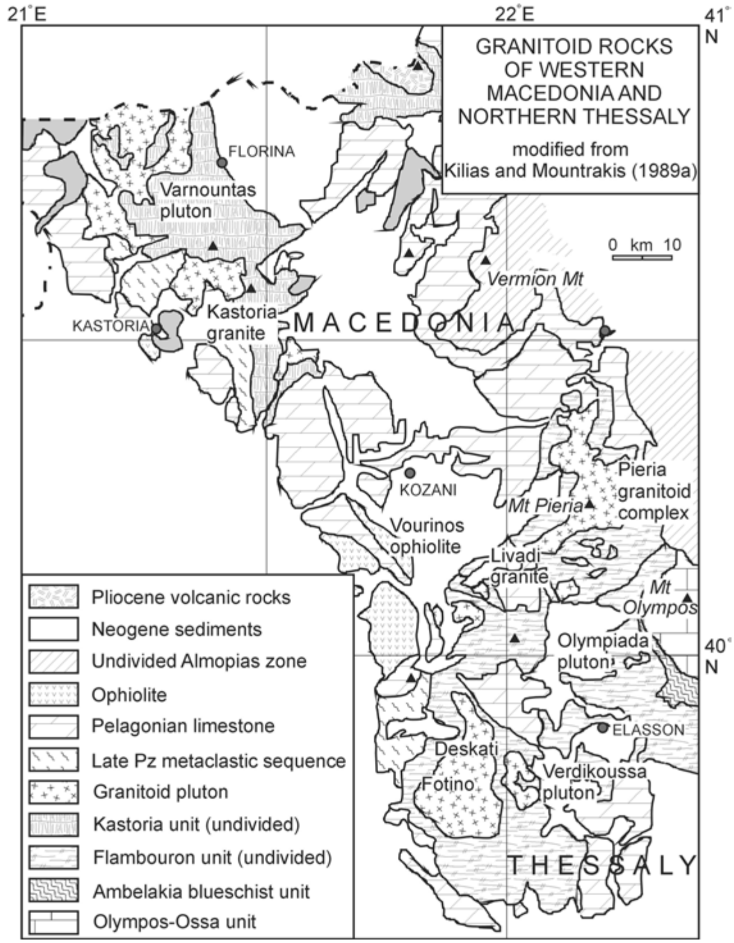

Fig. 21. Geological map of the Pelagonian nappes of northern Thessaly and Western Macedonia (modified from Mountrakis & Kilias 1989a) showing granitoid plutons and Hercynian basement of the Kastoria and Flambouron units.

gneisses (Yarwood & Dixon 1979). Composition ranges from granite to adamellite, with enclaves of tonalite and xenoliths of metamorphic schists. Pegmatites and aplites cut the granitoid: their orientations appear random in the less deformed parts of the pluton. Mafic sheets cutting the pluton are generally transposed into the foliation. No intrusive contacts are seen with country rock: the granitoid appears to occupy a synform. The nearby **Livadi** granite (Nance 1981) occurs in a structurally lower sheet of the Pelagonian basement nappe. It is intruded into a sequence of meta-sandstones overlying a conglomerate with quartzite pebbles. In the same area, another molassic terrigenous metasedimentary rock sequence of epidote-amphibolitic schists (greywacke protolith) and quartzose conglomerates appears derived from a source area with mafic igneous rocks and passes up

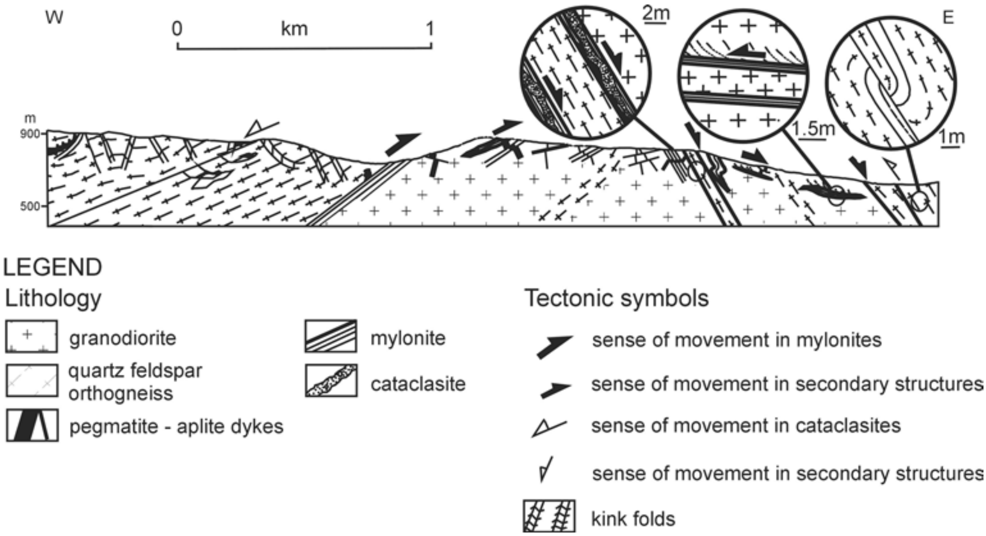

Fig. 22. Structural cross section through the Verdikoussa pluton (modified from Pe-Piper et al. 1993a).

into arkoses similar to those of the Aposkepos series described below. These metasedimentary rocks are cut by both granitic and mafic dykes (Nance 1977).

The **Olympiada** (=Diava) granitoid, described by Pingot (1988), Schermer et al. (1989), Pe-Piper et al. (1993b) and Katerinopoulos (1997), outcrops near the villages of Olympiada, Flambouron and Kriovrissi in the Pierian structural unit west of Kato Olympos and structurally overlies the Ambelakia unit. The main rock unit is granodiorite and tonalite, in places showing little deformation, but commonly deformed to orthogneiss. In parts of the granodiorite, a blueschist metamorphic assemblage is developed. The granodiorite is cut by aplite veins and locally by dykes of trachyandesite composition (see Ch. 3.5.1). It contains common mafic enclaves generally a few centimetres in size but exceptionally measuring as much as a metre.

The Deskati series of the Pelagonian basement of **Kamvounia** Mountain in northern Thessaly (Kilias & Mountrakis 1987) consists of granites and granodiorites, commonly deformed to gneiss and mylonite (Sfeikos & Frisch 1993). The plutonic rocks intrude amphibolites and amphibolitic schists that experienced Hercynian metamorphism (Sfeikos & Frisch 1993). The plutonic rocks are best known near the villages of Deskati and Fotino (Kilias & Mountrakis 1989b, Katerinopoulos et al. 1994, 1996).

The Kastoria unit lies to the north of the Flambouron unit (Fig. 21). It is intruded by the **Varnountas** (= Florina, Eastern Varnountas) granitoid (Kilias 1980, Katerinopoulos 1986, 1988a, b, Koroneos & Christofides 1990, Koroneos et al. 1991, 1994), which is a hornblende-biotite quartz monzonite to monzodiorite porphyry, with minor bodies of leucogranite, biotite granite to granodiorite porphyry, and hornblende-biotite quartz diorite to monzodiorite (IUGS modal nomenclature). These minor bodies form NNE-striking zones 0.5 to 3 km wide. The Varnountas granitoid appears to continue northward to the Baba Mountain granitoid of southern former Yugoslavia (Katerinopoulos & Kyriakopoulos 1989, Katerinopoulos et al. 1992). According to Kilias (1980), the Varnountas granitoid cuts the **Kastoria** gneissic granitoid (Mountrakis 1984), which outcrops to the south and reportedly resembles the Pieria granitoid. Both the Varnountas and Kastoria granitoids cross-cut foliation and show contact metamorphism against Hercynian paragneisses (Kilias 1980, Mountrakis 1984). The granites are unconformably overlain by the terrigenous metasedimentary rocks of the Aposkepos and Sidirochori Series (Mountrakis 1984), which pass stratigraphically upward into Middle Triassic limestones. The terrigenous metasedimentary rocks are of molassic affinity and include conglomerate, arkose, pelites and minor calc-schists. The arkose contains K-feldspar megacrysts similar to those in the Kastoria granite. The conglomerates contain rounded quartzite, gneiss and schist. In the Kastoria region, Plastiras (1979) described aplitic granite dykes cutting schists; a similar relationship to that seen near the Livadi granite.

2.3 Probable Hercynian plutonism in northeastern Greece

2.3.1 Vardar zone

Possible pre-Alpine basement rocks occur in various localities in the Vardar zone, Greek Serbo-Macedonian zone and Rhodope zone. In the Almopias subzone, the Likostoma gneiss outcropping northeast of Aridea (see Ch. 5.2.1) is interpreted as Hercynian (Kockel 1986c, Migiros & Galeos 1987). In the Païkon subzone, feldspar-chlorite-micaschist, interpreted as Hercynian basement, outcrops on Tzena mountain in the Kožuf Massif and is overlain by a thick Triassic to mid-Jurassic platform carbonate succession (Kockel 1986c; see also Ch. 5.2.7). Upper Jurassic rhyolites of the Païkon subzone contain xenoliths of mica- and sericite-epidote-schists (Mercier 1966a).

The so-called Štip-Axios massif of the Peonias subzone is best known north of the international border, where cordierite-sillimanite gneisses outcrop between Štip and Radoviš, at Gradeška Planina, Valandovo and Bogdanci (Kockel 1986c). Small outcrops in Greece include the Karatodoro Gneiss (Mercier 1966a, 1973) and the Pigi migmatite (Bébien & Mercier 1977, Zachariadou & Dimitriadis 1994; see also Ch. 5.2.8). The ages of these units are unknown, but they appear to be pre-Mesozoic and are interpreted as extended slivers of the rifted Serbo-Macedonian continental margin by Zachariadou & Dimitriadis (1994).

Two-mica gneiss and amphibolite at the southwestern margin of the Vavdos ultramafic complex of the Thessaloniki ophiolite was also assigned to the Štip-Axios massif by Kockel et al. (1971). Ricou & Godfriaux (1994), in contrast, interpret this and nearby high-grade metamorphic rocks as a klippe of the Vertiskos unit of the Serbo-Macedonian zone of probable Cretaceous age.

In the eastern part of the Peonias subzone, small granitic stocks associated with pegmatitic veins cut the Vaptisti schists and have yielded Rb-Sr ages of 290 Ma (Appendix 1; Mercier 1966a). The granites are fine grained with calc-alkaline affinities. The Vaptisti schists include basaltic flows 10–15 m thick, of unknown age (Mercier 1966a).

2.3.2 Serbo-Macedonian and Rhodope zones

Pegmatites in the Greek Serbo-Macedonian crystalline basement north of Thessaloniki have been dated at 280–300 Ma (Appendix 1; Borsi et al. 1965). Rb-Sr whole rock geochronology of biotite-paragneiss from Olympias in the eastern Serbo-Macedonian zone (Mantzos 1991) yielded an early Carboniferous (337 ± 5 Ma) age for the high grade metamorphic event and an early Cretaceous (113 ± 11 Ma) age for greenschist facies retrograde metamorphism. Elsewhere in the Serbo-Macedonian zone, there is a general eastward decrease in radiometric ages interpreted by Harre et al. (1968) and Kockel et al. (1977) as evidence of rejuvenation of a Hercynian or older basement.

The "Lower Terrane" of the Rhodope nappe pile (Burg et al. 1996) consists of the Pangeon unit of marble underlain by gneiss, in which the oldest deformation appears cut by granodioritic orthogneiss. The marble of the Pangeon unit locally contains indeterminate corals (Meyer & Kockel 1986) and is correlated with the Boz Dag marble of Bulgaria, which contains ?brachiopods of Devonian or older age (Ancirev et al. 1980). Further circumstantial evidence for a Paleozoic age for at least part of the Rhodope zone is the presence of Ordovician to Devonian acritarchs and chitinozoa and Devonian bivalves in the circum-Rhodope zone of Bulgaria (Lakova & Latcheva 1990). These data suggests that the orthogneiss may be of Hercynian age.

On the island of Thasos, pre-tectonic granodioritic orthogneiss was intruded into metasedimentary rocks of the Lower Gneiss Series (Atzori et al. 1990) and both now show mylonitic deformation (Schulz 1992). U-Pb chronology by Wawrzenitz et al. (1994) on clear prismatic zircons yielded an age of 299±1 Ma; turbid cataclastically deformed zircons show Pb loss. Rb-Sr chronology of white mica in an associated pegmatite yielded an age of 244 ± 9 Ma. Secondary minerals yielded Early Miocene (white micas) and Middle Miocene (biotites) ages.

The Kavala or Symvolon pluton, interpreted as an Early Miocene granodiorite pluton (Ch. 7.5.1), contains inherited zircons of probable Hercynian age (Kokkinakis 1980, Dinter et al. 1995). The pluton is most highly deformed near the Kavala-Komotini fault zone, where orthogneisses resembling deformed Kavala granodiorite are intercalated with paragneisses that yielded a poor Rb-Sr isochron of 260 ± 5 Ma (Del Moro et al. 1990).

2.3.3 Basement rocks of western Anatolia

The Sakarya Block of northwestern Anatolia is a complex Mesozoic microcontinent formed by collision of smaller blocks along the Triassic Karakaya suture zone of northern Anatolia. One of the units in the collisional melange, the Kalabak Formation (Okay et al. 1991) of low grade phyllites and quartzo-feldspathic schists, is intruded by the Camlik metagranodiorite, which has yielded an early Devonian U-Pb zircon age (Okay et al. 1997). Elsewhere, granites and granodiorites, with associated W, Mo, Fe, Zn, Cu, S and F mineralisation, intrude the melange in western Anatolia (van der Kaaden 1959), and have yielded an earliest Permian K-Ar biotite age of 278 ± 3 Ma (Cogulu & Krummenmacher 1967). Older continental crust is represented by the Kazdag massif core complex of quartzo-feldspathic gneiss, marble and amphibolite, intruded by an Oligocene-Miocene granodiorite. Radiometric dating of the Kazdag gneisses suggests pre-late Triassic protolith (Bingöl 1971). Although basement rocks of the Sakarya Block are not known in Greece, the Karakaya suture extends through Lesbos and Chios.

2.4 Hercynian basement rocks in southern Greece

2.4.1 Cyclades islands

Pre-Alpine plutonic rocks outcrop on the Cycladic islands of Ios, Sikinos, Naxos and Paros. On **Ios**, augen gneiss and garnet-mica-schist is intruded by tonalite to granodiorite with relict biotite (Maar 1980, Maar & Jansen 1983, Forster & Lister 1999). Minor mafic bodies interpreted as meta-lamprophyre (Maar 1980) also occur. Henjes-Kunst (1980) determined that the tonalite to granodiorite had S-type granitoid geochemistry and Rb-Sr whole rock dating suggested an intrusion age of about 500 Ma (Henjes-Kunst & Kreuzer 1982). They then experienced Hercynian (300 Ma) high-grade metamorphism. The old age in Ios may correlate with the Menderes nappes of southwestern Turkey (Ring et al. 1999a) which comprise Pan-African basement including 520–550 Ma orthogneisses (see below).

On **Sikinos**, a schist sequence similar to that on Ios is intruded by (meta)diorite (Maar et al. 1981), which yielded a Rb-Sr whole rock "errochron" of 275 ± 87 Ma (Andriessen et al. 1987), which although very imprecise indicates that the diorite is likely late Paleozoic or early Mesozoic in age.

The metamorphic core of the island of **Naxos** consists of repeated slices of Mesozoic metasedimentary rocks and minor volcanics resting on basement of granodiorite and local granite and diorite, in places deformed to orthogneiss or with migmatitic textures (Pe-Piper et al. 1997). Reischmann (1998) obtained average single-zircon ages from three granitic gneiss samples of 233, 275 and 316 Ma (Appendix 1). These older ages confirm earlier determinations by Andriessen et al. (1987) suggesting a Hercynian basement. The younger age is evidence of Triassic plutonism, similar to that recognised in the Menderes massif (Dannat 1997) (see Ch. 3.4.3). Gneissic basement on **Paros** locally includes orthogneiss comparable with that on Naxos (Robert 1982, Tarney et al. 1998) and single zircon dating yielded more coherent ages ranging from 302 to 325 Ma. Less deformed areas of the Naxos granitoids show many original igneous textures. For example, east of Kourounohori, somewhat foliated granodiorite contains metre-decimetre size diorite bodies cut by felsic veins. Some granodiorite contains abundant smaller mafic inclusions similar to those in the Olympiada granodiorite of Thessaly; elsewhere, more tonalitic bodies similar to those in the Verdikoussa pluton are seen. The granodiorite is cut by (now boudinaged) mafic dykes. Near the Skeponi dam-site, extensive diorite and granodiorite occur, together with net-veined hybrid mixtures of the two lithologies.

Possible Hercynian basement occurs on other islands of the Cyclades. The albite-K feldspar gneiss of Vari in southeast **Syros** was interpreted as Hercynian by Hecht (1984). The base of the lower tectonic unit on **Tinos** includes orthogneisses (Melidonis 1980a, Altherr et al. 1982) with an overlying sequence of metasedimentary and metavolcanic rocks similar to that on Naxos.

The **Menderes** Massif of western Turkey comprises a nappe pile that has some similarities to the Cycladic nappe pile. The basal unit of the Menderes Massif comprises three nappe units (Ring et al. 1999a). The Bozdag nappe was intruded by early and middle Triassic granitoids (Dannat 1997). The overlying two nappes comprise a Pan-African basement nappe and a Pan-African cover sequence, both of which contain ca. 550 Ma orthogneiss or granite (Hetzel & Reischmann 1996). This basal unit, therefore, may be correlative with the older rocks of Ios and rocks with Triassic granitoids in Naxos and Samos (see Chapter 3.4.3). The middle unit (Dilek nappe and Selçuk ophiolitic melange) correlate with the post-Hercynian cover sequence of the Cyclades (Ring et al. 1999a).

2.4.2 Crete, Dodecanese and the Peloponnese

In the lowest allochthonous nappe complex of eastern Crete (west of Sitia), the Phyllite-Quartzite unit is a tectonic melange with intercalations of Hercynian basement (Seidel 1978). Two units are recognised (Franz et al. 1990). The upper Chamezi unit consists of paragneisses, micaschists and quartzites. The lower Myrsini unit consists principally of amphibolites with subordinate marble, paragneiss and micaschist. The amphibolites have tholeiitic chemistry. K-Ar dating of hornblende from amphibolite and muscovite from gneiss and micaschist gave ages of 300–270 Ma; some muscovites show resetting by alpine metamorphism (Seidel et al. 1982).

Correlative Hercynian basement is present in the Dodecanese islands of Kalymnos, Leros and Lipsi. The lower Panormos unit consists of 100-m-thick amphibolites overlain by paragneisses, micaschists and marbles. The upper Emporios Unit is correlative with the Chamezi unit of Crete. The amphibolites also have a tholeiitic chemistry (Katagas & Sapountzis 1977). Peak Hercynian metamorphic conditions were estimated as 600 °C and 7 kbar (Franz 1991).

In the Phyllite-Quartzite unit of the southern Peloponnese (see Chapter 3.3.1), the Arna unit in the Taygetos Mountains (Skarpelis 1982, his figure 20) consists of slices of tholeiitic metabasalt, metapelite, metaconglomerate and serpentinized harzburgite. These may be correlative with the Hercynian amphibolites of Crete. Pebbles in Permian conglomerates of the Phyllite-Quartzite unit include metapsammite of probable Hercynian age (our unpublished data). Although Thiébault (1982) suggested that Hercynian metamorphic basement was present in the Lakkomata complex (Fig. 49), later authors have not confirmed this. In the Parnon mountains, a 200-m-thick leucogranite with aplite veins was interpreted as Hercynian plutonic basement (Thiébault 1982). Granite also occurs as a tectonic slice in the island of Kithira (C. Katagas and C. Kotopouli, pers. comm. 1997).

In southern Argolis, Gaitanakis & Photiades (1992) interpreted the small granodiorite of Thermissia as representing Hercynian basement, but it appears to be related to Jurassic ophiolites and is discussed in Chapter 4. The Jurassic ophiolitic melange of central Argolis (Photiades & Skourtsis-Coroneou 1994) contains clasts of marble, metaquartzite, micaschist and amphibolite that might have a Hercynian protolith or may be related to sole metamorphism (Clift & Dixon 1998).

2.4.3 Attiki, Othris, Evia and Chios

Basement rocks of Evia consist of highly altered, medium-grade, pelitic metamorphic rocks, intruded by felsic to intermediate igneous bodies (Stampfli et al. 1996). Medium-grained granites are in tectonic contact with Triassic limestone in central Evia and appear to intrude late Paleozoic schists (D. Mataranga, pers. comm. 1997). The basal Triassic conglomerate in central Evia contains abundant crystalline clasts (Guernet 1971, Kauffmann 1976). In Othris, the Pteleon Formation consists of metamorphic basement underlying Permian sedimentary rocks (Smith et al. 1975); although principally schists, it includes thin marbles and amphibolites (Ferrière 1982). Although no definite Hercynian basement rocks outcrop in Attiki and Chios, crystalline basement clasts are found in mid-Permian conglomerates in Pateras Mountain in Attiki (Spiliadis 1961) and in Chios (Herget 1968).

2.5 Geology of volcanic rocks of Karakaya suture zone

2.5.1 Introduction

The Karakaya suture zone of northern Anatolia (Fig. 23) includes metamorphosed oceanic lithosphere of blueschist, greenschist and amphibolite facies; and accretionary melange rocks of flysch with blocks of volcanics rocks and limestones that have been dated as late Carboniferous to late Triassic. In eastern Turkey, the closure of the Karakaya suture is constrained by overlying late Early Jurassic rocks (Tekeli 1981, Koçyiğit 1991). In western Turkey, the melange includes middle Triassic limestone blocks (Kaya & Mostler

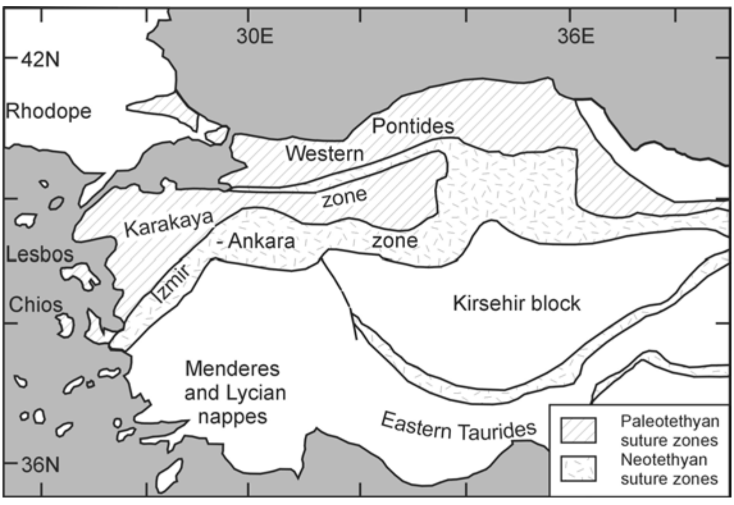

Fig. 23. Geological map of Turkey and the Aegean Sea showing the setting of the Karakaya zone (modified from Pickett & Robertson 1996). Suturing of the Karakaya zone produced the Sakarya continental block by Late Triassic.

1992) associated with alkaline oceanic basalts (Pickett & Robertson 1996) and is unconformably overlain by Late Triassic rocks. Şengör et al. (1984) extended the Karakaya suture as far east as Lesbos and northern Karaburun (Brinkmann et al. 1972). Granites and granodiorites, with associated W, Mo, Fe, Zn, Cu, S and F mineralisation, intrude the melange in western Anatolia (van der Kaaden 1959), and have yielded earliest Permian K-Ar biotite ages of 278 ± 3 Ma (Cogulu & Krummenacher 1967). However, the Camlik metagranodiorite has yielded an early Devonian U-Pb zircon age (Okay et al. 1997) and thus may be tectonically intercalated in the melange. Stampfli (2000) argued that the Karakaya suture formed by closure of a back-arc basin south of the Sakarya margin of Eurasia.

Volcanic rocks of western Chios and Lesbos are probably part of the Karakaya suture and are described below.

2.5.2 Geological setting of Chios

The western half of Chios ("autochthon") (Fig. 24) consists of a thick sequence of flysch sandstones and shales with some interbedded limestones, cherts and volcanic rocks that have yielded fossils of Silurian to lower Carboniferous age (Besenecker et al. 1968). This sequence is capped by late Permian shallow water carbonates (Baud et al. 1991) and then

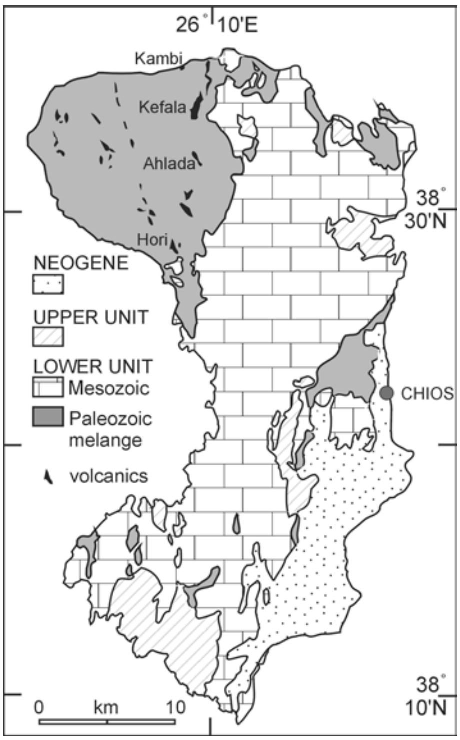

Fig. 24. Geological map of Chios showing the location of Paleozoic volcanic rocks.

unconformably overlain by basal Triassic clastics followed by thick neritic limestones, although the actual contact is obscured (Papanikolaou & Sideris 1983). Thrust over the autochthon is a Carboniferous to Cretaceous nappe in eastern Chios with affinities to the Pelagonian zone. Small Neogene volcanic centres occur throughout the island.

The Paleozoic of Chios has been interpreted by Papanikolaou & Sideris (1983) and Robertson & Pickett (2000) as a "wildflysch" or melange sequence with a ghost stratigraphy of olistostromes preserved within a predominantly flysch succession. Chromite grains occur in the flysch (Stattegger 1984). Silurian and Devonian blocks predominate in the upper units, lower Carboniferous blocks in the lower units. The general form of this melange indicates that it is the along-strike continuation of the accretionary melange of the Karakaya Complex.

Igneous rocks (Fig. 24) are found throughout the Chios accretionary melange, but are more common in the upper part, and were interpreted by Papanikolaou & Sideris (1983) as olistoliths. However, small stocks and sills show clear intrusive relationships with the flysch, with chilled margins and undeformed shale xenoliths (Pe-Piper & Kotopouli 1994). Pyroclastic sequences are locally several hundred metres thick, but contacts with flysch are either not exposed or are clearly faulted and were interpreted as melange blocks by Robertson & Pickett (2000) and possibly of Silurian age. The pyroclastic rocks are generally greenish in colour; no evidence has been seen for subaerial weathering.

Mineralisation has been described from two of the linear belts of volcanic rocks in northeastern Chios. The most easterly belt, from Kambi to Ahladi, has antimony mineralisation; the western belt north of Hori has sphalerite and galena (Melidonis 1978). The antimony of Keramos has been interpreted both as part of the more extensive late Paleozoic Sb-W-Hg mineralisation of western Turkey (Höll 1966), or related to Neogene igneous activity (Moussoulos 1948, Jankovic & Jelenkovic 1997, Skarpelis 1999). The common occurrence of brownish sulphide weathering in the volcanic rocks supports an interpretation of a late Paleozoic origin for the Pb-Zn mineralisation, as does the Pb isotopic composition of galena from Agrilia, in the eastern belt, which yielded a model age of about 320 Ma (Chalkias & Vavelidis 1989) (Fig. 47).

2.5.3 Geological setting of Lesbos

On the island of Lesbos, Katsikatsos et al. (1982a) and Migiros (1992) recognised a Carboniferous to Triassic metaclastite series that they consider autochthonous and lacking in igneous rocks. Thrust over this is a lower unit, hundreds of metres thick, of metabasites interbedded with schists and limestones, some of which contain middle-upper Triassic microfossils; and an upper ophiolitic unit. The metaclastite series is largely of flysch and contains lower Carboniferous sedimentary olistoliths (Hecht 1972): it is thus probably correlative with the accretionary melange in Chios. As in Chios, chromite grains occur within conglomerates. Hornblende-bearing meta-diabase and meta-tuffs were reported within this succession by Hecht (1972), but Katsikatsos et al. (1982a) interpreted all the metabasites as Triassic in age and separated by a major thrust from the Paleozoic section. However, Hecht (1972) described a bedded sequence of (?)lower to middle Permian limestones and shales, within which one 0.25 m thick keratophyric meta-tuff bed occurs. By analogy with adjacent areas of Anatolia (Pickett & Robertson 1996), at least all the Paleozoic rocks are probably allochthonous within an accretionary melange. The Triassic rocks appear to form a tectonic melange beneath the Lesbos ophiolite, emplaced about 153-158 Ma (see Ch. 5.2.20).

2.5.4 The possible extension of the Karakaya suture to Greece

The Karakaya suture in Anatolia appears to result from closure of a Paleotethyan ocean seaway and a back-arc basin during the Permian and Triassic (Şengör et al. 1984, Pickett & Robertson 1996, Stampfli 2000). This convergence created the Sakarya microcontinent, bounded to the south by the Neotethyan oceanic crust of the İzmir-Ankara zone.

In both Chios and Lesbos, late Paleozoic rocks form part of a northwest vergent accretionary melange that in Chios is demonstrably of pre-Triassic age. The Chios volcanic rocks appear to both cut and be intercalated with the accretionary wedge melange and may therefore be of various ages. Mineralisation in Chios resemble the early Permian mineralisation of the Karakaya complex in western Anatolia.

Several authors have attempted to extend the Karakaya suture through Greece. However, Stampfli (2000) argued that a Paleotethyan suture in Greece between Pelagonia and Apulia, represented by the Phyllite-Quartzite Group, was paleogeographically quite unrelated, lying near the Gondwanan margin rather than the Eurasian margin.

Chromite, which occurs in the flysch sequences of Chios and Lesbos, has also been found in Pelion and Skiathos in middle Permian to Triassic sedimentary rocks (Tataris 1975; D. Matarangas, pers. comm. 1992). This may indicate former outcrop of late Paleozoic ophiolitic complexes across the northern Aegean as a continuation of the ophiolites of the Karakaya complex (e.g. Tüysüz 1990) or might be derived from a Paleotethyan suture adjacent to Pelagonia. The ophiolite sequences of the Serbo-Macedonian zone (Dixon & Dimitriadis 1984), discussed in Chapter 5, have also been suggested to represent a Paleotethyan suture (Sakellariou & Dürr 1992).

Papanikolaou & Sideris (1983) compared the "wildflysch" of Chios with the upper Permian to Skythian rocks of Attiki and Salamis (Baud & Papanikolaou 1981), along the trend of the suture proposed by Stampfli (2000). Rather similar rocks occur in the Skiathos Series of the Northern Sporades and Skyros, where metavolcanic rocks interbed with red limestones and overlie metaclastites with minor dark upper Permian limestones (Marinos 1961, Harder et al. 1983). However, the volcanic rocks of the Skiathos Series are more likely to be associated with Permo-Triassic extension (see Ch. 3.5.1).

2.6 Petrography and geochemistry

2.6.1 Petrography of the granitoid rocks

Detailed petrographic and geochemical data is available from only a few of the Hercynian plutons, principally the Varnountas pluton (Koroneos et al. 1993), the Verdikoussa (Pe-Piper et al. 1993a), Olympiada (Pe-Piper et al. 1993b) and Pieria (Kotopouli et al.

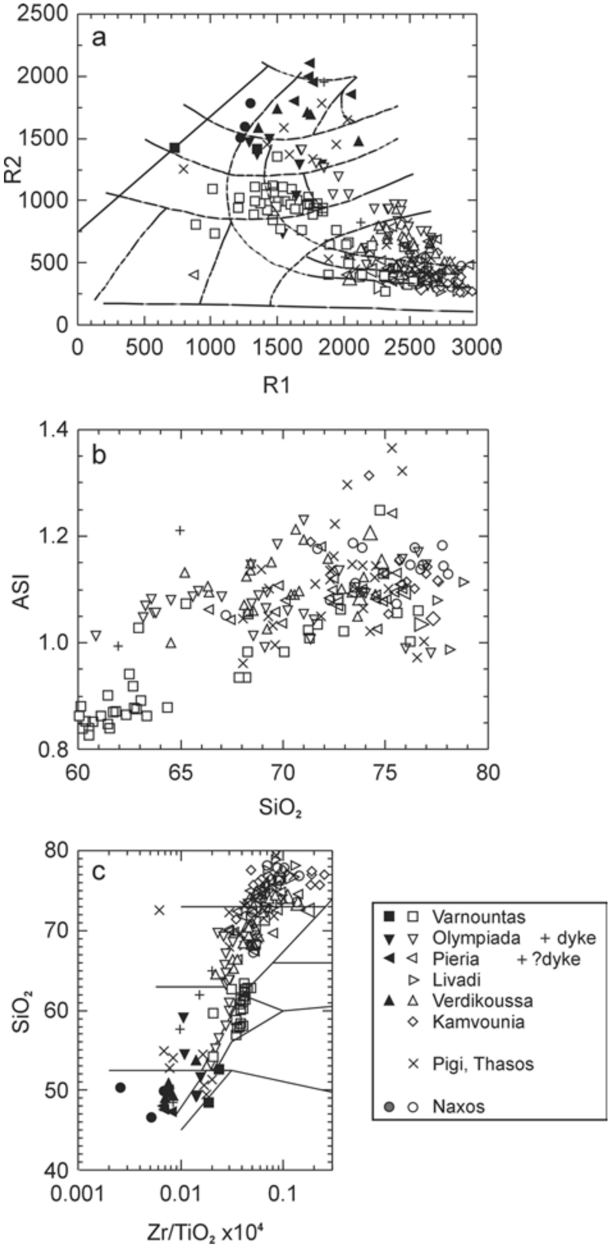

Fig. 25. (a) Chemical classification of Hercynian plutonic rocks (de la Roche et al. 1980, nomenclature: see Fig. 4b). (b) Plot of SiO_2 vs. aluminum saturation index (ASI). (c) Plot of SiO_2 vs. Zr/TiO_2 (after Winchester & Floyd 1977, see Fig. 3c). Solid symbols: mafic rocks, open symbols: felsic rocks. (Data sources in Appendix 3).

2000) plutons of Thessaly, and Naxos (Pe-Piper et al. 1997). Major element data is available from several other plutons (Fig. 25).

The typical granodiorites of Verdikoussa, Olympiada, Pieria, Varnountas and Naxos are highly porphyritic, with phenocrysts of K-feldspar, plagioclase, quartz and a ferromagnesian phase (Fig. 26). Hornblende is the main ferromagnesian mineral in the granitoid rocks of Olympiada, whereas biotite predominates at Verdikoussa, in Naxos and in all but the more tonalitic rocks at Varnountas and Pieria. Muscovite also occurs in some granites. Accessory minerals include allanite, epidote, titanite, Fe-Ti oxides, apatite, zircon and monazite. Zircon from the Deskati granites of Thessaly (Sfeikos & Frisch 1993) is of Pupin's (1980) morphological type D, with a single growth phase, attributed to a mantle origin. Three generations of zircon are recognised in the Varnountas pluton (Magganas et al. 1997b).

The K-feldspar generally occurs as zoned megacrysts up to many cm in length. In those at Verdikoussa and Olympiada, which have been studied in detail, four zones are distinguished (Fig. 27):

(1) the core, which may contain accidental quartz inclusions.
(2) a zone with layers of oriented inclusions of albite and lesser biotite, allanite and apatite.
(3) a zone with layers of quartz inclusions
(4) the rim with microperthite texture and quartz, plagioclase and myrmekite inclusions and microperthite textures.

Plagioclase crystals away from the K-feldspar megacrysts have an andesine-oligoclase composition (An_{32} to An_{27}), with both normal and reverse zoning. Plagioclase crystals in contact with the K-feldspar megacrysts have albitized rims and myrmekite overgrowths. Similar idiomorphic megacrysts of K-feldspar are known from other plutonic rocks (Hibbard 1979, Long & Luth 1986) and form over a wide range of pressure conditions

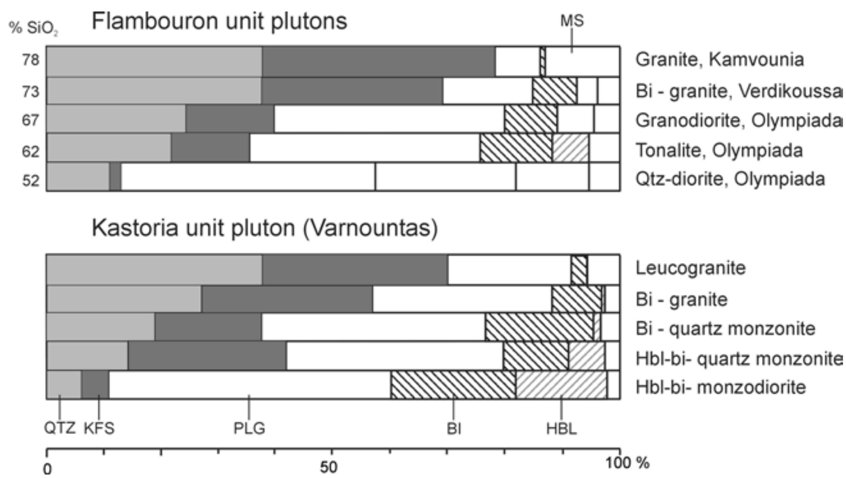

Fig. 26. Modal mineral compositions of representative plutonic rocks of Thessaly. Varnountas pluton from Koroneos et al. (1991); Flambouron unit plutons from Katerinopoulos et al. (1998).

(Mehnert & Büsch 1985). Ba distribution (Pe-Piper et al. 1993a) suggests steady K-feldspar crystallisation under magmatic conditions, with cotectic crystallisation of quartz and two feldspars in zone (3) followed by myrmekite crystallisation in zone (4) and creation of a hydrothermal phase represented by the common aplites and pegmatites. Higgins (1999) has suggested that such crystal coarsening is facilitated by slow cooling of granitoid plutons in proximity to mafic intrusions.

Much of the epidote in the granodiorite is magmatic, on the basis of the petrographic criteria of Zen & Hammarstrom (1984). Allanite and biotite occur as cores to epidote

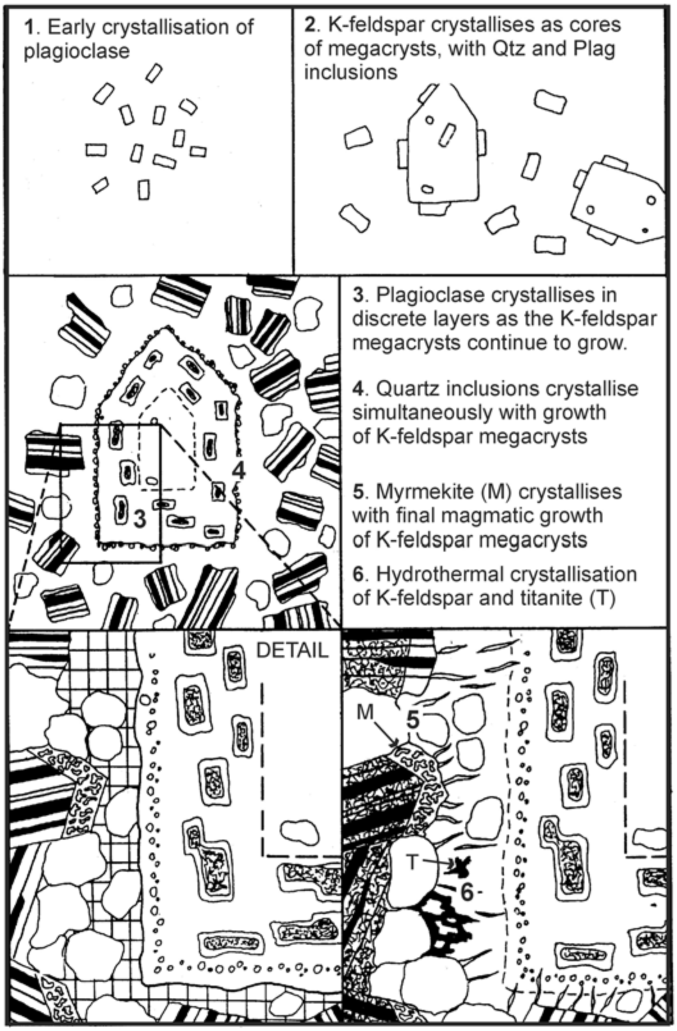

Fig. 27. Petrography and composition of feldspar megacrysts, Verdikoussa. (Modified from Pe-Piper et al. 1993a).

crystals and allanite also rims some epidote. At Verdikoussa, epidote is euhedral with respect to biotite but is highly embayed where in contact with quartz and feldspar. At Olympiada embayed hornblende is enclosed by epidote. Magmatic epidote has a pistacite content of 24–28%, whereas texturally secondary epidote has pistacite > 29%, as do the rims of larger crystals interpreted as secondary overgrowths on previously crystallized magmatic epidote.

2.6.2 Petrography of the mafic plutonic rocks

The mafic sheets within the Verdikoussa granodiorite are fine to medium grained diorites with an equigranular, panidiomorphic granular texture. They consist of amphibole (60–70%: both hornblende and actinolite), brown or brown-greenish mica (about 10%: mostly biotite, rarely phlogopite), and plagioclase and quartz (15–25%). Other minerals present include epidote, chlorite, apatite, titanite, Fe-Ti oxides and rutile. Diorite bodies at Olympiada and Varnountas are similar, consisting principally of amphibole and plagioclase, with minor quartz and biotite. Mafic sheets from the Pieria granites are generally highly deformed and have a metamorphic mineral assemblage. The diorites on Sikinos contain both hornblende and biotite, together with allanite (Maar et al. 1981).

Fine-grained mafic enclaves are common features of the Olympiada, Varnountas, and Naxos granitoids. The Olympiada granitoid contains many fine-grained enclaves (up to 1 m in diameter) generally with a round to ovoid shape. Many show chilled margins or lobate margins with the lobes convex towards the host rock. Many contain large isolated crystals of feldspar, some of which cross the contact between host rock and enclave. All these features indicate that these enclaves represent a contemporaneous hot mafic magma that chilled as it came in contact with the cooler felsic granitic magma with some melting of the granitic member and some mingling. The mafic enclaves are medium grained rocks consisting mainly of hornblende (~ 25%) and biotite (~ 15%) in a matrix of feldspar. In addition, quartz is an interstitial phase, titanite forms large crystals throughout the rock and magnetite is associated with hornblende. Individual samples may contain ilmenite, large quartz crystals rimmed with hornblende, and K-feldspar megacrysts.

Amphiboles (Fig. 28) in the Verdikoussa diorites cluster in the area of edenitic hornblende to hornblende; the hornblende gabbro has amphiboles that range from hornblende to actinolitic hornblende. Hornblende in the Olympiada granodiorite has a slightly lower $(Na+K)_{(A)}$ than that from the Verdikoussa diorites, clustering in the area of pargasitic hornblende – edenitic hornblende – hornblende. Primary amphiboles from Naxos are pargasites.

2.6.3 Geochemistry of the granitoid rocks and associated mafic rocks

Two major associations of granitoid rocks can be distinguished. (1) At Varnountas, in the Kastoria unit of the Pelagonian zone, monzonite, monzodiorite and tonalite predominate: diorite, granodiorite and syenogranite are of minor importance and the rocks are consistently less aluminous (Fig. 25b). (2) The plutons of the Flambouron unit of the

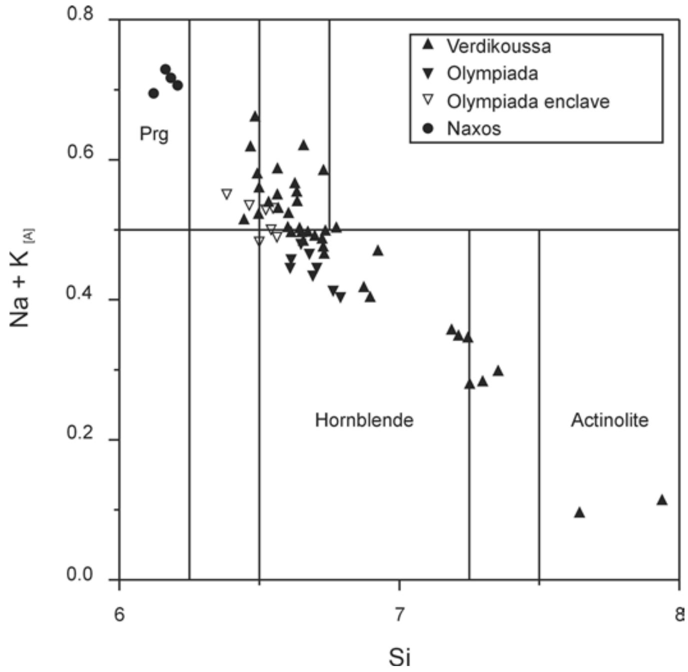

Fig. 28. Composition of amphiboles from Hercynian plutonic rocks. (Data from Pe-Piper et al. 1993a, 1993b; Pe-Piper & Kotopouli 1997).

Pelagonian zone (Verdikoussa, Olympiada, Pieria, Kamvounia), together with plutonic rocks from Naxos and possibly plutonic rocks from Thasos, consist principally of granitoid rocks that form a bimodal suite with minor diorite. At Verdikoussa and Olympiada, granodiorites are associated with minor tonalite, monzogranite and syenogranite (Fig. 25a) and are slightly peraluminous (Fig. 25b). Granitoid rocks from Pieria, Naxos and Thasos are similar, except that the rocks are principally granite rather than granodiorite. The late phase Livadi granitoids consist almost entirely of trondhjemitic granite with low Ba and Sr, but show a similar degree of peraluminous character and generally similar geochemical trends to the main plutons of the Flambouron unit.

The late diorite bodies at Verdikoussa and some diorites from Pieria and Naxos are olivine normative, with Mg# ranging from 0.53 to 0.64, and Cr contents generally > 100 ppm. Their trace element distribution (e.g., sample T51 in Fig. 29) resembles that of many continental flood basalts, with slightly elevated normalised LILE and LREE compared with Nb, Ta and the HFS elements, but overall a rather flat pattern. The high Ba/La ratio (> 5) is typical of island-arc tholeiites (Arculus & Powell 1986) and many samples lack the negative peak in Sr that is common in many continental tholeiites (Thompson et al. 1983). The REE pattern is quite flat (Fig. 30). The early diorite stock from Verdikoussa (sample T60) is similar, with a flat REE pattern, but has elevated LILE, probably as a result of metasomatism by the host granodiorite (Pe-Piper et al. 1993a).