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GREECE by

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Contents

GREECE

Greek capital is Athens, and modern Greece comprises three main parts.

The Greek (or Hellenic) Peninsula, which constitutes the S tip of the Balkans S of Albania and former Yugoslavia, extends into the Mediterranean between the Aegean Sea to the E and the Ionian Sea to the W. Included in the peninsula are to the W, Attica, Peloponnesus, Boeotia, Acarnania and Epirus; and to the E, Thessaly and Western Macedonia.

A strip of land along the N coast of the Aegean Sea includes the provinces of Eastern Macedonia and Thrace. It is bordered by Bulgaria on the N and touches Turkey on the E.

The third part takes in the Ionian Islands (Kerkyra, Cephallynia, Leukas, Zakynthos, and others), Crete, the Aegean Islands or Cyclades and, near the coast of Turkey, the S Dodecanese (Sporades) Islands, among which Rhodes, Chios and Mytilene (Lesvos) are the largest. Out of more than 1000 islands, only 50 are larger than 40 km^2.

The Greek coastline is over 15 000 km, with land borders amounting to 1200 km. Most of the country (80%), even the islands, is mountainous, generally bare and rugged; but in places (especially in the N) it is covered with forests of fir, birch or chestnut. The average elevation is 500 m, which is high compared with the European average of 350 m. The highest peak is Mount Olympus (2911 m). The only plains of importance are those of Thessaly and Macedonia (Salonica), which are up to 80 km wide. The plain of ancient Copais is in Thessaly in central Greece.

Greece has five universities, Athens, Salonica, Patras, Cretan and Aegean, with departments of geology and paleontology, mineralogy, geophysics, physical geography, etc., in the first three. Geology is also taught in the Miniap Faculty of the Athens Technical University, as well as in the technical universities of Salonica and Thrace. The geological survey is called the Institute of Geology and Mineral Research (IGME) and its head office is in Athens (70 Mesogion Street).

Mining activity goes back to early antiquity. Some centers, such as the Laurion (or Lavrion) silver–lead–zinc mines, date from before 1000 BC and were one of the sources of Athens' wealth, especially during the 5th century BC. The ancient State of Macedonia also benefited from the mining of gold and silver.

The ancient Greeks (notably Aristotle, Herodotus, Theophrastus, Xenophanes of Colophon and Strabo) had some sound ideas on geology, but geological research in Greece started mainly at the beginning of the 19th century. It must be noted that there exist some references to the geology and mineral wealth of Greece from the interval between 16th century and the beginning of the 19th century. These references are due mainly to the touristic impressions of several travelers.

In the year 1829, the French 'Scientific Mission in Morea' (Peloponnesus) came to Greece. Among other scientists, two geologists (E. Boblaye and Th. Virlet) took part in that expedition and studied the geology of Peloponnesus as also the geology of Cyclades and N Sporades, publishing their results at 1833.

During the years 1840–1880 an important geological–paleontological study of the Greek land took place. R. Wagner and A. Gaudry revealed the fossil vertebrates at Pikermi (Attica). F. Fouque studied systematically the volcano of Santorini Island, and several other researchers, such as H.E. Strickland, T. Spratt and V. Raulin, studied the geological structure of different areas (Zakynthos, Crete, etc.).

Between 1876 and 1878 the Austrian geologists N. Neumar, A. Bittner, F. Teller and L. Burgenstein visited continental Greece as well as several Aegean islands and published later on the results of their studies. Also the Germans H. Bucking and R. Lepsius (1880–1884) produced a geological map of Attica (scale 1: 25 000).

At the end of the 19th century and at the beginning of the 20th, a new age of the geological research in Greece started. A. Philippson studied the Peloponnesus during the years 1887–1889, central Greece in 1890, Zakynthos and parts of Epirus and Thessaly in 1893, and several Aegean islands in 1896.

C. Renz came to Greece and particularly to Kerkyra in 1903 (or 1902) for the first time, and carried on his research for the following 50 years. He published the results in a monograph entitled *Die Tektonik der Griechischen Gebirge*. In collaboration with the Greeks N. Liatsikas and H. Paraskevaidis (1955), Renz produced the first geological map of Greece (scale 1:500 000).

During World War I (1914–1918) the French C. Arambourg, J. Bourcart and J. Piveteau studied a large part of N Greece. In 1931 and 1933 M. Blumenthal published the first series of tectonic sections of N Peloponnesus.

An essential contribution has been made by many important Greek geologists who worked before or after World War II. Among them we can distinguish N. Liatsikas, H. Gardicas, G. Georgalas, M. Mitsopoulos, I. Trikkalinos, I. Papastamatiou, G. Marinos, G. Voreadis, P. Kokkoros, H. Paraskevaidis, P. Psarianos, A. Galanopoulos, D. Kiskyras and many others.

An important event in the geological research of Greece is represented by the foundation of the Institute of Underground Geological Researches (IGEY) in 1950, (now IGME), which is now one of the main scientific institutions carrying out geological studies in Greece.

After World War II the detailed studies of foreign geologists have contributed essential knowledge about the geological structure of Greece. These include the French geologists J. Brunn, J. Aubouin, P. Celet, J. Dercourt, I. Godfriaux, J. Mercier, C. Guernet, P. Vergely, G. Bizon, J. Fleury, M. Bonneau, F. Thiebault, X. Le Pichon, J. Angelier, J. Ferriere and B. Clemen, German geologists V. Jacobshagen, F. Kockel, H. Walther, H. Mollat, D. Richter, S. Durr, H. Bachann and E. Seidel, English scientists J. Dixon and A. Smith, the Dutch J. Meulenkamp and Italian geologists. Present-day Greek geologists are contributing further and more detailed research on Greece.

A general geological bibliography for Greece is provided by Haralambous (1975–1980). The Geological Society of Greece was established in 1950 and has published 15 volumes of its journal, *Deltion*. There is also a Greek Speleological Society (with its own *Deltion*).

Apart from the above-mentioned organizations, several other teams are active in geological research. The National Oil Company has explored for oil in several areas of Greece with promising results. Also, the Centre of Oceanography is actively working on marine geology in several Greek gulfs and has made contributions on the question of marine pollution.

Structural framework

The position of Greece within the collision area between the European and African plates, including the world famous Hellenic Arc (Fig. 261), has controlled its geotectonic evolution. Thus, Greece was between Laurasia and Gondwana during the convergence and collision of these two supercontinents.

The pre-Alpine crystalline basement is extensive in Greece. Throughout most of Greece (except for possibly the northeasternmost corner) Alpine sediments of Triassic age uncomformably overlie pre-Triassic crystalline deposits which were deformed during the Hercynian Orogeny (Carboniferous–Permain). Mesozoic Tethyan sediments of great thickness were deposited throughout Greece and subsequently underwent deformation during the Alpine Orogeny, which formed the Greek mountain chains. These mountains, apart from Alpine sediments, are composed of pre-Alpine deposits of Paleozoic and/or Precambrian age.

The Greek mountain chains or 'Hellenides' belong to the Dinaric branch of the Alpine system. The Hellenides have been subdivided into a binary series of tectonic or 'isopic zones' (of similar sedimentary facies) composed of a series of 'horsts' and 'graben' according to Auboin (1965) (Fig. 260). These zones are grouped into 'internal' and 'external' based on the timing of deformation and location. The internal zones are located N and E of the External ones. The internal Hellenides were affected by an early phase of orogeny during the Late Jurassic–Early Cretaceous times, whereas the external Hellenides were deformed only by the final phase of deformation during the Tertiary. On Figs 259 and 260, the boundary between the internal and external zones is generally along the belt of ophiolites. Thus the internal zones include the Rhodope, Serbo-Macedonian, Axios, Attic–Cycladic and Anatolikis–Ellados (or Pelagonian) zones. The extenal zones include the Preapulian, Ionian, Gavrovo–Tripolitsa, Pindos and Parnassos zones.

As shown in Fig. 262, all zones are in thrust contact with each other. The Paxos (Preapulian) Zone is the lowest unit in the tectonic stack, and the Serbo-Macedonian Zone is the highest. Thus although the 'horsts' and 'graben' terminology is useful for descriptive purposes, it is important to keep in mind that the various zones are not horst and graben structures in the classical sense. Rather they represent various 'terranes' whose original paleogeographic positions relative to one another are not so clear as formerly thought.

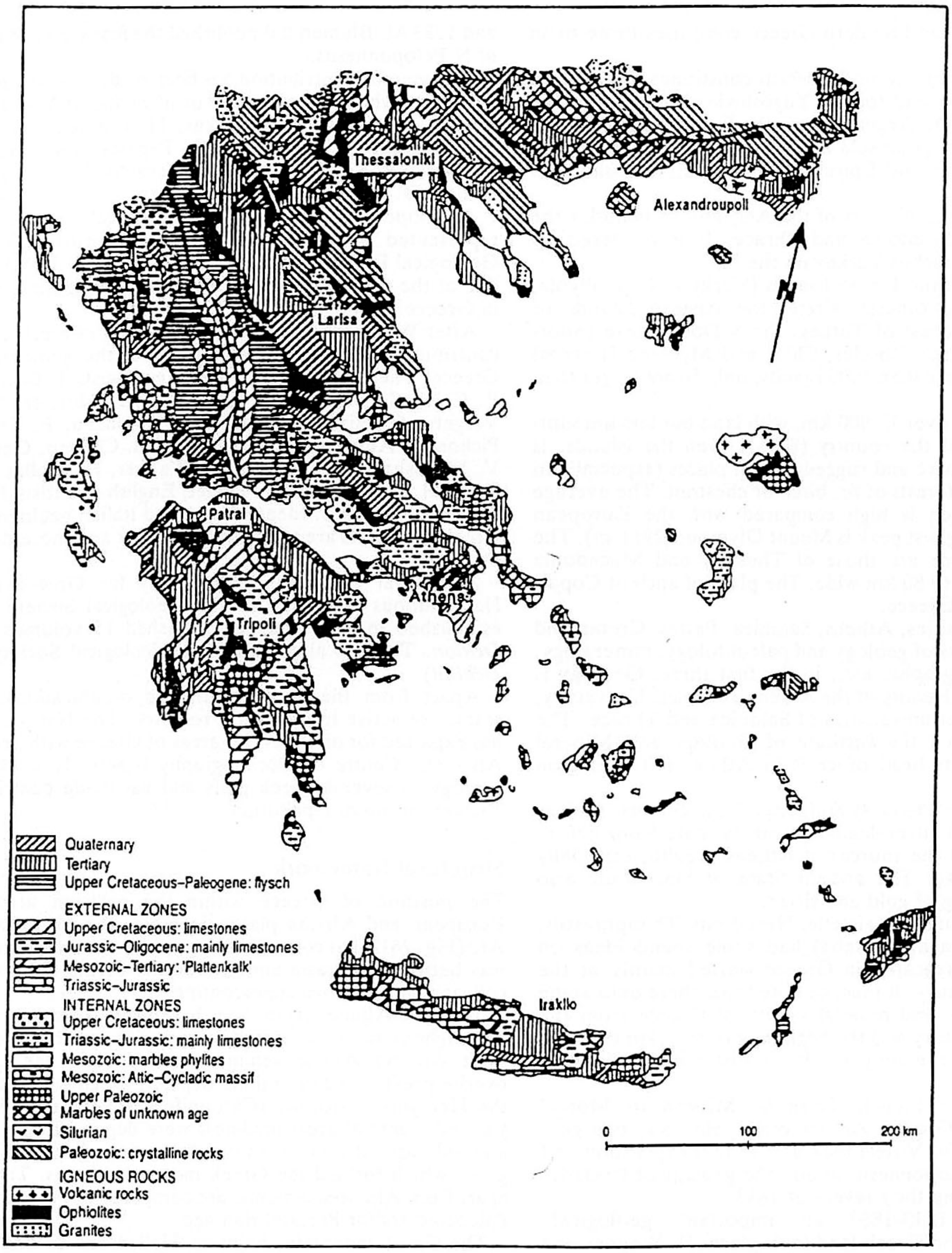


Fig. 259 Geological map of Greece (based on IGRS and IFP, 1966).

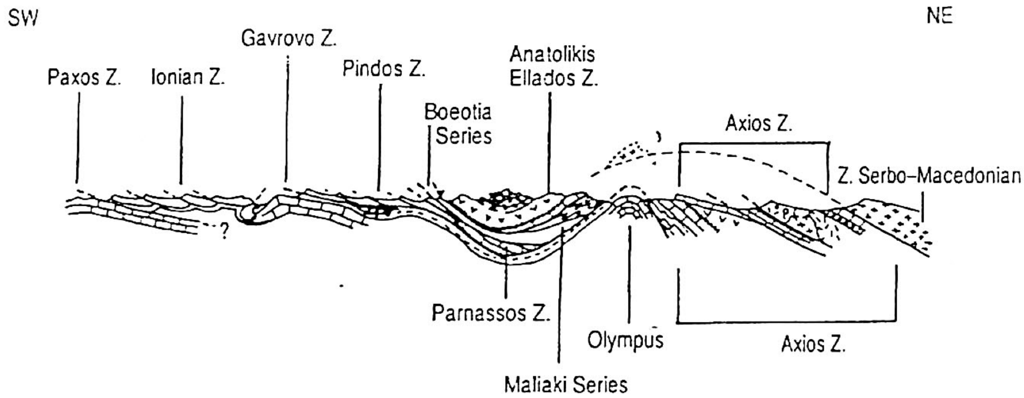


Fig. 262 Schematic section of the Hellenides (based on Aubouin *et al.*, 1977).

- The neotectonic period (the Aegean Arc) affected the Aegean Sea, where the numerous islands correspond to uplifted tectonic blocks, whereas the downthrown blocks (graben) are covered by the sea. The present arrangement in the Aegean Arc is associated with distension, which developed above the present Benioff zone. Thus the Aegean Sea corresponds to a marginal sea that started to form during the Middle Miocene and still is active today (Lort, 1971). The faults are very complex and are accompanied by rotation of blocks (Pucher, Bannert and Fromm, 1974).

Tectonic zones

The tectonic zones are described from SW to NE, i.e. from the external margin to the inner core of the mountain system.

External zones

Paxi (or Paxos or Preapulian) Zone

The Paxi Zone (Renz, 1940) occupies the W part of some of the islands of the Ionian Sea, including Zakynthos, Cefallinia, Leukas, Ithaki and Paxi. Stratigraphically, it is composed of a continuous neritic series of calcareous sediments (Fig. 263). The oldest sediments, known only from drilling, are composed of dolomites and dolomitized limestones of Middle and Upper Jurassic age. These are followed by Cretaceous to Miocene neritic calcareous sediments, marls and gypsum. The Pliocene is transgressive. In contrast to the more easterly zones, there are no flysch deposits (Eocene to Miocene), showing that clastic material originating from the inner zones did not reach this outer part. The Paxi Zone therefore must have had a submarine ridge at the time of the clastic invasion and most probably represented the foreland of the Hellenides. Folding is relatively slight and occurred relatively late (post-Miocene). By calling this zone 'Preapulian', Aubouin (1959) implied that it was linked with Apulia in SE Italy, i.e. that this sedimentary belt once had a shoreline in S Italy. Clastic sedimentation of pelites and sandstones began in the Upper Miocene and is characterized by certain authors as flysch. This feature is the most important characteristic of this isopic zone which, from the tectonic point of view, is considered autochthonous (see Albania; Bosnia; Croatia; Italy; Yugoslavia).

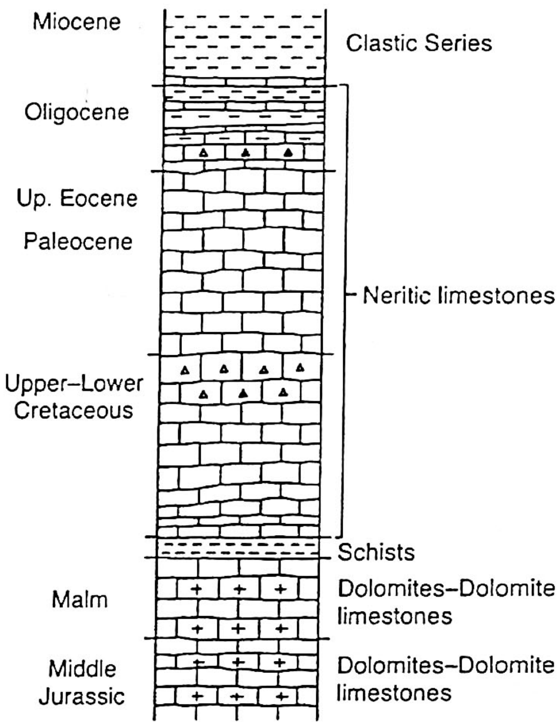


Fig. 263 Simplified stratigraphic column of the Paxi Zone (based on Bornovas, 1964).

Ionian (or Adriatic–Ionian) Zone

The Ionian Zone covers an extensive area in W mainland Greece and appears on the W edge of the Peloponnesus and in some of the Dodecanese Islands (Karpathos–Rhodes; e.g. Christodoulou, 1961).

Upper Triassic evaporites (Fig. 264) (anhydrite, gypsum and halite) are overlain by Lower Jurassic neritic limestones and dolomites known as the Pantocrator (Almighty) Limestone. During Carixian (upper Liassic) times, the initial shallow carbonate platform began to break up (Karakitsios, 1992). The first general deepening of the Ionian area is testified from the Siniais Limestones and their equivalent, the Louros limestones, in which ammonites and brachiopods indicate a Carixian to Domerian (upper Liassic) age. These two facies correspond to the first syn-rift sediments of the Ionian Series marking the beginning of Ionian differentiation.

From Domerian to late Toarcian, the continuation of distension, accompanied by halokinesis of the evaporitic substratum, was accompanied by intense block-faulting which led to internal differentiation of Ionian Basin. Listric normal faults associated with this phase caused the separation of the initial basin into a number of small paleogeographic units (on each tilted block) which underwent differential subsidence. Thus, in the deeper part of the half-graben, 'Ammonitico Rosso' or 'Lowermost Posidonia Beds' were deposited. The tops of the tilted blocks exhibit hiatuses, hardgrounds and sedimentary dikes. These tops constitute either submarine highs or (rarely) emergent areas.

During the Early Berriasian a general sinking of the entire basin is attested by the onset of the deposition of pelagic Vigla Limestones in the whole Ionian Zone. Apart from halokinetic movements, which probably provoked the variation in thickness of Vigla Limestones, the same conditions persisted until the Late Eocene times, when flysch sedimentation set in. Pelagic sediments interfinger with clastic deposits derived from the adjacent Gavrovo and Apulian platforms.

During the Tertiary, compressional phases of the Alpine Orogeny reactivated the pre-existing Jurassic extensional fault system. Listric normal faults were transformed in reverse faults, thrusts or transcurrent faults. This phenomenon was facilitated by diapiric movements and localization of tectonic surfaces along the evaporitic base of the Ionian sequence. The symmetry of the Ionian Basin associated with the Jurassic distensional phase is manifested in the double

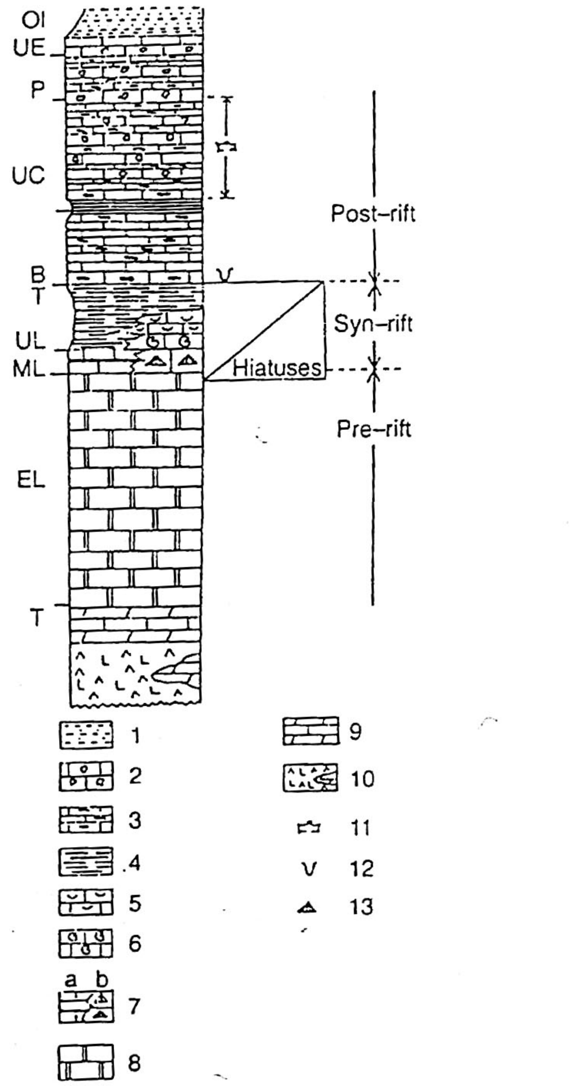


Fig. 264 Representative stratigraphical column of Ionian Zone (based on Karakitsios, 1992). 1, flysch; 2, clastic limestones; 3, Vigla limestones (cherty limestones); 4, non-differentiated *Posidonia* Beds (Upper and Lower *Posidonia* Beds); 5, limestones with filaments; 6, Ammonitico Rosso; 7, Sinias (a) and Louros (b) limestones; 8, Pantokrator Limestones; 9, Foustapidima Limestones; 10, evaporites; 11, Globotruncanidae; 12, calpionellids; 13, brachiopods; Ol, Lower Oligocene; UE, Upper Eocene; P, Paleocene; UC, Upper Cretaceous; B, Berriasian; T, Tithonian; UL, Upper Lias; ML, Middle Lias; EL, Early Lias; T, Triassic.

divergence of its compressional structure (westward in the W and eastward in the E). The Ionian Zone constitutes a good example of inversion tectonics of a basin.

The Ionian Zone appears to be overthrust to the W over the Preapulian Zone, with many folds and thrusts in the Burdigalian ;equence. Molasse facies and other Neogene sediments unconform- ıbly overlie the Alpine sediments of the Ionian Zone.

In the Peloponnesus, Crete and Rhodes, a metamorphosed equence, known as the Plattenkalk Series, displays great litho- tratigraphic similarities to the Ionian Zone; it may correspond to a ıetamorphic Ionian Zone. At the base there are Triassic clastic ediments, which are composed of sandstones, conglomerates and elites. Then carbonates appear in the Upper Triassic to Upper ocene sediments. In the Eocene sequence of south-central Pelopon- esus (Tagetos area) fossils have been found similar to those of the on-metamorphic Ionian Zone in mainland Greece. The Mesozoic is ell represented on Rhodes (Orombelli and Pozzi, 1967). Flysch dimentation began in Lower Oligocene time (Fig. 265); on Rhodes includes tuffs (Mutti, 1965; Mutti, Orombelli and Muzzi, 1965).

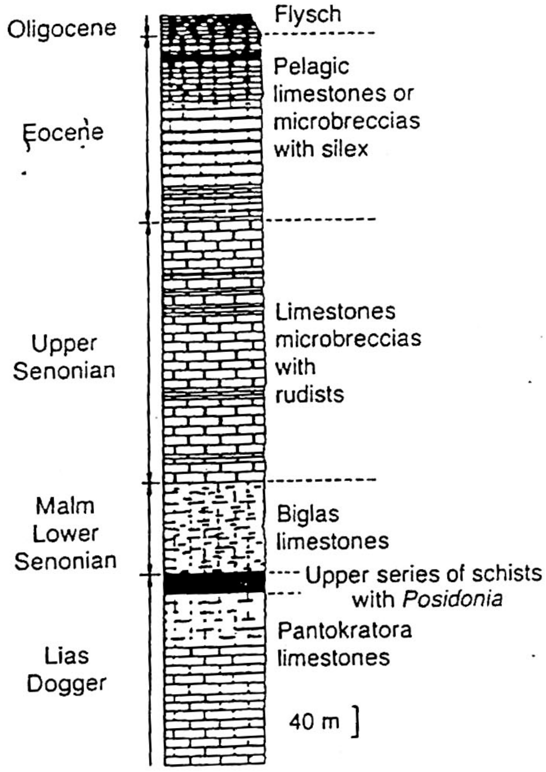


Fig. 265 Simplified stratigraphic column of the 'Plattenkalk Unit' (based on Thiebault, 1977).

In the Lower Miocene, metamorphism affected a thick sequence of schists. This metamorphic phase of the green to red schists occurred later than that of the marbles of the Plattenkalk Series; the schists occupy a large area in the Peloponnesus and in Crete. According to some authors they belong to the base of the Tripolis subzone (discussed next); to others the schists comprise an independent paleogeographic unit; still others believe that they represent the metamorphosed flysch of the Plattenkalk Series or the metamorphosed Ionian Zone.

Tectonically the Plattenkalk unit of the Peloponnesus and Crete is believed to be a parautochthonous zone that suffered multiple phases of deformation.

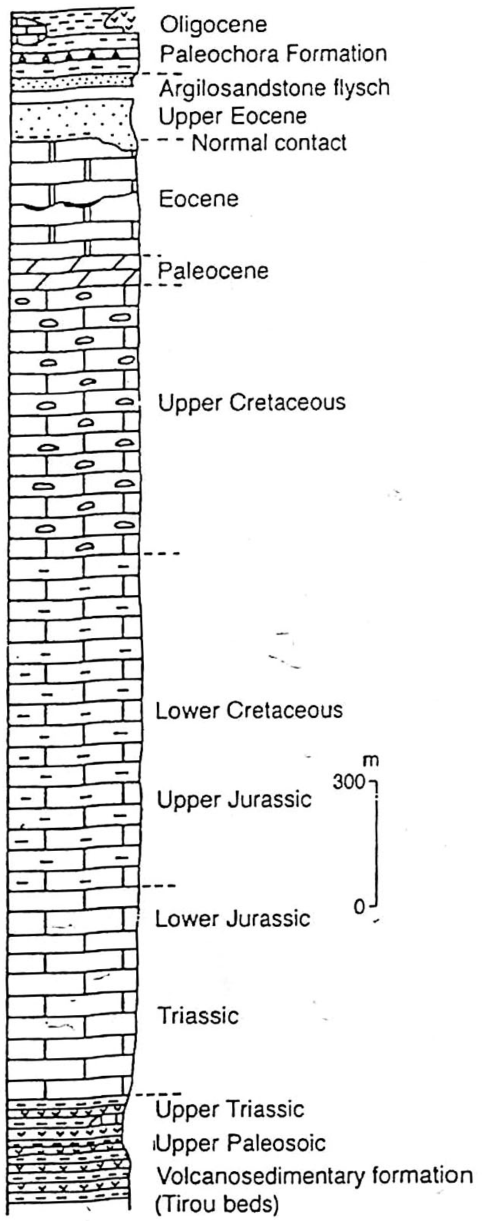
Gavrovo–Tripolis Zone

The Gavrovo–Tripolis Zone (Fig. 266) appears everywhere under the cover of the Olonos–Pindus Zone (see below), in W mainland Greece, the Peloponnesus, Crete and the Dodecanese (Astypalea, Karpathos and Rhodes). In addition, it is also found within the internal zone in the 'tectonic window' of Olympus (Godfriaux, 1970; Fig. 260). In W mainland Greece and in the W Peloponnesus, this zone is designated the Gavrovo subzone and is composed of thick neritic carbonates of Upper Jurassic to Upper Eocene age (Aubouin and Dercourt, 1962).

In the Peloponnesus and in Crete, the Tripolis subzone contains the so-called Tyros Beds – a volcano-sedimentary sequence with limestone intercalations of Upper Paleozoic–Triassic age. Carbonate sedimentation began in Upper Triassic time and continued up to the Upper Eocene. The whole carbonate sequence corresponds to a shallow sedimentation (sub- to supratidal) accompanied by extensive subsidence.

Locally, in the Middle Eocene interval, bauxites indicate a brief emergence. In some areas this emergence continued until Oligocene times, when flysch unconformably covered the carbonates; in other areas the neritic sedimentation grades up to pelagic, with deposition of shales (transitional sediments towards flysch), which in turn pass into flysch.

In the upper layers of the flysch there are olistholiths composed of limestone and igneous blocks. The igneous ones come from the Olonos–Pindus Zone and the limestone ones come from both the Olonos–Pindus Zone and the transitional series of the Maggassa type (see below). There are also conglomerates, the cobbles of which



:66 Simplified stratigraphic column of the Gavrovo–Tripolis (based on Lekkas, 1980).

ıate both from the Olonos–Pindos Zone and the Gavrovo–lis Zone. The whole formation comprises a type of 'wildflysch' 266). It resulted from the arrival of the Pindus cover, which have taken place during the flysch sedimentation in the upper ›ns of the Tripolis subzone.

tonically, in the Peloponnesus and Crete regions the Gavrovo–is Zone comprises an overthrust cover upon the relatively ıthonous Plattenkalk unit. Because of the nature of the thick-d sediments, the folds in the carbonate part of the series are with a large radius of curvature. Overfolds are abundant in the and lower parts of the series.

genic movements (Germanotype block faulting and folding) ›lace from the early Oligocene onward, but there are no ·essive molasse formations that might give more details on the story of the zone. The tectonic style of the Gavrovo subzone is ıaracteristic. Folding is slight, but great normal faults with up) m displacement occurred fairly early, while nappes and ·usts were developing in neighboring zones. The emerged of the Tripolitza Limestones were widely affected by karst , but this surface was buried by the younger flysch sediments. ıe main Miocene orogeny and subsequent block faulting, this ırst was gradually exposed (Richter and Mariolakos, 1972).

The NE limits of the Gavrovo subzone are not known. On the NE side it is always overthrust by the Olonos–Pindus Zone. Its original width may have been considerable, as it appears in tectonic windows of the Pindus Nappe some 50 km E of the overthrust front. The Gavrovo subzone reappears on Crete and Karpathos, where it is extensively overthrust by the Pindus and Subpelagonian nappes (Aubouin *et al.*, 1970).

Olonos–Pindus Zone

The Olonos–Pindus Zone extends from Montenegro (see Yugoslavia) to the island of Rhodes (Dodecanese), a distance of over 2000 km. It corresponds to the deepest trough in Greek territory. It is characterized by a consistently pelagic facies with sediments in thin and graded beds associated with deep-sea turbidity currents. The sediments are usually fine grained, except for occasional microbreccias.

The Olonos–Pindus Zone is wholly allochthonous and may be an allochthonous series emplaced by gravity slides (Temple, 1968). It is found in mainland Greece, the Peloponnesus, Crete and the Dodecanese (Fig. 260).

Stratigraphically it can be separated into six different formations (Fig. 267).

- Clastic sediments of the Triassic: alternating sandstone and shales with lenses and intercalations of limestone. The thickness is less than 60 m.
- Drymos Limestone: pelagic micritic limestones with nodules and intercalations of chert and flint. The upper layers are composed of

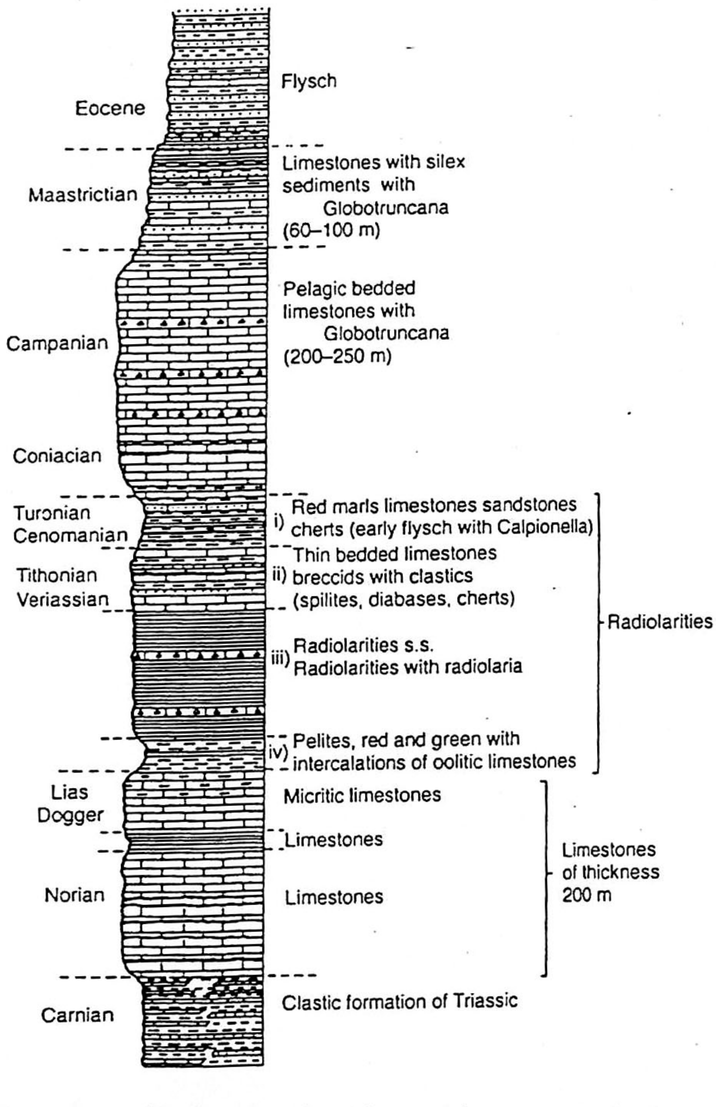


Fig. 267 Simplified stratigraphic column of the Olonos–Pindos Zone (based on Fleury, 1980).

granular limestones and are about 150 m thick, ranging in age from late Triassic to late Lias.

- Radiolarites (s.l.). This group has a thickness of 150–200 m and can be divided into separate formations.
  - Red and green shales with limestone intercalations of Lower Dogger (Middle Jurassic) age.
  - Radiolarites (s.s.): red and green cherts rich in *Actinozoa*, and limestones with *Capionella*. The red limestones with chert and shale intercalations are Upper Tithonian–Lower Berriasian in age.
  - The first Pindus flysch: intercalations of shales, limestones and sandstones, with some volcanic rocks.
- Platy limestones: pelagic limestones of Upper Cretaceous age having a thickness of 150–200 m. They are rich in *Globotruncana*, and along their margins there are breccias of neritic origin.
- Sediments (hemipalagic) that are transitional to flysch composed of alternating shales and pelagic limestones about 100 m thick. The age is Upper Maastrichtian to Paleocene.
- Flysch begins in the Paleocene segment and continues up to the Upper Eocene stratum. It becomes wildflysch because of the arrival of the next internal nappe.

Also notable are sediments having pelagic and neritic characteristics that are recognized as transitional between the Gavrovo–Tripolis Ridge and the Pindus Trench. The flysch in this sequence begins in Eocene time. The sediments are tectonically wedged between the Pindus Nappe and its base, the Gavrovo-Tripolis Zone. On mainland Greece this transitional sequence is known as the Megdova Series. In the Peloponnesus, Crete and the Dodecanesse it is designated the Magassa Series 'of Ethias and Agridakiou'.

Tectonically, the Olonos–Pindus Zone is a huge tectonic nappe, with many (often isoclinal) folds, thrust over the Gavrovo–Tripolis Zone. In the central Peloponnesus the Pindus Nappe is composed of the upper stratigraphic horizons, mainly the platy limestones covered locally by tectonic blocks, radiolarites and shales. This region has the characteristic name 'Arcadian Tableland'. No folding occurs until the late Eocene. The tectonic style is especially mobile, with considerable shortening and numerous overthrusts, nappes and imbrications, but the style is nevertheless relatively simple. Structural units may be followed for ten of kilometers.

The Pindus Trough may not have developed into real ocean, i.e. deep sea with oceanic crust, from which could come ophiolitic rocks. Rather, an extensive thinning of continental crust may have occurred without the actual formation of oceanic crust. The zone may be intermediate between the outer zones, which developed from a former continental marginal platform, and the inner zones, where more thoroughly tectonized (and more properly oceanic) features are present.

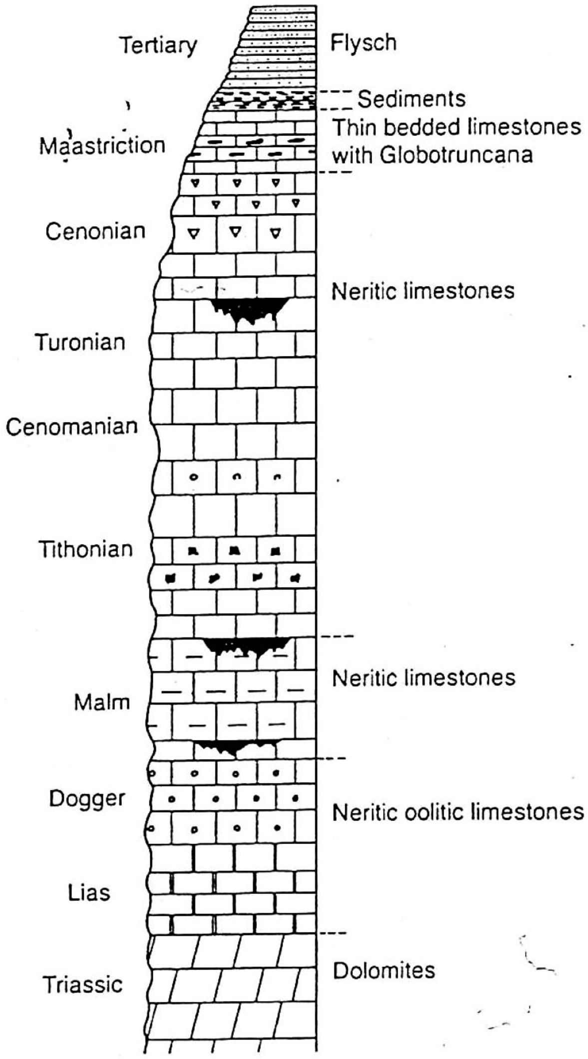


Fig. 268 Simplified stratigraphic column of the Parnassus-Giona Zone (based on Clement, 1977).

Parnassos–Giona Zone

The Parnassos–Giona Zone appears only in the S part of mainland Greece and in Peloponnesus (Fig. 260), but its disappearance northward may be due to tectonic overlap (the Olympus Window, see below). Here again, thick limestones and dolomites are interrupted only by bauxite horizons (Jurassic and Cretaceous in age) that indicate periods of emergence. There can be little doubt that this zone was another area of submarine shoals or ridges. Analogies with the Gavrovo subzone suggest that these two zones were originally one, and that they are today separated by a completely 'floating' Pindus Nappe. However, field evidence does not as yet provide support for this idea.

Geographically, the Parnassos–Giona Zone forms a topographic ridge. Stratigraphically, it is composed of neritic limestones and dolomitized limestones from late Triassic to late Cretaceous age (Senonian). This carbonate sedimentation was not continuous, but underwent three interruptions accompanied by erosion and the development of bauxites. The first bauxite horizon formed between the Dogger and the Malm, the second at the end of the Jurassic and the third between the Lower and Upper Cretaceous.

During the Upper Cretaceous (Cenomanian–Maastrichtian) pelagic limestones with *Globotruncana* were deposited. Paleocene red shales locally lie unconformably on the limestones. These shales grade gradually upward to flysch of Eocene age. The carbonate sediments exceed 2000 m in thickness and the flysch 1000 m (Fig. 268).

Southeast of the Parnassos Massif, some sediments are transitional towards the Pindus Trench; this feature is known as the Vardussia Series, a sequence of intermediate pelagic and neritic character.

Tectonically, the Parnassos–Giona Zone is thrust to the W over the Olonos–Pindus Zone, and to the E it is thrust over the Boeotia Series (discussed next) and the zones of E Greece. The overthrusts represent folds with a large radius of curvature resulting from the massive nature of the Parnassos–Giona Zone's neritic limestone sediments.

Boeotia Series

The tectonism of the internal zones during the Upper Jurassic–Lower Cretaceous resulted in clastic formations in the neighboring isopic zones; these formations are composed of shales, limestone intercalations, sandstones and conglomerates; all derived primarily from Upper Jurassic ophiolites in the more easterly regions. The formations formed the so-called Boeotian Flysch, the age of which, in some places, reaches up to the Lower Senonian. The base of this clastic formation varies from neritic limestones in the W to pelagic limestones and radiolarites in the E. The clastics of the Boetian Flysch grade uniformly upward to an Upper Cretaceous limestone series, which then gives way to Paleocene flysch (Fig. 269).

Internal zones

The characteristic feature of the internal zones is the presence of two tectonic phases. The first during the Upper Jurassic to Lower Cretaceous formed the Paleohellenides (the Paleohellenic or Eohellenic phase) and resulted in emplacement of ophiolites (Brunn, 1961; Moores, 1969; Zimmerman, 1971); the younger phase, during Eocene–Oligocene time, placed the internal zones in their present position over the external zones.

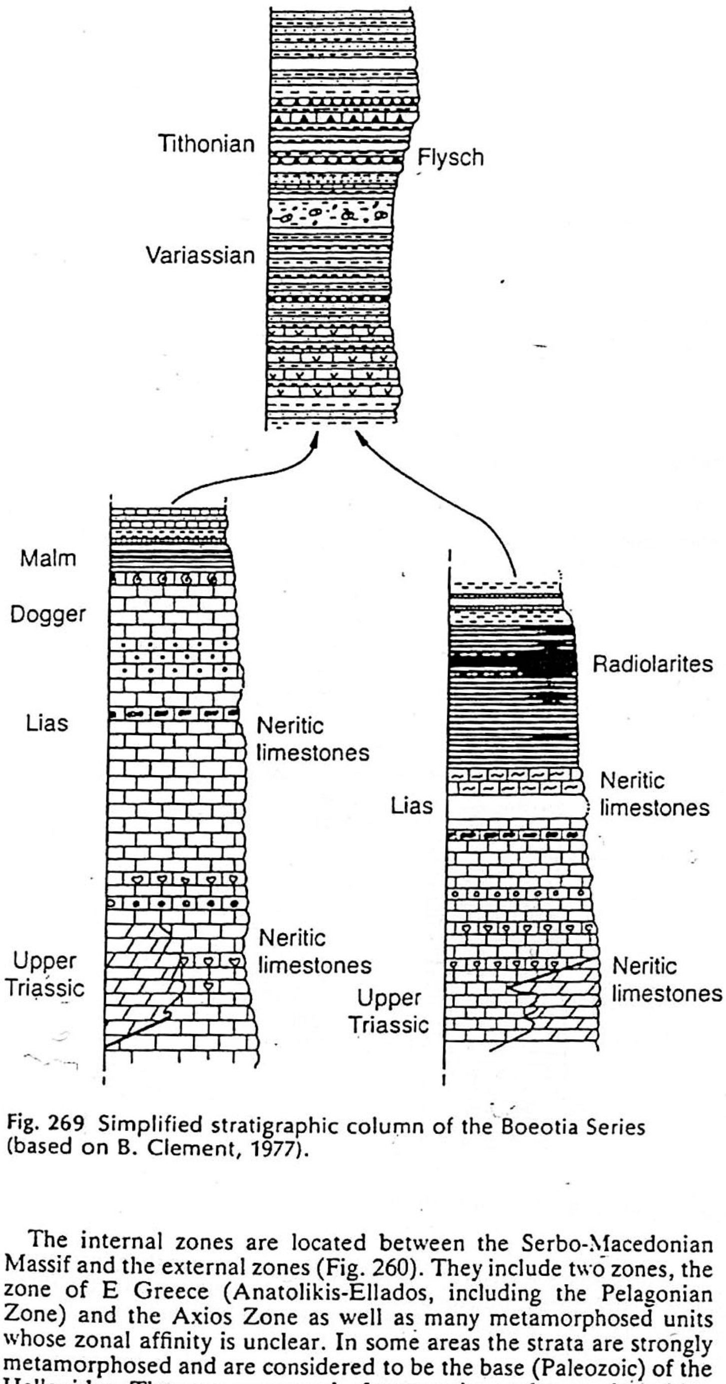


Fig. 269 Simplified stratigraphic column of the Boeotia Series (based on B. Clement, 1977).

The internal zones are located between the Serbo-Macedonian Massif and the external zones (Fig. 260). They include two zones, the zone of E Greece (Anatolikis-Ellados, including the Pelagonian Zone) and the Axios Zone as well as many metamorphosed units whose zonal affinity is unclear. In some areas the strata are strongly metamorphosed and are considered to be the base (Paleozoic) of the Hellenides. They are composed of magmatite gneisses, mica schists and marbles. This whole complex is invaded by granitic intrusions (Mercier, 1973).

Pelagonian Zone

The Pelagonian Zone consists of (Aubouin, 1959) (1) all the metamorphosed rocky 'masses' exposed in W Macedonia, E Thessaly, Euboea, Attica, Cyclades, etc. and (2) the overlying nappe representing sedimentary formations of late Paleozoic–early Mesozoic age. The Pelagonian Zone can be divided into non-metamorphosed formations (E Greece, N Attica, Argolis, Euboia and Cyclades) and metamorphosed zones (E and N Thessaly, W Macedonia; Katsikatsos, 1992).

On the outer side, primarily in the regions of central Euboea, Locride, Boeotia and N Attica, the zone of E Greece is metamorphosed. It is not known which metamorphosed masses correspond to the Paleohellenic phase. The metamorphic phases that have been dated are of Tertiary and, in some parts of the Aegean region, Miocene age.

Special attention is due to the Malliakos Series of the internal zones, which has clearly pelagic characteristics and represents the extension of the Olonos–Pindus Zone towards the interior. Stra graphically, this series begins with schist–sandstone series wi limestone lenses of Permian age, overlain by Triassic–Dogger dol mitized and brecciated limestones. The Upper Triassic limeston include intercalations of shales with pillow lavas. Shales and radiola ites characterize the Dogger sediments and nodular and brecciat limestones characterize the Malm sediments. Then shales gradua pass upwards to wildflysch with ultrabasic rock fragments th originated from the ophiolites, and which are overlain by t ophiolitic nappe.

Overlying the ophiolite in the Maliakos Series is an Upp Cretaceous transgressive sequence with conglomerates at the ba that develop into brecciated limestones in the Upper Cretaceo overlain by Maastrichtian pelagic limestones and Paleocene flysch

Thus the Maliakos stratigraphic sequence resembles the Olonc Pindus Zone. If the Maliakos Series had not been affected by t Paleohellenic tectonic phase, and if there was no ophiolitic nappe would be more accurate to talk about an internal Olonos–Pinc Zone.

Zone of Eastern Greece (Anatolikis Ellados; non-metamorphic equivalent of the Pelagonian Zone)

In places the Pelagonian Zone is covered by the Upper Cretace transgression and strongly metamorphosed, whereas elsewhere upper Cretaceous rocks are not metamorphosed. These regions designated the Zone of Eastern Greece.

The Upper Paleozoic deposits in the Zone of Eastern Gre include a clastic sequence with sandstones, siltstones, conglomer tuffs and limestones; the latter take the form of lenses and in calations with basalt layers.

During the Upper Triassic, carbonate neritic sedimentation beg and it continued up to the Lower Cretaceous. An emergence t place during the Malm and was accompanied by the depositio bauxites. The ophiolitic nappe, which is composed of peridoti gabbros and lavas (Fig. 270) was thrust over this sequence. Opin differ as to the direction of ophiolite emplacement. One inter tation is that it corresponds to the Paleohellenic or Eohellenic fol which resulted in the closing of the 'Axios Ocean' and the thrustin the ophiolitic nappe to the W. The other theory proposes that ophiolitic units overthrust the Eastern Greece Zone at its margins; the Axios ophiolites westward at the E margin, and Pindus–Vourinos ophiolite thrust eastward upon the W margin.

Unconformably overlying the ophiolites or neritic limestones Upper Cretaceous limestones, which pass upward into Maastrich pelagic facies and Paleocene.

Tectonically, the Zone of Eastern Greece is overthrust ontc external zones, which appear as tectonic windows. The most stri is the Olympus Window. Based on the age of the beginning of fl sedimentation and on a lithostratigraphic study of the limestones believed that this window corresponds to the Gavrovo–Tri Zone. Similar nappes are found in Attica and the Cyclades, but a lower plate there was strongly metamorphosed during Miocene t it cannot be correlated with certainty in a specific facies zone.

Axios Zone (Vardar Zone)

The Axios Zone is considered to mark the closing of the 'A Ocean and that it separates the Pelagonian zones towards the ext from the Serbo-Macedonian Massif towards the interior. In the A Zone the Upper Paleozoic is represented by clastic sediments an Mesozoic by sediments and ophiolites. The Upper Cretaceous t gression covers the older sediments, and its evolution is simi that in the Zone of Eastern Greece.

Paleogeographically, the Axios region can be divided into subzones from E to W (Mercier, 1966; Fig. 260): the Pae subzone, representing a 'trough'; the Paikon subzone, represen ridge; and the Almopias subzone, representing a 'trough'.

The Paeonias subzone was a trough or ocean basin starting i Triassic time, with calc-schists, diabase, cherts and ophiolit underwent at least three tectonic phases from the Upper Ju onward, each followed by marine transgressions (uppermost Ju Lower Cretaceous and Upper Eocene). Granites have been da 150 Ma (late Jurassic).

The Paikon subzone contains massive Triassic–Jurassic limes and with acidic volcanism and repeated emergences (Middle Ju Lower and uppermost Cretaceous, and Lower Oligocen undoubtedly represents a paleogeographic ridge (island arc?).

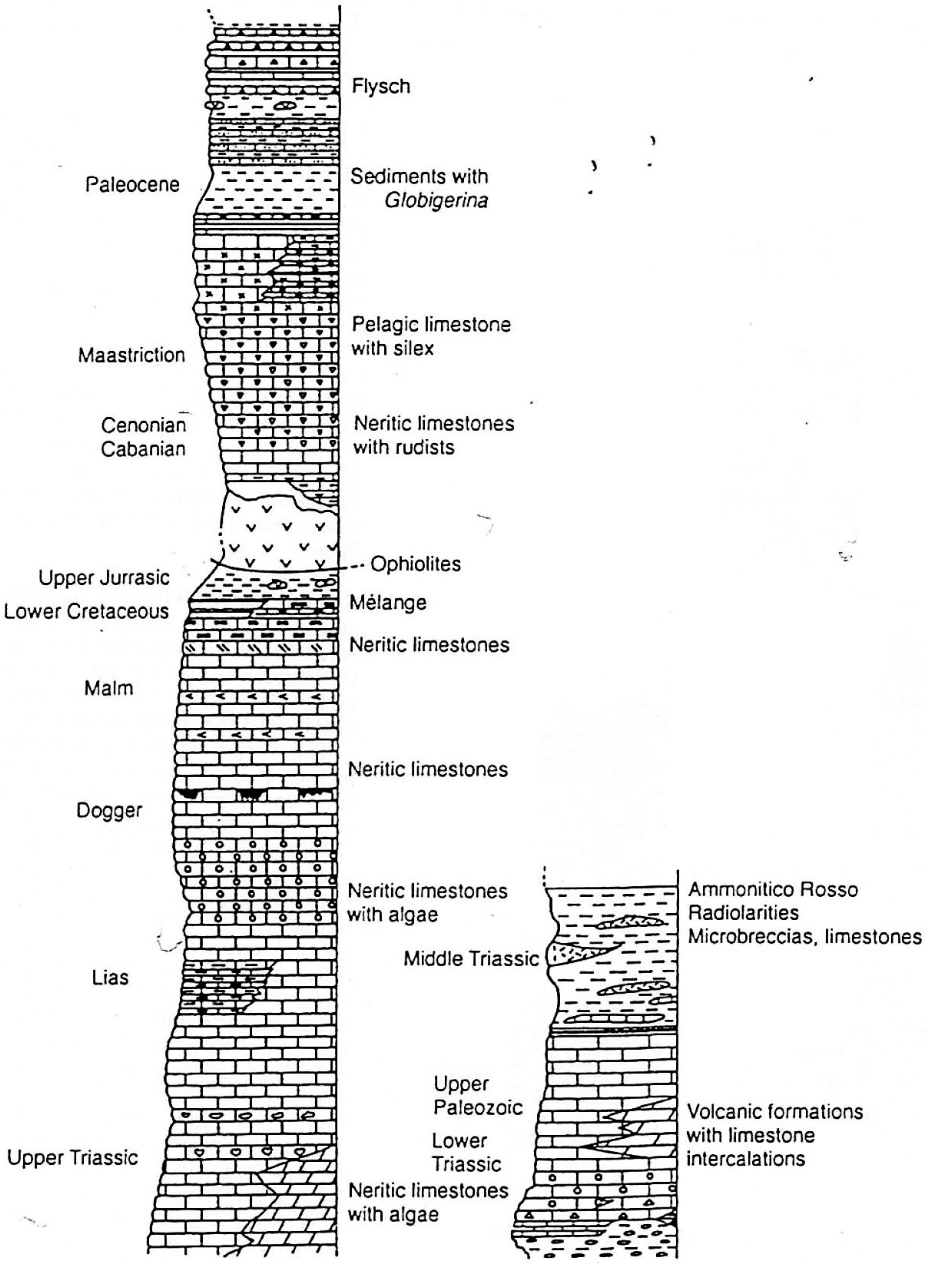


Fig. 270 Simplified stratigraphic column of the zone of E. Greece (based on B. Clement, 1977).

The Almopias subzone, bordering the Pelagonian zones to the NE, is rich in ophiolites (some highly silicified) and probably corresponds to a former trough or ocean basin. Successive tectonic phases associated with transverse and strike-slip movements have resulted in pronounced structural complexity. The main phase seems to have been Upper Jurassic to Lower Cretaceous and was accompanied by major transcurrent faulting, and the development of 'colored mélanges' (Mercier and Vergely, 1972). The mélanges are distinctly tectonic (not olistostromes).

Post-Alpine cycle

Neogene deposits

Distinguishing Alpine and post-Alpine deposits is rather difficult, even futile in very active areas like the Aegean, where deformation processes still go on, and there are regions (such as the Preapulian Zone and other domains further along the periphery of the arc) where this distinction is not possible. This difficulty is illustrated by the use of the terms 'molassic' and 'Neogene' to refer to sediments in the Aegean (Dermitzakis and Papanikolaou, 1979). The first molasse deposits transgressing the Alpine structures slightly postdate the last structures, but they are successively younger from the internal zones (Upper Eocene–Oligocene) to the external ones (Lower–Middle Miocene).

The post-Alpine deposits characterized as Neogene were not, in general, deposited during all the Neogene Period, but in most cases only during late Miocene–Pliocene. This type of Neogene representation is very common and it is clearly separated from the postorogenic molasse for two reasons: (1) the basin geometry is very different, with the molasse basins following the geometry of the neoformed Alpine chains, in contrast to the Upper Miocene–Pliocene basins, which have developed in a completely different pattern; and (2) an important event took place during Middle–Late Miocene and was marked by unconformities, the disappearance of molasse basins, the creation of discordant basins, and abrupt changes in the paleogeography with consequent sedimentological changes within the Miocene sequences.

However, this distinction between molassic and Neogene can be applied only in the more internal domains of the Aegean Arc. At its external border (external Ionian Zone and Preapulian Zone) the age of the Alpine tectonism is very young and neither molassic nor even Neogene characteristics can be distinguished.

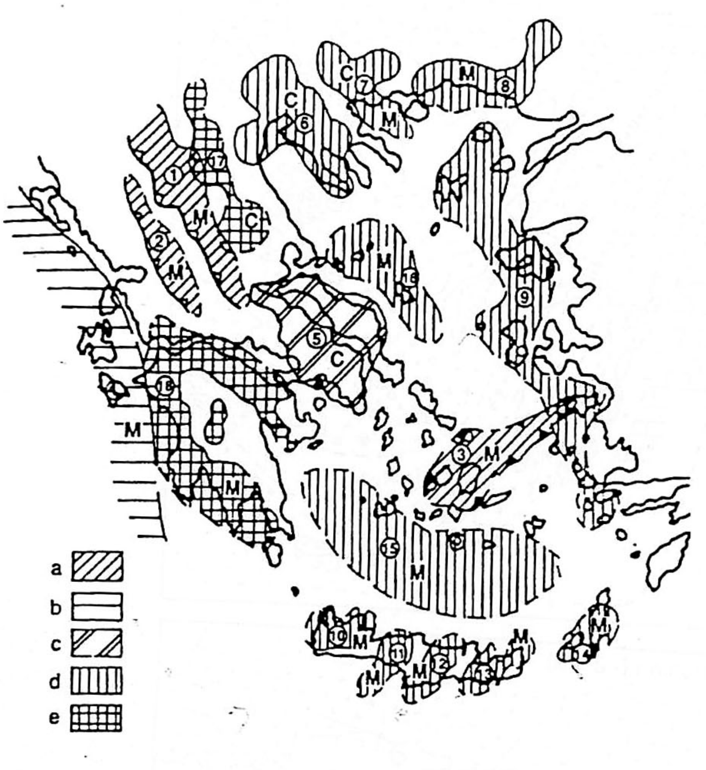


Fig. 271 Map showing the main Neogene basins of Greece (based on Dermitzakis and Papanikolaou, 1979). The main Neogene basins of the Aegean Region: a, Lower Miocene; b, Lower Miocene–Pliocene; c, basins of the Miocene and Pliocene; d, Upper Miocene–Pliocene; e, Pliocene. M, marine; C, continental; 1, Mesohellenic Trough; 2, Epirus; 3, Cycladic; 4, Ionian Islands; Attica–Viotia; 6, Thessaloniki; 7, Strymon; 8, Thraki; 9, E Aegean islands; 10, Chania; 11, Rethymno; 12, Iraklio; 13, Ierapetra; 14, Karpathos; 15, Cretan Basin; 16, N Sporades; 17, Ptolemais–Kozani; 18, Peloponnesus.

The outcrops of Neogene sediments in Greece are numerous and it is difficult to group them into distinct basins because (1) in the same area there may be sediments belonging to different basins that existed during different time spans (e.g. the Mesohellenic Trough. Evia and Kos); and (2) there are many small individual basins, each one with its own development, as is well illustrated in Crete (Symeonidis, 1967; Meulenkamp, Jonkers and Spaak, 1977). Nevertheless, identifying some large basins (or groups of basins; Fig. 271) is useful for understanding the variety of paleogeographic domains that existed during the geodynamic evolution of the area, which is characterized by a non-uniform regime.

Thus, there are three basins with Lower Miocene sediments of molassic facies: the Mesohellenic Trough (1 in Fig. 271), which is a typical molasse (Brunn, 1956); the basin of Epirus–Akarnania (2), which is not clearly distinguished from the underlying basin of the Ionian flysch (IGRS and IFP, 1966); and the Cycladic Basin (3), which is a molassic basin similar to the Mesohellenic Trough, except that it is allochthonous. The same is true for the basin of Epirus–Akarnania, which was transported, together with its basement, during the thrusting of the Ionian Zone onto the Preapulian Zone. The basin of the Ionian islands (4 in Fig. 271) is exceptional because it belongs to the upper part of the Preapulian Zone. Its sedimentary sequence has probably been interrupted only at the Miocene–Pliocene boundary (Horstmann, 1967; Dermitzakis and Sondaar, 1978). The basin of Attica–Viotia–Evia (5) is continental, with some important interruptions in its sequence. Many basins (6–16 in Fig. 271) comprise sediments of Upper Miocene–Pliocene age; most of them are marine, except for the basin of the E Aegean islands (9) and that of Thessaloniki (6), which are continental. It is remarkable that the sediments of basin 9 are usually deformed with intense folding (Angelier, 1979; Papanikolaou 1978). Finally, there are the basins of Western Macedonian–Thessaly (17, continental) and of Peloponnesus (18, marine), which are of Pliocene age.

Upper Miocene sediments are transgressive in some synclines of the Ionian Zone. West of the Pelagonian zones the Mesohellenic Trough is filled with Upper Eocene, Oligocene and Miocene marine detrital deposits up to 5000 m thick; they lie transgressively upon earlier folded rocks (Brunn, 1956).

In the Axios Zone, analogous formations are slightly older (Eocene–Oligocene). On the island of Rhodes they extend from the Oligocene to early Miocene layers as the Vati Group (Mutti, Orombelli and Pozzi, 1965). There were extensive connections with the Paratethys of the Balkans and the Black Sea (Sonnenfeld, 1974).

Paleogeographic evolution of the Aegean during the Neogene

The paleogeographic evolution of the Aegean region during the Neogene is indicated in four paleogeographic sketch maps that indicate the relative positions of various domains during certain critical periods (Dermitzakis and Papanikolaou, 1979; Fig. 272). In general, these maps resemble those presented earlier (e.g. Vinken, 1965; Keraudren, 1979) with the exception of the Cycladic domain's position. They are based on the present geography of the Aegean, but it should be stressed that the whole geometry must have been very different in former times because (1) nappe transport and thrusting was very important durng the Lower–Middle Miocene in the Ionian and Gavrovo–Tripolis zones, and even during the Upper Miocene–Lower Pliocene in the Preapulian Zone; and (2) considerable intracontinental deformation (with folding, thrusting and normal and strike-slip faulting) was manifest all over the Aegean during the Neogene.

For Aquitanian time we can distinguish an external domain where flysch sedimentation of the Ionian and Gavrovo–Tripolis zones was still taking place; carbonate sedimentation shows up in the more external areas of the Preapulian Zone. The Pindus Cordillera, so named from the main zone of the Hellenides, appears along the Preapulian line of sedimentation (Aubouin, 1959). Then a more or less continuous molassic trough runs from N Greece, through the E Peloponnesus and the Cretan Basin, to the SW part of Asia Minor. Further, there is the Pelagonian Cordillera, which consists mainly of the metamorphic rocks of the former Pelagonian zones and extends from W Macedonia, through E Thessaly, Attica, and the Cyclades to the area of Menderes in W Turkey (Van der Maar and Jansen, 1983). In the N Aegean Sea, along the S margin of the Rhodope Massif, remnants of another molassic basin of Upper Eocene–Oligocene age might still exist locally. A volcanic zone from E Thrace to N Samos is evident, running parallel to the present coastline of Asia Minor.

During the late Burdigalian–Langhian (Fig. 272) the molassic basin was filled up. Strong uplift along the Pindus Cordillera now expanded towards the exterior, including the Gavrovo–Tripolis and Ionian zones, and caused a gravity décollement nappe to slide onto the metamorphics of the Cyclades. At the same time, a general tilting of the molasse towards the ENE took place in the Mesohellenic Trough. Volcanism continued along the same N–S-trending zone as before, but also spread more to the S as far as the island of Kos. The main granite gneiss in the central Cyclades forms an almost E–W zone, having preceded the emplacement of the Cycladic Nappe. Gravity décollement also took place on the external side of the Pindus Cordillera in E Crete, where huge blocks of Alpine rocks are observed 'floating' within the Lower Serravallian sediments (Fortuin, 1978). In this case the décollement was smaller and the allochthonous rocks formed an olistolite terrane.

During the late Tortonian (Fig. 272) intense fracturing of the central part of the arc led to invasion by the sea and the beginning of a very pronounced subsidence of the Cretan Basin (Marinos *et al.*, 1970). Throughout the late Tortonian and the Messinian, temporary marine connections existed between the newly formed Cretan Basin and the N Aegean, but they did not create a 'barrier' major enough to prevent land-mammal migration from Asia Minor through Samos and Chios to Attica and Evia. Famous vertebrate fossil locations are also found in some of the numerous interior basins. One of them is Pikermi, the classic fossil location near Athens, where the marine Miocene is overlain by the late Miocene Pikermi Formation, which consists of continental clays with a rich mammalian fauna and is capped by marine Pliocene material (Heye, Schneider and Symeonides, 1981). The famous Pikermi fauna also appears in Samos, Chios, Psara, Evia, Salonica, etc.

The international orogenic-phase term Attican was proposed in 1924 by Stille, on the basis of century-old work by Gaudry and Fuchs,

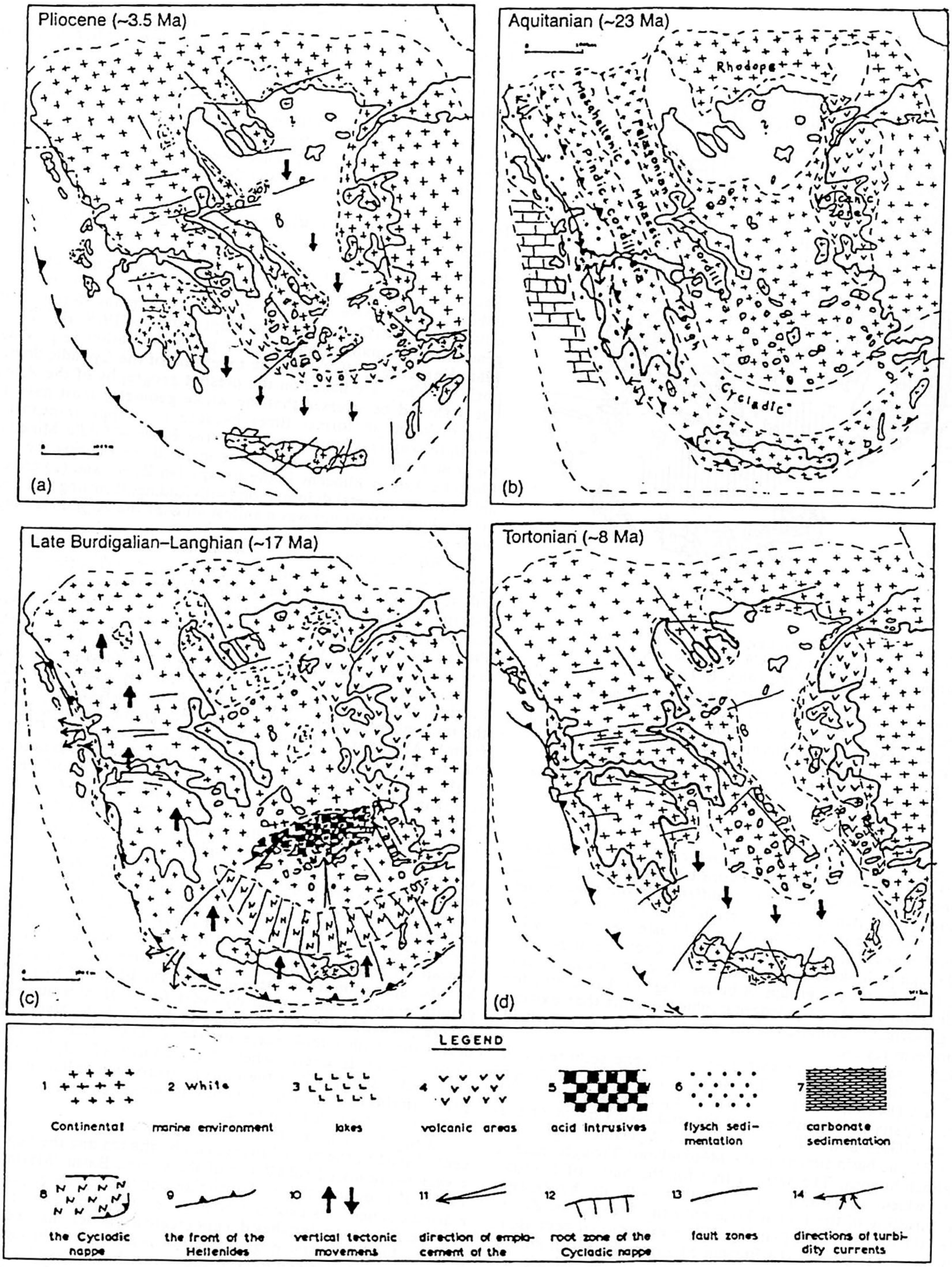


Fig. 272 Paleogeographic sketches of the Aegean region during the Neogene. (a) Pliocene (*c.* 3.5 Ma); (b) Aquitanian (*c.* 23 Ma); (c) Late Burdigalian–Langhian (*c.* 17 Ma); (d) Tortonian (*c.* 8 Ma). 1, continental, 2, marine environment, 3, lakes, 4, volcanic areas, 5, acid intrusives, 6, flysch sedimentation; 7, carbonate sedimentation; 8, the Cycladic Nappe; 9, the front of the Hellenides; 10, vertical tectonic movements; 11, direction of emplacement of the Cycladic Nappe; 12, root zone of the Cycladic Nappe, 13, fault zones; 14, directions of turbidity currents (modified from Dermitzakis and Papanikolaou, 1979).

as falling between the Sarmatian and Meotian (probably the Paratethyan equivalents of the Pikermi and Messinian). The setting is poorly established, however.

The Messinian regression led to extensive evaporites offshore, and locally they are uplifted (Sonnenfeld, 1974). Thick evaporitic masses are exposed in Crete and W Greece, with sparse outcrops in N Greece.

The Messinian drop in sea level also gave rise to giant fluvial cross-bedding in places, e.g. in E Crete (Gradstein and Van Gelder, 1971). It was probably this major regression that permitted a widespread immigration of savanna-type mammals to many of the now-isolated Aegean islands, although earlier writers spoke of an 'Aegean continent', and undoubtedly there has been much block-faulting of former land areas (Besenecker and Otte, 1972; Schroder, 1974). In the late Pliocene there was again a connection between the Aegean and brackish Black Sea–Caspian Paratethys bringing *Aspheron* facies as far as Corinth (von Freyberg, 1974).

Following the Messinian regression, the Pliocene was widely transgressive and coastal basins are in part marine. During the Pliocene, the Aegean Sea evolved with important subsidence in the Cretan Basin and to the N of the Cyclades. At the same time, the Pliocene channel of Corinthos separated the Peloponnesus from the mainland. In the Sperchios Valley the Pliocene sea did not invade far into the graben. From this period onward the volcanic Aegean Arc developed.

In conclusion, it seems that until the Burdigalian, the Aegean Arc was relatively homogeneous from NW Greece as far as the Dodecanese Islands. After the general structural revolution of the Langhian–Serravallian (caused by the final collision of Arabia and Europe), a remarkable differentiation started, with very important vertical and horizontal movements, especially in the arc's central concave part. The first period of high uplift, with consequent gravity tectonics, was manifested mainly in the area of the Cyclades. It was followed by intense fragmentation and block-faulting, with consequent invasion by the sea and extension of the arc during the Pliocene, when the creation of the tectonic axis in mainland Greece and the Peloponnesus became very important; this importance probably continues today.

Post-tectonic volcanic activity has been limited for most of mainland Greece, except for the Vardar Zone, where it was intense during the late Miocene. On the Aegean islands, during the Lower and Middle Miocene, there was widespread calc-alkaline volcanism with latite andesites, dacites and rhyolites, partly as ignimbrites. They seem to be related to a sinking slab of lithosphere such as that recognized in the Aegean today (Borsi *et al.*, 1973).

Volcanic activity

In the Aegean region and the adjacent areas of the Greek and Asia Minor coast, intense volcanic activity during the Neogene and Quaternary produced lavas of varied composition (Olausson, 1971). These periods of activity and the positions of the volcanoes in the Aegean region are directly connected with the tectonism of this area, which led to major faults in places such as the borders of crystalline massifs. Thus volcanoes of the S Aegean volcanic arc are situated on the border between the Attico-Cycladic Massif and the Menderes Massif. The volcanoes of W Thrace, Samothrace and Limnos are situated on the borders of the Rhodope Massif. On the borders of the Pelagonian Massif are the volcanoes of Thives, Thessalia, the Lichades islands, N Sporades, etc.

According to Ktenas and other geologists, the volcanoes of the Aegean region can be divided into three geographical areas, each with special geological and petrologic features: the S Aegean areas, with volcanoes that are situated S of the 38th and 39th parallels; and the N Aegean area, which takes in volcanoes N of the 39th parallel (Fig. 273). Another feature is the age of volcanic activity. Based on age, we can distinguish two large groups of volcanoes: paleovolcanoes active during the Eocene–Miocene and neovolcanoes active from the Pliocene until today (e.g. Borsi *et al.*, 1973).

Of all the volcanoes of the Aegean region, one group in particular has played a very important role in the civilization of the E Mediterranean: the volcanoes of Santorini (Thera). The Santorini volcanoes originated in the Neogene. The first volcanic eruption happened in the SE part of Thera Island in Profitis Ilias and Akrotiri. During the same period, another group of volcanoes named Peristeria was active in N Santorini and formed the mountains called Little Profitis Ilias, Megalo Vouno (big hill) and Kokkino Vouno (red

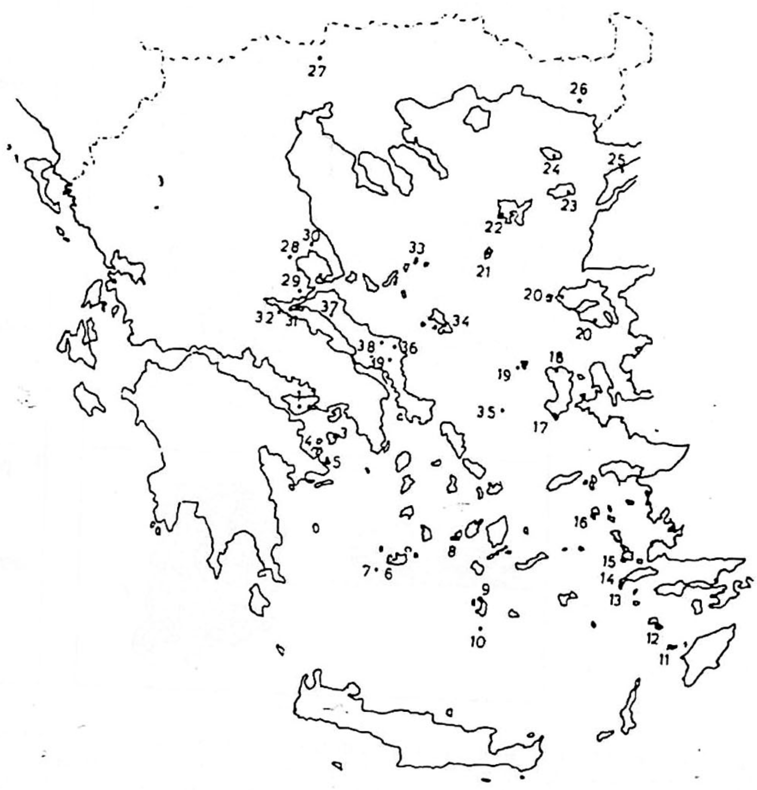


Fig. 273 The main Greek volcanic sites: 1, Kalamaki; 2, Sousaki; 3, Aegina; 4, Methana; 5, Poros; 6, Milos; 7, Ananes; 8, Antiparos; 9, Santorini; 10, Christiana; 11, Episkopi (Tilos); 12, Nisyros: 13, Giali; 14, Kos; 15, Kalymnos; 16, Patmos; 17, South Chios; 18, North Chios; 19, Psara; 20, Lesbos; 20a, Ordymnos; 21, Ag. Efstratios; 22, Limnos; 23, Imbros; 24, Samothraki; 25, Sgourafa; 26, Ferres; 27, Almopia; 28, Mikrothives; 29, Achillio; 30, Porfyrion; 31, Lichades; 32, Vromolimni; 33, Psathoura; 34, Skyros; 35, Kalogeri; 36, Oxylithos; 37, Thorio; 38, Metochi; 39, Orio.

hill). In the W part of Santorini another volcano, Simantiri, started later. In the S part the volcano of Thirasia has also been active as has the volcano of Skaros in the N (Fig. 274).

With these repeated eruptions of the volcanoes of Santorini, an island known as Strongyli (Fig. 274a) developed and was inhabited by prehistoric people who apparently knew the use of iron but used stone tools. Strongyli was possibly about 600–1000 m in altitude. Later, the inhabitants belonged to the Minoan civilization and knew how to cultivate olive trees, grain, etc. They also had commercial relations with the adjacent Aegean coasts. The Minoan installations were destroyed by catastrophic eruptions that happened around 1700–1500 BC (Fig. 274b).

The volcanic activity continued, and after a short interval dramatic tectonic events fragmented the crater and the sea invaded (Fig. 274c). With progressive subsidence, the caldera was formed (Fig. 274d). Fragmentation of the caldera followed, and the lateral fragments sank under sea level (Fig. 274e). After some hundred years, the volcanic action restarted in the center of the caldera and the islands of Kamenae were formed. The 'Santorini ash' has been dated in E Mediterranean deep-sea cores and in the cores of the DSDP. The major eruption that terminated the Minoan civilization left distinctive ash layers (Bond and Sparks, 1976) on many islands and Crete (Boekshoten, 1963), as well as offshore (Keller and Ninkovitch, 1972). The resultant tsunami carried floating pumice to the mainland (Rapp, Cooke and Henrikson, 1975) and left a beachline of broken shells on the coast of Israel.

Quaternary deposits

The Quaternary in Greece and the Aegean shows a remarkably wide and varied development that has long attracted attention (e.g. Bate, 1905). The interior basins contain thick lacustrine accumulations with

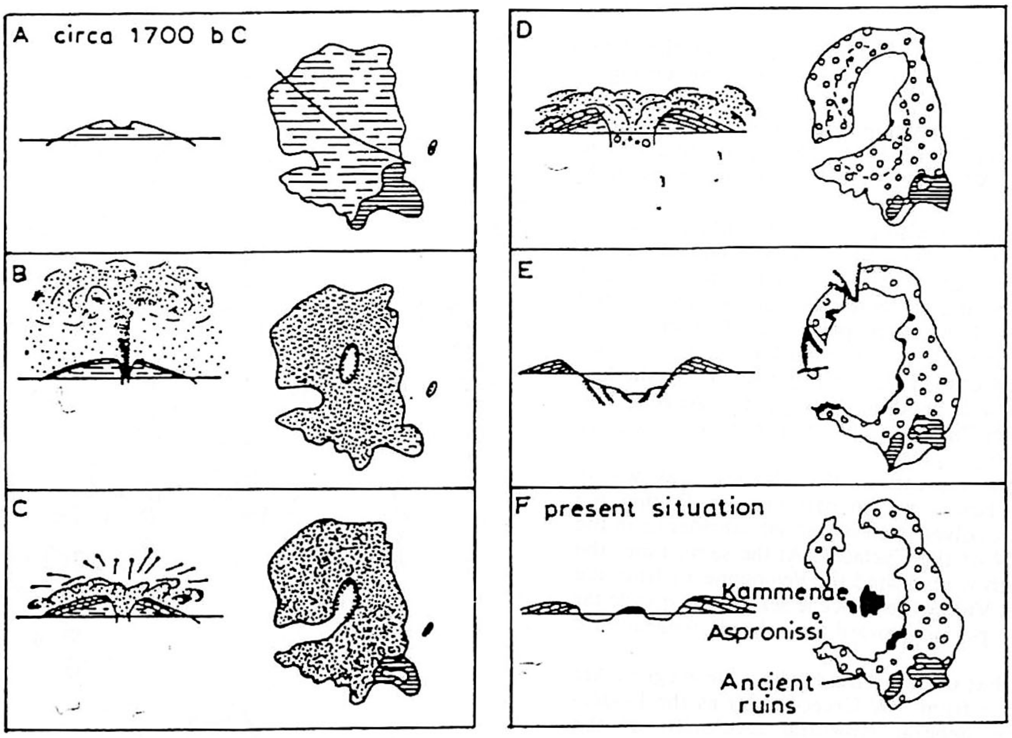


Fig. 274 The evolution stages of the caldera on the island of Santorini from 1700 BC to today (by H. Pichler and W.L. Friedrich, 1980). (a) During previous volcanic explosions the island of Strongyli was formed (500–600 m) and occupied with Minoan installations. (b) The explosion of c. 1700 BC made an opening at the center of Strongyli and destroyed the Minoan installations. (c) The volcanic action continues; the crater is shattered and the sea comes in. (d) The explosions continue, with bombs and ashes. At the same time the cone collapses and the caldera starts to form. (e) Tectonic fractures enlarge the caldera at the sides. (f) Years later, volcanic action restarts and the islands of Kammenae are formed inside the caldera.

lignites (Megalopolis, Ptolemais, Phillippi, etc). Pollen analysis has furnished long records of climatic change in the E Mediterranean. The pattern involves a repeated alternation from warm and seasonally wet (interglacial) to cold and semiarid (full glacial), with cool fluvial conditions in the oscillatory transitional phases (Luttig, 1963; Fairbridge, 1972). There was minor glaciation on the highest mountains (Mount Olympus). Periglacial activity was very extensive and greatly indurated calcareous talus and colluvial deposits occur in many areas, in places going below present sea level (Creutzburg, 1961). Travertine-spring deposits have furnished interesting fossil faunas and floras. Pollen sequences are present in the Ptolemais, Megalopolis, Vegora and Serres basins, etc. (Benda, Meulenkamp and Van de Weerd, 1979; Riegel *et al.*, 1988; Van der Hammen, Wijmstra and Zagwijn, 1971). Some pollen evidence indicates that a semi-tundra developed in the coldest times.

In the coastal regions there is commonly a stepped unconformable sequence of thin transgressive littoral deposits; sandy marls, sandstones and conglomerates include characteristic faunas as in the type areas of Calabria and Sicily. The highest and most deformed formations correspond to the Calabrian (Keraudren, 1967, 1970) with Sicilian and Tyrrhenian deposits at progressively lower levels (Dermitzakis, 1969, 1973; Dermitzakis and Symeonides, 1974). Because of complex neotectonics, no altitudinal correlation is reliable, but the formations are highly fossiliferous in places. For example, the early Pleistocene Vasfi Formation of the island of Rhodes contains the typical Calabrian cold-water fauna with the mollusk *Cyprina islandica* (Zaccaria, 1968) and the foraminifera *Hyalinea balthica* (Meulenkamp, De Mulder and Van De Weerd, 1972). The Vasfi is followed by the Rhodos Formation (probably Sicilian and Tyrrhenian), which in places reaches 200 m above sea level.

Another interesting fauna with mollusks and foraminifera and the characteristic presence of *Hyalinea balthica* has been identified in the S Peloponnesus (Neapolis) and on the island of Zakynthos (Gerakas; Keraudren 1970, 1973; Dermitzakis and Goedicke, 1977).

In the Mani peninsula in the S Peloponnesus, eight Holocene and late Pleistocene terraces above the Tyrrhenian have been described by Kelletat *et al.* (1978).

Tyrrhenian deposits are extensively represented, especially on Crete (Boekschoten, 1963; Keraudren, 1970, 1973, 1979). Tyrrhenian marine levels (Dermitzakis, 1972), with the characteristic species *Strombus bubonius*, are relatively clear and almost universal, with terrace I generally around 40 m above sea level, II at 20 m and III at 8 m. In mainland Greece *Strombus bubonius* was found in 1833 by Boblaye and Virlet in Nauplion. The same fossiliferous beds have been described by Keraudren (1970, 1973) as Neotyrrhenian in age. In the area of Corinth–Kalamaki, *Strombus bubonius* is found in many localities (Mitzopoulos, 1933), as also in the Peloponnesus (Dufaure, 1977).

An interesting Middle Pleistocene invasion by 'Caspi-brackish' Black Sea waters is recorded near Corinth (Chauda stage) according to Gillet (1961, 1963) and von Freyberg (1974). Also Krstic and Dermitzakis (1981) refer to a Middle Pleistocene fauna of the Corinth Canal, the ecology of which is similar to the Chaudian Beds in Moldova and the valley of the Dnieper River.

A wealth of data on fossil mammals of the Aegean islands has recently been recorded. These faunas range in age from Upper Miocene to Holocene. They are mostly found in fissure or cave deposits. Two main types are distinguishable (Dermitzakis and Sondaar, 1978): a fauna resembling that on the mainland and an unbalanced, mostly endemic fauna. Since the dispersal of mammals depended on paleogeography, and the sea was generally an insurmountable obstacle, the faunas can tell us something about former connections (Mitzopoulos, 1961; Greutzburg, 1966; Sondaar, 1971, 1977; Dermitzakis, 1977; Storch, 1977). Possible dispersal routes may have followed a corridor in which faunal interchange from one region to another was possible in various categories: the corrider, filter, sweepstake and pendulum methods (Dermitzakis and Sondaar, 1978).

In general, the Aegean's land–sea configuration in the Pleistocene did not differ essentially from that of today. The Peloponnesus was part of the mainland in the Pleistocene, as shown by findings of several species of elephants, hippopotamuses, carnivores, deer, rodents, etc. In Kythira the fossils of *Elephas antiquus, Cervus dama, Megaceros cretensis*, etc., show an unbalanced fauna; so Kythira was an island probably not far from the coast.

Crete has shown a balanced European Miocene fauna (De Bruijn and Meulenkamp, 1972). In the Upper Miocene and Pliocene, however, Crete was mainly submerged (Drooger and Meulenkamp, 1973). From the Pleistocene, 63 localities are known with unbalanced and endemic fauna, which points to a sweepstake dispersal at a time when Crete must have been an island. Cretan fauna consisted mainly of elephants, deer, hippopotamuses and micromammals. Changes in the size of the island, caused by glaciation and tectonic forces, have also played a role in the extinction of the endemic mammals.

During the Pliocene (Ruscinian), Karpathos was still connected to the mainland, as shown by deposits containing fossils of a Pliocene balanced rodent fauna (Daams and Van der Weerd, 1980). In the Pleistocene, Karpathos took on its present shape and was probably once connected to Kassos, because *Megaceros cretensis* is present on both islands. The deer in Karpathos came by a sweepstake route, although it is not clear when. The presence of *Mus* indicates that the fauna is Mindel or younger. As the deer of Karpathos must have evolved separately from those of Crete, a barrier between the two islands is implied.

On Rhodes, Turolian and Ruscinian mammalian fauna assemblages show a balanced continental-type fauna, which means a land connection with Asia Minor. In the Pleistocene, Rhodes became an island with local deer and dwarf elephants (*Paleoloxodon*). Important vertebrate fossil remains have been found in the Charkadio Cave on Tilos Island (Symeonides, 1972, Bachmeyer *et al.*, 1976).

Dermitzakis and Goedicke (1977) proposed a land connection in the Middle to Upper Pleistocene between the islands of Rhodes and Tilos. A structural ridge between the two islands is indicated at a depth of 420 m and could have formed a land bridge. The dwarf elephants of Tilos were dated by ^{14}C as 4390 and 7090 BP.

In the Cyclades Islands (Naxos, Delos, Seriphos, Kythnos, Amorgos and Milos) remains of dwarf elephants, deer, micromammals, etc., suggest that there were in this region Pleistocene islands separate from Crete and the Dodecanese.

Neotectonic movements

The shorelines of Greece underwent considerable vertical motion during the Holocene (Spratt, 1865; Pirazzoli *et al.*, 1982). This is particularly evident in Crete, Karpathos, Kassos, Rhodes, Antikythira and the Peloponnesus. The existence of ancient port cities, both submerged and uplifted, demonstrates that tectonic movements as well as eustatic changes have occurred. Some of these movements must have taken place during historical times. For example the geographer Strabo, and others, described an artificial port in Phalasserna which now lies above sea level and considerably inland. The radiocarbon datings of Pirazzoli *et al.* (1982) prove that the movements took place in historical times.

On the island of Rhodes, Marinos and Dermitzakis (1977) observed a succession of eight terraces along the E coastlines, while along the N coast only the higher ones can be distinguished. The old strandlines of the E part of the island are located at quite different heights. A combination of tectonic and eustatic movements are thus indicated.

Analogous observations of recent old standlines and terraces are reported from the Peloponnesus (Schroder, 1974), Kasos, Karpathos and from the Hellenic Arc (Pirazzoli *et al.*, 1982).

Economic geology

A large variety of economic minerals, both metalliferous and nonmetalliferous, occurs in Greece. The most important are as follows (Marinos, 1982; Fig. 275).

Antimony

Many small deposits of stibnite are found in Macedonia (Lachaw) and Thrace. On the island of Chios there are veins of low-grade stibnite in argillaceous rocks of Devonian age.

Asbestos

In Western Macedonia (Kozani) asbestos occurs in joints, as well as in the serpentinized peripheral zone of the main Vourinos Massif.

Barite

Most of the hydrothermal deposits of E Greece contain barite. The island of Mykonos exports it for use as a heavy medium for oil drilling; reserves are estimated at several million tonnes. In Milos, abundant barite is massive or is disseminated in lavas and tuffs. Reserves in Milos are several million tonnes. Barite also occurs on Kos and Polyvos.

Bauxite

Bauxite makes Greece a substantial producer of aluminum; there are more than 1000 deposits. They are most abundant in central Greece (Oeti, Giona, Parnassos and Kithairon mountains; Evia and Amorgos) and are there identified as the Mediterranean limestone type, so called because the deposits occur between limestone formations.

Bentonite

Greek bentonites belong to the calcium type; however, they can be transformed into the swelling, sodium type, after 'activation' treatment with sodium carbonate. Large bentonite deposits amounting to many millions of tonnes are located on Milos and Kimolos; intensive open-pit exploitation is carried out. Greece ranks second in the world after the United States as an exporter of bentonite.

Chromite

Ultrabasic rocks from ophiolites, such as peridotites and, particularly, dunites common host rocks for chromite deposits occur widely in Greece. Occurrences of chromite vary greatly in form, structure and size. In the Vourinos area, chromite spinels, associated with ultramafic rocks, occur in ore reserves of about 1 500 000 tonnes. In Thessalia at Volos, an ancient operating mine working massive refractory chromite has had a total production to date of about 500 000 tonnes. At Domokos a large deposit of refractory chromite in dunite has had a total production of over 500 000 tonnes. The total reserves of chromite in Greece approach 5 million tonnes.

Copper

The largest copper deposit in Greece is on the Chalkidiki Peninsula around Neochorion. The proved ore reserves there exceed 15 million tonnes. In Western Macedonia, in the area of Pella, there are veins and disseminations of copper sulfide minerals in diabase volcanic rocks. The largest deposit of cupriferous pyrite is in the E Peloponnesus near Hermioni, and in ancient Krokeas of Sparta many veins of copper minerals are found in diabase rocks.

Natural pozzolana ('earth of Thyra')

In Greece natural pozzolana (siliceous sediment) occurs in extensive deposits in Thera (Santorini) and nearby islands. The extraction of the 'earth of Thyra' is by open-pit mining.

Hydrocarbons

The history of hydrocarbons in Greece started in ancient times wher Herodotus described the oil bitumen from a well at Limin Keri, in S Zakynthos Island. Continental exploration started in 1938, but afte extensive drilling, results have been disappointing. Offshore explora tion began in 1970 with unsuccessful drilling in the Thermaicos Gul in the N Aegean. In 1970 the 'Oceanic' Company started exploratio in the Thrakikon Pelagos (N Aegean) and found oil in the Princ Anticline, where considerable drilling has followed in clastic an evaporitic sediments. These sediments have a maximum thickness c 5800 m and are of Miocene age and younger. The evaporitic basi has its maximum extension in the Messinian (Pollak, 1979).

Iron

'Soft' iron ores occur in more than 200 locations throughout Greece. 'Hard' iron–nickel (lateritic) ores occur in Larymna (130 k N of Athens) and include about 200 million tonnes of ore over 1.0' Ni.

Geothermal energy

In the program for the development of new sources of energy, start in 1970, geothermal sources appeared promising. The Public Pow Corporation (PPC), in cooperation with foreign specialists, regard several areas as favorable, including the islands of Milos, Lesb

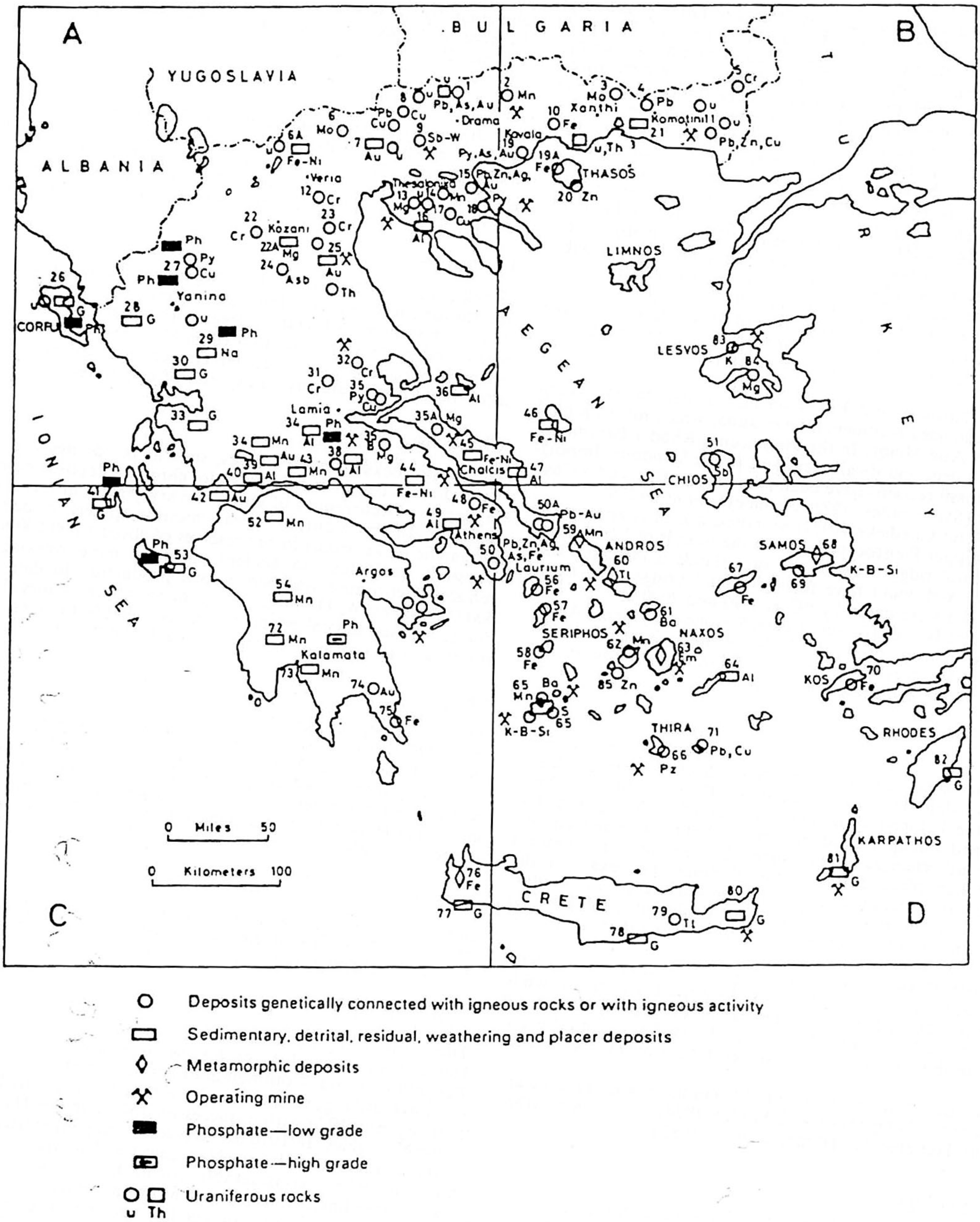


Fig. 275 The principal mineral deposits of Greece. Al, aluminum; As, arsenic; Asb, asbestos; Au, gold; B, bentonite; Ba, barite, Cr, chromite; Cu, copper; Em, emery; Fe, iron; G, gypsum; K, potash; Mg, magnesite; Mn, manganese; Mo, molybdenum; Na, salt; Ni, nickel; Pb, lead; Ph, phosphate; Py, pyrite; Pz, pozzolana; Sb, antimony; Si, amorphous silica; Th, thorium; Te, talc; U, uranium; W, tungsten; Zn, zinc.

Nissiros, the Aghioi Theodori area (Sousaki), the Methana Peninsula, the Sperhios Valley and North Evia. On the island of Milos, the PPC drilled to 1100 and 1160 m; they found temperatures over 300°C and during trials obtained a production of 50 t h^{-1} of mixed steam and hot water.

Gold

Gold in greatly varied conditions occurs in Macedonia, Thrace, Laurium and in the Cyclades Islands. In some river gravels gold may exceed 150 g t^{-1}. In ancient Greece, particularly Macedonia, the exploitation of gold produced about 300 000 kg during the period from 1200 BC to AD 50.

Gypsum, anhydrite and salt

Large deposits of sedimentary gypsum and anhydrite occur mainly in W Greece, including the mainland and the islands of the Ionian Sea, and in Crete and the Dodecanesse. Gypsum also occurs at Kavalla in Macedonia.

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